

Chapter 2: Changing state of the climate system

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Date of Draft:

29 April 2019

Note: TSU compiled version.

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Executive Summary

Changes in historical drivers of climate change

Changes in drivers of radiative forcing since the late 19th Century are dominated by greenhouse gases and aerosols. The net effect is a positive imbalance of the radiative budget, which is accelerating. Present-day concentrations of atmospheric CO₂ are higher than at any point in at least the past two million years. {2.2}

Changes in total solar irradiance during the industrial era are not unusual in the context of at least the past 9000 years (*medium confidence*). Changes in volcanic forcing by aerosol loading during the industrial era are also not unusual in the context of at least the past 2500 years (*medium confidence*). {2.2.2, 2.2.3}

The mixing ratio of CO₂ has slowly fluctuated by at least 2000 ppm over the last 500 million years (*medium confidence*). The last time CO₂ levels were similar to the present-day was over 2 million years ago (*high confidence*). While, mixing ratios of CH₄, and N₂O prior to 800 ka are not known, it is certain that pre-industrial WMGHG mixing ratios were lower than present day levels and that current rates of change in the WMGHGs are unprecedented in at least the last 800 ka. {2.2.4.1}

There is robust evidence that the concentrations of all three major long-lived greenhouse gases have continued to increase since 2011 by 15 ppm for CO₂, 47 ppb for CH₄ and 5.6 ppb for N₂O. Their abundances are now 118.1 ± 1.6 ppm, 1043 ± 6 ppb and 59 ± 4 ppb respectively above pre-industrial levels (*very high confidence*). {2.2.4.2}

Atmospheric abundances of most CFCs, regulated under the Montreal protocol, are further declining, with a radiative forcing of 0.25 W m^{-2} in 2017. Direct radiative forcing from HCFCs and HFCs were 0.059 and 0.03 W m^{-2} in 2017, respectively, with emission restrictions imposed under the Kigali Amendment of the Montreal protocol. {2.2.4.2}

Frost point hygrometer measurements over Boulder, Colorado depict a net increase of Stratospheric Water Vapour (SWV) of ~ 1.0 ppm ($\sim 25\%$) during 1980-2010 most of which occurred prior to 1993. Conversely, several sets of merged satellite data created since AR5 show no evidence of a net increase since the late 1980s. There thus continues to be *low confidence* in a long-term trend of SWV. {2.2.5.1}

Near global (60S-60N) stratospheric ozone has declined from pre-1980 values, with estimated declines of about 2.5% over the period 1980-1995. Present day (2014-2017) near global annual column average is about 2.2% below the 1964-1980 average (*high confidence*). {2.2.5.2}

Since the mid-1970s, free tropospheric ozone has increased by 2-10% per decade in most regions of the northern mid-latitudes and tropics sampled (*high confidence*), while surface trends are variable. It has also generally increased at the surface and in the free troposphere of the more poorly sampled Southern Hemisphere (*medium confidence*). When compared to limited historical data representative for the mid-20th Century, tropospheric ozone has increased by 30-70% in the northern hemisphere and tropics (*medium confidence*). {2.2.5.3}

Satellite data and in-situ records indicate that Aerosol Optical Depth exhibits predominantly decreasing trends since 2001 in most regions (*high confidence*) but in some regions, in particular Southern Asia, is increasing. Trends are most pronounced in fine-mode AOD pointing with *high confidence* to a primarily anthropogenic cause that is decreasing significantly in the global mean. Atmospheric aerosols sampled by glacier ice show increasing trends for the industrial era until the last quarter of the 20th century (*medium confidence*). {2.2.6}

There is *high confidence* that land use activities increased global albedo with an equivalent radiative forcing of $-0.15 \pm 0.10 \text{ W m}^{-2}$ since the mid-19th century. Other factors related to land use change have a potentially large impact on the radiative balance, but there is a substantial range in estimates of the radiative forcing, and no agreement on the sign of the effect, and hence *low confidence*. {2.2.7}

The net radiative effect of all observed changes in climate forcings is positive since 1850, except for short periods (up to a few years in duration) following moderate to large volcanic eruptions (*high confidence*), and the rate of change has increased in the last 30 years (*high confidence*). {2.2.8, chapter 7}

Changes in key indicators of climate change

Directly observed changes in key large-scale atmospheric, oceanic, cryospheric and biospheric indicators of climate are unequivocal. Many of the assessed indicators are now in states not seen for centuries, and some have exceeded conditions over many millennia. The climate system as a whole has shifted significantly and unusually rapidly since industrialization. {2.3}

Temperature changes

Global Mean Surface Temperatures of the deep past were at times both higher and lower than has been observed during the instrumental period. GMST relative to pre-industrial ranged from at least 13°C warmer during the early Eocene climate optimum (around 50 Ma) to $6^\circ \pm 2^\circ\text{C}$ colder for the last glacial maximum (around 21-19 ka) (*medium confidence*). During the last inter-glacial, GMST averaged over the warmest millennium was $1^\circ \pm 0.5^\circ\text{C}$ warmer than pre-industrial (*medium confidence*) and *very likely* did not exceed 2°C higher. {2.3.1.1.1}

Surface temperatures over the Northern Hemisphere, and possibly also the Southern Hemisphere, have tended to gradually decrease over the past several thousand years (*medium confidence*) culminating in the Little Ice Age, which was globally the coldest multi-century interval of the Holocene (*high confidence*). This cooling trend was reversed during the mid 19th Century (*high confidence*). The GMST of the past 50 years was likely warmer than any 50 year mean of the Medieval Climate Anomaly (*high confidence*), and *more likely than not* was the warmest 50-year period of the Common Era (*medium confidence*). {2.3.1.1.2}

It is certain that global mean surface temperatures have increased since the late 19th century. Each of the last four decades has been warmer than any decade that preceded it. The total estimated warming of GMST from 1850-1900 to 1995-2014, is 0.83 °C (0.72 – 0.97 °C). The estimated warming of GMST from 1850-1900 to 2009-2018 is 0.99 °C (0.86 – 1.14 °C), and from 1850-1900 to 2002-2018 is 0.95 °C (0.81 – 1.09 °C). Estimated total change from a linear trend fit over the 1880-2018 period is 1.01 °C (0.73 – 1.28 °C), with 1980-2018 trends of 0.70 °C (0.56 – 0.88 °C). {2.3.1.1.3}

Estimating changes in Global Surface Air Temperature requires application of an adjustment factor to available GMST estimates to account for physically well-understood differences to maintain marine boundary layer stability. This adjustment factor is estimated as 1.05 (1.03-1.07), increases uncertainty and thus lowers confidence. Estimated GSAT changes are 0.88 °C (0.75 – 1.04 °C) over 1850-1900 to 1995-2014 and 1°C (0.83-1.17°C) for 1850-1900 to 2009-2018. Estimated total changes from a linear trend fit in GSAT are 1.07 °C (0.75-1.37°C) for 1880-2018, and 0.74 °C (0.58-0.94°C) for 1980-2018. {2.3.1.1.3}

It is *virtually certain* that diurnal temperature range decreased over the 1951-2012 period. It is *likely* that most of the decrease occurred over 1960-1980 with little change since. Change in DTR is small relative to increases in land surface air temperatures, and hence it is *virtually certain* that both daily maximum and daily minimum temperatures have increased globally since 1950. {2.3.1.1.3}

The troposphere has warmed since the mid-20th century. Lower tropospheric temperature has warmed at a rate of 0.13-0.23°C per decade over both the 1961-2018 and 1980-2018 periods. There is *medium confidence* that temperatures in the tropical upper troposphere have warmed faster than those at the surface since 2001, but *low confidence* in the magnitude and vertical extent of the difference, and in changes prior to 2001. It is *virtually certain* that the tropopause height has risen over the 1980-2018 period but there is *low confidence* in the size of the change, or whether the rate of change has changed in the later part of the period. {2.3.1.1.4}

It is *virtually certain* that the lower stratosphere has cooled since the mid-20th century, with a linear trend-based cooling rate of 0.22-0.29°C per decade over both 1960-2018 and 1980-2018. However, most data sets show that lower stratospheric temperatures have stabilized since the mid-1990s with no significant cooling over the last 20 years. It is *likely* that middle and upper stratospheric temperatures have decreased since 1980, but there is *low confidence* in the magnitude. {2.3.1.1.4}

Hydrological cycle changes

Positive trends in global land precipitation have been balanced by evaporation losses associated with the increase in global temperature, implying a precipitation-evaporation (P-E) trend close to zero. This is consistent with the observed global streamflow trends (a commonly used proxy for P-E over land) and the observed increases in specific humidity and total column water vapour. {2.3.1.2}

Prior to the Common Era the available data on hydroclimate change and variability are sparse and there is *low confidence* in large-scale reconstructions. Hydroclimate during the Common Era is dominated by regional heterogeneity, generally precluding definitive statements about long-term changes at continental and larger scales. {2.3.1.2.1}

Observations show a *very likely* increase in near surface specific humidity over both land and ocean since the 1970s. A *very likely* decrease in relative humidity was observed over land areas since 2000, particularly over midlatitude regions of the Northern Hemisphere. {2.3.1.2.2}

Positive trends in global total column water vapour are *likely* during the satellite era, although changes in the observing system bring uncertainty and, therefore, *medium confidence* in estimation of the magnitude. *Low confidence* in longer-term trends arises from uncertainties in the sea surface temperature (SST)-total column water vapour relationship and in the reliability of current surface-only reanalyses, particularly during the first half of the 20th Century. {2.3.1.2.3}

Global land precipitation has *likely* increased since the beginning of the 20th century (*medium confidence*), with a faster increase in global precipitation since 1980s (*medium confidence*). {2.3.1.2.4}

Observed trends in globally averaged P-E are small. Trends based upon reanalysis products are not consistent with P-E changes implied from observations and proxy data, giving *low confidence* in the assessment of global P-E trends. {2.3.1.2.5}

It is *likely* that there has been no significant trend in globally averaged streamflow over the last century. The magnitude of the observed trends varies regionally, with slightly more globally gauged rivers experiencing significantly decreasing flows than significantly increasing flows (*medium confidence*). {2.3.1.2.6}

Atmospheric circulation changes

There has been a *very likely* widening of the Hadley Circulation since the 1980s, although there is only *medium confidence* in the magnitude. This has been accompanied by a *likely* strengthening of the

Hadley Circulation, particularly for the northern hemisphere cell (*medium confidence*). {2.3.1.3.1}

Global Monsoon precipitation has increased over the last 40 years, mainly due to a significant increasing trend in the Northern Hemisphere Summer Monsoon precipitation (*medium confidence*). However, GM precipitation has exhibited large multi-decadal variability over the last century, particularly for the Southern Hemisphere (*low confidence*). {2.3.1.3.2}

It is *likely* the extratropical jet has been shifting poleward since the 1980s and that the mid-latitude jet meandering has intensified in the Northern Hemisphere during summer and autumn since the 1990s (*medium confidence*). The extratropical jets shifted poleward in the Northern Hemisphere during the mid-Holocene (*low confidence*) and the Medieval Climate Anomaly (*medium confidence*) to latitudes that *likely* were similar to those of recent decades. {2.3.1.3.3}

Since the 1970s a worldwide weakening of near surface wind has *likely* occurred over land. Over oceans, near-surface winds *likely* strengthened over 1980-2000, but divergent estimates lead to *low confidence* in globally aggregated behaviour of ocean surface winds thereafter. {2.3.1.3.4}

It is *likely* that the northern stratospheric vortex has weakened since the 1980s, and its location has shifted more frequently toward the Eurasian continent. {2.3.1.4.1}

The balance of evidence from observations suggests a strengthening of the stratospheric Brewer Dobson Circulation in last few decades of somewhat uncertain magnitude (*low confidence*). {2.3.1.4.2}

In recent decades the QBO amplitude has *likely* decreased in the lower stratosphere. There is *low confidence* owing to limited studies on its long-term changes. {2.3.1.4.3}

Changes in the cryosphere

Since the late 1970s, Arctic sea ice extent has decreased in both summer ($12.9 \pm 2.4\%$ per decade) and winter ($2.7 \pm 0.5\%$ per decade). Arctic sea-ice has become, younger, thinner, and faster moving (*very high confidence*). Snow thickness on sea ice has also decreased in the western and central Arctic (*medium confidence*). Proxy indicators show that Arctic sea ice has fluctuated on multiple time scales. Arctic Sea-ice area was *about as likely as not* last as low as today in the mid-holocene. {2.3.2.1.1}

Antarctic sea ice extent has experienced an increase of $1.7\% \pm 0.2\%$ per decade over 1978-2015, followed by significantly lower levels since 2016 (*high confidence*). There remains *low confidence* in all aspects of Antarctic sea-ice prior to the satellite era remains *low* owing to a paucity of evidence that is highly regional in nature and often contradictory. {2.3.2.1.2}

Substantial reductions in snow cover extent, particularly in the spring, have occurred in the Northern Hemisphere since 1978 (*high confidence*). These reductions continue a long-term decline seen since the early 20th Century in many locations (*medium confidence*). {2.3.2.2}

There is *very high confidence* that, with few exceptions, glaciers worldwide are currently shrinking. In most places, mountain glaciers are currently more extensive than during the early-middle Holocene (*high confidence*). Some Arctic mountain glaciers however are now smaller than they have been at any time during the present inter-glacial period (*medium confidence*). {2.3.2.3}

Greenland Ice sheet mass loss began at the end of the Little Ice Age and the rate of loss has increased in the 21st century (*high confidence*). The total mass loss from Greenland reached a historical record value during summer 2012 (-627 ± 89 Gt), which was *likely* the greatest one-season loss since at least 1840. The Greenland ice sheet was smaller than present during both the last interglaciation (*high confidence*) and the middle Holocene (*medium confidence*). {2.3.2.4.1}

No significant changes of the ice sheet mass balance away from zero in major parts of eastern Antarctica have been observed since the onset of modern measurements, while in western Antarctica a rapid decrease of mass has occurred in the Amundsen Sea Embayment sector and the Antarctic Peninsula (*medium confidence*). {2.3.2.4.2}

There is an observed warming of permafrost in all permafrost regions (*high confidence*) and evidence of permafrost thawing over the last 250 years in many places. {2.3.2.5}

Oceans

It is *virtually certain* that the global ocean warmed substantially between 1971 and present. There is *high confidence* that layer contributions are approximately 60% (0-700m), 30% (700-2000m) and 10% (2000-6000m). Current rates of heat content change are greater than at any point since the last deglaciation (*medium confidence*). Confidence in this assessment is strengthened by consistent closure of both global sea level and energy budgets. {2.3.3.1, 7.2.2.2, Cross-Chapter Box 9.2}.

It is *extremely likely* that near-surface and subsurface salinity changes have occurred across the global ocean since the mid-20th century. The salinity pattern implying that fresh regions becoming fresher, and salty regions saltier, is supported by all available observational studies that have considered salinity change since the modern instrumental record began in the 1950s. {2.3.3.2}

There is *high confidence* that the rate of Global Mean Sea Level rise is accelerating since the 19th century and that the rate of GMSL rise for the 20th century is higher than at any point in at least the past two millennia. The GMSL linear rate of increase is 1.6 (1.0-2.1) mm per year for 1901-2015 (with a *likely* acceleration from -0.002 - 0.019 mm yr⁻²); 1.5 (0.9-2.1) mm per year for 1901-1990; 2.1 (1.8 -2.4) mm per year for 1971-2015; 3.2 (2.8-3.6) mm per year for 1993-2018; and 4.0 (3.7 -4.4) mm per year for 2006-2018. {2.3.3.3}

There is robust agreement from a limited number of direct observations for a rapid weakening of the AMOC beginning around 2005-2008 (*high confidence*) but records are of insufficient length to determine whether this is a significant trend or a manifestation of multi-decadal variability. There is agreement from proxy evidence for a longer-term decline in the AMOC, however, uncertainties in the robustness of the proxies and inconsistencies in the timing and duration of the associated AMOC decline means that there is *low confidence* in a weakened AMOC on the basis of proxy data. There is no evidence for a long term weakening in the AMOC on the basis of observed source water variability. {2.3.3.4}

The response of wind-driven circulation to changing wind patterns is inconsistent and thus there is *low confidence*. There is little evidence for a predicted strengthening of western boundary currents in response to strengthening winds. There is some evidence that increased wind forcing is leading to increased eddying of WBCs. {2.3.3.4}

Biosphere

There is *high confidence* that the amplitude of the seasonal cycle of atmospheric CO₂ has increased by as much as 50% poleward of 45°N since the early 1960s. {2.3.4.1}

All ocean basins are storing anthropogenic CO₂ and declining pH since the preindustrial era. Rapid deoxygenation is occurring across the Oceans with commensurate increases in oxygen minimum zones. It is *virtually certain* that the current rate of change in both Ocean pH and Ocean oxygenation is unprecedented in the available geological record. {2.3.4.2}

There is *high confidence* that the timing of lifecycle events in marine species has advanced over the

past several decades. {2.3.4.3}

Based on consistency between ecological and instrumental records, there is *high confidence* in observed changes in land ecosystem composition and species distributions over much of the globe during the past century, with species generally moving poleward and upslope. The accelerated rates of change during the late 20th and early 21st century are unusual in the context of the current Holocene interglacial (*medium confidence*). {2.3.4.4}

There is *medium confidence* that terrestrial vegetation across the globe has increased since the early 1980s. {2.3.4.5}

Based on multiple independent analyses of in situ, satellite, and phenological data, there is *high confidence* that the length of the growing season has increased over much of the extratropical Northern Hemisphere since at least the mid-20th century. There are no widely available and reliable sources of proxy information that permit a longer-term context for growing season length changes. {2.3.4.6}

Changes in modes of variability

Most assessed modes of internal climate variability modulate inter-annual to decadal climate, but the majority do not exhibit substantial long-term trends since the early 20th Century (*high confidence*). Uncertainty in millennial scale behaviour of most modes reduces confidence in assessments of the significance of observed changes {2.4}

There is no clear evidence for a recent sustained shift in ENSO or associated features such as the Walker Circulation, or in teleconnections associated with these, beyond the range of variability on decadal to millennial timescales. There is *medium confidence* that both ENSO amplitude and the frequency of high-magnitude events since 1950 is higher than over the pre-industrial period from 1850 as far back as 1400, but *low confidence* that it is outside the range of variability over periods prior to 1400. A high proportion of El Niño events in the last 20-30 years have been centred in the central, rather than eastern Pacific (with an associated westward shift in the Walker Circulation) but there is *low confidence* that this represents a long-term change. {2.4.1}

There is *low confidence* in any changes to long-term Indian Ocean Dipole variability. Understanding of the Indian Ocean Basin mode of variability trends during the instrumental period is also limited due to underlying Indian Ocean warming trends. {2.4.2}

It is *very likely* that the Atlantic Multidecadal Oscillation/Variability index has not exhibited any significant trend during the instrumental period (*high confidence*). However, instrumental records may not be long enough to distinguish any oscillatory tendency from trends. There is *low confidence* in paleo-AMO/V reconstructions due to a paucity of high-resolution records. {2.4.3}

The PDO is dominated by large decadal variability with persistence of positive and negative phases over the last 150 years with no overall trend (*high confidence*). Since AR5 a shift from positive to negative phase of the PDO has occurred (*high confidence*). There is *low confidence* in paleo-PDO reconstructions due to poor agreement among the various available estimates. {2.4.4}

Strong positive trends for the NAM/NAO/AO indices were observed since 1960, which have weakened since the 1980s (*high confidence*). The NAO variability in the instrumental record was *likely* not unusual in the millennial and multi-centennial context. {2.4.5.1}

Over the instrumental period there has been a robust positive trend in the SAM index, particularly since 1970 (*high confidence*). It is *likely* that the recent trend in the SAM is unprecedented in the past several centuries, although *medium confidence* arises due to the differences between reanalyses and

1 proxy records before the 1950s. {2.4.5.2}

2

3 It is *likely* that the Atl3 index has exhibited a significant positive trend since the beginning of
4 instrumental period, but this trend has become weaker since the 1970s. For the AMM, no significant
5 trend was identified. {2.4.6}

6

7

2.1 Introduction

This chapter performs an assessment of the observational evidence for large scale changes in the climate system across all of its components. As such, it combines many aspects covered in Chapters 2 through 5 of the AR5 WGI contribution (IPCC, 2013) that, taken together, supported a finding of unequivocal warming of the climate system. The adopted AR6 structure differs substantively from that in AR5, with Chapters 5 through 9 considering key climate processes, and Chapters 10 through 12 regional aspects and extreme events (Section 1.8). These assessments are performed in a holistic manner that integrates observations, models, theory and projections. To avoid duplication the current chapter hence focuses upon observed changes in climate system drivers and changes in key selected indicators of climate change and in large-scale modes of variability (Cross-Chapter Box 2.1) which, taken together, allow for an assessment of important changes in the global climate system in an integrated manner. This Chapter supports Chapters 3 and 4 which respectively consider model assessment / detection and attribution, and future climate projections for subsets of these same indicators and scales (Cross-Chapter Box 2.1). This chapter does not consider changes in observed extremes, as they are assessed in Chapter 11.

The chapter starts out in Section 2.2 by assessing current knowledge of historical changes in key natural and anthropogenic forcing agents. This provides a contextual basis for the following assessment in Section 2.3 of changes in key indicators of the atmosphere (Section 2.3.1), cryosphere (Section 2.3.2), oceans (Section 2.3.3) and biosphere (Section 2.3.4). Section 2.4 provides an updated assessment of our knowledge of modes of variability and their changes. Section 2.5 closes by highlighting key remaining uncertainties in the characterisation of large-scale observed changes in the climate system.

Throughout the chapter recourse is made to paleoclimate, in-situ, satellite and reanalysis data products to the extent possible, which is dependent upon the variable being considered (Section 1.4). Use of common periods ranging from the deep past through to the modern era is applied to the extent permitted by available data records (Cross-Chapter Box 1.3). In all cases the narrative proceeds from the deep past through to the present. Each sub-section starts by highlighting the key findings from AR5 and, where relevant, the three AR6 special reports, proceeds to outline the new evidence arising from a combination of: i) new findings reported in the literature; and ii) recently observed changes, and then closes with a new assessment.

Where they are directly calculated as part of the assessment, all trends have been calculated using a common metric following the approach adopted in box 2.2 of Chapter 2 of AR5 (Hartmann et al., 2013). Statistical significance of trends is assessed at the two-tailed 90 or 95% confidence level. Limited recourse is also made to published analyses which have employed a range of methodological choices to determine trends. In each case the method / metric is clearly stated. In addition to trends, consideration is also made of changes between various timeslices / periods in performing the assessments throughout the chapter. These timeslices and their short names are summarised in Table 2.1.

[START TABLE 2.1 HERE]

Table 2.1: Summary of the range of timeslice periods considered in the present chapter to assess the unusualness or otherwise of aspects of the present-day climate state. Note that not all periods are considered for all variables assessed as consideration is driven by the availability of evidence. (Cross-Chapter Box 1.3)

Period name	Short acronym	timeslice
Paleocene-Eocene Thermal Maximum	PETM	55.5-55.3Ma
Early Eocene Climate Optimum	EECO	52-50Ma

Mid-Pliocene Warm Period	MPWP	3.3-3Ma
Last Interglaciation	LIG	129-116ka
Last glacial maximum	LGM	21-19ka
Last deglaciation		18-11ka
Post-glacial		8.2-0ka
Mid-holocene		6.0 +/-0.2ka
Medieval Climate Anomaly	MCA	950-1250CE
Little Ice Age	LIA	1450-1850CE
Pre-Industrial / early industrial	PI	1750/1850-1900CE
Present		1995-2014CE

[END TABLE 2.1 HERE]

2.2 Changes in climate drivers

2.2.1 Orbital forcing

Slow periodic changes in the Earth's orbit around the Sun cause well-known cycles in the seasonal and latitudinal distribution of insolation at the top of the atmosphere, as summarized in AR4, Box 6.1 (Jansen et al., 2007). Precise calculations (*high confidence*) of insolation based on orbital parameters are available for each latitude and month back for the past several millions of years (Berger and Loutre, 1991). Orbital forcing involves multi-millennial-scale periodic changes in the obliquity (tilt) of the Earth's axis, the eccentricity of its orbit, and the precession of equinoxes relative to its annual revolution around the Sun. Slow changes in orbital configuration result in unique combinations of the duration and intensity of seasons and their latitudinal distribution through time. Over the last millennium it has had a small impact on seasonality and essentially no discernible effect on annual mean global radiative forcing.

2.2.2 Solar forcing

AR5 presented evidence for solar variability and its influence on climate over multiple time scales. For variability over the entire Holocene, AR5 concluded multi-millennial fluctuations generally are $< 1 \text{ W m}^{-2}$, but AR5 did not include an assessment of the confidence in reconstructions of multi-millennial TSI fluctuations. For multi-decadal to centennial variability over the last millennium, AR5 emphasized TSI reconstructions that show little difference ($< 0.1\%$) between the late 20th Century and the Maunder Minimum (1675-1715) when the Sun was very quiet, but again did not provide a confidence level for these reconstructions. Over recent decades, AR5 further concluded that the best estimate of radiative forcing due to TSI changes for period of 1750 to 2011 is 0.05 to 0.10 W m^{-2} and that TSI *very likely* changed by -0.04 (-0.08 - 0.00 W m^{-2}) between 1986 and 2008. These values were assigned a *medium confidence* level, and *high confidence* for the last three decades.

Since AR5, new multi-millennial reconstructions of solar irradiance extending back 9000 years have emerged (Wu, C.-J. et al., 2018; Wu, C. J. et al., 2018). These are based on updated cosmogenic isotope datasets and their improved production and deposition models (Polunin et al., 2016). Long-term changes with inferred amplitudes up to 1.5 (1.4 - 2.1) W m^{-2} have been identified in the 11-year averaged irradiance reconstructions from the cosmogenic time series. This range includes the changes due to the varying amplitude of the 11-year activity cycle (up to about 0.5 - 0.6 W m^{-2}) as well as the quasi-periodic, relatively stable, roughly 2400-year-period Hallstatt cycle (Usoskin et al., 2016) and the more intermittent Suess/de Vries cycle of about 210 years (Usoskin, 2017), expressed as recurrence of Grand Minima clusters within the Hallstatt cycle.

For the past millennium, the reconstructed TSI time series recommended for CMIP6/PMIP4

simulations (Figure 2.1a) is based on the radiocarbon dataset scaled to the CMIP6 historical forcing (Jungclaus et al., 2017). It features little difference in TSI averaged over the MCA (1360.91 W m^{-2}) compared with the LIA (1360.63 W m^{-2}) and PI periods (1360.69 W m^{-2}), and due to the prescribed scaling, probably underestimates the variability in total and near-UV irradiance (Jungclaus et al., 2017; Lean, 2018a; Wu, C.-J. et al., 2018; Wu, C. J. et al., 2018). In addition, for sensitivity tests, the PMIP4 recommendation includes the TSI reconstruction by Egorova et al. (2018), with a larger TSI range of $3\text{--}7 \text{ W m}^{-2}$.

Evidence for TSI changes during the industrial era (Figure 2.1b) has strengthened since AR5, and confirms a small contribution to global forcing (Section 7.3.4.4). Variations in solar activity are known from ground-based observations, mostly since the late 19th Century, and more accurately from direct space-borne observations since the 1970s. On the basis of the recommended CMIP6 forcing (Matthes et al., 2017), the long-term trend in TSI averaged over solar cycles was increasing during the first seven decades of the 20th Century, by 0.8 W m^{-2} from 1900–1980, and decreasing since, by -0.04 W m^{-2} between 1986 and 2009 [to update value through 2018 in SOD].

[START FIGURE 2.1 HERE]

Figure 2.1: Time series of solar and volcanic forcing for two periods: (a, c) past 2500 years, and (b,d) past 165 years. (a) Total solar irradiance (TSI) reconstruction recommend for CMIP6/PMIP4 millennial experiments based on the radiocarbon dataset before 1850 (blue) scaled to the CMIP6 historical forcing after 1850 (black) (Jungclaus et al., 2017). (b) TSI time series (6-month running averages) from CMIP6 *historical* forcing as inferred from sunspot numbers (black) (Matthes et al., 2017), with CMIP5 forcing based on (Lean, 2000) and (Wang et al., 2005) for comparison (red). (c) Volcanic forcing represented as reconstructed stratospheric aerosol optical depth (SAOD) at 550 nm. Estimates covering 500 BCE to 1900 CE (blue) are from Toohey and Sigl (2017) and those for the period 1850–2015 (black) are from (Luo et al., 2018). SAOD was translated to global volcanic forcing based on the scaling of $\text{RF} = -25 \times \text{SAOD}$, as in AR5, which differs somewhat from the formulation for effective radiative forcing presented in Section 7.3.4.4. (d) SAOD reconstruction from CMIP6 (v 4) (black) (Luo et al., 2018), with CMIP5 forcing for comparison (red) (Sato et al., 1993). Note the change in y-axis range between panels c and d.

[END FIGURE 2.1 HERE]

Despite progress in the physical understanding and modelling of solar irradiance variations (Shapiro et al., 2017; Yeo et al., 2017b), there remain significant uncertainties in reconstructions of solar forcing prior to satellite measurements (Krivova, 2019; Lean, 2018b). One critical issue is the magnitude of the secular variability (i.e., the entire potential range of the variability between the grand-minima episodes, such as the Maunder Minimum, and periods of higher solar activity, such as the recent Modern Grand Maximum). Another issue is the variability of the UV band, typically from 200–400 nm, because of the difficult inter-calibration of spectral solar irradiance instruments that are generally short-lived and lack stability (Ermolli et al., 2013; Haberreiter et al., 2017; Yeo et al., 2015). Because its effect on ozone photochemistry in the middle atmosphere can be transferred to the troposphere and surface, the solar forcing recommendations for CMIP6 include a 16% stronger contribution to TSI variability (i.e., 0.19 W m^{-2}) in this wavelength range for recent solar cycles compared to CMIP5 (Matthes et al., 2017).

Another uncertainty arises from energetic particles, mainly protons and electrons, originating from the Sun, the magnetosphere or from beyond the solar system in the case of galactic cosmic rays. The flux of the energetic particles can affect the ionization levels in the polar middle and upper atmosphere leading to changes in the chemical composition, such as odd nitrogen and odd hydrogen causing catalytic ozone destruction (Sinnhuber et al., 2012). They can also affect ionization in the lower stratosphere and troposphere where the connection to cloud production is still under debate, although

laboratory-based studies have confirmed the existence of ion-mediated aerosol formation and growth rates (Dunne et al., 2016).

To conclude, changes in TSI during the industrial era are not unusual in context of at least the past 9000 years (*medium confidence*).

2.2.3 Volcanic aerosol forcing

AR5 concluded that, on inter-annual time scales, the radiative impacts of volcanic aerosols are a dominant natural driver of climate variability, with the greatest impact occurring within the first few years following an eruption. Reconstructions of radiative forcing by volcanic aerosols used in PMIP3 simulations and in AR5 featured short-lived perturbations of a range of magnitudes, with events greater than 1 W m^{-2} (annual mean) occurring on average every 35–40 years, although *confidence* levels were not assigned for long-term volcanic aerosol reconstructions. AR5 found *robust evidence* from satellite observations of the solar irradiance and volcanic aerosols for a near-zero ($-0.1 \pm 0.1 \text{ W m}^{-2}$) change in forcing from 1980 to 2011.

Since AR5, improved dating and synchronization of sulphate records from polar ice sheets, has resulted in a more complete and continuous representation of volcanic eruptions reaching back 2500 years (Sigl et al., 2015). This updated eruption history has been used to estimate volcanic stratospheric sulphur injections and comprehensive aerosol optical properties (Toohey and Sigl, 2017) (Figure 2.1c), the latter recommended for use in PMIP4 experiments (Jungclaus et al., 2017). This volcanic forcing reconstruction extends those available for AR5 by 1000 years and features multiple strong events that were previously misdated, underestimated or absent, particularly before ~1500 CE. It shows that, over the past 2500 years, the length of time between eruptions large enough to decrease global radiative forcing by more than 1 W m^{-2} averaged around 30 years, whereas 100-year-long periods that lack such large eruptions occurred once every 400 years on average. The reconstructed stratospheric aerosol optical depth averaged over the MCA (0.012) was lower than for the LIA (0.017) and similar to the PI period (0.011), although the estimated uncertainty is large: $\pm 30\%$ 1SD (Toohey and Sigl, 2017). Over the past 100 years, loading of volcanic aerosols averaged 14% lower than the mean of the previous 24 centuries (back to 500 BCE), and well within the range of centennial-scale variability. Isotopic analysis of sulphate from Antarctic glacier ice generally corroborates a stratospheric pathway for aerosols from strong tropical eruptions in volcanic forcing reconstructions (Gautier et al., 2019).

Direct observations of volcanic gas-phase sulphur emissions (mostly SO_2), sulphate aerosols, and their radiative impacts are available from a variety of sources (Kremser et al., 2016). Since AR5, SO_2 emissions from explosive eruptions have been estimated from satellite and in-situ measurements (Brühl, 2018; Carn et al., 2016; Höpfner et al., 2015; Neely and Schmidt, 2016). Satellite observations of volcanic aerosol are available from 1978 to present, albeit with considerable gaps. Complementary in situ measurements (Deshler, 2008) provide information on aerosol composition and size as well as vertical distribution. Merged datasets (Thomason et al., 2018) and sparse ground-based measurements (Stothers, 1997) form the basis for volcanic forcing estimates back to 1850. In contrast to the volcanic forcing datasets recommended for CMIP5 historical simulations (Ammann et al., 2003; Sato et al., 1993), the volcanic forcing time series used in CMIP6 simulations (Luo et al., 2018; Figure 2.1d) includes a more comprehensive set of optical properties including latitude-, height- and wavelength-dependent aerosol extinction, single scattering albedo and asymmetry factor. While there have been no major eruptions since that of Mt. Pinatubo in 1991, a series of small-to-moderate eruptions since 2000 produced significant negative radiative forcing (Andersson et al., 2015; Schmidt et al., 2018; Solomon et al., 2011; Vernier et al., 2011), which was discussed in the AR5 (Myhre et al., 2013), but not included in CMIP5 simulations (Fyfe et al., 2013; Santer et al., 2014; Schmidt et al., 2014).

Uncertainty in the radiative forcing caused by volcanic eruptions arises from a number of sources.

Satellite observations of aerosol extinction during recent eruptions have uncertainties on the order of 10-15% (Bourassa et al., 2012; Vernier et al., 2011), with additional uncertainties due to saturation effects following large eruptions (Thomason et al., 2018). Uncertainty in the timing of eruptions in the proxy record is ± 2 years back 1500 years and at most ± 5 years before. Systematic uncertainties related to the scaling of sulphate abundance in glacier ice to radiative forcing have been estimated to be $\sim 35\%$ (1SD) (Hegerl et al., 2006).

To conclude, changes in volcanic forcing by aerosol loading during the industrial era are not unusual in the context of at least the past 2500 years (*medium confidence*).

2.2.4 Well-mixed Greenhouse gases

2.2.4.1 Atmospheric greenhouse gases in geologic and glacier-ice archives - deep past to common era

AR5 assigned *medium confidence* to the values of atmospheric CO₂ concentrations (mixing ratios) during the warm geological periods of the early Eocene and Pliocene, and it quantified the fluctuations in CO₂, CH₄, and N₂O in ancient atmosphere trapped glacier ice. AR5 concluded with *very high confidence* that, by 2011, the concentrations of these well-mixed gasses in the modern atmosphere exceeded the range of those in ice cores, which extend back 800,000 years. AR5 also concluded with *very high confidence* that the observed rates of increase of the greenhouse gasses were unprecedented over at least the past 22,000 years.

2.2.4.1.1 CO₂ in the deep past (500 Ma to 800 ka)

New evidence (Figure 2.2) since AR5 has reinforced the understanding of long-term changes in atmospheric CO₂ during the past 500 million years (Ma) (Anagnostou et al., 2016; Chalk et al., 2017; Foster et al., 2017; Martínez-Botí et al., 2015; Sosdian et al., 2018; Super et al., 2018; Witkowski et al., 2018; Anagnostou et al., 2019; Bolton et al., 2016; de la Vega et al., 2019). Atmospheric CO₂ has gradually declined over the last 50 million years at a rate of ~ 16 ppm/Ma. During the Triassic atmospheric CO₂ mixing ratios reached a maximum of $\sim 3700 \pm 1600$ ppm (200-220 Ma), and were near PI levels during at least one protracted interval (350-250 Ma). The last time atmospheric CO₂ mixing ratio was as high as present was very likely over 2 million years ago during the late Pliocene. The last time CO₂ mixing ratio was as high as 1000 ppm (reached by pessimistic SSP scenarios by 2100) was around 33.5 million years ago, immediately prior to the Eocene-Oligocene transition that was associated with the first major advance of the Antarctic Ice Sheet (Anagnostou et al., 2016; Hollis et al., 2019a; Pagani et al., 2011; Pearson et al., 2009). The highest CO₂ of the last 60 million years was >1100 ppm and likely occurred during the Early Eocene Climatic Optimum around 50 Ma. Since AR5, several studies have focused on the rate of CO₂ emissions that drove the PETM - a period of rapid climate change 56 million years ago. During the PETM CO₂ rapidly rose from ~ 800 to ~ 2200 ppm in perhaps as little as 4000 years although there remains *low confidence* in the details (Gutjahr et al., 2017a; Kirtland Turner, 2018; Zeebe et al., 2016a). Estimated rates of CO₂ emission during this event range from 0.3 to 1.5 PgC per year (Gingerich, 2019) – around 7-35 times slower than current anthropogenic rates (Chapter 5). Following the MPWP, atmospheric CO₂ mixing ratio generally decreased at a rate of ~ 40 ppm/Ma. Related to the shift of glacial-interglacial cycle frequency from 40 to 100 thousand years during the Mid Pleistocene Transition (MPT) at 0.8-1.2 million years ago, marine sediment records show no significant changes in atmospheric CO₂ during interglacials, but a decrease of glacial CO₂ across the MPT (Chalk et al., 2017; Hönlisch et al., 2009). However, these results are not confirmed by ice core records due to a lack of sufficiently ancient ice with continuous stratigraphy (Brook and Buizert, 2018; Higgins et al., 2015).

[START FIGURE 2.2 HERE]

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Figure 2.2: The evolution of atmospheric CO₂ through the last 450 million years reconstructed from ice core and marine sediment records. (a) 0 to 450 million years ago, data from (Foster et al., 2017) and references therein (and as detailed in subsequent panels). Stomatal based estimates in grey circles, paleosol carbonate $\delta^{13}\text{C}$ based estimates as blue crosses, $\delta^{13}\text{C}$ of phytane as brown circles. Dark green line is non-parametric LOESS fit (see Foster et al., 2017) through the data presented here, red band is uncertainty envelope at 68% uncertainty. See (Royer et al., 2001; Witkowski et al., 2018) for a description of the basis of these CO₂ proxies. (b) 0 to 58 million years ago, boron isotope based CO₂ in blue (Anagnostou et al., 2016, 2019; Sosdian et al., 2018), $\delta^{13}\text{C}$ of phytane in brown (Witkowski et al., 2018), $\delta^{13}\text{C}$ of alkenones in black (Pagani et al., 2005, 2011; Super et al., 2018). The basics of these CO₂ proxies are described in (Foster and Rae, 2016a; Pagani, 2002; Witkowski et al., 2018). (c) 0 to 3500 thousand years ago, ice core data shown in black (Bereiter et al., 2015) and boron isotope based CO₂ in blue (Chalk et al., 2017a; de la Vega et al., 2019; Hönlisch et al., 2009; Martínez-Botí et al., 2015a; Sosdian et al., 2018) (d) 0 to 450 thousand years ago, ice core data shown in black (Bereiter et al., 2015) and boron isotope based CO₂ in blue (Chalk et al., 2017a; de la Vega et al., 2019).

[END FIGURE 2.2 HERE]

2.2.4.1.2 Glacial-interglacial greenhouse gas fluctuations (CO₂, CH₄, N₂O) from 800 ka

Since AR5, the number of ice cores and their temporal resolution have improved (Figure 2.3), especially for the last 60 ka. Combined with analyses of firn air, they increase confidence in the quantification of greenhouse gas concentrations prior to direct measurements, which begin only in the mid-20th Century.

CO₂

Since AR5, the accuracy of CO₂ records for the last 800 ka has improved (Bereiter et al., 2015). High-resolution analyses of CO₂ from glacier ice formed during the last glacial maximum and the last deglaciation show multiple century-scale fluctuations of up to 10-13 ppm over intervals of Centuries (Schmitt et al., 2012; Marcott et al., 2014; Ahn and Brook, 2014; Bauska et al., 2015). Although these rates are an order of magnitude lower increases observed in recent decades, they provide information on non-linear responses or abrupt climate-biogeochemical cycle feedbacks (Chapter 5). During warm interglacial periods, the rate of concentration change is much slower than those of glacial periods. After removing the multi-millennial CO₂ trend, millennial variations during the early Holocene are less than ~2 ppm, implying that natural CO₂ change rates during interglacial conditions are several orders of magnitude slower than recently observed changes (Ahn and Brook, 2014; Marcott et al., 2014).

Multiple high-resolution records for the last 2000 years prior to 1850 show CO₂ mixing ratios of 274-285 ppm. Offsets among the records are about 1%, but the long-term trends agree well and show coherent multi-centennial variations of about 10 ppm (MacFarling Meure et al., 2006; Ahn et al., 2012; Bauska et al., 2015; Rubino et al., 2016). Multiple records show CO₂ concentrations of 278.6 ± 2.0 ppm in 1750 and 286.9 ± 1.6 ppm in 1850. During the MCA CO₂ concentration increased by 5.0 ± 0.8 ppm, while CO₂ decreased by 4.6 ± 1.7 ppm during the LIA. The fastest rate of change confirmed by multiple studies is about 1-2 ppm per decade prior to 1850 (Ahn et al., 2012, 2019; Bauska et al., 2015).

CH₄

High-precision records of GHGs from glacier ice show that CH₄ concentrations over the past 110 ka are higher in the NH than in the SH, but closely correlated on centennial and millennial timescales (Brook et al., 1996). On timescales of tens of thousands of years, CH₄ concentrations are correlated with precession cycles of the earth's orbit (Brook et al., 1996). On millennial timescales, most abrupt climate changes observed in Greenland and other regions coincided with abrupt CH₄ changes (Rhodes et al., 2015, 2017; WAIS Divide Project Members., 2015). The variability of CH₄ on centennial

timescales during the early Holocene is not significantly different from that of the pre-industrial part of the late Holocene (Rhodes et al., 2013; Ana et al., 2018) (Yang et al., 2017). The global mean concentrations during 0-1850 CE range between 625 and 807 ppb. High-resolution ice core records from Antarctica and Greenland exhibit the same trends with an inter-polar gradient of 36-47 ppb (Mitchell et al., 2013b; Sapart et al., 2012). There is a long-term positive trend with rate of ~ 0.05 ppb per year during the last 2000 years which constitutes a continuation of the previous longer-term trend. The most rapid CH_4 changes prior to industrialisation are 30-50 ppb and take place over several decades. Global mean CH_4 concentrations estimated from Antarctic and Greenland ice cores are 726 ± 10 ppb in 1750 and 807 ± 10 ppb in 1850. The industrial era values are higher than at anytime during the last 800ka.

N₂O

New high-resolution records from glacier ice show N_2O concentration changes associated with abrupt climate changes during glacial periods and glacial terminations on millennial timescales (Flückiger et al., 2004; Schilt et al., 2010; Schilt et al., 2014). Among the most rapid changes is during the last glacial termination, when N_2O increased 30 ppb in 200 years (Schilt et al., 2014), which is an order of magnitude smaller than the modern rate (Section 2.2.4.2.3). During the Holocene, N_2O reached a minimum of 257 ± 4 ppb during 6-8 ka, but millennial variation is not clearly defined due to analytical uncertainty and/or ice core quality (Flückiger et al., 2002; Schilt et al., 2010). Recently acquired high-resolution records from Greenland and Antarctica for the last 2000 years commonly show multi-centennial variations, although the magnitudes differ (Ryu et al., 2019b). Common Era concentrations range from 260-270 ppb with centennial variations of ~ 5 -10 ppb (Figure 2.3). At least three ice cores exhibit a short-term minimum at ~ 600 CE (MacFarling Meure et al., 2006; Ryu et al., 2019a). It is *very likely* that industrial-era N_2O increase started before 1900 CE (MacFarling Meure et al., 2006; Machida et al., 1995; Ryu et al., 2019b; Sowers, 2001). Multiple ice cores show N_2O concentrations of 270.4 ± 3.4 ppb in 1750 and 271.2 ± 4.0 ppb in 1850. The industrial era values are higher than at anytime during the last 800ka.

[START FIGURE 2.3 HERE]

Figure 2.3: Atmospheric greenhouse gas concentrations from ice cores. a) Low-resolution records during the last 800ka (Bereiter et al., 2015; Louergue et al., 2008; Lüthi et al., 2008; Schilt et al., 2010). b) High-resolution records during the last glacial termination (Köhler et al., 2017; Rhodes et al., 2015, 2017; WAIS Divide Project Members., 2015). c) Multiple high-resolution records over the common era. CO_2 data from (SIEGENTHALER et al., 2005; MacFarling Meure et al., 2006; Lüthi et al., 2008; Ahn et al., 2012; Ahn et al., in prep.). Antarctic ice core CH_4 records from (MacFarling Meure et al., 2006; Mitchell et al., 2013b); N_2O concentrations from Greenland and Antarctic ice cores (Machida et al., 1995; Sowers, 2001; Flückiger et al., 2002; MacFarling Meure et al., 2006; Schilt et al., 2010; Ryu et al., 2019b).

[END FIGURE 2.3 HERE]

In summary, the mixing ratio of CO_2 has slowly fluctuated by at least 2000 ppm over the last 500 million years (*medium confidence*). The last time CO_2 levels were similar to the present-day was over 2 million years ago (*high confidence*). While, mixing ratios of CH_4 , and N_2O prior to 800 ka are not known, it is certain that pre-industrial WMGHG mixing ratios were lower than present day levels and that current rates of change in the WMGHGs are unprecedented in at least the last 800 ka.

2.2.4.2 Modern measurements of well mixed GHGs

AR5 reported that it is certain that from 2005 to 2011 atmospheric burdens of a subset of well mixed greenhouse gases targeted by the Kyoto protocol (CO_2 , CH_4 , and N_2O) increased, reaching in 2011

levels of 390.5 ppm, 1803.2 ppb and 324.2 ppt. Growth rates of CO₂ and N₂O from 2005-2011 were comparable to those in 1996-2005, while CH₄ resumed increasing in 2007, after remaining nearly constant from 1999-2006.

In this section and for calculation of RF, surface global averages are used, since most observations are at the surface. Modern global means presented herein are determined from measurements representative of large well-mixed volumes of the lower troposphere. Measurements from sites subject to local or regional influence are excluded. Global averages that include sites subject to significant anthropogenic activities or influenced by strong regional biospheric emissions will typically be larger than those based distant from such sources, and require weighting accordingly (Table 2.2). This section focusses on global mean mole fractions estimated from global networks, and a historical dataset (Meinshausen et al., 2017a) for periods prior to the existence of global networks.

[START TABLE 2.2 HERE]

Table 2.2: Atmospheric global annual mean mixing ratios (dry-air mole fraction) of radiatively important gases estimated from various measurement networks or assimilations, in parts-per-million (ppm) (CO₂), parts-per-billion (ppb) (CH₄ and N₂O), or parts-per-trillion (ppt) (all others). Note that AGAGE, NOAA, and UCI global averages are calculated from measurements in the remote, unpolluted troposphere, while CMIP6 and WMO-GAW global averages include observations subject to regional and local influence.

Species	Rad. Eff.	2011	unc.	2017	unc.	Change	Change (%)	Network/Dataset
	W m ⁻² ppb ⁻¹	2011-2017						
CO ₂	1.38E-05	390.44	0.16	404.98	0.13	14.54	3.7%	NOAA
		390.20	0.16	405.12	0.16	14.92	3.8%	CSIRO
		390.9	0.2	405.5	0.2	14.6	3.7%	WMO-GAW
		390.94						CMIP6
CH ₄	3.63E-04	1803.2	1.2	1849.6	1.1	46.4	2.6%	NOAA
		1803.5	4.8	1849.4		45.9	2.5%	AGAGE
		1791.8	0.6	1842.2	3.0	50.4	2.8%	UCI
		1802.3	3.8	1847.2	3.8	44.9	2.5%	CSIRO
		1813	3	1859	3	46	2.5%	WMO-GAW
N ₂ O	3.0E-03	1813.1						CMIP6
		324.2	0.1	329.7	0.1	5.5	1.7%	NOAA
		324.7	0.1	330.3	0.1	5.6	1.7%	AGAGE
		324.0	0.5	329.6	0.5	5.6	1.7%	CSIRO
		324.3	0.2	329.9	0.2	5.6	1.7%	WMO-GAW
CFC-12	0.32	324.2						CMIP6
		527.0	0.4	508.4	0.4	-18.5	-3.5%	NOAA
		530.4	0.2	512.9	0.2	-17.5	-3.3%	AGAGE
		525.3	0.8	514.3	1.6	-11.0	-2.1%	UCI
		528.5						CMIP6
CFC-11	0.26	237.2	0.2	228.9	0.2	-8.3	-3.5%	NOAA
		237.5	0.1	228.5	0.1	-9.0	-3.8%	AGAGE
		237.9	0.8	226.1	1.0	-11.8	-5.0%	UCI
		238.3						CMIP6
CFC-113	0.3	74.5	0.1	70.9	0.1	-3.6	-4.8%	NOAA
		74.6	<0.1	71.0	<0.1	-3.6	-4.8%	AGAGE
		74.9	0.6	71.3	0.5	-3.6	-4.8%	UCI
		74.6						CMIP6
CFC-114	0.31	16.36		16.27		-0.09	-0.6%	AGAGE
		16.36						CMIP6
CFC-115	0.2	8.39		8.56		0.17	2.0%	AGAGE
		8.39						CMIP6
CH ₃ CCl ₃	0.07	6.34	0.1	2.20	0.1	-4.14	-65%	NOAA
		6.25	0.1	2.23	0.1	-4.02	-64%	AGAGE
		6.8	0.6	2.49	<0.1	-4.30	-63%	UCI
		6.31						CMIP6
CCl ₄	0.174	86.9	0.3	80.2	0.3	-6.67	-7.7%	NOAA ²

		85.3	0.1	79.0	0.1	-6.34	-7.4%	AGAGE
		87.8	0.6	79.8	0.6	-8.00	-9.1%	UCI
		86.1						CMIP6
halon-1211	0.29	4.00		3.43		-0.57	-14.3%	NOAA
		4.10		3.49		-0.61	-14.9%	AGAGE
		4.05		2.49		-0.37	-9.1%	UCI
		4.05						CMIP6
halon-1301	0.30	3.18		3.26		0.08	2.5%	NOAA
		3.27		3.36		0.09	2.8%	AGAGE
		3.23						CMIP6
HCFC-22	0.21	212.6	1.2	240.8	1.2	28.2	13.3%	NOAA
		213.7	0.8	241.2	0.8	27.5	12.9%	AGAGE
		209.0	1.2	248.3	2.2	39.3	18.8%	UCI
		214.6						CMIP6
HCFC-141b	0.16	21.3	0.2	24.5	0.2	3.2	14.9%	NOAA
		21.4	0.1	24.4	0.1	3.0	14.1%	AGAGE
		20.8	0.5	26.0	0.5	5.2	25.0%	UCI
		21.6						CMIP6
HCFC-142b	0.19	20.9	0.1	22.1	0.1	1.2	5.7%	NOAA
		21.5	<0.1	22.6	<0.1	1.1	5.2%	AGAGE
		21.0	0.5	24.4	0.4	3.4	16.2%	UCI
		21.4						CMIP6
HFC-23	0.18	24.1	0.3	29.9	0.3	5.8	24%	AGAGE
		24.1						CMIP6
HFC-134a	0.16	62.7	0.6	95.7	0.6	33.1	53%	NOAA
		62.8	0.3	95.6	0.3	32.8	52%	AGAGE
		63.4	0.9	97.2	1.6	33.8	53%	UCI
		62.9						CMIP6
HFC-152a	0.1	6.47		6.79		0.32	4.9%	NOAA
		6.62	0.1	6.80	0.1	0.18	2.7%	AGAGE
		7.89						CMIP6
HFC-143a	0.16	11.9		20.6		8.7	73%	NOAA
		12.1	0.1	21.0	0.1	8.9	74%	AGAGE
		11.9						CMIP6
HFC-125	0.32	10.1		22.8		12.70	125%	NOAA
		10.5	<0.1	23.5	<0.1	13.01	124%	AGAGE
		10.5						CMIP6
CF₄	0.09	79.0	0.1	83.6	0.1	4.59	5.8%	AGAGE
		79.0						CMIP6
NF₃	0.2	0.83		1.62		0.79	95.2%	AGAGE
		0.83						CMIP6
SF₆	0.57	7.32	0.02	9.26	0.02	1.94	26.5%	NOAA
		7.28	0.02	9.24	0.02	1.96	26.9%	AGAGE
		7.31						CMIP6

AGAGE = Advanced Global Atmospheric Gases Experiment; NOAA = National Oceanic and Atmospheric Administration, Global Monitoring Division; UCI = University of California, Irvine; CSIRO = Commonwealth Scientific and Industrial Research Organization, GasLab, Aspendale, Australia; CMIP6: Climate Model Intercomparison Project – Phase 6 (Meinshausen et al., 2017b). Radiative efficiencies are from the 2018 Scientific Assessment of Stratospheric Ozone (WMO, 2018b). NOAA CO₂ and CH₄ are derived from flasks collected at 43 marine boundary layer sites, updated from (Conway et al., 1994; Dlugokencky et al., 1994; Masarie and Tans, 2004). NOAA N₂O and SF₆ data are derived from flask and in situ measurements at 11 sites, updated from (Hall et al., 2011). NOAA CFC-11 and CFC-12 are derived from flask and in situ measurements at 11 sites, and include both GC-ECD and GC-MS analysis. All other NOAA data are derived from flask measurements using GC-MS analysis, updated from (Montzka et al., 2009). AGAGE data are derived from in situ measurements at five sites, updated from (Prinn et al., 2018; Rigby et al., 2014). UCI data are derived from flask measurements at over 20 sites, updated from (Simpson et al., 2012). CSIRO data are derived from flask measurements at 9 sites, updated from (Kirschke et al., 2013; Langenfelds et al., 2002). NOAA, CSIRO, and WMO-GAW data are reported on WMO-GAW scales for CO₂, CH₄, N₂O, and SF₆. Other NOAA data are reported on NOAA scales. AGAGE and UCI data are reported on independent scales. WMO-GAW data were downloaded from ["https://gaw.kishou.go.jp/publications/global_mean_mole_fractions#content1"](https://gaw.kishou.go.jp/publications/global_mean_mole_fractions#content1) (last access 04Feb2019). Uncertainties are estimated at 90% confidence intervals. Networks use different methods to estimate uncertainties. Some uncertainties, such as HCFC-142b and HFC-125, have been rounded up to be consistent with the number of decimal places shown. Minor differences between 2011 values reported here and in the previous Assessment Report (AR5) are due to updates in calibration and data processing. [Note1: some uncertainties assumed to be the same as reported in AR5, and will be updated in subsequent drafts. Note 2: updates for next drafts foreseen to include observations for the year 2018 (SOD) and 2019 (Final report)]

[END TABLE 2.2 HERE]

2.2.4.2.1 Carbon Dioxide (CO₂)

Globally averaged surface CO₂ calculated by smoothing direct measurements starting in 1958, shows an overall positive trend of 1.5 ppm yr⁻¹ (Figure 2.4a), reflecting the net balance of fossil fuel and cement production and uptake in oceans and terrestrial vegetation (Chapter 5). Since 2011, globally averaged annual mean CO₂ increased by 14.54 ± 0.15 ppm to 404.98 ± 0.13 ppm in 2017, based on measurements from the NOAA network. Other global networks show similar changes (Table 2.2). Instantaneous growth rates, (Figure 2.4d), have varied substantially, e.g., rapid growth during the El Niño of 2015-2016 (Bastos et al., 2013; Betts et al., 2016). Further assessment of changing seasonality is undertaken in Section 2.3.4.1.

[START FIGURE 2.4 HERE]

Figure 2.4: Globally averaged dry-air mole fractions of (a) CO₂ from Scripps Institution of Oceanography (SIO) at annual time resolution based on measurements from Mauna Loa, Hawaii and South Pole (orange), CSIRO at monthly time resolution (cyan), and NOAA/ESRL/GMD at quasi-weekly time resolution (blue). SIO values are deseasonalized. (b) CH₄ from NOAA (blue), AGAGE (red), CSIRO (cyan), and UCI (green); and (c) N₂O from NOAA (blue), AGAGE (red), and CSIRO (cyan). Annual growth rates, calculated as the time derivative of the deseasonalized global averages (Dlugokencky et al., 1994) for globally averaged atmospheric mole fractions are shown in corresponding right panels, (d) CO₂, (e) CH₄, and (f) N₂O using the same color code as in (a, b, c). Note that the annual growth rate of CO₂ was consistently below 2 ppm yr⁻¹ between 1960 and 1980 (not shown). Insufficient and noisy data prevent calculation of growth rates prior to 1995. For abbreviations see Table 2.2.

[END FIGURE 2.4 HERE]

2.2.4.2.2 Methane (CH₄)

The globally averaged surface concentration of CH₄ in 2017 was 1849.6 ± 1.1 ppb, an increase of 46.4 ± 1.2 ppb since 2011 based on measurements from the NOAA network. Figure 2.4b shows broad agreement in global averages and growth rates among global means derived from multiple networks. Over the period of direct observations, atmospheric CH₄ shows marked growth rate changes. From the late-1970s through the early-2000s, CH₄'s growth rate (Figure 2.4d) was decreasing, and its atmospheric burden was approximately constant from 1999-2006. Atmospheric CH₄ fluctuations result from complex variations of sources and sinks by mainly the OH radical. A detailed discussion of recent methane trends and our understanding of their causes is given in (Cross-Chapter Box 5.1).

2.2.4.2.3 Nitrous Oxide (N₂O)

AR5 reported 324.2 ± 0.1 ppb for global surface annual mean N₂O in 2011; since then, it increased by 5.5 ± 0.2 ppb to 329.7 ± 0.2 ppb in 2017 (NOAA measurements). Globally averaged N₂O and its instantaneous growth rate are plotted in Figure 2.4cf. From 2008-2017, the N₂O global mean annual increase averaged $0.9 \text{ ppb} \pm 0.2 \text{ ppb}$, and it is increasing. The Pre-Industrial atmospheric burden of N₂O was a balance between natural emissions and loss processes (photolysis and reaction with O(¹D)) in the stratosphere. Atmospheric N₂O global annual averages combined with an estimate of its lifetime (123 yr, with most likely range 104-152 yr), (SPARC, 2013), show its budget of emissions and sinks is out of balance by ~40%.

Observed N₂O background spatial gradients are small (e.g., ~1.7 ppb between the northern subtropics and high southern latitudes), so excellent precision and internal network consistency are extremely important. WMO's Global Atmosphere Watch (GAW) program recommends inter-network compatibility of 0.1 ppb (WMO, 2018a). Achieving this scientific goal is difficult with historically-

used methods (gas chromatography with electron capture detection), and rarely achieved among GAW participants. While current measurement quality is sufficient to quantify the atmospheric N₂O budget at large spatial scales, determining emissions at policy relevant scales is more challenging.

In summary, there is robust evidence that the concentrations of all three major long-lived greenhouse gases have continued to increase since 2011 by 15 ppm for CO₂, 47 ppb for CH₄ and 5.6 ppb for N₂O. Their abundances are now 118.1 ± 1.6 ppm, 1043 ± 6 ppb and 59 ± 4 ppb respectively above pre-industrial levels (*very high confidence*).

2.2.4.3 Synthetic Greenhouse Gases (CFCs, HCFCs, HFCs, PFCs, SF₆ and others)

AR5 reported that atmospheric abundances of CFCs until 2011 were decreasing in response to controls on production and consumption mandated by the Montreal Protocol on Substances that Deplete the Ozone Layer. Abundances of HCFCs (replacements for CFCs) were increasing as their production was not yet capped and emissions exceeded losses. Atmospheric abundances of HFCs (replacements for HCFCs) were also increasing.

Further details on ozone-depleting substances and other minor greenhouse gases can be found in the 2018 Scientific Assessment of Stratospheric Ozone (WMO, 2018a). All man-made greenhouse gases have been produced since the industrial revolution and thus there is no long-term historical context or natural process to assess.

2.2.4.3.1 CFCs

Atmospheric abundances of most chlorofluorocarbons (CFCs) continue to decline as a result of the Montreal Protocol. The globally-averaged abundance of CFC-12 decreased by 16 ppt (3%) from 2011 to 2017, while CFC-11 decreased 10 ppt (4%) over the same period (Table 2.2, Figure 2.5). Atmospheric abundances of some minor CFCs (CFC-13, CFC-115, CFC-113a) are increasing, possibly related to HFC production. Overall, as of 2017 the radiative forcing from CFCs has declined by nearly 8% from its maximum in 2000.

While reported production of CFCs has been essentially zero since 2010 and the abundance of CFC-11 is still decreasing, a change in CFC-11 growth rate, along with an observed increase in the inter-hemispheric gradient of CFC-11, suggests that unreported production of CFC-11 is occurring (Montzka et al., 2018). The global lifetimes of several ozone depleting substances, including CFC-11, have been updated since AR5 (SPARC, 2013). The global lifetime of CFC-11 was revised from 45 to 52 years, resulting in slightly longer lifetimes for some other gases, since CFC-11 has been used as a reference to derive stratospheric lifetimes of other gases (Volk et al.).

2.2.4.3.2 HCFCs

The atmospheric abundances of the major HCFCs, primarily used in refrigeration and foam blowing, are increasing, but rates of increase have slowed in recent years (Figure 2.5). Global mean mole fraction estimated by three global networks (NOAA, AGAGE, UCI) (Table 2.2) showed good agreement at the time of AR5 for the period 2005-2011. For the period 2011-2017, the UCI network detected larger increases in HCFC-22, HCFC-141b, and HCFC-142b compared to NOAA and AGAGE networks. Reasons for the discrepancy are unknown, but could be related to differences in sampling locations (the UCI network is limited to the Pacific region, whereas NOAA and AGAGE sample in both the Atlantic and Pacific regions) (Simpson et al., 2012). HCFC-22 increased 30-40 ppt (~15%) from 2011 to 2017. Emissions of HCFC-22, derived from atmospheric data, have remained relatively constant since 2012, while those of HCFC-141b and HCFC-142b have declined (Engel et al., 2018).

2.2.4.3.3 HFCs, PFCs, SF₆ and other radiatively important gases

Hydrofluorocarbons (HFCs) are the replacements for HCFCs. The atmospheric abundance of HFC-134a (mobile air conditioning, foam blowing, and domestic refrigerators) increased from 63 ppt in 2011 to 96 ppt in 2017 (52% increase) (Figure 2.5). All three global networks detected a similar increase. HFC-23, which is emitted as a by-product of HCFC-22 production, increased by 5.8 ppt (24%) over the same time period. Most other HFCs are also increasing. Only HFC-152a is showing signs of stable (steady-state) abundance. There is good agreement among NOAA and AGAGE global networks on the global abundance of HCFC-152a. A higher mole fraction indicated in the CMIP6 dataset may be related to the inclusion of measurements subject to regional and local influence. While the direct radiative forcing of HFCs is currently small (0.03 W m^{-2} in 2017), concern about rapid future growth in the use of HFCs led to controls on HFC production under the Kigali Amendment of the Montreal Protocol entered into force at the start of 2019, and will involve a gradual phase-out of HFCs in favor of more climate-friendly alternatives, such as hydrocarbons and hydrofluoroolefins (HFOs).

Other radiatively important gases with predominantly anthropogenic sources also continue to increase in abundance. SF₆, which is used in electrical distribution systems, has increased from 7.3 ppt in 2011 to 9.2 ppt in 2017 (+27%). The global lifetime of SF₆ has been revised downward from 3200 years to around 1000 years (Kovács et al., 2017; Ray et al., 2017). NF₃, which is used in the semi-conductor industry, increased 95% over the same period. Its contribution to radiative forcing remains small but significant, at 0.3 mW m^{-2} . SO₂F₂ is used as a fumigant in place of ozone-depleting methyl bromine, and has a radiative forcing comparable to that of NF₃, at 0.45 mW m^{-2} . The atmospheric abundance of SO₂F₂ increased 34% to 2.3 ppt from 2011 to 2017. CF₄ and C₂F₆, which have exceedingly long global lifetimes, showed modest increases from 2011 to 2017. CF₄ increased 5.8% to 83.6 ppt, and C₂F₆ increased 11.7% to 4.66 ppt.

In summary, atmospheric abundances of most CFCs, regulated under the Montreal protocol, are further declining, with a radiative forcing of 0.25 W m^{-2} in 2017. Direct radiative forcing from HCFCs and HFCs were 0.059 and 0.03 W m^{-2} in 2017, respectively, with emission restrictions imposed under the Kigali Amendment of the Montreal protocol.

[START FIGURE 2.5 HERE]

Figure 2.5: Global mean atmospheric mole fractions of select ozone-depleting substances and other greenhouse gases, based on the CMIP6 historical dataset (Meinshausen et al., 2017b) and data from NOAA and AGAGE global networks. See Table 2.2 for information on lifetime and radiative effectiveness.

[END FIGURE 2.5 HERE]

2.2.5 Short-lived gases

2.2.5.1 Stratospheric water vapor

AR5 assessed *low confidence* in long-term stratospheric water vapour (SWV) trends based on substantial seasonal and interannual variability in satellite data from 1992-2011. The 1980-2010 record of balloon-borne frost point hygrometer measurements at a single NH mid-latitude location showed an average net increase of $1.0 \pm 0.2 \text{ ppm}$ in the lower stratosphere (16-26 km), most of which occurred prior to 1993. A RF of $0.07 [0.02 \text{ to } 0.12] \text{ W m}^{-2}$ was estimated for SWV produced by CH₄ oxidation.

Recent studies (Diallo et al., 2018; Garfinkel et al., 2018; Konopka et al., 2016; Tao et al., 2015) have

demonstrated how the QBO and ENSO can significantly influence SWV abundance and the tropical cold point tropopause temperatures that control it. It has also been shown that the convective lofting of ice can moisten the lower stratosphere over large regions (Anderson et al., 2017; Avery et al., 2017; Dessler et al., 2016). Several mechanisms have been proposed that link climate change with increasing SWV, through higher tropical tropopause temperatures (Lin et al., 2017), increased natural CH₄ emissions, and an enhanced frequency of deep convection events. The additional SWV would induce further warming, creating a positive feedback (Dessler et al., 2013).

Near-global observations of SWV since AR5 have revealed unusually strong and abrupt changes, especially in the tropical lower stratosphere. Between Dec 2015 and Nov 2016, the tropical mean SWV anomaly at 82 hPa dropped from +0.9 ppm to –1.0 ppm, accompanied by highly anomalous tropical stratospheric QBO dynamics (Newman et al., 2016c; Tweedy et al., 2017b) and the transition of ENSO from strong El Niño to La Niña conditions (Davis et al., 2017). The tropical mean anomaly then rose sharply to +0.7 ppm in June 2017 as warm westerlies returned to the tropical lower stratosphere and ENSO neutral conditions prevailed (Davis et al., 2018).

Since AR5, spatially comprehensive sets of satellite-based SWV measurements have been merged into multi-decade records after efforts to reduce biases between instruments (Davis et al., 2016; Froidevaux et al., 2015; Hegglin et al., 2014). These composite records indicate no net increase of SWV in the lower stratosphere since their inception in the late 1980s. In contrast, monthly SWV measurements by balloon-borne frost point hygrometers over Boulder, Colorado, do show an average 1.0 ± 0.2 ppm ($25 \pm 5\%$) net increase from 1980 to 2010 (Hurst et al., 2011), with 50-100% of the altitude-dependent increase occurring before 1993. Since 2010, a 0.1 to 0.2 ppm net decrease in SWV has been observed over Boulder. Overall, the satellite and Boulder, Colorado (USA) frost point hygrometer records have been in reasonable agreement since 1992. While Hegglin et al (2014) suggested that the increasing trend over Boulder should not be considered representative of the global stratosphere, Losow et al. (2018) showed insignificant differences between SWV trends at Boulder (40°N) and for the 35-45°N zonal mean from 1980 to 2010 using model simulations and satellite observations.

In summary, Frost point hygrometer measurements over Boulder, Colorado depict a net increase of Stratospheric Water Vapour of ~1.0 ppm (~25%) during 1980-2010 most of which occurred prior to 1993. Conversely, several sets of merged satellite data created since AR5 show no evidence of a net increase since the late 1980s. There thus continues to be *low confidence* in a long-term trend of SWV.

2.2.5.2 Stratospheric ozone

AR5 assessed that it was certain that global stratospheric ozone from the mid-1990s to 2011 was nearly constant and about 3.5% lower than in the reference period 1964–1980. Most of the declines occurred prior to the mid-1990s. The AR5 estimated RF was -0.10 ± 0.15 W m⁻² for 2005.

Annually averaged total ozone (Figure 2.6a-d) significantly declined during the 1980s and the early 1990s by about 3.5% for global (90°S-90°N) and by 2.5% for non-polar near-global (60°S-60°N) total ozone and increased slightly, but not significantly, during 2000-2017. Near-global 2014-2017 mean total ozone concentration is about 2.2% below the pre-ozone depletion 1964–1980 average. At southern and northern mid-latitudes declines, related to anthropogenic ODS, are 5.5% and 3.0% below the 1964–1980 average respectively. Total ozone remained practically unchanged in the tropics (Braesicke et al., 2018). Emission of these gases started before 1980 and it is estimated that as much as 40% of the long-term ozone loss occurred between 1960 and 1980 (Shepherd et al., 2014), lowering the 1964-1980 baseline values by about 1% (outside the polar regions), a value close to observational uncertainties. The world's longest series of total ozone measurements from Arosa, Switzerland, started in 1926, and does not show any substantial long-term changes before ~1980 (Staehelin et al., 2018).

[START FIGURE 2.6 HERE]

Figure 2.6: Time-series of annual mean total ozone (a good proxy for stratospheric ozone) from 1970-2017 in four zonal bands (a-d) and polar (60°-90°) total ozone in March (NH) and October (SH) (e). March (NH) and October (SH) are the months when polar ozone losses usually are largest. The values are given in Dobson units, where 100DU correspond to a 1mm thick slab of pure ozone gas at standard conditions. Adopted from Weber et al (2018).

[END FIGURE 2.6 HERE]

Radiative forcing depends strongly on the altitude of ozone changes. Two stratospheric regions are mainly responsible for long-term changes outside the polar regions. In the upper stratosphere (35 to 45 km), there was a strong decline (about 10%) up to the mid-1990s and a subsequent increase by about 4% to present. In the lower stratosphere (20 to 25 km), there also was a statistically significant decline (7 to 8%) up to the mid-1990s, followed by stabilization.

Springtime averages of total NH and SH polar ozone are shown in Figure 2.6e. By far the strongest ozone loss in the stratosphere continues to occur in austral spring over Antarctica (ozone hole), with emergent signs of recovery in recent years (WMO, 2018b). Interannual variability in polar stratospheric ozone is driven by variability in meteorological conditions (temperature and winds), and, to a lesser extent by the stratospheric aerosol loading and the solar cycle. This variability is particularly large in the Arctic, where the largest depletion event, comparable to a typical event in the Antarctic, occurred in 2011 (Langematz et al., 2018).

In summary, near global (60S-60N) stratospheric ozone has declined from pre-1980 values, with estimated declines of about 2.5% over the period 1980-1995. Present day (2014-2017) near global annual column average is about 2.2% below the 1964-1980 average (*high confidence*).

2.2.5.3 Tropospheric ozone

AR5 assessed *medium confidence* in large-scale increases of tropospheric ozone at rural surface sites across the Northern Hemisphere (1970-2010), and a doubling of European surface ozone during the 20th century (*medium confidence*). *Low confidence* was attributed to increases in the Southern Hemisphere. Surface ozone *likely* increased in East Asia, but leveled off or decreased in the eastern USA and Western Europe. Free tropospheric trends (1971-2010) from ozonesondes and aircraft showed positive trends in most, but not all, assessed regions, and for most seasons and altitudes. The estimated RF was $0.40 \pm 0.20 \text{ W m}^{-2}$.

Since AR5 the Tropospheric Ozone Assessment Report (TOAR) activity has collated and analysed many more long-term ozone records, and re-evaluated pre-industrial ozone observations. It concludes that data prior to 1896 are either quantitatively unreliable or not representative of ozone on regional or hemispheric scales (Tarasick et al., 2019). Re-assessment of the Montsouris (Paris) record (1876-1910) concluded that these observations were probably lowered by emissions of air pollutants and therefore not representative of late 19th-century ozone. Tarasick et al (2019) identified 60 surface ozone time series at rural locations worldwide between 1896-1975, taken with a range of methods with potentially large uncertainties. Compared to present day (1990-2014) rural observations, a quality screened data selection indicates a surface ozone increase since the mid-20th century of 30-70% across the northern extra-tropics, comparable to a tropospheric column ozone increase of $48 \pm 30\%$ between 1934-1944 and 1990-2012. This analysis revises downward the 100 % 20th century increase over Europe reported by AR5, which relied on far fewer historical data sets.

In the northern tropics and the southern hemisphere limited historical data (1954-1975) provide no clear evidence for ozone increases (Tarasick et al., 2019), except for Mauna Loa, Hawaii, USA, where lower-free tropospheric ozone increased by 50% from the late 1950s to present (Cooper et al., 2019). This increase, on the northern edge of the tropics, is similar to the northern mid-latitude increases.

In the period of more modern instrumentation (1970-1980s), ozonesonde records (Table 2.SM.1) show no significant changes in the free troposphere above the Arctic region since the mid to late 20th Century, while surface sites show small increases of 2-3% per decade. In the northern mid-latitudes, aircraft observations show free tropospheric ozone increases ranging from 2-10 % over 2/3 of the sampled regions with no significant changes elsewhere. Northern mid-latitude surface ozone trends are mixed (Figure 2.7). Given their limited spatial representativeness no firm conclusions can be drawn regarding hemispheric scale surface ozone trends. The largest ozone increases (7-21% decade⁻¹) are in the northern tropics in the vicinity of Asia, as observed by commercial aircraft. A composite satellite product shows increases in zonally averaged ozone concentrations in the northern (3% decade⁻¹) and southern tropics (2% decade⁻¹) since 1979. Observations in the southern mid-latitudes are limited but indicate surface and free tropospheric ozone increases of 2-8% decade⁻¹, except for decreasing ozone in the upper troposphere above New Zealand. In the Antarctic region there is no indication of a significant ozone increase at the South Pole since 1975. However, ozone has increased at the surface since 1995 and in the mid-troposphere since the late 20th Century (Gaudel et al., 2018).

[START FIGURE 2.7 HERE]

Figure 2.7: Decadal ozone trends based on monthly anomalies (relative to 1995-2010) at 28 remote surface sites, plus IAGOS aircraft trends at 650 hPa (∇) above Europe the northeastern USA, NE China, SE Asia, southern India, the Persian Gulf and Malaysia/Indonesia (as reported by (Gaudel et al., 2018). High elevation sites (>1500 m a.s.l.) are indicated by (Δ). Sites below 1500 m a.s.l. are indicated by unaccompanied circles or arrows. All surface trends end with the most recently available year, but begin in 1995. Mauna Loa, Hawaii trends are shown for dry mid-latitude (top arrow), all (middle arrow) and moist tropical (bottom arrow) air masses. [Note a known need to tidy up the title to the graphic]

[END FIGURE 2.7 HERE]

To conclude, since the mid-1970s, free tropospheric ozone has increased by 2-10% per decade in most regions of the northern mid-latitudes and tropics sampled (*high confidence*), while surface trends are variable. It has also generally increased at the surface and in the free troposphere of the more poorly sampled Southern Hemisphere (*medium confidence*). When compared to limited historical data representative for the mid-20th Century, tropospheric ozone has increased by 30-70% in the northern hemisphere and tropics (*medium confidence*).

2.2.6 Aerosols

AR5 assessed large-scale Aerosol Optical Depth (AOD) trends for the 2000–2009 period, and concluded that there was *low confidence* in a global trend, but that there were *very likely* significant positive and negative trends regionally. The effective radiative forcing associated with aerosol-radiation interactions for 2011 (relative to 1750) was estimated to be -0.45 W m^{-2} and of aerosol-cloud interaction estimated as -0.90 W m^{-2} . Aerosol radiative forcing uncertainty was assessed as the largest contributor to the overall radiative forcing uncertainty since pre-industrial times.

This section assesses the observed large-scale temporal evolution of aerosols. Aerosol processes,

chemical and physical properties, and links to air quality, are discussed in Chapter 6. An in-depth assessment of aerosol interactions with radiation and clouds is provided in Chapters 7 and 8. A large spatial heterogeneity in aerosol abundance and trends (Figure 2.8), is determined by a variety of anthropogenic (mainly fossil fuel and bio-fuel) and natural (sea salt, mineral dust, volcanoes, wildfires) sources, formation and removal processes (direct emission or formed from gas phase).

Since AR5, more proxy aerosol records have become available and there have also been advances in the temporal resolution and quality in some of the emerging records in terms of dating and estimation of proxy uncertainties from different types of archives (e.g. Jacobel et al., 2017; Kylander et al., 2016; Middleton et al., 2018a; Stevens et al., 2016, 2018), which also aids synthesis of new global compilations (Albani et al., 2015a; Kienast et al., 2016). Estimates of the glacial/interglacial ratio in global dust deposition remain within 2 to 4 (Albani et al., 2015b; Lambert et al., 2015). Notably more data indicates a magnitude of 3 to 5 for the glacial/interglacial ratio for mid and high latitude oceans of both hemispheres (Lamy et al., 2014; Martínez-García et al., 2014; Serno et al., 2015a), and additional evidence suggests synchronicity of variations in dust deposition in Eurasia, the North Pacific, and Greenland over the same time scale (Rousseau et al., 2017; Serno et al., 2015b; Újvári et al., 2017). In addition, a broader scale quantification of changes in dust deposition from North Africa as seen in North Atlantic sediments records confirms dust deposition rates lower by a factor 2 to 5 during the African Humid Period (10-5 ka BP) compared to the late Holocene (Albani et al., 2015b; McGee et al., 2013; Middleton et al., 2018b). On multi-millennial time scales, biogenic and volcanic emissions drive significant variability in sulphate concentrations (Schüpbach et al., 2018).

Aerosols contained in glacier ice provide indicative regional information on multi-centennial mid- and-high latitude aerosol concentration trends, including for sulphate and black carbon (Figure 2.8a, b). During the 20th Century, sulphate in ice cores experienced increases by a factor of 8 from the end of the 19th century to the 1970s in Europe, by a factor of 4 from the 1940s to the 1970s in Russia, and by a factor of 3 from the end of the 19th century to 1950 in the Arctic (Svalbard). In all regions studied, concentrations have declined by a factor of approximately 2 following the peak (1970 in Europe and Russia, and 1950 in the Arctic). For black carbon (BC), strong increases are observed in the 20th century over Europe, Russia, Greenland (representing emissions from North America), and in the Arctic (Svalbard). South America exhibits a small long-term increasing trend.

In-situ aerosol concentrations have been measured by regional ground-based networks since the 1980s in North America and Europe. Although the observational network capable of monitoring absorbing aerosol has expanded since AR5, data coverage is still not sufficient to assess reliable large-scale trends (Samset et al., 2018). Truly global information on large-scale aerosol column distribution is only available from satellite remote sensing. Ground-based solar attenuation networks help to constrain and improve the retrieval potential of AOD. The trends derived from the AERONET network (Figure 2.8c, d) corroborate satellite measurements everywhere except for the Southern Asian region where insufficient ground-based data is available. Several studies have investigated trends from the AERONET network and come to conclusions for Europe (Cherian et al., 2014; Li et al., 2014; Stjern et al., 2011) and for the USA (Jongeward et al., 2016; Li et al., 2014) which corroborate the satellite results (Cherian et al., 2014; Li et al., 2014; Stjern et al., 2011) (Jongeward et al., 2016; Li et al., 2014), strengthening findings available at the time of AR5. Li et al. (2014) analysed AERONET stations until 2014 and documented a trend over Southern Asia. Confidence in satellite trends is further supported by in-situ surface concentration measurements of sulphate and PM_{2.5}, respectively, from observations networks over Europe (Stjern et al., 2011) and over North America (Jongeward et al., 2016).

Over 2001 to 2017, in addition to Europe and North America, declining trends in AOD are also found over East Asia, where the trend reversed from positive to negative around 2009. Over southern Asia, however, AOD continues to increase (also supported by AERONET data; (Li et al., 2014). The globally averaged AOD from MODIS/MISR exhibits over 2001-2017 a slight decrease which is not statistically significant. Aerosol retrievals from the Ozone Mapping Instrument (OMI) that measures

in the ultraviolet spectral range show similar trends (Dahutia et al., 2018; Hammer et al., 2018; Zhao et al., 2017). A comparison of MODIS and MISR radiometric observations with the broadband CERES satellite instrument (Corbett and Loeb, 2015) showed that drifts in calibration are unlikely to affect the trends derived from the satellite retrievals.

Satellite AOD retrievals are sensitive to aerosol hygroscopic growth at elevated relative humidity. Thus, while AOD is a key variable to assess the direct impact of aerosols on radiation, it is only an indirect proxy for the cloud-active aerosol (Stier, 2016). With AOD retrievals at different solar wavelengths, such as blue, red and green light, it is possible to extract the AOD attributed to sub-micrometer ‘fine-mode’ aerosol sizes (AOD_f). As anthropogenic aerosol is almost entirely found in the fine-mode aerosol fraction, AOD_f can be used to corroborate evidence for trends related to anthropogenic emissions (Kinne, 2019) (Figure 2.8d). A widespread significant decline of more than -1.5 % per year is observed over Europe and North America. In turn, upward trends of about +1.5 % per year are observed over the Southern Asia. The global-scale negative trend in AOD_f of -0.03 % per year is significant at the 90 % level. Trends in cloudiness or cloud radiative properties co-incident with aerosol trends are currently inconclusive (Norris et al., 2016).

[START FIGURE 2.8 HERE]

Figure 2.8: Ice core concentrations of (a) sulphate (ng g⁻¹) and (b) black carbon (ng g⁻¹) from Europe (reproduced from (Engardt et al., 2017; Sigl et al., 2018), Russia (reproduced from Oliver et al. (2006), Arctic (Svalbard, reproduced from Wendl et al., 2015 and Osmont et al., 2018), South America (ex-sulphate, corrected for mineral dust input, reproduced from Kellerhals et al. (2010), and BC from (Osmont et al., 2019), Eastern Europe (reproduced from Lim et al. 2017), and Greenland (stacked rBC record from Greenland including the four ice cores NEEM-2011-S1, D4, TUNU2013), and Summit2010 (Keegan et al., 2014; McConnell et al., 2007; Mernild et al., 2015; Sigl et al., 2013, 2015) reproduced from Osmont et al. (2019) and Sigl et al. (2018). Given are 10-year averages. Sulphate concentrations were not corrected for sea-salt input, which is negligible at the presented ice core locations. (c) geographical distribution of the linear trend in annual mean AOD retrieved from satellite data for the 2001 – 2017 period (% per year). The average trend from MODerate Imaging Spectroradiometer (MODIS; (Levy et al., 2010)) and Multi-Angle Imaging Spectroradiometer (MISR, (Garay et al., 2017)) is shown. Hatched areas show a trend that is significant at the 0.9 level (Kendall’s rank correlation test) according to both MISR and MODIS retrievals, with the two trends agreeing to within 50%. Superimposed are the trends in annual-mean AOD from AERONET surface sunphotometer network sites (Giles et al., 2019; Holben et al., 1998) for 2000 – 2018 where significant at the 0.9 level and where at least 10 years of data are available. (d) Trend in 2001-2016 AOD_f derived from MISR. Hatched areas show trends significant at the 0.9 level (Kendall’s rank correlation test), AERONET trends for 2000 – 2018 superimposed where statistically significant.

[END FIGURE 2.8 HERE]

To conclude, satellite data and in-situ records indicate that Aerosol Optical Depth exhibits predominantly decreasing trends since 2001 in most regions (*high confidence*) but in some regions, in particular Southern Asia, is increasing. Trends are most pronounced in fine-mode AOD pointing with *high confidence* to a primarily anthropogenic cause that is decreasing significantly in the global mean. Atmospheric aerosols sampled by glacier ice show increasing trends for the industrial era until the last quarter of the 20th century (*medium confidence*).

2.2.7 Land use and land cover

AR5 assessed historical changes in land cover up through the 2000s and their impact on surface climate, with an equivalent radiative forcing of $-0.15 \pm 0.10 \text{ W m}^{-2}$ (*very likely*). Non-radiative

changes in land use (such as in the hydrologic cycle) can impact surface temperature in opposite ways than albedo, with large uncertainties. Consequently, AR5 assessed a *low agreement* on the sign of the net change in global mean temperature as a result of land use change.

More than half of the global land surface has been modified or managed in some way by human activities, particularly over the past three centuries (Hurt et al., 2011). Since AR5, new land cover reconstructions and scenarios augmented with pollen data, archeological records, and other historical information (Gaillard et al., 2018; Goldewijk et al., 2017; Koch et al., 2019; Krausmann et al., 2013) have improved understanding of long-term changes in land use from agricultural applications (such as cropland and pasture) and other historical drivers of variability (such as human disease and conflict). For instance, recent studies suggest that vegetation on Earth was strongly transformed through deforestation since ca. 6ka BP (Early Neolithic) and earlier in some regions of the world (Goldewijk et al., 2017). Despite these improvements, however, significant uncertainty remains in the depiction of historical changes in global land cover (Kaplan et al., 2017; Marquer et al., 2017). Since the early 1980s satellite remote sensing has provided unprecedented accuracy in quantifying contemporary land cover use and change (Hansen et al., 2013b). Recent estimates suggest that 60% of all changes from 1982–2016 were associated with direct human activities (Song et al., 2018), with spatial patterns emphasizing the regional importance of land use, including tropical deforestation, temperate reforestation, cropland intensification, and increased urbanization. At present, about 75% of the Earth’s ice-free terrestrial surface is impacted by human endeavours, mainly in the form of agricultural applications and forest management (SRCCL).

There is a significant range in post-AR5 estimates of radiative forcing from anthropogenic changes in land cover. In general, recent estimates of associated albedo forcing fall within the wide span given in AR5 ($-0.15 \pm 0.10 \text{ W m}^{-2}$). For instance, a modelling study by Unger (2014) identified a radiative forcing estimate from cropland expansion of -0.11 W m^{-2} for the year of 2000 (relative to 1850), while an empirical analysis by Ghimire (Ghimire et al., 2014) found an albedo forcing of -0.15 W m^{-2} for 2005 (relative to 1700). Other recent estimates fall decidedly outside the AR5 range because they consider factors in addition to albedo. For instance, a modelling study by Andrews (Andrews et al., 2017) identified an effective radiative forcing from the biogeophysical effects of land use change of $-0.4 \pm 0.5 \text{ W m}^{-2}$ for 2005 (relative to 1860), ascribing much of the impact to changes in albedo, dust, and clouds. A modelling study by Ward et al. (2014) found a land cover radiative forcing of $0.9 \pm 0.5 \text{ W m}^{-2}$ for 2010 (relative to 1850), driven by increases in CO_2 and CH_4 concentrations that were likely attributable to local agricultural applications and other direct land-surface modifications (Ward and Mahowald, 2015). An empirical analysis by Bright et al. (2017) corroborated AR5 findings that non-radiative mechanisms dominated the local response to a land-cover change in most regions.

In summary, there is *high confidence* that land use activities increased global albedo with an equivalent radiative forcing of $-0.15 \pm 0.10 \text{ W m}^{-2}$ since the mid-19th century. Other factors related to land use change have a potentially large impact on the radiative balance, but there is a substantial range in estimates of the radiative forcing, and no agreement on the sign of the effect, and hence *low confidence*.

2.2.8 Changes in radiative budget

AR5 concluded that changes in the radiative budget over the industrial period corresponded to a positive radiative forcing which increased more rapidly after 1970 than before. It concluded that there was *very high confidence* in the RF due to WMGHG, and CO_2 was the single largest contributor. AR5 concluded that there was *high confidence* that aerosols have offset a substantial portion of the WMGHG forcing. It introduced the concept of effective radiative forcing that is the main forcing metric used in AR6.

This section assesses changes in the Earth radiation budget over the industrial era since 1750 CE. This

assessment relies on the observed changes as assessed in the remainder of Section 2.2 where possible, making use of the methods and details as described in Chapter 7, including, where possible the rapid adjustments especially of clouds. For short-lived climate forcers, for which details of the spatiotemporal distribution matter for their radiative impact, this section largely relies also on modelling results described in Chapter 6.

Total solar irradiance (TSI) (Section 2.2.2) showed an increase implying a radiative forcing of less than 0.1 W m^{-2} over the 1900 – 1980 period. Between 1850 and 1900, and after 1980, TSI was subject to weak multi-decadal trends (*medium confidence*). TSI varies cyclically over the 11-year solar cycle with an amplitude of order of 0.1 W m^{-2} . Strong volcanic eruptions (Section 2.2.3) capable of cooling the surface occurred in the late 19th and early 20th Centuries, then a relatively quiescent period ensued between ~1920 and 1960, followed by three strong eruptions in 1963, 1982 and 1991, and only small-to-moderate eruptions thereafter.

The atmospheric concentrations of well-mixed greenhouse gases (Section 2.2.4) have continuously increased since the industrial revolution, as has their positive influence on the radiation budget. CO_2 contributes the largest share. The increase in atmospheric concentrations of CO_2 and N_2O accelerated in the 21st century, while the increase of CH_4 accelerated in the last decade; CFCs and HFCs concentrations mostly declined or their rate of increase slowed down (Section 2.2.4). The ERF for CO_2 alone is stronger than for all the other anthropogenic well-mixed greenhouse gases taken together throughout the industrial period, and its relative importance increased in recent years (Figures 2.9, 7.11 and 7.14).

[START FIGURE 2.9 HERE]

Figure 2.9: Changes in radiative forcing estimates related to the factors assessed in Section 2.2. effective radiative forcings are based upon the calculations described in Chapter 7. The global annual mean temporal evolution since 1850 is shown as the central assessment value. Gaseous changes to the atmospheric composition are shown for carbon dioxide (CO_2), methane (CH_4), nitrous oxide (N_2O), Halogens, and tropospheric ozone (O_3). Aerosol changes are the sum of the effective radiative forcing due to aerosol – radiation and aerosol – cloud interactions. Other anthropogenic forcings include stratospheric ozone, stratospheric water vapour, land use/land cover changes, black carbon deposition on snow, and contrails. The two other key forcings shown are variability in TSI and in volcanic forcing. Shading shows the 5-95% uncertainty range as assessed in Chapter 7 (Section 7.3). The sum of all forcings is shown as the total forcing.

[END FIGURE 2.9 HERE]

Among the gaseous short-lived climate forcers (Section 2.2.5, Chapter 6) tropospheric ozone (O_3) is the most important component. Tropospheric ozone is assessed herein to be increasing in recent decades.

The effect of aerosols (Section 2.2.6, Chapter 6) on the radiation budget, including their effect on clouds, and cloud adjustments, as well as the deposition of black carbon on snow (Section 7.3), was negative throughout the industrial period (*high confidence*), plateauing to -0.35 Wm^{-2} in recent decades. Therefore, the relative importance of aerosol forcing compared to other forcing agents has decreased at the global scale in the most recent 30 years (*medium confidence*).

Land use and land cover changes (Section 2.2.7) over the industrial period introduce a radiative forcing by increasing the surface albedo. This effect increased monotonically over the industrial period, reaching current values of about -0.15 Wm^{-2} (*medium confidence*), with *low confidence* in the ERF due to large and highly uncertain feedbacks of other land-use related factors.

The net radiative effect of all observed changes in climate forcings is positive since 1850, except for short periods (up to a few years in duration) following moderate to large volcanic eruptions (*high confidence*), and the rate of change has increased in the last 30 years (*high confidence*).

2.3 Changes in large-scale climate

[START CROSS-CHAPTER BOX 2.1 HERE]

Cross-Chapter Box 2.1: Large-scale indicators of climate change

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The Working Group I contribution to the sixth assessment cycle includes a substantially modified structure compared to prior assessments (Section 1.8). Chapters 2 through 4 are charged with undertaking an assessment of past large-scale climate changes (Chapter 2), our ability to model and understand the causes of these changes (Chapter 3), and projections of changes in both the short- and long-term (Chapter 4). These chapters underpin and support a range of process-based chapters (Chapters 5 through 9) and regionally-focused chapters (Chapters 10 through 12), and inform the Atlas. It is necessary at this juncture to define both what is meant by large-scale, and the rationale of choosing the key-indicators indicators in support of Chapters 2, 3 and 4.

Defining large-scale

The climate system is an interaction of processes across a range of scales from the microscale to the global scale. Any scale separation point is, ultimately, both subjective and a compromise. The WGI structure for AR6 requires that we define a separation between large-scale (Chapters 2, 3 and 4) and process-relevant / regional scales (subsequent chapters). Understanding the large-scale climate variability and change requires knowledge of both the response to external forcings and the role of internal variability. Many external forcings have substantial hemispheric or continental scale variations. Modes of climate variability, which are assessed in Chapters 2-4, are driven by ocean basin scale processes. For example, the El Nino phenomenon spans the tropical Pacific basin. Thus, within Chapters 2, 3 and 4 we define large-scale to constitute ocean basin and continental scales as well as hemispheric and global scales. Although the assessment shall be at these scales, where appropriate global maps etc. shall be shown and discussed.

Defining a key set of indicators

Chapters 2, 3 and 4 are charged with looking at key indicators of a changing climate system. Key indicators must enable an assessment of changes in the climate system as a whole, but avoid overt overlap with later chapters. The adopted approach has been to select a limited set of observable key indicators across the atmospheric, oceanic, cryospheric and biospheric domains which are overall indicators of the changing state of the climate system. Taken together, these indicators would be expected to both have changed and continue to change in the future in a coherent and consistent manner in a changing climate. Chapters 2, 3 and 4 shall each consider a subset of these indicators. These indicators are summarized in Cross-Chapter Box 2.1, Table 1.

[START CROSS-CHAPTER BOX 2.1, TABLE 1 HERE]

Cross-Chapter Box 2.1, Table 1: Summary of the large-scale indicators used across Chapters 2 through 4 and their principal applications in remaining chapters. Indicators are split by domain and inclusion in each of Chapters 2 through 4 is indicated by a solid fill of the relevant table cell. The list of remaining chapters is limited to those where the variable is a principal consideration of that chapter and is not intended to be exhaustive.

Selected large-scale indicator of climate	Used in Chapters	Additional regional or
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change	2	3	4	process-based assessments undertaken in chapters
<i>Atmosphere</i>				
Surface and upper air temperatures				7, 10, 11, 12, Atlas
Hydrological cycle components (surface humidity, PWV, Precipitation, runoff, P-E)				8, 10, 11, 12, Atlas
Tropospheric circulation (Sea level pressure and winds, Hadley / Walker circulation, global monsoons, blocking, storm tracks and jets, MJO)				10, 11
Stratospheric circulation (sudden stratospheric warmings, QBO, Brewer Dobson circulation, zonal winds)				11
<i>Cryosphere</i>				
Sea-ice extent/area and thickness				9, 12
Seasonal snow cover (extent, duration, depth)				9, 12
Glacier mass and extent				9
Ice sheet mass and extent				9
Permafrost				5, 9, 12
<i>Oceans</i>				
Temperature / Ocean Heat Content				5, 7, 9, 12, Atlas
Salinity				9
Sea-level				9, 12
Circulation				5
<i>Biosphere</i>				
Atmospheric CO ₂ including changes in seasonal cycles of growth rates				5, 12
Ocean CO ₂ / O ₂				5, Atlas
Ocean pH				5, 12, Atlas
Ocean phenology				
Land community assemblages and species ranges				5
Terrestrial vegetation (global greening)				5
Growing season onset and length				5, 12

[END CROSS-CHAPTER BOX 2.1, TABLE 1 HERE]

Atmospheric indicators

Limiting global surface temperature increases is integral to the negotiations of UNFCCC and thus essential to include. Upper-air temperatures are a key indicator of different causal mechanisms underlying climate change and underpinned the very first conclusion in the SAR of a ‘discernible human influence’. An overall intensification in the global hydrological cycle across all domains may be expected and is assessed in detail in Chapter 8. To ascertain large-scale changes of and human influence on the global hydrological cycle, a small subset of indicators across the cycle are chosen: ocean and land P-E, global land precipitation, total column water vapour, surface humidity, soil moisture, and global river runoff. Finally, a warming world may be accompanied by a change in large-scale circulation patterns such as the extent and strength of the Hadley Circulation, Monsoon systems, or the position and strength of the sub-tropical polar jets. Large-scale circulation changes in the troposphere and the stratosphere are thus also considered a key atmospheric indicator.

Cryospheric indicators

All aspects of the cryospheric system would be expected to change in a warming world. Changes in

ice sheets are key to constraining the longest-term impacts associated with changes in global and regional sea-level, although future changes are best captured with process-based modelling and expert judgement (Chapter 9). Changes in sea-ice extent/area and thickness have potential impacts for hemispheric-scale circulation in addition to local effects in the polar regions (Cross-Chapter Box 10.1). Changes in glacier mass balance contribute to changes in sea-level but also have substantial implications for water supply for a substantial proportion of the global population. Seasonal snow-cover has many implications for mid- to high-latitude regions with knock on effects on biospheric components of the system. Finally, changes in permafrost and the seasonally thawed active layer have substantial implications in mid- to high-latitudes and have been posited to be important in potential feedbacks through degassing of greenhouse gases were the permafrost to thaw.

Oceanic indicators

Most of the energy imbalance in the climate system is taken up by changes in Ocean Heat Content (OHC), which constitutes the largest energy storage component in the climate system. Changes in OHC are associated with changes in salinity and surface currents. As shown in Cross-Chapter Box 9.2 sea-level change is key to building confidence in the closure of the global energy budget. The oceanic overturning circulation is key to the redistribution of heat and salinity within the oceans. For glacial cycles changes in the efficacy of the circulation have portended large-scale changes in global and regional surface climate.

Biospheric indicators

A warming global climate would be expected to cause changes in a broad range of biospheric indicators. A lengthening of the growing season would be expected in most of the globe. Changes in plant assemblages may occur with increases in altitude and latitude of the tree-line. Aspects of productivity such as net primary productivity or fractional absorption of photosynthetically active radiation may be expected to change. Similarly changes in ocean pH and oxygenation as well as physical state may portend changes in marine ecosystems.

Defining a selection of modes of variability

There are many modes of variability in the climate system. These modes of variability affect global, hemispheric and regional climate across a range of timescales. A selection of important modes has been undertaken under the criteria that they: i) have impacts at the large scales of interest to chapters 2 to 4; and ii) have substantial potential to modify interannual to multidecadal climate system properties. The selection and relationship to latter chapters is summarised in Cross-Chapter Box 2.1, Table 2.

[START CROSS-CHAPTER BOX 2.1, TABLE 2 HERE]

Cross-Chapter Box 2.1, Table 2: Summary of the modes of variability used across Chapters 2 through 4 and their principal applications in remaining chapters. Inclusion in each of Chapters 2 through 4 is indicated by a solid fill of the relevant table cell. The list of remaining chapters is limited to those where the mode of variability is a principal consideration of that chapter and is not intended to be exhaustive.

Selected mode of variability	Used in Chapters			Additional regional or process-based assessments undertaken in chapters
	2	3	4	
El Nino Southern Oscillation				
Indian Ocean Basin and dipole modes				10
Atlantic Multidecadal Variability				10
Pacific Decadal Variability				10
Annular modes (NAO, NAM, SAM)				10
Atlantic Meridional and Zonal Modes				

[END CROSS-CHAPTER BOX 2.1, TABLE 2 HERE]

Summary

Chapters 2, 3 and 4 assess our current knowledge around the evidence basis for climatic changes, their causes, and their potential future under different possible emissions pathways using a combination of observations and state-of-the-art earth system models. To facilitate this assessment we have defined what is meant by large-scale for the purpose of these chapters and then gone on to identify a key subset of indicators and modes of variability to consider. These indicators and modes of variability both characterise overall changes to the climate system as a whole and have direct public policy relevance.

[END CROSS-CHAPTER BOX 2.1 HERE]

2.3.1 Atmosphere

2.3.1.1 Surface and upper-air temperatures

2.3.1.1.1 Temperatures of the deep past

AR5 assessed the surface temperatures during six well-studied intervals of the deep past including three pre-Quaternary warm periods of the Cenozoic Era. AR5 assigned a *low confidence* level to the reconstructed temperature during the PETM (4° to 7°C warmer than pre-PETM mean climate), and a *medium confidence* to those of the EECO and MPWP (9° to 14°C and 1.9° to 3.6°C warmer than pre-industrial, respectively). AR5 assigned a *medium confidence* to the temperature of the LIG (no more than 2.0°C warmer than pre-industrial), and probabilities of *very likely* to the LGM temperature estimate (3-8°C lower than pre-industrial), and *likely* to the maximum rate of global warming during the subsequent deglaciation (1° to 1.5°C per thousand years).

The deep past periods in Table 2.1 attest to different states of the evolving climate system, and thereby provide the long-term context and broader perspective for recent climate changes. This assessment draws from studies based mostly or entirely on indirect observational evidence from geological archives (i.e., proxy records) rather than reconstructions relying heavily on modelled parameters (e.g., Friedrich et al., 2016; Köhler et al., 2015). For the PETM, reconstructions compiled in a recent data synthesis (Hollis et al., 2019b) indicate that global surface temperatures rose by 4° to 12°C compared to the pre-PETM mean state, with the greatest increases at high latitudes. For the EECO, a recent global summary (Caballero and Huber, 2013) indicates that GMST averaged over multiple millennia was 13° ± 3°C (±2 SE) warmer than modern, whereas (Hollis et al., 2019) suggests that it was likely warmer still. For the MPWP (Figure 2.10a), a recent summary (Martínez-Botí et al., 2015) indicates that sea-surface temperatures averaged across the 300,000-year-long interval were about 2.3°C warmer than the Holocene mean. Applying the scaling factor of 1.6 derived by Snyder (2016) to SST yields a GMST of 3.7° ± 0.8°C (±1SD) during the MPWP compared to pre-industrial. These new GMST estimates for the three oldest reference warm periods of the Cenozoic Era are consistent with prior estimates.

[START FIGURE 2.10 HERE]

Figure 2.10: Global mean surface temperature reconstructed from various proxy records over four time scales. (a) 60Ma to present based on a scaling of deep-ocean temperature inferred from oxygen isotopes (grey line) (Hansen et al., 2013), with temperatures adjusted (Δ temperature = 0°C) relative to mean of last 1600 years (youngest data point). (b) 4Ma to present based on the same data in panel a, along with a composite of multi-proxy sea-surface temperature reconstructions (blue line) (Snyder, 2016) adjusted to the mean of the last 1000 years (youngest data point). Note that the

temperature estimated for the last inter-glacial in these reconstructions ($+2^{\circ}\text{C}$ relative) is at the high end of the estimated range discussed in the text. (c) 20ka to present from the data syntheses of (Shakun et al., 2012) for the last deglaciation (black line) and (Marcott et al., 2013) for the Holocene (orange line), with 2.5% and 97.5% uncertainty bands. Temperatures were increased by 0.31°C from published values to adjust them to relative to pre-industrial (1850-1900), rather than relative to 1961-1990. (d) 2ka to present (aka, Common Era) relative to pre-industrial from a synthesis of seven different reconstruction methods (PAGES 2k Consortium, in revision) all based on the proxy data compilation of the (PAGES 2k Consortium et al., 2017), showing the median (solid line) and the 2.5th and 97.5th percentiles of the ensemble members (bands), smoothed with a 31-year Butterworth low-pass filter. Instrumental data (1850-2017) from (Cowtan and Way, 2014), smoothed like the reconstruction (red line).

[END FIGURE 2.10 HERE]

For the LIG (Figure 2.10b), new temperature reconstructions since AR5 are based on proxy data syntheses that yield harmonized time series from multiple sites, recognizing that peak warmth was not necessarily globally synchronous. A recent reconstruction (Hoffman et al., 2017) indicates that global average sea-surface temperature around 125 ka was $0.5^{\circ} \pm 0.3^{\circ}\text{C}$ higher compared to pre-industrial. Fischer et al. (2018) applied a scaling factor of 1.6 to this SST reconstruction to infer a GMST of $0.8^{\circ} \pm 0.5^{\circ}\text{C}$, which is on the lower bound of the 95% confidence range of 0.4° to 3.6°C (relative to the late Holocene) derived from Snyder's (2016) two-million-year-long GMST reconstruction. New reconstructions for high-latitude sea-surface temperatures (Capron et al., 2017) indicate that regional warming was amplified relative to the global mean by about a factor of approximately two in polar regions.

During the LGM (Figure 2.10b) temperature reconstructions show that the ocean surface generally cooled less than on land. A new reconstruction indicates GMST was 6.2°C [4.5° - 8.1°C 95% CI] cooler than the late Holocene average (Snyder, 2016), which is similar to area-weighted reconstructions from exclusively land-based sites ($6^{\circ} \pm 1^{\circ}\text{C}$ colder) (Harrison et al., 2015), despite clear evidence that the ocean surface cooled less. New reconstructions indicate that sea-surface temperature was $2.2^{\circ} \pm 0.4^{\circ}\text{C}$ cooler than the Holocene average (Martínez-Botí et al., 2015), which agrees with a network of area-weighted temperature reconstructions (Harrison et al., 2015). LGM temperature variability on century scale was about four times higher than during the Holocene, especially at high latitudes (Rehfeld et al., 2018). For the last deglaciation (Figure 2.10c), no new large-scale studies have been published since AR5 to further assess the rate of GMST change during this period of rapid global warming.

The level of uncertainties associated with large-scale temperature estimates for periods of the deep past has not changed substantially since AR5. While uncertainties generally increase for periods further back in time, along with the sparseness of the data coverage, the magnitude of temperature change also increases. Reconstructions of GMST are hampered by incomplete sampling and the influence of seasonality on the climate proxies. The reconstructions for the deep past rely mainly, or exclusively, on marine proxies, but uncertainties associated with inferring GMST from sea-surface temperatures are large. Because of geochronological uncertainties and limited replication, published uncertainty estimates, which are usually statistically derived, do not account for all sources of uncertainty.

In conclusion, GMST of the deep past were at times both higher and lower than has been observed during the instrumental period. GMST relative to pre-industrial ranged from at least 13°C warmer during the early Eocene climate optimum (around 50 Ma) to $6^{\circ} \pm 2^{\circ}\text{C}$ colder for the last glacial maximum (around 21-19 ka) (*medium confidence*). During the last inter-glacial, GMST averaged over the warmest millennium was $1^{\circ} \pm 0.5^{\circ}\text{C}$ warmer than pre-industrial (*medium confidence*) and *very likely* did not exceed 2°C higher.

2.3.1.1.2 Temperatures of the post-glacial period

Assessments of paleo temperatures in AR5 focused more on the Common Era (CE) rather than on any other past period. AR5 found that, for average annual Northern Hemisphere temperatures, the period 1983–2012 was *very likely* the warmest 30-year period of the past 800 years (*high confidence*) and *likely* the warmest 30-year period of the past 1400 years (*medium confidence*); the warm multi-decadal periods prior to the 20th Century were not as synchronous across regions as was the warming since the mid-20th century (*high confidence*). AR5 did not include an assessment of spatially integrated, large-scale temperature estimates for 6 ka, although it assigned *high confidence* to the long-term cooling trend that affected mid-to-high latitudes of the Northern Hemisphere during the 5000 years that preceded 20th Century warming. SR1.5 highlighted the Holocene thermal maximum, the period between around 10 and 5 ka when GMST is inferred to have been up to 1°C higher than pre-industrial, but which occurred at different places at different times and therefore overestimates the GMST for any single period. Instead of the Holocene thermal maximum, this section focuses on the long-standing PMIP time slice during the middle Holocene at 6 ka.

No new GMST reconstructions have been published for 6 ka, but Harrison et al. (2014) expanded a benchmarking dataset for 6 ka to include proxy indicators beyond the terrestrial pollen dataset (Bartlein et al., 2011) that was featured prominently in AR5 (Flato et al., 2013). Area-weighted mean temperature at 6 ka is indistinguishable from the pre-industrial in a network of land-based reconstructions, and slightly higher (about $0.2^{\circ} \pm 0.2^{\circ}\text{C}$) for sea-surface reconstructions (Harrison et al., 2015). A new compilation of multiproxy, predominantly mean annual proxy temperature data from marine and terrestrial sites primarily across the Northern Hemisphere (north of 10°S) adds context to the 6 ka PMIP time slice (Routson et al., in press) and supports the global and hemispheric reconstruction featured in AR5 (Marcott et al., 2013) (Figure 2.10c). Zonally averaged, the dataset shows that Holocene temperature evolved distinctly by latitude, with greater warming at 6 ka in the northern high latitudes than to the south. In aggregate, it shows the period between 8 and 5 ka was generally warmer in the Northern Hemisphere by about 0.5°C relative to pre-industrial, although the warming was likely greater regionally, especially when averaged over shorter periods. Numerous new site- and regional-level studies generally agree with the trends summarized in AR5. For example, a new pollen-based temperature reconstruction for North America and Europe shows that mean annual temperature was 0.5° to 0.7°C higher during the warmest millennium of the Holocene, which was centred around 5 ka (Marsicek et al., 2018).

The temperature history of the last millennium and the methods used to reconstruct it have been studied extensively, as summarized recently by Christiansen and Ljungqvist (2017) and Smerdon and Pollack (2016). The last millennium largely coincides with the PMIP millennium transient simulations from 850 to 1859 CE. Despite its well documented oversimplification (e.g., Dätwyler et al., 2017; Neukom et al., in press; Steiger et al., 2018), this report follows AR5, which assigned the mostly warm conditions from about 950 to about 1250 to the MCA and the mostly colder conditions from about 1450 to about 1850 to the LIA (Table 2.1). These epochs were followed by industrial-era warming, which commenced as early as the mid-19th century over the Northern Hemisphere continents (Abram et al., 2016).

Since AR5, new regional (e.g., Stenni et al., 2017; Werner et al., 2018), quasi-hemispheric (Anchukaitis et al., 2017; Neukom et al., 2014; Schneider et al., 2015), and global (Tardif et al., 2018) temperature reconstructions have been published based primarily on evidence from tree rings, which reflect growing-season conditions and rarely extend back over the entire CE. In addition, a major new global compilation of multiproxy paleo-temperature records for the CE (PAGES 2k Consortium, 2017) has been analysed using a variety of statistical methods for reconstructing temperature (Neukom et al., in press; PAGES 2k Consortium, 2017). Unlike previously reported proxy data networks, which yield more method-dependent reconstructions (Wang et al., 2015), the new data synthesis shows that different reconstruction methods result in similar patterns of multi-decadal temperature variability. The median of the multi-method ensemble of GMST reconstruction from this

synthesis (Figure 2.10d) generally agrees with the patterns reported in AR5, while affording more robust estimates (*medium confidence*) of the following major features of GMST during the CE: (1) an overall millennial-scale cooling trend of -0.23°C/ka (-0.31 – -0.11) prior to 1900; (2) multi-centennial intervals of generally higher and lower temperatures during the MCA and LIA, respectively (MCA = 0.14°C (-0.05 – 0.26), LIA = -0.03°C (-0.30 – 0.06), relative to pre-industrial); (3) a warmest 50-year period, which occurred during the final 50 years of the reconstruction; and (4) a rate of global warming during the last 50 years (1968–2017, based on instrumental evidence) that exceeds the rate over any other 50 year period.

The amount and variety of publically accessible proxy temperature data, and its temporal and spatial resolution have improved since AR5. Nonetheless, reconstructions of GMST are hampered by the influence of seasonality, the sometimes confounding influences of moisture changes on the temperature proxies, and the likely bias of biological indicators to represent summer conditions, which can confound reconstructions and annual means. The availability of different proxy archives is limited by their natural spatial and temporal coverage. For example, the vast majority of data used to reconstruct temperature of the CE are from tree rings and as such only a small fraction extends back through the first millennium CE. Climate field reconstruction methods that impose a space-time structure based on a modern calibration period assume that such patterns are stationary and may therefore implicitly consider remote teleconnections to be stable. Accurately calibrating lower-resolution and time-uncertain proxy records to specific climate variables remains challenging, with published uncertainties unlikely to include all sources of uncertainty.

To conclude, surface temperatures over the Northern Hemisphere, and possibly also the Southern Hemisphere, have tended to gradually decrease over the past several thousand years (*medium confidence*) culminating in the LIA, which was globally the coldest multi-century interval of the Holocene (*high confidence*). This cooling trend was reversed during the mid 19th Century (*high confidence*). The GMST of the past 50 years was *likely* warmer than any 50 year mean of the MCA (*high confidence*), and *more likely than not* was the warmest 50-year period of the CE (*medium confidence*).

2.3.1.1.3 Temperatures during the instrumental period – surface

AR5 found that it was certain that global mean surface temperatures had risen since the late 19th century. Total warming in global mean surface temperatures was assessed at 0.85 (0.65 – 1.06) $^{\circ}\text{C}$ over the period 1880–2012, whilst the change from 1850–1900 to 2003–2012 was assessed at 0.78 (0.72 – 0.85) $^{\circ}\text{C}$. SR1.5 reported warming from 1850–1900 to 2006–2015 of 0.87°C , with an 1880–2012 trend of 0.86°C and an 1880–2015 trend of 0.92°C .

Since AR5, there have been substantial improvements in the availability of instrumental archive data both on land and over the ocean. The International Surface Temperature Initiative have released a much improved collection of fundamental land surface air temperature records (Rennie et al., 2014) comprising in excess of 35,000 station records. A new version of the International Comprehensive Ocean-Atmosphere Data Set (ICOADS Release 3.0, Freeman et al., 2017) comprises over 450 million in situ marine reports and incorporates recently digitised data, increasing coverage in data sparse regions and times (e.g. polar oceans and WWI). These advances have permitted enhanced consideration of both land and marine data quality.

For SST analyses, three long-term products – HadSST.3.1.1.0 (1850–present, Kennedy et al., 2011a, 2011b), ERSSTv5 (1880–present, Huang et al., 2017) and COBE SST2 (1880–present, Hirahara et al., 2014) – now have bias adjustments applied throughout the record, as does a complementary data set of marine air temperatures, HadNMAT2 (1890–2010, Kent et al., 2013). The SST adjustments account for two major issues previously identified in AR5: that globally averaged buoy SSTs are about 0.12°C cooler than ship SSTs (Huang et al., 2015; Kennedy et al., 2011a, 2011b), and that SSTs measured at ship engine room intakes can be biased warm or cold (Kent and Kaplan, 2006). The first issue

1 primarily affects data since 2000, when buoys began to play an increasing role in the observation
2 network, and the second issue primarily affects data from the 1940s to the 1970s. From the standpoint
3 of uncertainty, ERSSTv4 (Huang et al., 2016; Liu et al., 2015a) and HadSST.3.1.1.0 now have
4 estimates presented as ensembles. Comparisons between these independently-derived analyses and
5 the assessed uncertainties (Kennedy, 2014; Kent et al., 2017) support the conclusion that global mean
6 SST increased since the start of the 20th century, a conclusion that is insensitive to the method used to
7 treat gaps in data coverage (Kennedy, 2014).

8
9 A number of recent studies also corroborate important components of the long-term SST record
10 (Cowtan et al., 2018; Hausfather et al., 2017; Kent et al., 2017); in particular, ATSR SST retrievals
11 (Berry et al., 2018; Merchant et al., 2012), oceanographic profiles (Gouretski et al., 2012; Huang et
12 al., 2018), and coastal observations (Cowtan et al., 2018) have all been shown to be consistent with
13 the homogenized analyses. Hausfather et al. (2017a) also affirmed the higher rate of warming at a
14 global scale seen in ERSST since the late 1990s. Nevertheless, dataset differences still exist in the
15 mid-20th century when there were large, poorly-documented changes in instrumentation (Kent et al.,
16 2017), particularly during the WWII period, when the number of available ship observations was
17 much lower and these measurements disproportionately arise from US naval sources. Kennedy et al.
18 (submitted) also identify differences between the new HadSST4 dataset and other SST datasets in the
19 1990s, indicating that some level of structural uncertainty remains during this period. [Note that
20 HadSST4 data are not yet publicly available and are not used in FOD analyses. A fuller integration of
21 HadSST4 findings and associated redrafting will occur prior to SOD].

22
23 A new version of the Global Historical Climatology Network (GHCN, Menne et al., 2018) was
24 released, which includes many more land stations, and calculates a parametric uncertainty ensemble.
25 This ensemble draws upon the benchmarking analysis of (Williams et al., 2012) to define a draw of
26 100 plausible set of parameter settings for the homogenization processing. It in addition accounts for
27 sampling effects. The CRUTEM dataset has undergone annual updates which have served to increase
28 data completeness and quality. A new global land data set, the China Land Surface Air Temperature
29 (CLSAT) dataset (Xu et al., 2018) has higher network density in some regions (particularly Asia) than
30 previously existing datasets. Global trends in CLSAT are generally consistent with other land datasets
31 through 2014 (Xu et al., 2018).

32
33 An ongoing limitation of conventional data sets is lack of coverage of data-sparse areas, especially
34 high-latitude areas. In particular, the CRUTEM4 (Jones et al., 2012; Morice et al., 2012) and NOAA
35 (Vose et al., 2012) land data sets used in AR5 had limited polar coverage. Cowtan and Way (2014)
36 applied a kriging-based method to extend existing data sets to polar regions. As the Arctic has been
37 warming faster than the global average (e.g. Davy et al. (2018)), under-sampling of the Arctic will
38 result in an under-estimation of recent warming trends. Results from other approaches (Hausfather et
39 al., 2017) and comparison with reanalyses (Simmons et al., 2017) also find that data sets with full
40 polar coverage show stronger recent warming than those which do not. [Text to be substantially
41 modified here in later versions as new data products likely to be spatially complete are published by
42 the major analysis centres prior to subsequent drafts]

43
44 AR5 identified diurnal temperature range (DTR) as a substantial knowledge gap, with no new
45 comprehensive global analysis published since that of Vose et al. (2005). The most recent analysis of
46 Thorne et al. (2016a, 2016b) compared a broad range of gridded estimates of change in DTR,
47 including a new estimate derived from the ISTI databank release using the pairwise homogenization
48 algorithm used to create GHCNv4, and grids derived from the CRU TS (Harris et al., 2014), Berkeley
49 Earth (Rohde et al., 2013), HadEX2 (Donat et al., 2013a), GHCNDEX (Donat et al., 2013b), Vose et
50 al., (2005) and HadGHCND (Donat et al., 2013a) data sets. The analysis highlighted substantial
51 ambiguity in pre-1950 estimates arising from sparse data availability. This made estimates highly
52 sensitive to averaging and gridding choices. Post-1950 there was broad agreement that DTR had
53 decreased globally with most of that decrease occurring over the period 1960-1980. Regionally,
54 changes were more mixed in timing with DTR in North America and Europe increasing in recent

decades while over SE Asia they continued to decrease. A second analysis by Sun et al. (2018), using the global CLSAT data set (Xu et al., 2018) found increasing DTR from 1901 to 1950, mostly in the northern hemisphere, and confirmed decreasing DTR after 1950, with most of that decrease occurring between 1950 and 1980.

No recent evidence has emerged to alter the AR5 finding that it is *unlikely* that any uncorrected effects from urbanization, or from changes in land use or land cover, have raised global surface air temperature trends on land by more than 10%, although larger signals have been identified in some specific regions, especially rapidly urbanizing areas such as eastern China (Li et al., 2013; Liao et al., 2017). There is also no clear evidence that site-specific data homogeneity issues have had any significant impact on trends at global scale since the early 20th century, with some uncertainty prior to then, mainly arising from a lack of standardization of instrument shelters, which has been largely accounted for in data from central Europe (Jones et al., 2012) but less so elsewhere. Hawkins et al. (2017) found differences of up to 0.1 °C in the warming from 1850-1900 to 1986-2005 between the three data sets (HadCRUT4, Cowtan and Way (2014), Berkeley Earth) which extend back to 1850, which they concluded was at least partially indicative of structural uncertainties in the pre-1900 period.

Recent warming has been strongest at high latitudes in the northern hemisphere in winter; 1995-2015 trends north of 70 °N were 2-4 times those for the northern hemisphere as a whole (depending on the data set), reaching 6-9 times the hemispheric average in winter (Davy et al., 2018). More generally, most land areas in the extratropical northern hemisphere have warmed faster than the GMST average over both the 1900-2018 and 1980-2018 periods (Figure 2.11), although at more local scale considerable uncertainty is introduced by large differences in trends between different LSAT data sets (Rao et al., 2018). The five conventional surface temperature products are in robust agreement that each of the last four decades has been the warmest globally since the beginning of the records (Figure 2.12, Table 2.3).

[START FIGURE 2.11 HERE]

Figure 2.11: Trends in a range of global data sets over the 1900–2018 (left) and 1980–2018 (right) periods [significance overlay to be added in SOD], calculated using ordinary least squares, with time series (bottom left) of annual anomalies from a 1981–2010 base period. Trends are shown where data are available for all 12 months in at least 70% of years during the trend period, and at least 20% of years during the first and last decile of the trend period.

[END FIGURE 2.11 HERE]

[START FIGURE 2.12 HERE]

Figure 2.12: Global mean timeseries at annual and decadal scales for the five conventional surface products from 1850 to date (as available). The HadCRUT4 estimates include uncertainties (grey shading) calculated per (Morice et al., 2012). All timeseries are rebased relative to a 1961-90 climatology

[END FIGURE 2.12 HERE]

Another significant development since AR5 has been the incorporation of reanalyses as a standard part of operational GMST monitoring. It was reported in AR5 that various reanalyses were broadly consistent with conventional surface data sets in the representation of trends since the mid-20th century. Since that time, Simmons et al. (2017) found that the ERA-Interim (Dee et al., 2011) and JRA-55 (Ebata et al., 2011; Harada et al., 2016) reanalyses continued to be consistent, over the last 20 years, with those surface data sets which fully represented the polar regions. However, the MERRA-2

reanalysis (Gelaro et al., 2017) GMST spuriously cooled sharply relative to ERA-Interim and JRA-55 over recent years. Although biases in reanalysis mean temperatures have been found in regions such as Antarctica (Bracegirdle & Marshall, 2012; Jones & Lister, 2015; Zhang et al., 2018), GMST trends from several modern reanalyses are consistent with those from conventional surface data sets. Since the early 2000s, analyses of surface and near-surface temperature have also been available from various satellite-based platforms (Famiglietti et al., 2018; Prakash et al., 2018; Susskind et al., 2019), which have the potential to be of value in monitoring data-sparse areas into the future.

Finally, to aid later chapters it is necessary to provide an estimate of GSAT from the available GMST estimates. Surface temperatures consist of a range of inter-related parameters which may differ at both high and low frequencies (Merchant et al., 2013). GMST is the combination of air temperatures over land with temperatures of the sea at a nominal depth below the surface. GSAT is the combination of air temperature over both surfaces. Conversion can be achieved by applying an adjustment factor with uncertainty and that is subject to assumptions around spatio-temporal invariance. As such there will inevitably be *lower confidence* in estimates of GSAT than of GMST. Cowtan et al. (2015), Huang et al. (2015) and Richardson et al. (2018) have variously addressed the question of systematic differences between GSAT and GMST, finding that models indicate that air temperature over the oceans would be expected to warm slightly faster than SSTs based upon adiabatic theory. Model simulations with historical forcings (Richardson et al., 2018) found that using air temperature over oceans, rather than SST, increased the rate of warming from 1861-1880 to 2007-2016 by 5.8% (3.2%-7.3%), whilst a GSAT calculated with full global coverage warmed 16.2% (5.2% - 28.7%) faster than a GMST masked to approximate the coverage in the HadCRUT4 data set. However, comparisons reported in AR5 between the HadNMAT2 data set (Kent et al., 2013b) for night marine air temperature (NMAT) and the HadSST3 SST data set show little evidence to date of systematic differences between the two in observed data at global scale, except in the period during and immediately after World War II. Also, those SST data sets (such as ERSSTv5) which use NMAT as a reference for adjustments to SST may alias the GSAT-GMST difference into those adjustments to some extent. Given the quantified uncertainties in NMAT and SST it is *likely* that it would take some decades of further warming for systematic differences between GMST and GSAT to become detectable in the observations, but given the robust theoretical basis the effect is *very likely* real. Considering the available evidence at this time an inflation of 5.8% to the trend of GMST with a range of 3.2 to 7.3% is applied to the globally complete GMST records to provide a GSAT estimate for later chapters. [by time of final publication all estimates are likely to be globally complete]

In summary, it is certain that global mean surface temperatures have increased since the late 19th century. Each of the last four decades has been warmer than any decade that preceded it. The total estimated warming of GMST from 1850-1900 to 1995-2014, is 0.83 °C (0.72 – 0.97 °C). The estimated warming of GMST from 1850-1900 to 2009-2018 is 0.99 °C (0.86 – 1.14 °C), and from 1850-1900 to 2002-2018 is 0.95 °C (0.81 – 1.09 °C). Estimated total change from a linear trend fit over the 1880-2018 period is 1.01 °C (0.73 – 1.28 °C), with 1980-2018 trends of 0.70 °C (0.56 – 0.88 °C).

Estimating changes in Global Surface Air Temperature requires application of an adjustment factor to available GMST estimates to account for physically well-understood differences to maintain marine boundary layer stability. This adjustment factor is estimated as 1.05 (1.03-1.07), increases uncertainty and thus lowers confidence. Estimated GSAT changes are 0.88 °C (0.75 – 1.04 °C) over 1850-1900 to 1995-2014 and 1°C (0.83-1.17°C) for 1850-1900 to 2009-2018. Estimated total changes from a linear trend fit in GSAT are 1.07 °C (0.75-1.37°C) for 1880-2018, and 0.74 °C (0.58-0.94°C) for 1980-2018.

It is *virtually certain* that diurnal temperature range decreased over the 1951-2012 period. It is *likely* that most of the decrease occurred over 1960-1980 with little change since. Change in DTR is small relative to increases in land surface air temperatures, and hence it is *virtually certain* that both daily maximum and daily minimum temperatures have increased globally since 1950.

[START TABLE 2.3 HERE]

Table 2.3: Observed increase (°C) in global mean surface temperature (GMST) in various datasets. Numbers in square brackets indicate 5–95% confidence range in trend magnitude. Trend values are calculated with ordinary least squares with AR(1) correction.

Diagnostic/ dataset	1850–1900 to 1995–2014	1850–1900 to 2009–2018	Trend 1880–2018	Trend 1960–2018	Trend 1980–2018
HadCRUT	0.76	0.90	0.94 [0.76–1.12]	0.85 [0.72–0.98]	0.68 [0.57–0.78]
NOAAGlobalTemp			0.98 [0.73–1.22]	0.93 [0.83–1.03]	0.66 [0.56–0.76]
GISTemp			1.02 [0.78–1.25]	0.97 [0.86–1.07]	0.68 [0.57–0.80]
Cowtan-Way	0.82	0.99	0.99 [0.81–1.17]	0.90 [0.77–1.03]	0.73 [0.63–0.84]
Berkeley Earth	0.92	1.09	1.11 [0.94–1.28]	0.93 [0.81–1.05]	0.74 [0.64–0.85]
Average	0.83	0.99	1.01	0.91	0.70
ERA-Interim					0.73 [0.59–0.88]
JRA-55				0.95 [0.82–1.08]	0.71 [0.59–0.82]

[END TABLE 2.3 HERE]

2.3.1.1.4 Temperatures during the instrumental period – free atmosphere

AR5 reported that it was *virtually certain* that tropospheric temperatures have risen, and stratospheric temperatures fallen, since the mid-20th century, but that assessments of the rate of change and its vertical structure had only *medium confidence* in the northern hemisphere extratropics and *low confidence* elsewhere. In particular there was *low confidence* in the vertical structure of temperature trends in the upper tropical troposphere, which were observed to have been overestimated by most CMIP5 models over the 1979–2012 period.

Radiosonde data with sufficient coverage for a quasi-global assessment exist back to 1958. The major radiosonde data sets with updates through 2018 are RAOBCORE, RICH (Haimberger et al., 2012) and University of New South Wales (UNSW) (Sherwood and Nishant, 2015). Other radiosonde data sets reported in AR5 have had limited or no updates since. All radiosonde data sets have limited coverage in the tropics. Reanalyses products have identified limitations near the 300 hPa level where the proportion of observations which are made by aircraft has increased rapidly in recent years (Dee et al., 2011; Gelaro et al., 2017) but are generally well aligned with radiosonde and satellite observations in the middle and lower troposphere and lower stratosphere.

Microwave retrieval satellite data extend back to 1979. The most recent versions of satellite-based data sets, using the Microwave Sounding Unit (MSU) and Advanced Microwave Sounding Unit (AMSU), are UAHv6.0 (Spencer et al., 2017), RSSv4.0 (Mears and Wentz, 2017) and NOAA STAR (Zou and Wang, 2011). There are three layers which are used for common reporting, the lower stratosphere (LS), mid-troposphere (MT) and lower troposphere (LT) (UAH and RSS only). The LT product is derived from recombination of primary sensor channels and there are subtle, but potentially

important, differences in how the two groups perform this. Updates to the data sets have led to some level of convergence (Maycock et al., 2018); in particular, the move to UAHv6.0 has addressed homogeneity issues identified by Seidel et al. (2016).

Global Navigation Satellite Systems (GNSS, such as e.g. GPS) radio occultation (RO) limb soundings provide a self-calibrated SI traceable satellite measurement (Anthes, 2011; Ho et al., 2010), yielding globally distributed, high vertical resolution profile information in the upper troposphere and stratosphere from the early 2000s (e.g. Angerer et al., 2017a; EUMETSAT, 2019). The measured time delay can be converted to temperature, water vapour density and other parameters (Kursinski et al., 1997; Scherllin-Pirscher et al., 2017); this processing does however add structural uncertainty to the data. CHAMP RO instrument data have primarily been used to determine this uncertainty, confirming that temperature trends agree within $<0.06^{\circ}\text{C}$ over the 7 year mission between 50°S and 50°N , between 8 km to 25 km, and that the uncertainty is larger at high latitudes and at higher altitudes (Ho et al., 2012; Steiner et al., 2013). Data sets have been compared against MSU/AMSU channel 4/9 data records, as well as against several AMSU channels from the Aqua satellite, finding almost identical trends (Khaykin et al., 2017). Comparison of RO with collocated radiosondes, Vaisala RS90/92 and GRUAN data product (RS92-GDP), showed very good agreement with global annual mean temperature differences of less than 0.2°C in the upper troposphere and lower stratosphere. Radiosonde daytime radiation biases were identified at higher altitudes (Ladstädter et al., 2015; Ho et al., 2017). The stability and high accuracy of RO also makes this data type a useful calibration reference for AMSU (Chen and Zou, 2014) and radiosondes (Ho et al., 2017; Tradowsky et al., 2017). The effective vertical resolution of RO measurements in the upper troposphere and lower stratosphere was recently investigated (Zeng et al., 2019) and is found to be up to 100 m at the tropical tropopause.

Warming continues in the lower troposphere in all radiosonde, reanalyses and MSU data sets, with a rate over the 1980-2018 period generally comparable with surface warming rates over the same period (Figure 2.13). Radiosonde-based products generally show faster warming rates than satellite-based products and reanalyses. The warmest year on record in all lower tropospheric products assessed was 2016. Sherwood and Nishant (2015) identify a region of enhanced warming in the tropical upper troposphere, with a warming trend of 0.25°C per decade in homogenized radiosonde data for 1979-2012 at 300 hPa in the region from 30°S to 30°N , compared with 0.14°C per decade near the surface. Data reported in Christy et al. (2018) also indicate faster warming rates near the 300 hPa level than at the surface in the tropics for the 1979-2016 period in the UNSW, RAOBCORE and RICH data sets. Earlier analyses by Allen and Sherwood (2008), who inferred tropical upper tropospheric temperatures through the relationship between temperature gradients and wind shear, also found evidence of enhanced warming in the tropical upper troposphere over the 1979-2005 period. Vertically resolved trends in radio occultation data, supported by radiosonde data sets, for the period 2002 to 2018 show a significant warming trend in the mid- to upper troposphere over all latitudes (Figure 2.13); in the tropics, also the lowermost stratosphere is warming while at higher levels a cooling is visible after QBO and ENSO effects are removed. MSU tropical MT show that warming rates are near or above those in LT, but the MSU data are measuring much broader layers which greatly complicates interpretation. Direct comparison with reanalysis data sets is difficult, due to biases in reanalysis datasets near the 300 hPa level owing to ambiguities arising from aircraft observations (Dee et al., 2011; Gelaro et al., 2017) and high uncertainty owing to the sparse observational constraint in the deep tropics.

A rise in the height of the tropopause of 40 to 120 m per decade between 1981 and 2015 was determined from both radiosonde and reanalysis data sets (Xian and Homeyer 2018). Local studies (e.g. Tang et al., 2017; Chen et al., 2019) found stronger trends in some regions near the subtropical jet linked to tropical expansion (Section 2.3.1.3.1). Whilst Seidel and Randel (2006) found that the tropopause height was more closely coupled with temperatures in the stratosphere than the troposphere, it is not yet clear whether the rate of increase in tropopause height has experienced a similar recent slowdown to that of the rate of cooling of the lower stratosphere. Radio occultation data (Gao et al., 2015) indicate little change in tropopause height over the period from 2006 to 2014, but a

significant warming is observed over the period 2002 to 2018 (Figure 2.13).

Lower stratospheric temperatures have cooled over the 1979-2018 period in all data products (Figure 2.13), even if the influence of the El Chichon (1982) and Pinatubo (1991) eruptions is removed. The rate of cooling has slowed sharply since the late 1990s with no data set showing a significant trend over the 2000-2018 period, although 2016 was the coldest year on record in all data sets.

Since AR5, which provided no new update of changes in the middle and upper stratosphere, a new operational mid- and upper-stratospheric data set has been developed by Zou & Qian (2016), merging the previous 1979-2006 SSU data set (Zou et al., 2014) with a data set from 1998 onwards drawn from relevant AMSU channels (Wang and Zou, 2014). This allows assessment of trends in the middle and upper stratosphere from 1980 to present. Further satellite-based data sets covering fixed periods have been developed by Randel et al. (2016) and McLandress et al. (2015). Reanalysis data are of limited value in the middle and upper stratosphere because of known biases (Rienecker et al., 2011; Long et al., 2017; Wright and Hindley, 2018; Karami, 2019), due to a combination of a lack of radiosonde observations at those levels to facilitate transitions between observations from different satellites and sparse vertical model resolution which inhibits capturing relevant dynamical processes. The data set of Zou & Qian (2016) shows cooling in the middle and upper stratosphere since 1979 with a trend of $-0.56^{\circ}\text{C} \pm 0.15^{\circ}\text{C}$ per decade for the mid-stratosphere and $-0.62^{\circ}\text{C} \pm 0.14^{\circ}\text{C}$ per decade for the upper stratosphere over the 1979–2018 period, although both cooling rates have slowed substantially since the mid-1990s, and the overall trend is reduced in magnitude by about 0.10°C per decade at both levels if the influence of the El Chichon and Pinatubo eruptions is removed (Zou and Qian, 2016). The results obtained by Randel et al. (2016) for 1979-2015, and McLandress et al. (2015) for 1980-2012, are broadly consistent with this.

In summary, the troposphere has warmed since the mid-20th century. Lower tropospheric temperature has warmed at a rate of $0.13\text{-}0.23^{\circ}\text{C}$ per decade over both the 1961-2018 and 1980-2018 periods. There is *medium confidence* that temperatures in the tropical upper troposphere have warmed faster than those at the surface since 2001, but *low confidence* in the magnitude and vertical extent of the difference, and in changes prior to 2001. It is *virtually certain* that the tropopause height has risen over the 1980-2018 period but there is *low confidence* in the size of the change, or whether the rate of change has changed in the later part of the period.

It is *virtually certain* that the lower stratosphere has cooled since the mid-20th century, with a linear trend-based cooling rate of $0.22\text{-}0.29^{\circ}\text{C}$ per decade over both 1960-2018 and 1980-2018. However, most data sets show that lower stratospheric temperatures have stabilized since the mid-1990s with no significant cooling over the last 20 years. It is *likely* that middle and upper stratospheric temperatures have decreased since 1980, but there is *low confidence* in the magnitude.

[START FIGURE 2.13 HERE]

Figure 2.13: (Top left) Trends in temperature at various heights for 1980-2018 (left) and 2002-2018 (centre), from the RAOBCORE and RICH radiosonde data sets, the ERA-Interim reanalysis and, for 2002-2018, the Wegener Centre radio occultation (RO) data set (Angerer et al., 2017b), for near-global (70°N to 70°S) (upper) and tropical (20°N to 20°S) (lower) regions. (Top right) Latitude/altitude cross-section of temperature trends for 2002-2018 (with ENSO and QBO influences removed by multiple regression) for the Wegener Centre radio occultation data set (upper) and the ERA-Interim reanalysis (lower). (Bottom) Time series of globally-averaged (upper) lower troposphere and (lower) lower stratosphere temperatures, using a 2002-2016 baseline to match the availability of radio occultation data. [Note known need to better label panels and improve description accordingly in SOD. Will likely also flip the order of lower two panels so stratosphere above troposphere for physical intuition]

[END FIGURE 2.13 HERE]

2.3.1.2 Global hydrological cycle

This section focuses on global and large-scale change in a subset of components of the hydrological cycle (Cross-Chapter Box 2.1). The reader should note that Chapter 8 undertakes a holistic assessment of changes in the hydrological cycle integrating observations, modelling and theoretical understanding. Readers interested in regional scale changes and a more in-depth treatment of critical aspects such as monsoon characteristics and changing seasonality should thus refer to Chapter 8.

2.3.1.2.1 Paleo perspective of the global hydrological cycle

AR5 assessed large-scale (continental to global) indicators of terrestrial paleo hydroclimate in several sections, including as part of its assessment of model performance and paleo floods and droughts. Paleo evidence indicated that droughts and floods of greater magnitudes and longer durations than those recorded in the 20th century occurred during the past five to ten centuries over many regions (*high confidence*). The paleoclimate evidence presented therein was broadly consistent with global hydroclimate scaling with temperature such that warmer periods are wetter (e.g., the Pliocene; increased P-E) and colder drier (e.g., the LGM; decreased precipitation and precipitable water vapour content). However, there were few statements on continental scale hydroclimate variability and change, or the large-scale hydroclimate state during the Holocene and CE. This is, in part, because of the complexity of hydroclimate, which tends to be regional and involves feedbacks between multiple components of the climate system. Indicative of this complexity are the range of metrics by which hydroclimate is defined—these include inherent climate variables like precipitation, precipitation minus evaporation (P-E), salinity, soil moisture at varying depths, evapotranspiration and multivariate metrics like the Palmer Drought Severity Index (PDSI) and the Standardized Precipitation Evaporation Index (SPEI).

While consistency is expected between some of these metrics, research since AR5 has established that they can decouple on paleoclimate timescales. For instance, while past research has suggested that the LGM was drier than present (decreased precipitation and precipitable water vapor), based on evidence of sparse vegetation and high dust deposition (Williams, 2014), lake levels appear to have been generally higher (Putnam and Broecker, 2017; Scheff et al., 2017) likely indicating greater runoff and wetter topsoils (increased soil moisture and streamflow). Likewise, large scale changes during the LGM mask continental and regional variability that was even larger in magnitude (e.g., Lowry and Morrill, 2018; Oster et al., 2015) and likely driven by dynamic rather than thermodynamic considerations. Reconciling the hydroclimate of the Pliocene also appears to require dynamic considerations—while it was both globally wet and warm, the subtropics and midlatitudes of both hemispheres were much wetter than present (increased P-E) despite similar CO₂ levels (Burls and Fedorov, 2017; Prentice et al., 2017).

Quantifying paleo hydroclimate variability requires spatial and temporal sampling that is not yet available for periods prior to the in the Holocene. Research since AR5 suggests that, during the Holocene, North America became increasingly wet (increasing P-E), with those conditions peaking at present (Shuman and Marsicek, 2016), and that there was considerable hydroclimate variability on centennial to millennial timescales (Newby et al., 2014; Shuman and Marsicek, 2016). This is consistent with recent evidence that the Northern Hemisphere mid-latitudes as a whole became increasingly wet during the Holocene (increasing P-E), in phase with the strength of the latitudinal temperature and insolation gradients (Routson et al., 2019). There is sufficient sampling during the CE to determine the “unusualness” of hydroclimate variability during the observational interval, given relatively stationary boundary conditions. The MCA, for instance, was broadly drier than present in North America, as inferred by hydroclimate proxy synthesis that revealed the occurrence of megadroughts and pan-continental droughts (Cook et al., 2016). These records exhibit large scale multidecadal to centennial variations during the hydroclimate transition from the MCA into the LIA,

with no clear trend in the past 900 years.

To summarise, prior to the Common Era the available data on hydroclimate change and variability are sparse and there is *low confidence* in large-scale reconstructions. Hydroclimate during the Common Era is dominated by regional heterogeneity, generally precluding definitive statements about long-term changes at continental and larger scales.

2.3.1.2.2 Surface humidity

AR5 reported *very likely* widespread increases in near surface air specific humidity since the 1970s, abating from around 2000 to 2012 (*medium confidence*). This latter period resulted in a decline in relative humidity consistent with the greater warming of the land surface compared to the ocean surface.

Near surface humidity over land and ocean can be monitored using in situ data (HadISDH; Willett et al., 2014, Willett et al., in prep) and reanalyses (ERA5, ERA-Interim, JRA-55 and MERRA-2). There is generally good agreement between these products (and older, now static products) in the year-to-year variability and long-term trend directions giving *high confidence* in increasing specific humidity since the 1970s and decreasing relative humidity since 2000. Since 2012, specific humidity over land and ocean has remained well above the long-term average and reached record or near-record values, with the strong 2016 El Niño event boosting surface moisture levels. The abatement mentioned in AR5 is no longer apparent. This is consistent with the behaviour of tropospheric moisture in the form of total column water vapour (TCWV, (Section 2.3.1.2.3)) and a resumption of rapid warming in surface temperatures (Section 2.3.1.1.3). The relative humidity has remained well below average over land.

From 1973 to 2018 increases in specific humidity have been widespread and significant across the majority of the land surface where observations are available (Figure 2.14). In contrast, trends in relative humidity from 1973 to 2018 show distinct spatial patterns with generally increasing saturation over the higher latitudes and tropics and generally increasing aridity over the extratropics and mid-latitudes (Figure 2.14). Hence, regions can be experiencing increasing absolute surface moisture yet still experiencing decreasing relative humidity, therefore becoming more arid.

Trend analyses highlight discrepancies across the various monitoring products, especially over ocean and for the decline in relative humidity. Sources of uncertainty include the initial measurement, climatology representativeness, homogenisation over land, ship height and instrument bias adjustment over ocean, whole number reporting, sparse spatio-temporal sampling over a gridbox and incomplete coverage uncertainty. For reanalyses there is also a model uncertainty component which is largest in regions with few or no observations. The largest source of uncertainty appears to be incomplete coverage; notably, sampling ERA-Interim to the HadISDH mask gives much closer agreement (Dunn et al., 2017).

In summary, observations show a *very likely* increase in near surface specific humidity over both land and ocean since the 1970s. A *very likely* decrease in relative humidity was observed over land areas since 2000, particularly over midlatitude regions of the Northern Hemisphere.

[START FIGURE 2.14 HERE]

Figure 2.14: Decadal trends in a) surface specific humidity and b) surface relative humidity over 1973 to 2018 from HadISDH in situ monitoring product. Trends are fitted using the median of pairwise slopes. Trend significance is assigned to gridboxes (identified with crosses) where the 5th to 95th percent confidence intervals are both above or below zero.

[END FIGURE 2.14 HERE]

2.3.1.2.3 Changes in total column water vapour

AR5 concluded that increases in total column water vapour were *very likely* since the 1970s, occurring at a rate that is generally consistent with the Clausius-Clapeyron relationship (about 7% per degree Celsius) and the observed increase in atmospheric temperature.

Records prior to the instigation of global radiosondes require the use of surface-only reanalysis products or statistically-based estimates based upon relationships with SSTs (Smith and Arkin 2015). The resulting statistically-based estimates of global-mean oceanic TCWW have two periods of increasing trend strongly linked to SST variations, one during 1910-40 and the other beginning 1975. The leading modes of global water vapour variations over the twentieth century are controlled by global temperatures, the Atlantic multidecadal oscillation (AMO), and ENSO (Mieruch et al., 2014; Shi et al., 2018; Zhang et al., 2013). Potential sources of errors in the SST-based estimation of TCWW include uncertainties in historical SST, uncertainties in the parameters that define the relationship between both variables and the influence of other factors like water vapour advection (Smith and Arkin, 2015). Trends based on surface-only reanalyses indicate an increase in TCWV over much of the globe, particularly since the beginning of the 20th century and over the tropics (Bordi et al., 2015) (Smith and Arkin, 2015). TCWV trends estimated since the middle of the twentieth Century from radiosonde observations show positive significant trends over North America and large portions of Eurasia, while negative trends are mostly located over Oceania (Chen and Liu, 2016; Zhang et al., 2018d).

Since the latter part of the 20th Century a range of satellite-based techniques allow quasi-global assessment. Several products provide water vapour measurements based upon several spectral domains, radiosondes, reanalysis products and GPS radio occultation. The GEWEX Water Vapour Assessment (G-VAP) provided an intercomparison of several total column water vapour (TCWV) data records, with global coverage but limited temporal resolution (Schröder et al., 2016, 2018). The various products generally exhibit a positive trend since 1979, which is particularly evident over the tropical oceans (Chen and Liu, 2016; Gu and Adler, 2013; Mieruch et al., 2014; Wang et al., 2016a). The existence of apparent breakpoints in available TCWV records, which in most cases coincide temporally with changes in the observing system, affected trend estimates based on satellite and merged products over Central Africa, the Sahara and South America (Schröder et al., 2016, 2018; Wang et al., 2016a). Moreover, data gaps over these regions in observations from ground-based GPS receivers and the radiosonde archive create *low confidence* in the TCWV estimations in these regions.

In summary, positive trends in global total column water vapour are *likely* during the satellite era, although changes in the observing system bring uncertainty and, therefore, *medium confidence* in estimation of the magnitude. *Low confidence* in longer-term trends arises from uncertainties in the SST-total column water vapour relationship and in the reliability of current surface-only reanalyses, particularly during the first half of the 20th Century.

2.3.1.2.4 Changes in global land precipitation

AR5 concluded that there was *low confidence* in precipitation change averaged over global land areas prior to 1950, being *medium confidence* afterwards with mixed and non-significant global trends. Nevertheless, observed trends assessed in AR5 were concluded to show a *likely* overall increase in precipitation in the well sampled northern hemisphere mid-latitudes, with *high confidence* after 1951.

In-situ precipitation rain gauge records over land extend back for centuries locally, and to the early to mid-20th Century quasi-globally. For this assessment, we use the latest versions of the Global Historical Climatology Network (GHCN) V2 (updated from Vose et al., 1992), the Global Precipitation Climatology Centre (GPCC) V8 (Schneider et al. 2018) and the Climate Research Unit

(CRU) TS 4.02 (Harris and Jones, 2019). These products differ in their input data, completeness of records, and their processing choices, most prominent of which is their approach to interpolation of fields which, given spatial clustering of instruments and the small spatial scales of precipitation, results in differences in global and regional estimates. The resulting temporal evolution of global annual land precipitation anomalies show increases, which are significant over 1901-2017 in both CRU TS 4.02 and GHCN V2, but not in GPCC V8 (Figure 2.15, Table 2.4). Differences between these datasets before the 1940s are most plausibly associated with limitations in data coverage and potentially homogeneity in the early records. On an interannual timescale precipitation anomalies globally are highly correlated with ENSO state (positive in La Nina, negative in El Nino). Consistency in decadal variations between the products is only observed since the 1940s, with primarily positive precipitation anomalies during the 1950s, 1970s and after the mid-1990s. Global trends for the more recent period (1980-2016) show a marked increase in annual precipitation (Table 2.4), generally consistent with tendencies reported by AR5 for the period 1980-2010, while contrasting with the negative trends observed in AR5 for the 1951-2008 period, showing the sensitivity of short-term trends in precipitation to the start and end-point.

[START FIGURE 2.15 HERE]

Figure 2.15: Annual precipitation anomalies averaged over global land areas from three global precipitation data sets (GPCC V8; CRU TS 4.02 and GHCN V2) relative to a 1981-2010 climatology. The resulting average of the annual precipitation anomalies from the three global precipitation data sets was smoothed with an 11-year Gaussian-weighted filter. The weights used, from the first through to the 11th years in the running window, are 1/24, 1/24, 1/12, 1/8, 1/8, 1/6, 1/8, 1/8, 1/12, 1/24, and 1/24 respectively.

[END FIGURE 2.15 HERE]

The spatial variability of observed long-term trends (1901-2016) based on GPCC V8 (Figure 2.16) shows that there have been increases in precipitation mainly over eastern North America, northern Eurasia southeastern South America, and much of Australia. Negative trends are strongest across tropical western equatorial Africa, western equatorial South America and moderate over the Arabian Peninsula. Recent trends (1980-2016) show increases over most areas except western North America, central South America, the Arabian Peninsula, and parts of Southern Asia. Trend reversals towards wet conditions are evident over much of Africa and Oceania, while a recent drying trend was observed over much of southern South America (Figure 2.16). Centennial trends and trends over 1960-2017 hint on the “wet wetter – dry drier” pattern in annual precipitation totals which is not the case for the recent trends.

[START FIGURE 2.16 HERE]

Figure 2.16: Spatial variability of observed GPCC precipitation change over different subperiods: a) 1901-2016, b) 1960-2016 and c) 1980-2016. Trends have been calculated using OLS regression and the confidence intervals in all plots have been corrected to account for serial correlation following the method of Santer et al. (2008).

[END FIGURE 2.16 HERE]

[START TABLE 2.4 HERE]

Table 2.4: Globally averaged trend estimates and 95% confidence intervals for annual precipitation for each time series in Figure 2.15 over three common periods of record all ending in 2017 [will be updated

to as near as possible to present day]. Trends have been calculated using OLS regression and the confidence intervals in all plots have been corrected to account for serial correlation following the method of Santer et al. (2008).

Data set	Trends in annual precipitation (mm) per decade		
	1901-2017	1960-2017	1980-2017
GPCC V8++	0.985±1.072	1.679±3.710	7.035±7.611
CRU TS 4.02	1.977***±0.940	1.552±3.211	6.922*±5.876
GHCN+	2.166**±1.437	1.686±4.318	10.149**±7.659

* Trend values significant at the 10% level.

** Trend values significant at the 5% level.

*** Trend values significant at the 1% level.

+ Values available until 2014 [to be updated].

++ Values available until 2016 [to be updated].

[END TABLE 2.4 HERE]

Several satellite-based precipitation datasets have been developed which improve the representation of the spatio-temporal variability of precipitation since the late twentieth Century. Some of these are based exclusively on satellite data (e.g. CMORPH, (Joyce et al., 2004), GSMaP, (Okamoto et al., 2005)), with others being a combination of in situ observations, reanalyses and satellite retrievals (e.g. PERSIANN-CDR, (Ashouri et al., 2015), GPCP, (Adler et al., 2018), CMAP, (Xie and Arkin, 1997), TRMM 3B42 V7 (Huffman et al., 2007), CHIRPS, (Funk et al., 2015)). Most of these products show increases in global precipitation during the last ~40 years, consistent with in-situ records, although with substantial interannual variability, likely related to ENSO, and large spread among different products (Li et al., 2016; Adler et al., 2017, 2018; Gehne et al., 2016; Li et al., 2015). [maybe new figure in SOD].

In summary, global land precipitation has *likely* increased since the beginning of the 20th century (*medium confidence*), with a faster increase in global precipitation since 1980s (*medium confidence*).

2.3.1.2.5 Changes in Precipitation minus Evaporation

AR5 concluded that the pattern of precipitation minus evaporation over the oceans had been enhanced since the 1950s (*medium confidence*). The inferred changes in P-E were consistent with the observed increased water vapour content of the warmer air.

Estimating global-scale trends in P-E using direct observations alone is challenging due to limited evaporation measurements and inhomogeneities in precipitation data derived from satellites and rain gauges (Hegerl et al., 2015). Hence, assessment of P-E trends is generally performed by recourse to reanalyses, although changes in the observing system brings uncertainties to the estimation of long-term trends (Skirris et al., 2014). Over land, trends in P-E from ERA-Interim show significant positive trends in Northwestern South America, Southern Africa and much of the Maritime Continent (Figure 2.17). Drying trends are observed over Tropical Africa, Southeastern South America and eastern Asia. Changes in continental runoff are a proxy for P-E and the majority of rivers show no significant trend over the last century (Section 2.3.1.2.6). During the second half of the 20th century, several reanalysis products and observational datasets have shown increases in P-E over global land, although 75% of land area have non-significant changes, which is primarily a result of data uncertainty although interannual and decadal variability may play a role (Greve et al., 2014; Robertson et al., 2016).

Over the oceans, negative global trends of P-E are observed since 1979 (Bosilovich et al., 2017). Regional negative trends are observed along the Tropical Pacific, large portions of the Indian Ocean and over the South Atlantic, while positive trends are located over the eastern equatorial Atlantic Ocean, the Northern Indian Ocean and the areas adjacent to the Maritime Continent (Figure 2.17). Strong negative P-E trends over most of the tropical ocean are partially offset by moderately positive trends in the subpolar NH oceans and in the Southern Ocean. An alternative method to estimate P-E changes is based on near-surface ocean salinity, which is partially controlled by the freshwater flux at the ocean surface (Yu, 2011). The surface salinity trends are found to be more spatially coherent than from P-E reanalyses, with an intensification of the water cycle over oceans specially in subtropical regions (Durack et al., 2012; Skliris et al., 2014). The precise rate of water cycle intensification implied by salinity trends is being actively debated in the literature (e.g., Skliris et al., 2016; Zika et al., 2018). Considering remotely sensed P-E, Robertson et al. (2014) found considerable differences in the variability of the estimations since 1979, with a reasonable coherence with ENSO phases.

[START FIGURE 2.17 HERE]

Figure 2.17: Linear trends in precipitation minus evaporation between 1980 and 2018 from the ERA-Interim reanalysis. Blue shading shows regions that have moistened [$\delta(P-E) > 0$] and red shading shows regions that have dried [$\delta(P-E) < 0$]. The stippling indicates regions where the trends are significant at the $p = 0.1$ level. Trends have been calculated using OLS regression and the confidence intervals in all plots have been corrected to account for serial correlation following the method of Santer et al. (2008).

[END FIGURE 2.17 HERE]

In conclusion, observed trends in globally averaged P-E are small. Trends based upon reanalysis products are not consistent with P-E changes implied from observations and proxy data, giving *low confidence* in the assessment of global P-E trends.

2.3.1.2.6 Changes in streamflow

AR5 concluded that there was *low confidence* for a positive trend in global river discharge during the 20th century. It noted that caution must be taken while interpreting results from most of the largest rivers with long term streamflow records, given that many of them have been impacted by non-climatic human influences such as dam construction or land use change.

River discharge is monitored widely around the world, although gaps remain at a subcontinental scale over Asia and Africa. Since AR5 substantial efforts have been made to generate new global streamflow datasets. Barbarossa et al. (2018) developed a 1 km gridded annual flow metrics (FLO1K) considering 6600 monitoring stations over the period 1960-2015. The Global Streamflow Indices and Metadata Archive (GSIM) collected daily streamflow observations from more than 30,000 stations worldwide, being most of the data available for the second half of the 20th century (Do et al., 2018; Gudmundsson et al., 2018). Using reanalysis data, Suzuki et al. (2018) developed a dataset of continental river discharge based on JRA-55, with 0.25° horizontal resolution from 1958 to 2016. These new datasets provide an updated basis for an assessment of global streamflow magnitude and variability.

Recent research examining changes in streamflow has primarily considered regional changes, with few studies analysing global streamflow variability. Available global studies for the period after 1950 generally concur that there have been more rivers experiencing decreases in runoff than increases in runoff (Do et al., 2017; Gudmundsson et al., 2019; Su et al., 2018). Most of the rivers have not experienced statistically significant changes according to these studies and when globally aggregated

there is no significant change (Dai and Zhao, 2017). Uncertainties arise predominantly from the global role of human-controlled flow regulation and irrigation, which impacts most of the large hydrological basins of the world (Jaramillo and Destouni, 2015). Regional streamflow trends are dominated by precipitation variations, with low (high) land precipitation and continental runoff during El Niño (La Niña) events (Dai, 2016; Dai and Zhao, 2017). Increases in streamflow were found mainly over mid- to high-latitude areas of the northern hemisphere, in southern South America and northern Oceania (Dai, 2016; Gudmundsson et al., 2019; Su et al., 2018). Decreases in streamflow over 1971-2010 were identified in low- and mid-latitude basins in western and eastern North America, northern South America, Africa, southern Asia, the Mediterranean basins, and southwestern Australia (Dai and Zhao, 2017; Gudmundsson et al., 2019). In many regions these decreases are contrasting the upward tendencies identified for 1950-1990 and 1960-2000 (Gudmundsson et al., 2019) hinting at important interdecadal variability signals in the streamflow. No significant trends are found for the reanalysis-based discharges, while significant increasing trends are observed for the satellite-based estimations during the 1993-2015 period (Chandanpurkar et al., 2017). Discharge variation patterns during the past two decades might be associated to observed ENSO changes from the Eastern Pacific El Niño to the Central Pacific El Niño, although these patterns tend to be geographically dependent (Liang et al., 2016).

In summary, it is *likely* that there has been no significant trend in globally averaged streamflow over the last century. The magnitude of the observed trends varies regionally, with slightly more globally gauged rivers experiencing significantly decreasing flows than significantly increasing flows (*medium confidence*).

2.3.1.2.7 Hydrological cycle summary

To summarize the findings of the hydrological cycle changes, positive trends in global land precipitation (Section 2.3.1.2.4) have been balanced by evaporation losses associated with the increase in global temperature, implying a P-E trend close to zero (Section 2.3.1.2.5). This is consistent with the observed global streamflow trends (a commonly used proxy for P-E over land) (Section 2.3.1.2.6) and the observed increases in specific humidity (Section 2.3.1.2.2) and total column water vapour (Section 2.3.1.2.3). The reader should refer to Chapter 8 for the regional scale changes in the components of the hydrological cycle and its theoretical rationale.

2.3.1.3 Tropospheric circulation

2.3.1.3.1 Hadley circulation extent and intensity

AR5 reported *low confidence* in trends in the strength of the Hadley circulation due to uncertainties in the reanalysis data sets, with large interannual-to-decadal variability. However, evidence indicated a *likely* widening of the tropical belt since the 1970s.

Only certain components of the Hadley Circulation (HC) are represented in current proxy records using available techniques. Evidence for the strength and position of trade winds based on proxy records from marine and lake deposits show that an intensification of the Northern Hemisphere HC and associated trade winds was observed during Heinrich stadials, in parallel with a weakening of the Southern Hemisphere HC (McGee et al., 2018a) and wetter conditions in the western US (McGee et al., 2018b). An intensification of the HC and climatic conditions similar to a La Niña-like state resulting in a more northerly position of both the Westerlies and the ITCZ were identified for the mid-Holocene using records of productivity and nitrogen isotopes covering the Holocene and the last deglaciation (Mollier-Vogel et al., 2019). Tree ring chronologies from mid-latitude regions in the Northern Hemisphere show that the northern edge of the HC contracted (expanded) during positive (negative) phases of the ENSO and PNA teleconnection patterns over the last 800 years (Alfaro-Sánchez et al., 2018). The HC expansion during the LIA was consistent with occurrence of drought conditions in several Northern Hemisphere regions (Alfaro-Sánchez et al., 2018).

Based on the reanalyses for the 20th century, Liu et al., (2012) and D’Agostino and Lionello (2017) found divergent results regarding the HC width, while an apparent intensification of the southern hemisphere HC was revealed. However, the substantial discrepancy in recent HC behaviour between these surface-only reanalysis products and ERA-Interim (D’Agostino and Lionello, 2017) yields significant questions regarding their ability to capture changes in HC behaviour. Taken together with the existence of apparent non-climatic artefacts related to assimilation of in situ surface observations by centennial reanalysis products (Nguyen et al., 2015) this implies *low confidence* in surface-only reanalysis centennial-scale changes in the extent and intensity of HC.

Using a set of several tropical width metrics, Staten et al. (2018) reviewed the rates of tropical expansion since the beginning of the satellite era, showing strong evidence for a tropical expansion of 0.5° lat per decade. Similar results were found by Davis and Birner (2017) based on modern reanalyses and commonly used tropical belt width metrics –including the latitudes of the HC edges and the latitudes of maximum downwelling in the HC-. The HC extent has shifted poleward at an approximate rate of 0.2° to 0.3° lat per decade in both Hemispheres during the last ~40 years (Grise et al., 2018; Staten et al., 2018; Studholme and Gulev, 2018) (Figure 2.18). Large discrepancies in the location of the poleward edges of the HC are found between several reanalyses (Feng et al., 2016), leading to uncertainties in the estimation of trends (Garfinkel et al., 2015b). Nevertheless, dynamical reanalysis products that assimilate radiosonde profiles and satellite data all agree on the sign and broadly the magnitude (Davis and Birner, 2017; Studholme and Gulev, 2018). Large-scale features of the HC based on reanalysis data show high resemblance with data from the Integrated Global Radiosonde Archive (IGRA) during 1979-2012 (Lucas and Nguyen, 2015; Mathew et al., 2016). The cells tend to shrink (expand) during the warm (cold) phase of ENSO and during the low (high) phase of the annular modes (Hu et al., 2018b; Nguyen et al., 2013). In terms of seasonality, recent trends based on reanalysis data indicate a larger widening in the HC for summer and autumn seasons for each hemisphere, although the magnitude is strongly dependent on dataset and metrics used (Grise et al., 2018; Hu et al., 2018b; Staten et al., 2018).

[START FIGURE 2.18 HERE]

Figure 2.18: Time series of the annual mean (a) NH and (b) SH Hadley Cell extent and (c) NH and (d) SH Hadley Cell intensity computed with ERA-Interim, JRA-55 and MERRA-2 reanalysis. The edge of the HC is taken as the average position of the weakened overturning peak value between 800 and 400 hPa. HC intensity is taken as the vertically averaged maximum value of the meridional stream function between 900 and 200 hPa in each overturning cell.

[END FIGURE 2.18 HERE]

Trends in the HC intensity since 1979 show inconsistencies among several reanalyses datasets, although there is a tendency toward HC intensification (Chen et al., 2014; D’Agostino and Lionello, 2017; Nguyen et al., 2013). Annual mean northern hemisphere HC strength shows persistent intensification, in contrast to only slight strengthening observed in the southern hemisphere HC (Studholme and Gulev, 2018).

In summary, there has been a *very likely* widening of the Hadley Circulation since the 1980s, although there is only *medium confidence* in the magnitude. This has been accompanied by a *likely* strengthening of the Hadley Circulation, particularly for the northern hemisphere cell (*medium confidence*).

2.3.1.3.2 Global monsoon changes

AR5 reported a weakening of the global monsoon (GM) circulations as well as a decrease of global land monsoon rainfall over the past 40 to 50 years. Nevertheless, trends in precipitation were

dependent on the definition of the monsoon area. Moreover, there was *low confidence* in the tendency of weakening of the East Asian monsoon since the early 1900s, due to a large uncertainty inherent in wind observations and circulation trends from reanalysis data. From a paleo perspective, AR5 indicated with *medium confidence* that more megadroughts occurred in monsoon Asia and wetter conditions prevailed in arid Central Asia and the South American monsoon region during the LIA compared to the MCA.

New research based on high-resolution proxy evidence shows that orbital cycles influenced the GM, with a ~20 ka precession signal that is generally out of phase between the hemispheres and an eccentricity signal modulated by a 100 ka forcing that shows no hemispheric contrast, with stronger monsoons during interglacial periods (Mohtadi et al., 2016; Wang et al., 2014a, 2017b; Zhisheng et al., 2015). During the Holocene, the northern hemisphere monsoons weakened and southern hemisphere monsoons strengthened between 8-3 ka BP, concurrently with a weakened AMOC and ITCZ shift (Band et al., 2018; Sachs et al., 2018; Wang et al., 2017c). Links with ENSO and PDO/AMO are posited to account for interannual and interdecadal GM variability, respectively (Wang et al., 2014a; Zhisheng et al., 2015).

Direct observational evidence shows that the GM circulation index has experienced large-amplitude interdecadal fluctuations since 1871, in phase with the occurrences of extreme ENSO events and AMO/V-related SST anomalies (Wang et al., 2013). Large decadal variations have occurred also for GM land precipitation based on several datasets, with a slight increase from 1900 through the early 1940s, followed by an overall decrease from the 1940s to the 1980s, then an increase to present (Kitoh et al., 2013). Several precipitation products from reanalyses and in-situ data indicate an upward trend in summer monsoon precipitation in the Northern Hemisphere (NHSM) since 1979. This conclusion is based on the analysis of GPCP precipitation for the extended boreal summer (JJAS) (Deng et al. 2018) and supported by analysis of CMAP precipitation for 1980-2010 by Jiang et al. (2016) and Han et al. (2018) over an extended boreal summer period (MJJAS). Upward trends in the Southern Hemisphere (SHSM) and GM for recent decades, superimposed with a strong interannual variability (Deng et al., 2018; Han et al., 2018; Jiang et al., 2016; Kamae et al., 2017; Kitoh et al., 2013; Lin et al., 2014) were less evident and mostly dominated by trends in the South African monsoon (Han et al., 2018). Uncertainty predominantly arises from the GM precipitation over the oceanic portion of the domain, leading to the large apparent spread between datasets (Kamae et al., 2017).

In summary, Global Monsoon precipitation has increased over the last 40 years, mainly due to a significant increasing trend in the Northern Hemisphere Summer Monsoon precipitation (*medium confidence*). However, GM precipitation has exhibited large multi-decadal variability over the last century, particularly for the Southern Hemisphere (*low confidence*).

2.3.1.3.3 Extratropical jets, storm tracks, and blocking

Both AR5 and SREX did not find statistically significant changes in the strength of the extratropical storm tracks since the late 19th century (*low confidence*). AR5 reported a poleward shift of storm tracks and jet streams since the 1970s based on consistent trends from different data sets, variables, and approaches. These trends were consistent with the widening of the tropical belt and the poleward shifting of the circulation features since the 1970s. There was *low confidence* in characterizing the global nature of any change in blocking.

Although not explicitly considered by AR5, changes in the position of the extratropical jets/westerlies during the Holocene can now be inferred from a variety of proxy evidence. Recent syntheses of moisture-sensitive proxy records indicate drier-than-present conditions over mid-latitudes of North America (Hermann et al., 2018) and elsewhere (Routson et al., 2019) during the middle Holocene, which together with a weakened Aleutian Low (Bailey et al., 2018), suggests that the winter jet over the North Pacific was shifted northward of its current position. A synthesis of evidence from the Southern Hemisphere indicates that the westerly winds were stronger 14-5ka and zonally symmetric,

1 followed by regional asymmetry after 5ka (Fletcher and Moreno, 2012). Whether the Southern
2 Hemisphere westerlies moved poleward (Anderson et al., 2018; Lamy et al., 2010; Turney et al.,
3 2017; Voigt et al., 2015) or equatorward (Kaplan et al., 2016) in the mid Holocene is not clear,
4 however.

5
6 During the MCA, a poleward shift of the extratropical circulation systems in both hemispheres is
7 indicated by various proxy records. Over North America, a poleward shift of Pacific storm tracks
8 (Cook et al. 2014; McCabe-Glynn et al., 2013; Steinman et al., 2014) was comparable in magnitude to
9 the interval 1979-2015 (Wang et al., 2017a). Storm tracks over the North Atlantic-Europe sector
10 moved northward as indicated by multi-proxy evidence for shifted moisture over the North Atlantic
11 (Giraudeau et al., 2010; Graham et al., 2011; Orme et al., 2017; Wirth et al., 2013), Mediterranean
12 (Roberts et al., 2012) and central Asia (Chen et al., 2015a). Reconstructed westerly winds in the
13 Southern Hemisphere suggest a poleward shift (Goodwin et al., 2014; Koffman et al., 2014; Lamy et
14 al., 2010; Schimpf et al., 2011), probably of a magnitude comparable with the shift during recent
15 decades (Swart and Fyfe, 2012).

16
17 Since AR5, observational evidence shows that the Northern Hemisphere extratropical zonal winds in
18 the middle troposphere have likely weakened, especially in the summer half year (Coumou et al.,
19 2014, 2015; Francis and Vavrus, 2012; Haimberger and Mayer, 2017). Although inferences are
20 sensitive to the choice of methodology (Barnes, 2013; Screen and Simmonds, 2013), multiple studies
21 indicate an enhanced meandering. The most robust changes are detected for boreal autumn, with a
22 pronounced increase in meandering at the northern hemisphere scale and over the Eurasian sector (Di
23 Capua and Coumou, 2016; Francis and Vavrus, 2015), while in summer a downward trend in
24 meandering was observed (Di Capua and Coumou, 2016). These are consistent with the weakening
25 trend in meridional geopotential height gradients since the early 2000s (Cohen et al., 2014).

26
27 Extratropical storm tracks can be identified by quantifying SLP or geopotential height variances for
28 synoptic scale range (different bands within typically 1 to 8 days range) or directly tracking individual
29 cyclones and counting their occurrences. Multiple reanalysis and radiosonde observations show a
30 10% or more increase in Pacific storm track activity, and <10% increase in Atlantic storm track
31 activity between 1959 and 2010 during boreal winter (Chang and Yau, 2016). In recent decades the
32 number of deep cyclones in Northern Hemisphere has decreased in winter (Neu et al., 2013; Wang et
33 al., 2017a) and summer (Chang et al., 2016). In the Atlantic a strong and consistent decrease across
34 modern reanalysis products after 1990 was noted (Tilina et al., 2013). In the Southern Hemisphere a
35 significant positive trend is found in the strong extratropical cyclones in NCEP/NCAR reanalysis
36 during 1980-2012 (Reboita et al., 2015). Given that changes in the cyclone occurrences are sensitive
37 to choice of reanalysis datasets (Hodges et al., 2011; Lucas et al., 2014; Pepler et al., 2018b; Rohrer
38 et al., 2018; Tilina et al., 2013; Wang et al., 2016d) and also climatologies of cyclone track densities
39 are sensitive to the cyclone identification/tracking methods ((Neu et al. 2013, Grieger et al., 2018),
40 there is overall *low confidence* for recent changes in global extratropical storm tracks.

41
42 Since AR5 a poleward shifting trend in the tropospheric extratropical jets has been further reported
43 based on multiple reanalysis datasets (Davis and Birner, 2013; Davis and Rosenlof, 2012; Manney
44 and Hegglin, 2018; Pena-Ortiz et al., 2013), radiosonde winds (Allen et al., 2012), and satellite
45 observations (Davis and Rosenlof, 2012; Fu and Lin, 2011). Although regional differences exist in
46 North Atlantic and Asia-North Pacific sectors (Ma and Zhang, 2018; Melamed-Turkish et al., 2018;
47 Norris et al., 2016; Wang et al., 2017a; Woollings et al., 2014; Xue and Zhang, 2018), consistency of
48 poleward deflection of storm tracks in winter in both North Atlantic and North Pacific during 1979-
49 2010 was reported (Tilina et al., 2013). Over the globe as a whole the jets' latitude is shifting
50 poleward (Lucas et al., 2014). Such shifting in the subtropical jet stream is consistent with the
51 poleward expansion of the tropical circulation (Section 2.3.1.3.1). The changes of extratropical jets
52 and westerlies are also related to the modes of variability such as the NAO, NAM and SAM (Section
53 2.4.5).

Robust trends in the blocking have only been found in isolated regions and seasons during recent decades. It is reported that blocking frequency increased in low-latitudes over the Atlantic (Davini et al., 2012) and in the Greenland region in boreal summer (Hanna et al., 2018). Over Siberia a general reduction has been observed (Davini et al., 2012), while over Ural regions the blocking events have tended to be longer-lived during 2000–2013 than during 1979–99 and more frequently at higher latitudes (Luo et al., 2016). During austral summer, trends in blocking frequency are positive in the South Atlantic with no significant trend in the South Pacific in ERA-Interim (Dennison et al., 2016). Inter-annual variations in the total number of blocking events over the Southern hemisphere (Oliveira et al., 2014) and Atlantic (Kim and Pai, 2015) regions have enhanced. Hemispheric or global features in the blocking show diverse trends among studies and they are also sensitive to datasets and methods (Barnes et al., 2014; Cheung et al., 2013; Pepler et al., 2018a; Rohrer et al., 2018), and thus have a *low confidence*.

In summary, it is *likely* the extratropical jet has been shifting poleward since the 1980s and that the mid-latitude jet meandering has intensified in the Northern Hemisphere during summer and autumn since the 1990s (medium confidence). The extratropical jets shifted poleward in the Northern Hemisphere during the mid-Holocene (*low confidence*) and the MCA (*medium confidence*) to latitudes that *likely* were similar to those of recent decades.

2.3.1.3.4 Surface wind and sea level pressure

AR5 concluded that surface wind speeds over land have generally weakened, and the changes over oceans showed large-scale trend patterns with opposing signs that largely cancelled globally. There was *low confidence* owing to uncertainties in data sets and measures used. Sea level pressure (SLP) was assessed to have *likely* decreased from 1979–2012 over the tropical Atlantic and increased over large regions of the Pacific and South Atlantic, but trends were sensitive to the time period analysed.

Since AR5, terrestrial in situ datasets have been updated and the quality-control procedures have been improved, with particular attention paid to homogeneity and better retaining true extreme values (Dunn et al., 2014). Global mean land wind speed using 2584 observational stations (excluding Australia) from HadISD2 for 1978–2016 shows a negative trend of $-0.070 \text{ m s}^{-1} \text{ decade}^{-1}$ (Azorin-Molina et al., 2017). Trends are broadly insensitive to station selection and time period. Although the meteorological stations are unevenly distributed worldwide and scarce in South America and Africa, the majority exhibit negative trends, and the decline for strong winds is larger than for weak winds. The largest negative trends are reported in Central Asia and North America with values of -0.132 and $-0.093 \text{ m s}^{-1} \text{ decade}^{-1}$ during 1979–2016 (Figure 2.19) (Azorin-Molina et al., 2017; McVicar et al., 2012; Vautard et al., 2012; Wu et al., 2018). Global and some large-scale regional mean surface wind speeds have strengthened after 2010, though the robustness of such recovery is hard to assess owing to the short period and high inter-annual/decadal variability (Azorin-Molina et al., 2017; Kim and Pai, 2015; Kousari et al., 2013; Wu et al., 2018).

[START FIGURE 2.19 HERE]

Figure 2.19: Trends in surface wind speed during 1988–2017. (a) Cross-Calibrated Multi-Platform wind product (Atlas et al., 2011); (b) wind speed from the Objectively Analyzed Air-Sea Heat Fluxes data set, release 3 (Yu et al., 2008); (c) ERA5, and (d) Station observation wind from the Integrated Surface Database (HadISD) v2.0.2.2017f (Smith et al., 2011). Trends for HadISD were computed only if at least 36 years had values and each year has at least 3 seasons of observations available. White areas indicate incomplete or missing data. The symbol of ‘+’ indicates grid boxes where trends are significant (i.e. a trend of zero lies outside the 90% confidence interval) using OLS with AR(1) correction per Santer et al. (2008).

[END FIGURE 2.19 HERE]

Over oceans, multiple datasets demonstrate disagreement over trends in surface wind speed with discernible discrepancies in trends and spatial features among datasets (Kent et al., 2013a). Global ocean surface winds assessed by observations from WASwind and NOCv2.0 climatology demonstrate upward trends of $\sim 0.11 \text{ ms}^{-1}$ per decade (NOC, 1979-2015) and somewhat smaller from WASwind for 1979-2011 (Azorin-Molina et al., 2017; Dunn et al., 2016). The two datasets are consistent with each other until 1998 and after that only NOC shows continuing increases. ERA-Interim and JRA55 show consistently increasing global marine winds with trends being $0.05\text{--}0.06 \text{ ms}^{-1}$ per decade over 1979-2015. MERRA is in agreement with these until 1998 and then shows no trend in the last two decades. The satellite-borne microwave radiometer observations of SSM/I and SSMIS during 1988-2016 show no evident trend in global ocean winds, while satellite altimeter measurements indicate a general global trend of increasing wind speeds for the period 1985-2008 (Young et al. 2011). Most products give positive trends over the Southern Ocean, western North Atlantic and the tropical western Pacific.

Surface wind changes are tightly linked to variations in the lower tropospheric pressure gradient. The modern era reanalyses show for 1979-2018 SLP increases over the south hemisphere subtropics with stronger upward changes in austral winter. Over the northern hemisphere, SLP increased over the mid-latitude Pacific in boreal winter and decreased over the eastern subtropical and mid-latitude North Atlantic in boreal summer (Figure 2.20). Discrepancies in the low-frequency variations exist in the surface-data only 20th Century reanalysis products, particularly during the first half of the 20th century (Befort et al., 2016). The modern era reanalyses demonstrate good agreement in SLP trends during 1979-2018. Updated reanalysis datasets support the AR5 conclusion that there is no clear evidence for trends in the strength and position of the permanent and semi-permanent pressure centres of action since the 1950s. Instead, they highlight multi-decadal variations.

[START FIGURE 2.20 HERE]

Figure 2.20: Changes in the ERA5 SLP since 1979. Colors denote linear trends estimated by the ordinary least squares fitting and only significant values (i.e. a trend of zero lies outside the 90% confidence interval) are plotted. Trend calculation is via OLS accounting for AR(1) effects as per Santer et al. (2008). Contour lines are locations of the decadal mean SLP for 1004hPa and 1020.5hPa.

[END FIGURE 2.20 HERE]

In summary, since the 1970s a worldwide weakening of near surface wind has *likely* occurred over land. Over oceans, near-surface winds *likely* strengthened over 1980-2000, but divergent estimates lead to *low confidence* in globally aggregated behaviour of ocean surface winds thereafter.

2.3.1.4 Stratospheric circulation

2.3.1.4.1 Stratospheric polar vortex and sudden warming

AR5 assessed changes in the polar vortices and reported a *likely* decrease in the lower-stratospheric geopotential heights over Antarctica in spring and summer at least since 1979.

Since AR5 multiple definitions for the polar vortex and sudden stratosphere warming (SSW) have been proposed and compared (Butler et al., 2015; Butler and Gerber, 2018; Palmeiro et al., 2015; Waugh et al., 2017), and techniques identifying daily vortex patterns and events have been developed (Kretschmer et al., 2018; Mitchell et al., 2013a). Errors in reanalysis stratospheric winds were assessed and discrepancies in stratospheric atmospheric circulation and temperatures between reanalyses, satellites, and balloon observations have been reported (Duruiseau et al., 2017; Mitchell et al., 2013a).

The MERRA, ERA-Interim, JRA55 reanalysis and radiosonde datasets show that the midwinter lower stratospheric geopotential height over the northern polar region north of 70°N has increased significantly since the early 1980s (Bohlinger et al., 2014; Garfinkel et al., 2017). This behaviour extends to the middle and upper stratosphere. In January–February zonal winds north of 60°N at 10hPa have been significantly weakening (Kretschmer et al., 2018). Different trends have been reported depending on season, altitude, and data period highlighting sensitivity to the choice of metrics and the importance of variability. For example, the northern polar vortex has weakened in early winter but strengthened during early spring (Bohlinger et al., 2014; Garfinkel et al., 2015a, 2017; Ivy et al., 2016; Kretschmer et al., 2018; Seviour, 2017). In the middle and upper stratosphere, a strengthening trend of the northern polar vortex during DJF is found since 1998 in contrast to the weakening trend beforehand (Hu et al., 2018a). The position of the polar vortex also has long-term variations, exhibiting a persistent shift toward the Siberian continent and away from North America in February over the period 1979–2015 (Zhang et al., 2016, 2018a). Multiple measures show similar location changes (Seviour, 2017). Regional climate anomalies/extremes related to the polar vortex variations can be found in (Cross-Chapter Box 10.1).

Clustering techniques applied to the daily circulation in the northern stratosphere poleward of 60°N from ERA-Interim and MERRA2 reanalysis datasets show that the frequency of strong vortex patterns has significantly dropped, and the frequency of weak vortex events has increased and become more persistent. These changes in frequency largely explain the observed significant weakening of the vortex during 1979–2015 (Kretschmer et al., 2018).

The SSW, a phenomenon of rapid air temperature warming (sometimes more than 50°C in 1–2 days), may cause breakdown of stratosphere polar vortices (Butler et al., 2015; Butler and Gerber, 2018). On average, it occurs approximately 6 times/decade in the northern hemisphere winter (Butler et al., 2015; Charlton et al., 2007). Although the SSW frequency is sensitive to the definition and threshold, the SSW record from all modern era reanalyses is very consistent. NCEP–NCAR reanalysis shows higher occurrence of the major midwinter SSWs in the 1980s and 2000s with no SSW event during 1990–1997 (Reichler et al., 2012; Butler et al., 2015). An assessment of multidecadal variability and change in SSW events is sensitive to the definition and the method used (Palmeiro et al., 2015). Due to the lack of assimilation of upper air data, the 20th century reanalyses do not capture the SSW events, even for the most recent decades (Butler et al., 2015, 2017).

Study of trends in the southern hemispheric stratosphere vortex and SSW is limited. In MERRA and HadAT2 datasets the mean air temperature between 60–90S at 50hPa has a cooling trend for all seasonal means from 1980–2000 (Ivy et al., 2016).

In summary, it is *likely* that the northern stratospheric vortex has weakened since the 1980s, and its location has shifted more frequently toward the Eurasian continent.

2.3.1.4.2 Brewer-Dobson circulation

AR5 assessed the changes in Brewer-Dobson circulation during recent decades and pointed out that there is *low confidence* in any trend due to considerable uncertainties in measurements.

Recent balloon-borne measurements of CO₂, CH₄ and partly CO at Northern Hemisphere midlatitudes show that the mean age of air (AoA, the time it takes for the atmosphere to transport air from the tropical tropopause to a given place in the stratosphere) has a small, statistically insignificant positive trend of 0.15±0.18 years/decade during the last ~4 decades (Engel et al. 2017). This trend is smaller than that implied from balloon-borne observations of CO₂ and SF₆ (0.24±0.22 years/decade for 1975–2006) (Engel et al., 2009). The shorter ENVISAT satellite observations of SF₆ during 2002–2012 show a positive AoA trend in the middle stratosphere of the Northern Hemisphere but negative trends in the Southern Hemisphere (Haenel et al., 2015; Stiller et al., 2012). Changes in mean AoA represents a combination of mixing and residual transport (Dietmüller et al., 2017; Garny et al., 2014;

Ploeger et al., 2015; Ray et al., 2014). Different AoA trends in low and high latitudes might be related to the decoupling of the deep and the shallow branches of the BDC (Birner and Bönisch, 2011; Lin and Fu, 2013; Ploeger et al., 2015) and the recent poleward shift of the circulation pattern (Stiller et al. 2017). After adjusting for this shift, the observations reveal a more hemispherical symmetric decrease of AoA in the lower to middle stratosphere up to -0.25 years over the 2002–2012 period.

Tropical lower stratosphere Ozone (O_3), a sensitive proxy of vertical transport, shows negative trend in the range of -3% to -6% per decade from ~1985 onwards (Bourassa et al., 2014; Randel and Thompson, 2011; Sioris et al., 2014). Wargan et al. (2018) suggests that the decline of ozone concentrations in the lowermost stratosphere during 1998-2016 is due to an intensification of transport of ozone-poor air from the tropics (20°S–20°N) into the extratropics, therefore, indicating an increase of tropical upwelling and poleward transport in the stratosphere. A combined analysis of satellite MSU/AMSU lower stratospheric temperatures, ERA-Interim reanalysis data, and observed estimates of changes in ozone, water vapor, well-mixed greenhouse gases, and stratospheric aerosols showed that the annual mean BDC has accelerated in the last 30 years (at the 90% confidence interval), with most of the change coming from the Southern Hemisphere. By establishing a relationship between tropical residual vertical velocity at 70 hPa and lower stratospheric temperatures, it was shown that the relative strengthening of the annual mean BDC is about 2.1% per decade for 1980–2009 (Fu et al., 2015).

In summary, the balance of evidence from observations suggests a strengthening of the stratospheric Brewer Dobson Circulation in last few decades of somewhat uncertain magnitude (*low confidence*).

2.3.1.4.3 Quasi-biennial Oscillation

Beyond the seasonal cycle the quasi-biennial oscillation (QBO) represents the most regular mode of internal stratospheric climate variability. Characterised by successive eastward and westward wind shear zones descending from the top of the tropical stratosphere, winds switch sign upon reaching the tropical tropopause (15km), before appearing again in the upper stratosphere (40km). The observed mean period of the QBO wind-switching is 28.2 ± 4.5 months with winds ranging in strength from $+10 \text{ m s}^{-1}$ to -30 m s^{-1} (Schenzinger et al., 2017). Recent observed changes of the QBO were not assessed in AR5.

The tropical radiosonde network has been a mainstay in observing the tropical stratosphere winds since the early 1950s. The distribution of this network constrains reanalyses most strongly in the low-mid stratosphere (70-50hPa) near where the network is most dense (South-East Asia, Central and South America). At higher levels (30-10hPa) the reduced number of balloon measurements results in a greater spread between reanalyses especially during those times when eastward winds are appearing (Kawatani et al., 2016). The spread between reanalyses has been reduced over recent decades by the steady increase in the number and quality of radiosonde measurements and, more latterly, the availability of Radio Occultation measurements.

The amplitude of the QBO has decreased over recent decades, most notably in the lower stratosphere (Kawatani and Hamilton, 2013). The weakening is consistent with a strengthening of the tropical lower stratosphere branch of the BDC due to rising tropospheric temperatures. The regular descent of the QBO winds was unexpectedly disrupted in February 2016 when a thin westward jet grew within a core of eastward wind at 25km (40hPa). This temporarily reversed the descent of the eastward winds above 25km (Newman et al., 2016b). This disruption had the proximal cause of dissipating Rossby waves originating from higher latitudes (Osprey et al., 2016). The remote drivers for the QBO disruption have been linked with the strong 2015/2016 El Nino and anomalous Arctic sea-ice loss (Hirota et al., 2018). The QBO is known to modify the BDC in the sub-tropics (Coy et al., 2016). Consequently the QBO disruption led to record low ozone column concentration in the sub-tropics (Tweedy et al., 2017a).

In summary, in recent decades the QBO amplitude has *likely* decreased in the lower stratosphere. There is *low confidence* owing to limited studies on its long-term changes.

2.3.2 Cryosphere

This section focuses on global and large-scale change in a subset of components of the cryosphere (Cross-Chapter Box 2.1). The reader should note that Chapter 9 undertakes a holistic assessment of changes in the cryosphere integrating observations, modelling and theoretical understanding. Readers interested in regional scale changes and a more in-depth treatment should thus refer to Chapter 9.

2.3.2.1 Sea-ice extent/area and thickness

2.3.2.1.1 Arctic Sea Ice

AR5 reported that the annual mean Arctic sea ice extent *very likely* decreased by 3.5 to 4.1% per decade between 1979 and 2012 with the summer sea ice minimum (perennial sea ice) *very likely* decreasing 9.4 to 13.6% per decade. The spatial extent had decreased in every season, and in every successive decade since 1979, with the largest decrease in summer (*high confidence*). AR5 also concluded with *high confidence* that the average winter sea ice thickness within the Arctic Basin *likely* decreased between 1.3 m and 2.3 m from 1980 to 2008. This was consistent with the decline in multi-year and perennial ice extent. Lower sea ice volume in 2010–2012 compared to 2003–2008 was also documented (*medium confidence*). There was *high confidence* that, where the sea ice thickness had decreased, the sea ice drift speed had increased. SROCC reported that Arctic summer sea ice decline continued over the decade 2009–2018 (*high confidence*).

Despite the paucity of proxy records, for the period preceding direct observational estimates, co-variability of the available regional proxy records allows for *high confidence* in multi-millennial-scale estimates of relative changes of pan-Arctic sea-ice coverage throughout the past 13,000 years. These indicate extensive sea-ice coverage during the Younger Dryas, comparably low sea-ice coverage during the Early Holocene, and increasing sea-ice coverage from the Mid-Holocene until the Little Ice Age (Armand et al., 2017; Belt, 2018; Belt et al., 2015; De Vernal et al., 2013) (Section 9.3.1.2).

In agreement with SROCC, new data and analyses show a continuing decline in Arctic sea ice area and extent in recent years, being superimposed with significant interannual variability. For reducing grid-geometry associated biases and uncertainties (Ivanova et al., 2016; Meier and Stewart, 2019; Notz, 2014) in the following, sea ice area is used in addition to and/or instead of sea ice extent (SIE). A record low Arctic sea ice extent and area since the start of the satellite era (1979) occurred in September 2012 (Figure 2.21). Since then, minimum extent has been variable, but overall for 1979–2017, the trend in September extent is estimated as $-83200 \pm 15400 \text{ km}^2/\text{year}$ ($-12.9 \pm 2.4 \text{ \%/decade}$) relative to 1981–2010, based on passive microwave satellite data from the National Snow and Ice Data Center (Fetterer et al., 2017). In March (annual maximum Arctic sea ice extent and area), the trend is $-41400 \pm 7400 \text{ km}^2 \text{ yr}^{-1}$ or $-2.7\% \pm 0.5\%$ per decade. These estimates are supported by other passive microwave products (e.g., OSISAF/CCI, (Lavergne et al., 2019)). A longer baseline, using sea ice charts for pre-satellite era (prior to 1979) information (Walsh et al., 2017), indicates no significant trend before the 1990s (Figure 2.21). Most of the decline in area has occurred since 2000. The longterm trend is superimposed by notable interannual variability. The sharp decline in Arctic summer sea ice area is reflected by earlier surface melt onset (Bliss et al., 2017), (Mortin et al., 2016), later freeze-up, and thus a longer ice retreat and open water season (Howell et al., 2009; Parkinson, 2014; Peng et al., 2018).

[START FIGURE 2.21 HERE]

Figure 2.21: Arctic sea ice area for March and September since 1953 (Walsh dataset) and 1979 (passive microwave satellite era). Sea-ice area values have been calculated from raw sea-ice concentration fields provided by OSISAF/CCI, NASA Team and Bootstrap from NOAA CDR 3.0, and Walsh NSIDC G10010 data sets. Linear trend fits for NASA Team (dashed lines) are added for visual aid.

[END FIGURE 2.21 HERE]

Over the past two decades, first year sea ice (FYI) has become more dominant (for sea ice terminology see WMO 1970). The oldest multiyear ice (MYI, >4 years old) once made up 20-30% of the ice cover, but has nearly disappeared, now making up less than 3%. The loss of older ice is indicative of ice cover thinning (Tschudi et al., 2016). Direct thickness observations are less complete, but negative trends are observed since the mid-1970s, with a particularly steep decline during the 2000s, and little change since 2010 (Kwok, 2018; Kwok and Cunningham, 2015; Lindsay and Schweiger, 2015) (Figure 2.22). Altimeter-derived ice thickness data are subject to significant uncertainties due to limited information about snow cover (e.g. (King et al., 2018; Ricker et al., 2015). Nevertheless negative trends in ice thickness over recent decades are also reported from the Atlantic sector (Hansen et al., 2013a; King et al., 2018; Renner et al., 2014; Rösel et al., 2018). It should be noted that data collected in Fram Strait are connected to a larger geographical area upstream of the transpolar drift. The thinner and on average younger ice has less resistance to dynamic forcing, resulting in a faster-moving ice cover (Kwok et al., 2013). This has a potential feedback effect by moving sea ice faster around and out of the Arctic (through Fram Strait) and via more openings in the ice cover that can speed up melting during the summer.

[START FIGURE 2.22 HERE]

Figure 2.22: Arctic sea ice thickness changes (means) for winter and fall from submarine transects (Rothrock et al., 2008), ICESat (Kwok et al., 2009) and CryoSat-2 retrievals (Kwok and Cunningham, 2015). Shadings (blue and red) show expected residuals in the regression analysis. Inset shows the data release area of submarine data. [Placeholder from (Kwok and Kacimi, 2018). It is intended to update the figure for the SOD and add uncertainty ranges for satellite-based ice thickness levels.]

[END FIGURE 2.22 HERE]

In the Atlantic sector, where fewer data are available, much higher snowfall has been observed than in the Western Arctic, resulting in snow-ice formation. Comparison of recent observations (2014, 2015, 2017) with data from the 1970s, 1990s and 2000s shows that snow on sea ice are available exclusively from sparse in-situ observations, including those from the Soviet drifting stations and airplane landings (>37 years) in the Pacific sector (Warren et al., 1999) and Canadian Ice Service records of snow in the Canadian Archipelago going back back to the 1950s (Howell et al., 2016). The Warren et al. (1999) climatology indicates weak trends toward declining snow depth, while Howell et al., (2016) report trends with varying sign and marginal statistical significance for the Oct-May period. Recent observations of snow depth on sea ice collected by passive microwave remote sensing sensors (Cavalieri et al., 2012); (Brucker and Markus, 2013), airborne radars flown as part of Operation IceBridge and other missions (Kurtz et al., 2013; Panzer et al., 2013), and space-based radar deployed on CryoSat-2 (Laxon et al., 2013) when combined with in-situ data indicate a substantial thinning of the spring snowpack in the western Arctic, particularly in the Beaufort and Chukchi seas (Webster et al., 2014, 2018). However, available observations from drifting buoys do not show significant trends in snowfall rate (Webster et al., 2014). Instead, there is a connection between snow depth and ice type, with up to double the snow depths on MYI compared to FYI (Blanchard-Wrigglesworth et al., 2015; Kurtz and Farrell, 2011). thicknesses around 0.4 m or more have also been observed earlier (Rösel et al., 2018). Snow cover in the Arctic has, in the past, been generally thought of as too low to

load the ice to the point where negative freeboard occurs and surface flooding can result in snow ice formation (Granskog et al., 2017; Haapala et al., 2013).

In summary, since the late 1970s, Arctic sea ice extent has decreased in both summer ($12.9 \pm 2.4\%$ per decade) and winter ($2.7 \pm 0.5\%$ per decade). Arctic sea-ice has become, younger, thinner, and faster moving (*very high confidence*). Snow thickness on sea ice has also decreased in the western and central Arctic (*medium confidence*). Proxy indicators show that Arctic sea ice has fluctuated on multiple time scales. Arctic Sea-ice area was *as likely as not* last as low as today in the mid-holocene.

2.3.2.1.2 Antarctic Sea Ice

AR5 reported that changes in many Antarctic sea ice characteristics were unknown. AR5 further reported a small but significant increase in the total annual mean sea ice extent that was *very likely* in the range of 1.2 to 1.8% per decade between 1979 and 2012 (0.13 to 0.20 million km² per decade) (*very high confidence*). There was also a greater increase in ice area associated with an increase in ice concentration. This overall signal was associated with strong regional differences in sea ice extent/area and also in the length of ice-free season since 1979 (*high confidence*). Data were too inadequate to assess changes in sea ice thickness and volume. SROCC reported that sea ice extent increased between 1979 and 2017 at an annual-mean rate of 0.2 ± 0.04 million km² per decade, but with strong negative departures in 2016 and 2017 (*very high confidence*).

The satellite passive microwave record of Antarctic SIE has been extended back into the 1960s using snapshots from early Nimbus satellite visible and infrared imagery (Gallaher et al., 2014; Meier et al., 2013) and proxy reconstructions based on different recordings and techniques, such as use of whaling records (Cotté and Guinet, 2007; De La Mare, 1997, 2009), old ship logbooks (Ackley et al., 2003; Edinburgh and Day, 2016), and ice core records (Abram et al., 2010; Curran et al., 2003; Sinclair et al., 2014), amongst other methods (King and Harangozo, 1998; Murphy et al., 1995, 2014), all of which vary in coverage and reliability (Hobbs et al., 2016b, 2016a; Holland, 2014). The early Nimbus satellite snapshots indicate higher overall SIE in the 1960s compared to 1979–2013, in line with previous studies (Hobbs et al., 2016b, 2016a). The proxy reconstructions to varying degrees indicate: i) a decrease in summer SIE across all Antarctic sectors since the early- to mid-Twentieth Century; ii) a decrease in winter SIE in the Amundsen, Bellingshausen and East Antarctic and Amundsen-Bellingshausen seas sectors starting in the 1960s; and iii) small changes in winter SIE in the Weddell Sea over the 20th Century (Hobbs et al., 2016a, 2016b). There is also ice-core evidence that the pronounced Ross Sea increase dates back to the mid-1960s (Sinclair et al., 2014; Thomas and Abram, 2016). While there is reasonable broad-scale agreement across these estimates, the uncertainties are large and reconstructions require further validation (Hobbs et al., 2016a, 2016b).

Data from the passive-microwave satellites show that from 1978 to 2015 there was a modest increase in overall (net) Antarctic SIE of $1.7\% \pm 0.2\%$ per decade and in sea-ice area of $2.5\% \pm 0.2\%$ per decade (Comiso et al., 2017) (Figure 2.23). Both pan-Antarctic and regional trends in SIE and sea ice area exhibit a marked seasonal dependence with net positive long-term trends most pronounced in austral autumn (the season of sea ice growth/advance), more moderate in summer and winter, and lowest in spring (Comiso et al., 2017; Hobbs et al., 2016a, 2016b; Holland, 2014; Turner et al., 2015). Behaviour since AR5 is characterised by record high and low values of sea ice extent/area in quick succession. Record maxima were reached in the three successive winters of 2012, 2013 and 2014, with the magnitude exceeding 20×10^6 km² for the first time since 1979 (20.156×10^6 km² on 21 September 2014, or ~7% larger than the 1981–2010 average) (Comiso et al., 2017; Massonnet et al., 2015; Reid and Massom, 2015; Reid et al., 2015). Substantively Lower values of SIE followed this string of record maxima with the second-lowest maximum extent value in the modern era occurring in October 2017 (Reid et al., 2018). The record low minimum set in March 2017, was about 2×10^6 km² or 27% below the long-term mean since 1978 (Stuecker et al., 2017; Turner et al., 2017). These recent minima serve to reduce trends estimated to 2018 and their significance.

[START FIGURE 2.23 HERE]

Figure 2.23: Antarctic sea ice area since 1979. Sea-ice area values have been calculated from raw sea-ice concentration fields provided by OSISAF/CCI, NASA Team and Bootstrap from NOAA CDR 3.0. Linear trend fit for the NASA Team (dashed lines) is added for visual aid.

[END FIGURE 2.23 HERE]

Information on Antarctic sea ice thickness is sparse. New airborne surveys and radar satellite altimetry data show thickness levels for the Weddell Sea of 2.40 ± 1.07 (2011) to 2.60 ± 1.15 m (2014), and snow depth from 30.0 ± 8.51 (2016) to 43.6 ± 10.2 cm (2014) (Kwok and Kacimi, 2018). Xie et al., (2013) found ice thickness levels for the Bellingshausen and Amundsen Seas from satellite laser altimetry with seasonal mean values of 1.0–1.6 m, and an overall positive (but statistically insignificant) trend in mean ice thickness of 0.03 m/year from 2003 to 2009. Observations of snow on sea ice in the Antarctic are extremely sparse.

In summary, Antarctic sea ice extent has experienced an increase of $1.7\% \pm 0.2\%$ per decade over 1978–2015, followed by significantly lower levels since 2016 (*high confidence*). There remains *low confidence* in all aspects of Antarctic sea-ice prior to the satellite era owing to a paucity of evidence that is highly regional in nature and often contradictory.

2.3.2.2 Terrestrial snow cover

AR5 concluded with *high confidence* that the Northern Hemisphere had experienced significant reductions in snow cover extent (SCE) over the previous 90 years with much of this reduction occurring in the 1980s. SROCC concluded with *high confidence* that declines in spring snow cover have continued over the last decade. AR5 identified no definite trends for the Southern Hemisphere due to limited records and large variability.

Analysis of the combined in situ observations (Brown, 2002) and data from NOAA indicates that since 1922, April SCE in the Northern Hemisphere has declined by 0.43 million km² per decade (Figure 2.24). Most of the decline in spring SCE has occurred over the last 40 years with a 9% reduction (significant at 95% confidence level) in SCE in March and April relative to pre-1970 values or about 0.8 million km² per decade (Rupp et al., 2013). Satellite derived SCE products show a decrease in SCE since 1978 in the Northern Hemisphere for all seasons (Derksen et al., 2018; Hernández-Henríquez et al., 2015; Hori et al., 2017; Kunkel et al., 2016; Mudryk et al., 2018). The greatest declines occur during spring and summer with decline in SCE during the spring melt period being three times the pace of the winter trend (Bokhorst et al., 2016; Estilow et al., 2015). Larger significant decrease in SCE is observed over western Eurasia compared to North America. In the Arctic (north of 60°N), there have been reductions in SCE since 2005, particularly in late spring (Derksen et al., 2018). Weaker trends exist in autumn and winter snow cover extent. Reductions in October SCE from 1982 to 2011, have been observed which is consistent with warming autumn surface temperatures observed during this period (Derksen et al., 2015; Estilow et al., 2015).

[START FIGURE 2.24 HERE]

Figure 2.24: April SCE anomaly (departure from 1995–2014 mean) and linear least squares trend over the entire period 1921–2018 for the Northern Hemisphere. Data are from NOAA (<https://www.ncdc.noaa.gov/snow-and-ice/extent/>) for 1967–2018 with the earlier part of record based on in situ data (Brown, 2002 <https://nsidc.org/data/g02131>).

[END FIGURE 2.24 HERE]

Overall this has led to a decrease in the snow cover duration and persistence (Hammond et al., 2018; Hori et al., 2017), particularly in higher latitudes (Chen et al., 2015b). Arctic snow-cover duration is decreasing rapidly (~3–5 days per decade), particularly due to earlier spring melt and, in some cases, later onset of snow cover (Derksen et al., 2015). Larger declines have occurred in the Eurasian Arctic region (12.6 days), compared to the North American Arctic region (6.2 days) over the period 1982 to 2011 (Barichivich et al., 2013).

For the northern hemisphere, maximum snow depth has generally decreased since the 1960s (Kunkel et al., 2016). Several studies report declines in snow water equivalent (SWE) over the last three decades from satellite based passive microwave data sets and ground based measurement (Jeong et al., 2017), with the decrease in SWE in spring being strongest and more significant in the higher latitudes (Brown et al., 2017; Derksen et al., 2018; Mudryk et al., 2018; Thompson et al., 2017a).

In summary, substantial reductions in snow cover extent, particularly in the spring, have occurred in the Northern Hemisphere since 1978 (*high confidence*). These reductions continue a long-term decline seen since the early 20th Century in many locations (*medium confidence*).

2.3.2.3 Glacier mass and extent

AR5 concluded with *high confidence* that, during the Holocene, mountain glaciers and ice caps (collectively, “glaciers”) were at times smaller than at the end of the 20th century, and with *very high confidence* that most glaciers are shrinking and that the total mass loss from all glaciers worldwide *very likely* increased from 1970 to 2009. SROCC reported that satellite and in-situ observations of changes in glacier area, length and mass show a globally coherent picture of continued mountain glacier recession in the last three decades (*very high confidence*) with only a few exceptions. The global trend is significant despite considerable inter-annual and regional variations. Departures from this trend occurred in some regions, but were generally locally restricted. In SROCC it is stated further that it is *very likely* that Arctic glaciers have lost significant mass between 1961 and 2016, despite considerable inter-annual variations with evidence that the rate and magnitude of loss is larger than any time during the past 4000 years.

Two recent reviews of literature on Holocene mountain glacier fluctuations systematically summarized much of the available information from numerous previous studies to generate a global perspective on changes over the Holocene (Solomina et al., 2015) and during the past 2000 years (Solomina et al., 2016). These reviews provide recent, extensive and systematic summaries of 189 and 275 (respectively) local glacier chronologies, based on in total 376 peer reviewed studies, and they reinforce earlier inferences based on geologic evidence for the former extent of glaciers. They show that glaciers retreated during the global deglaciation and generally retracted to their minimum extent sometime between 9000 and 6000 years ago (Figure 2.25 a). Glaciers worldwide expanded thereafter reaching their maximum post-glacial position during the LIA, following the lull in glacier advances during the MCA (Figure 2.25 a). Glaciers have essentially all retreated from their LIA maximum position, but the areal extent of the modern glaciers in most places is still larger than those of the early to middle Holocene minimum. In the northeastern Canadian Arctic, however, new evidence based on radiocarbon ages of *in situ* plants that were previously overridden by glaciers and are now emerging, along with an inventory of cosmogenic nuclides in rock surfaces recently exposed near retreating glacier margins, demonstrate that these glaciers are now smaller than they have been anytime during the present interglacial period (Gilbert et al., 2017; Miller et al., 2017; Pendleton et al., 2019).

[START FIGURE 2.25 HERE]

Figure 2.25: (a) Mountain glacier relative extent over the Holocene. Mean qualitative status of mountain glacier length summarized from a recent global review (Solomina et al., 2015). The relative length of glaciers studied around the world was averaged for four major regions in 100-year intervals and smoothed with a 300-year running mean, where 4 = Holocene maximum position, 3 = larger than modern (approximately the past 30 years) and advancing, 2 = glaciers present in the study area to some unknown extent, 1 = larger than modern and retreating, 0 = near the present extent. Intervals lacking information were not included in the mean values (set to null). Inset shows the total number of glacier advances reported in a recent global compilation of mountain glacier fluctuations during the past 2000 years, summed in 50-year intervals (Solomina et al., 2016). (b) Annual mass balance for a set of global reference glaciers with more than 30 years of continuous glaciological measurements. Annual mass change values are given on the vertical axis in meter water equivalent (m w.e.) which corresponds to tons per square meter (t m⁻²). Source: WGMS (2017, updated and earlier reports; <https://wgms.ch/>). Global values are calculated using only one single value (averaged) for each of 19 mountain regions in order to avoid a bias to well observed regions.

[END FIGURE 2.25 HERE]

New global compilations (e.g. Pfeffer et al., 2014; RGI Consortium, 2017) reinforce, refine and update the quantification of glacier mass (Farinotti et al., 2019) and areal changes based on direct observations. Globally, post-LIA glacier shrinkage started, with regional differences of several decades, around the 1850s to 1900s, and the rate of mass loss has accelerated since the 1980s (Zemp et al., 2015) (Figure 2.25 a and b). Observations beginning in the 1930s indicate that the greatest decadal mass loss occurred during 2000-2010, the most recent decade of records compiled by (Zemp et al., 2015), when the mass balance averaged -0.5 m w.e. per year based on *in-situ* measurements (glaciological method) and -0.8 m w.e. per year based on volume changes (geodetic method). Observed glacier mass balances for 2010-2017 are similar or even more negative (e.g. for Asian high mountain glaciers (Brun et al., 2017, 2018) [and WGMS, to be updated from SROCC]).

Insufficient evidence is available to determine the last time glaciers retreated at a rate comparable to that which has been observed recently (Solomina et al., 2015). During the last deglaciation, glaciers retracted to a small fraction of their former much larger ice-age extent, but glaciers have overridden the evidence of previous retreat, and the resolution of the glacial-geologic record is generally too low to make robust comparisons with the changes that have occurred since industrialization (Solomina et al., 2016). Inter- and intra-regional differences in glacier mass changes are well documented and can be attributed to multiple climatic and non-climatic factors (Section 9.5.2.2). Glacier mass balance and flow also undergo strong seasonal and short-term inter-annual variability (Figure 2.25b), thus implying seasonality in the overall negative trends (Zemp et al., 2015).

In summary, there is *very high confidence* that, with few exceptions, glaciers worldwide are currently shrinking. In most places, mountain glaciers are currently more extensive than during the early-middle Holocene (*high confidence*). Some Arctic mountain glaciers however are now smaller than they have been at any time during the present inter-glacial period (*medium confidence*).

2.3.2.4 Ice sheet mass and extent

2.3.2.4.1 Greenland ice sheet

AR5 reported that the Greenland ice sheet has lost ice during the prior two decades (*very high confidence*), that the ice loss had occurred in several sectors, and that high rates of mass loss had both expanded (*high confidence*) and *very likely* accelerated since 1992. Ice was lost approximately equally between surface melt and outlet glacier discharge (*medium confidence*). SROCC also concluded that since 2000 it is *virtually certain* that the ice sheet lost mass (*high confidence*).

Since AR5, the history of Greenland ice sheet fluctuations during the last interglacial and Holocene continues to emerge. For the last interglacial, recent reviews elucidate a complicated spatio-temporal pattern of ice recession, and suggest faster and earlier retreat in marine-dominated ice margin sectors (Sinclair et al., 2016; Vasskog et al., 2015). These studies agree that the total Greenland ice sheet extent / volume was lower in the last interglacial than present day. For the Holocene, recent studies of marine and lake sediments, glacier ice, and geomorphic features (Briner et al., 2016) show that the ice sheet retreated rapidly during the early Holocene but halted periodically and at different times around the ice sheet. For example, in West Greenland, the ice sheet retreated as much as ~200 km between about 11 and ~7 ka (Lesnek et al. (in prep); Young et al. (submitted)). It then retreated more slowly behind its present-day margin during the middle Holocene, but at different times and to different extents in different sectors, and reached its minimum Holocene extent between 5 and 2 ka, although observational constraints on the timing and configuration of minimum ice-sheet size remain sparse. The ice sheet re-advanced beginning around 4 ka, as indicated by glacially transported sediment in marine, lake and terrestrial settings, and reached its maximum extent in most places during the Little Ice Age, although the timing and extent of this fluctuation differed by sector (Briner et al., 2016).

Little is known about the rates of mass change over the Greenland ice sheet as a whole prior to direct observations. A new study integrating proxy evidence of climate forcing and ice-sheet size with modelling indicates that the last time the rate of mass change was similar to 20th Century rates occurred during the early Holocene (Briner et al., in prep), generally consistent with (Brook and Buizert, 2018).

Since AR5, a combination of remote sensing, in situ observations and modelling have provided new insights regarding surface processes and their contribution to recent mass changes (AMAP, 2017; van den Broeke et al., 2017). Recent studies based on analysis of aerial photos and geomorphic information (e.g. (Kjeldsen et al., 2015)) have estimated total ice loss since the LIA and show that the rate of loss has increased in the 21st century (Figure 2.26). Greenland ice sheet mass loss for 1900–1983 was found to be 75 ± 29 Gt per year (Kjeldsen et al., 2015). The rate of ice loss has since accelerated from an average of 71 ± 67 Gt per year during 1993–2004 (sea level equivalent, $+0.20 \pm 0.19$ mm per year), to 273 ± 11 Gt per year, ($+0.75 \pm 0.03$ mm per year) over 2005–2015 (WCRP; Cazenave et al., 2018). The total annual mass loss from Greenland reached a historical record value during summer 2012 (-627 ± 89 Gt), which was the greatest one-season loss since the end of the LIA (Box and Colgan, 2013; Kjeldsen et al., 2015; Tedesco et al., 2013). The short-term mass loss trend decreased slightly following 2012 (Forsberg et al., 2017).

[START FIGURE 2.26 HERE]

Figure 2.26: Greenland ice sheet total mass balance (gigatons per year, left axis) since the end of the Little Ice Age, combining a dataset spanning from 1840 to 2010 (based on a combination of in situ and simulated climate data reconstruction of surface mass balance and parameterized marine ice loss; (Box and Colgan, 2013), and 1993–2016 (mainly based on GRACE satellite data; (Cazenave et al., 2018). Shown are annual levels and Gaussian running average filter for the first dataset, and annual levels and error estimates for the second. The right axis shows the equivalent eustatic sea level change in mm.

[END FIGURE 2.26 HERE]

Ice discharge measurements derived from extensive ice-thickness surveys by NASA's Operation Ice Bridge, have revealed widespread acceleration of marine terminating ice-sheet outlet glaciers since 2000. Despite this acceleration, only 15 (of the 178 total) glaciers accounted for 77% of the total ice loss due to acceleration since 2000 (Enderlin et al., 2014). However, the total mass loss over the last two decades has been dominated by a decrease in surface mass balance especially after 2009

(Enderlin et al. 2014; (van den Broeke et al., 2017).

In summary, Greenland Ice sheet mass loss began at the end of the Little Ice Age and the rate of loss has increased in the 21st century (*high confidence*). The total mass loss from Greenland reached a historical record value during summer 2012 (-627 ± 89 Gt), which was likely the greatest one-season loss since at least 1840. The Greenland ice sheet was smaller than present during both the last interglaciation (*high confidence*) and the middle Holocene (*medium confidence*).

2.3.2.4.2 Antarctic ice sheet

AR5 reported that there was *high confidence* that the Antarctic ice sheet was losing mass. The average ice mass change to Antarctica had been -97 (-135 to -58) Gt per year (a sea level equivalent of 0.27 (0.16 - 0.37) mm per year) over the period 1993–2010, and -147 (-221 - -74) Gt per year (0.41 (0.20 - 0.61) mm per year) over the period 2005–2010. These assessments included the Antarctic peripheral glaciers. SROCC concluded the ice sheet *very likely* lost mass since 2000 (*medium confidence*).

Paleoclimate records show that Antarctic Ice Sheet changes can be abrupt and irreversible on centennial timescales (within 2000 years or less). The West and East Antarctica experienced rapid retreat about 14,600 years ago and again about 8,000 years ago, Pine Island Glacier experienced abrupt thinning with grounding line retreat, with rates comparable with contemporary thinning rates (Johnson et al., 2014; Wise et al., 2017). During the early to mid-Holocene, outlet glaciers near Ross and Weddell Sea embayments experienced thinning (Hein et al., 2016; Jones et al., 2015; Spector et al., 2017).

Since AR5, estimates of surface mass balance (SMB) in Antarctica have been improved. A total of 67 ice core records investigated do not show any clear general trend in accumulation over most of Antarctica during the last 800 years (Frezzotti et al., 2013). On a centennial timescale, the decadal growth rate in Antarctica is estimated to be 7 ± 1.3 Gt between 1800 and 2010 CE and 14 ± 1.8 Gt since 1900CE (Marquer et al., 2017; Thomas et al., 2017). In coastal and inland Dronning Maud Land (East Antarctica), no trend in accumulation between 1950 and 2010 was reported (Altnau et al., 2015).

The mass balance of East Antarctica is not significantly different from zero, with the exception of the Wilkes Land sector, where glacier changes have occurred (Mohajerani et al., 2018). In West Antarctica, a rapid decrease of mass was observed in the Amundsen Sea Embayment (ASE) sector and the Antarctic Peninsula. Updates to the analysis of (Rignot et al., 2011) indicate an accelerated mass balance decrease from -55.6 ± 5 Gt per year in 1999-2009 to -158.7 ± 8 Gt/yr in 2009-2017 (Figure 2.27).

In summary, no significant changes of the ice sheet mass balance away from zero in major parts of eastern Antarctica have been observed since the onset of modern measurements, while in western Antarctica a rapid decrease of mass has occurred in the Amundsen Sea Embayment sector and the Antarctic Peninsula (*medium confidence*).

[START FIGURE 2.27 HERE]

Figure 2.27: Cumulative Antarctic Ice Sheet mass change (gigatons) from satellite-based measurements (The IMBIE Team, 2018). The cumulative ice-sheet mass changes (lines) are determined from the integral of monthly measurement-class averages for each part of the Antarctic ice sheet (see colored labels). The estimated 1σ uncertainties for the respective cumulative changes are shaded.

[END FIGURE 2.27 HERE]

2.3.2.5 Permafrost

AR5 concluded that in most regions and at most sites permafrost temperatures since the 1980s had increased (*high confidence*). Negligible change was observed at a few sites, mainly where permafrost temperatures were close to 0°C, with slight cooling at a limited number of sites. AR5 also noted a significant degradation of permafrost during 1975–2005 in northern European Russia (*medium confidence*). SROCC also concluded permafrost temperatures have continued to increase to record high levels. AR5 concluded that since the 1990s, trends in active layer (the seasonally thawed layer above the permafrost) thickness (ALT) were positive for many high latitude sites (*medium confidence*), with significant regional variability.

Trends and timing of permafrost aggradation and thaw over the last 6000 years in peatlands of the Northern Hemisphere were recently summarized (Hiemstra, 2018; Treat and Jones, 2018). Three multi-Century periods of permafrost aggradation are inferred, beginning after 3000 BP and resulting in more extensive permafrost in peatlands of the present-day discontinuous permafrost zone. Thawing occurred after 250 years ago as post LIA warming began (Treat and Jones, 2018). Although permafrost persists in peatlands at the southern extent of the permafrost zone there has been thawing over the past several decades (James et al., 2013; Mitchell et al., 2013a).

Permafrost temperature records acquired from several boreholes indicate, similar to AR5, that permafrost has generally warmed over the last 3–4 decades across the northern polar regions (Figure 2.28), with moderately strong regional variations in the warming (Biskaborn et al., 2019; Romanovsky et al., 2017a; Romanovsky et al., 2017b, 2018). Recent (2016–17) permafrost temperatures in the upper 20–30 m (at depths where seasonal variation is minimal) at most sites were the highest observed (Romanovsky et al., 2018), with temperatures in colder permafrost of northern North America being more than 1°C higher than they were in 1978. In agreement with AR5 reported trends, increases in the temperatures of colder permafrost are greater (average 0.4–0.6° per decade) compared to warm permafrost (average 0.17°C per decade) of sub-Arctic regions (Figure 2.28). During the 21st century, permafrost temperature has increased between 0.4 and 0.9°C per decade in the Arctic (Romanovsky et al., 2018), almost twice that of the entire record since 1978 for high Arctic Canada.

[START FIGURE 2.28 HERE]

Figure 2.28: Average departures of permafrost temperature (measured near the depth where seasonal variation is negligible) from a baseline established during International Polar Year (IPY 2007–09). Regions are those described in (Romanovsky et al., 2017; 2018). Barents: continuous-discontinuous permafrost Nordic region and Russia/Siberia 1974–2017 (11 sites with 6 starting 1998 or later). Baffin Bay Davis Strait: Continuous permafrost high and eastern Canadian Arctic 1978–2017 (7 sites with 4 initiated 2008). Beaufort-Chukchi Sea: continuous permafrost northern Alaska, Northwest Territories and eastern Siberia 1978–2017 (9 sites 4 initiated 2003 or later). Interior Alaska & Central Mackenzie Valley: discontinuous permafrost 1983–2017 (10 sites, 1 initiated 2001).

[END FIGURE 2.28 HERE]

Warming of permafrost over the last 1–3 decades has been observed in several high elevated regions in the northern hemisphere (European Alps, the Tibetan Plateau and some other high elevation areas in Asia) (Noetzli et al., 2018). Increases in permafrost temperature of up to 0.3°C per decade have been documented at depths of about 20 m (Cao et al., 2018; Liu et al., 2017a; Noetzli et al., 2018). In the Antarctic records are more limited and too short (less than 10 years) to document long-term changes in permafrost temperature.

The active layer response tends to be influenced more by interannual variability in climate compared to the deeper permafrost temperature. For example, in northwestern North America during the extreme warm year of 1998, ALT was greater than in prior years. Although ALT decreased over the following few years, it has generally increased since the late 2000s (Duchesne et al., 2015; Romanovsky et al., 2017; 2018). In the European and Russian Arctic there has been a general increase in ALT during 21st century (Romanovsky et al., 2018; Streletskiy et al., 2015).

In high elevation areas in Europe and Asia, increases in ALT have occurred since the mid 1990s (Cao et al., 2018; Liu et al., 2018; Noetzli et al., 2018). Limited and shorter records for the Antarctic show marked interannual variability and no definite trend with ALT being relatively stable or decreasing at some sites since 2006 (Hrbáček et al., 2018).

Observations of landscape changes from airborne and satellite imagery analysis provide additional evidence of thawing permafrost (e.g. thermokarst, slope instability). AR5 noted observed geomorphic change associated with ground thaw in various regions and loss of thin permafrost. Since the middle of the 20th century, loss of permafrost mounds, collapse of palsas and lithalsas, increases in thermokarst pond size and changes in lake area have been observed in the Arctic (Beck et al., 2015; Borge et al., 2017; Bouchard et al., 2014; Jolivel and Allard, 2017; Lantz and Turner, 2015; Mamet et al., 2017; Nitze et al., 2017). Widespread ice wedge melting has been documented over recent decades at several Arctic sites (Liljedahl et al., 2016; Necsoiu et al., 2013; Séjourné et al., 2015). Recent increases in thaw slumping associated with ground ice melting have been documented in northern Canada and elsewhere in the Arctic (Kokelj et al., 2017; Rudy et al., 2017; Segal et al., 2016; Séjourné et al., 2015). In mountain areas, destabilization and acceleration of rock glacier complexes that may be associated with warming permafrost have also been observed (Eriksen et al., 2018; Marcer et al., 2018).

In summary, there is an observed warming of permafrost in all permafrost regions (*high confidence*) and evidence of permafrost thawing over the last 250 years in many places.

2.3.3 Oceans

2.3.3.1 Ocean Heat Content (OHC)

Evidence in AR5 showed that there is *high confidence* that the majority (more than 90%) of the extra thermal energy in Earth's climate system since 1970 is stored in the global ocean. The upper ocean (0–700 m, *virtually certain*) and intermediate ocean (700–2000 m, *likely*) have warmed since the 1970s. According to IPCC SROCC, ocean heat uptake has continued unabated since AR5.

Long-term reconstructions of mean ocean temperature (MOT) arise from noble gases in ice cores (Bereiter et al., 2018), estimates of intermediate water temperatures (IWT; typically spanning depth range 0–1000 m) from sediment cores (Rosenthal et al., 2017; Rosenthal et al., 2013), and model simulations driven from the changes in surface temperatures (e.g., Crowley et al., 2003; Gebbie and Huybers, 2019). The largest documented heat uptake in the paleo-record occurred during the last deglaciation (18,000–8,000 years BP), when MOT increased by $\sim 2.5 \pm 0.2^\circ\text{C}$, (equivalent to c.13,000 ZJ) at a rate of $\sim 0.03^\circ\text{C}$ (13 ZJ)/Century. Notably, OHC uptake rate reached modern levels during a short period of rapid warming at the end of the Younger Dryas (12,750–11,550 years B.P.), following a period of ocean cooling and heat loss during the Antarctic Cold Reversal (ACR=14,580–12,750 yr BP) (Bereiter et al., 2018).

Ice core measurements exhibit a small decrease in MOT throughout the Holocene ($0.4 \pm 0.4^\circ\text{C}$), whereas IWT records (0–1000 m) from the equatorial Pacific and northeast Atlantic Ocean suggest stronger cooling of $\sim 1.8 \pm 0.4^\circ\text{C}$ and $1.0 \pm 0.7^\circ\text{C}$, respectively, from the early/mid-Holocene to the pre-

industrial (Kalansky et al., 2015; Morley et al., 2014; Rosenthal et al., 2017; 2013). The relatively large cooling of Pacific IWT might reflect the stronger sensitivity of intermediate waters to changes in high-latitude winter surface temperature influenced by surface heat fluxes and wind stress, especially in the Southern Hemisphere, to small climate perturbations that can amplify mid-depth temperature variability (Roemmich et al., 2007; 2015; Rosenthal et al., 2017).

Sediment records of ocean heat content from the 0-700 meters range (OHC₇₀₀) in the western equatorial Pacific through the CE suggest that IWT were $0.8 \pm 0.1^\circ\text{C}$ warmer during the MCA than the LIA. Within the uncertainties of the method, this estimate is consistent with estimates derived from model simulations based on global surface temperature reconstructions (Crowley and Lowery, 2000; Gebbie and Huybers, 2019). Collectively, the records suggest a negative global OHC₇₀₀ anomaly of $\sim 400 \pm 70 \text{ ZJ}$ ($\sim 170 \pm 100 \text{ ZJ}$ in the Pacific) (Gebbie and Huybers, 2019; Rosenthal et al., 2013).

Direct observations show that ocean temperatures in the upper layers of the ocean (0-700m in depth) have increased since the 1970s (Figure 2.29a) (e.g. Abraham et al. (2013) and Wjiffels et al. (2016). Since the mid 2000s there is *high agreement* that 30-40% of the ocean warming is stored in the 700-2000m depth layer (*low evidence*) (Roemmich et al., 2015; von Schuckmann et al., 2018). The global 1991-2010 temperature trend below 2000 m is derived from repeated hydrographic sections is equivalent to $0.065 \pm 0.040 \text{ Wm}^{-2}$ applied over the Earth's surface area (Desbruyères et al., 2016). An acceleration in the global ocean warming signal has been proposed recently (Cheng et al., 2019). All ocean basins have experienced significant warming over recent decades (Figure 2.29b), with the greatest warming in the southern oceans (Roemmich et al., 2015; Trenberth et al., 2016) and the tropical and subtropical basins (Figure 2.29a). There is, however, *low confidence* of regional ocean warming signals in areas of known pronounced interannual to decadal climate variability modes such as ENSO in the Pacific Ocean (Figure 2.29). In the subpolar northern hemisphere, the subsurface ocean warming signal is superposed by strong year-to-year to decadal variations (e.g. Häkkinen et al. 2015, Josey et al., 2017, Dubois et al., 2018) leading to a cooling on an inter-annual to decadal scale in this area.

[START FIGURE 2.29 HERE]

Figure 2.29: Rate of change over the period 1993-2017 a) of global zonal mean ocean subsurface temperature anomalies as a function of latitude, b) of ocean heat content (0-700m) anomalies relative to the 1993-2014 reference period based on the ensemble mean of the Copernicus Marine Service multi-product approach (3 global ocean reanalyses: GLORYS, CGLOR, OSRS5; 2 in situ based reprocessed products: CORA, ARMOR-3D, von Schuckmann et al., 2018; <http://marine.copernicus.eu/science-learning/ocean-monitoring-indicators/catalogue/>). Black dots indicate areas where the noise (ensemble standard deviation of trends) exceeds the signal (ensemble mean of trend), indicating areas of most robust signatures from the multi-product approach.

[END FIGURE 2.29 HERE]

The global ocean continues to be the main sink of accumulated heat in the Earth system arising from positive radiative forcing at the top of the atmosphere (*very high confidence*) (e.g. Cheng et al., 2017; Trenberth et al., 2016; von Schuckmann et al., 2016) (Section 7.2, 7.3). The Earth energy imbalance can be quite well estimated to a very good approximation from changes in ocean heat content, complemented by satellite measurements of variations in the top-of-atmosphere radiative fluxes (Section 7.2.2.1). However, long-term estimates of ocean heat content change reaching back in time are challenged by gaps and instrumentation changes in the global ocean observing system, leading to considerable differences in the global estimates (Figure 2.30a, b) (Section 7.2.2.2). The most widely-used approach used to estimate ocean heat content changes relies on gridding the available historical subsurface temperature profiles and interpolating across data gaps using a statistical mapping method (Abraham et al, 2013). The choice of climatology, the type and resolution of the grid, vertical

interpolation, bias corrections, and the mapping method all contribute to the uncertainty estimates of OHC change (Boyer et al., 2016). Uncertainties are reduced during the era of Argo profiling floats from 2005/6 onwards (Figure 2.30c) (Riser et al., 2016; Wjiffels et al., 2016). Estimates of OHC from space can be also obtained from the sea level budget approach (Chambers et al., 2016), but still suffer from large uncertainties (Lowell et al., 2014). However, although differences at interannual to decadal scale ocean heat content estimates exist owing to observing system challenges, there is *high confidence* that all estimates show a multi-decadal increase in global ocean heat content in both, upper and deep ocean layers (Figure 2.30) (e.g. Abraham et al., 2013; Boyer et al., 2016).

[START FIGURE 2.30 HERE]

Figure 2.30: a) Ensemble (see legend of panel b)) mean of yearly averaged near-global (60°S-60°S latitude) ocean heat content integrated over different depth layers (green: 0-300m; blue: 0-700m; yellow: 700-2000m; red: 0-2000m) over the period 1993-2016, together with the ensemble spread (shaded areas). b) Rate of change for near global ocean heat content over the period 1993-2016 as derived from different data sources, which are listed in the figure panel, and integrated over different depth layers. c) Same as b), but over the time period 2005-2016, which allows the assessment of a larger variety of data product (see legend). Adopted from von Schuckmann (in prep). Data sources linked to legend abbreviations include: Ishii: Ishii et al., 2009; Levitus: Levitus et al., 2012; CMEMS: von Schuckmann et al., 2018; CHENG: Cheng et al., 2017a; NORCPM: Counillon et al., 2014; GOURETSKI: Gouretski and Koltermann (2007); LYMAN: Lyman and Johnson, 2014; ISAS: Gaillard et al., 2015; IPRC: Yoshinari et al., 2006; JAMSTEC: Hosoda et al., 2008; NOAA: Levitus et al., 2012; Scripps: Roemmich and Gilson, 2009).

[END FIGURE 2.30 HERE]

In summary, it is *virtually certain* that the global ocean warmed substantially between 1971 and present. There is *high confidence* that layer contributions are approximately 60% (0-700m), 30% (700-2000m) and 10% (2000-6000m). Current rates of heat content change are greater than at any point since the last deglaciation (*medium confidence*). Confidence in this assessment is strengthened by consistent closure of both global sea level and energy budgets (Section 7.2.2.2, Cross-Chapter Box 9.2).

2.3.3.2 Salinity

AR5 concluded that it was *very likely* that regional trends enhanced mean geographical salinity contrasts, which have led to reported surface and subsurface ocean salinity changes across the globe since the 1950s and 1960s. SROCC supports these conclusions, and notes that the regional high-latitude freshening (part of the pattern of observed salinity change), combined with observed warming has also led to a surface-intensified ocean stratification.

Recent analyses confirm the AR5 reported broad-scale patterns of long-term near-surface and subsurface salinity changes. Global-scale analyses show the “fresh get fresher and salty getting saltier” paradigm reproduced in observations and reanalysis (Durack, 2015; Grist et al., 2016a; Skliris et al., 2014; Vinogradova and Ponte, 2017a), as well as in model-based assessments (Section 3.5.2). Investigations using reanalysis products have linked the salinity changes to P-E changes over the ocean, with larger rates of change reported for the more modern period (1979 to present) when contrasted to changes extending back to the 1950s (Grist et al., 2016b; Skliris et al., 2014; Vinogradova and Ponte, 2017b) (Section 2.3.1.2.5). A particularly long Atlantic Ocean assessment reports coherent changes across the subpolar (freshening) and low-latitude (enhanced salinity) regions for 1896 to 2013, with their analysis carefully considering Atlantic basin variability modes over this period (AMV/O, NAO; (Friedman et al., 2017).

Multi-decadal observed changes to ocean salinity are compelling evidence for coherent water cycle change (Durack, 2015). The spatial pattern of near-surface salinity changes, ascertained across four independent assessments (Boyer et al., 2005; Durack and Wijffels, 2010a; Good et al. 2014; Hosoda et al., 2009), is similar to the climatological distribution of salinity itself, as shown for the two observational datasets engaging ARGO data for the last decade in Figure 2.31a-b. Across all analyses, a coherent pattern of freshening in regions of low salinity and increasing salinity in salty regions is expressed. Freshening is particularly evident in the Western Pacific Warm (and fresh) Pool and North Pacific, and enhanced salinities are seen in the Atlantic in the subtropical salinity maxima regions (Figure 2.31a-b). Changes in the ocean subsurface mirror those of the near-surface, with freshening along subduction pathways, particularly for Antarctic Intermediate Water (20-70°S regions within the 34.5 isohaline), and enhanced salinity in the subtropical gyre bowls straddling the equator (Figure 2.31c-d). Different datasets show broad-scale similarities in the subsurface salinity changes, which suggests decreasing salinity in ocean waters fresher than the global average, and increasing salinity in waters saltier than the global average. Some regional differences, particularly in the high-latitude regions are apparent and are due to limited data sources, temporal periods of analysis, and analysis methodologies.

Summarizing, it is *extremely likely* that near-surface and subsurface salinity changes have occurred across the global ocean since the mid-20th century. The salinity pattern implying that fresh regions becoming fresher, and salty regions saltier, is supported by all available observational studies that have considered salinity change since the modern instrumental record began in the 1950s.

[START FIGURE 2.31 HERE]

Figure 2.31: Long-term estimates of global near-surface salinity (SSS) change after (A) Durack and Wijffels (Durack and Wijffels, 2010b) (analysis period 1950–2008), (B) Good et al. (2014); analysis period 1950–2012). Black contours show the associated climatological mean SSS for the analysis period. (C) and (D) Three long-term estimates of global zonal mean subsurface salinity changes after (A) Durack and Wijffels (Durack and Wijffels, 2010b) and (B) (Marcer et al., 2018) (Good et al., 2013). Both estimates in (A)-(D) are scaled to represent equivalent magnitude changes over a 50-year period (Practical Salinity Scale 1978 [PSS-78], 50 yr⁻¹). Black contours in maps (A, B) and cross-sections (C, D) show the associated climatological mean subsurface salinity for the analysis period from each corresponding climatological assessment. In all panels blue denotes freshening regions and red denotes regions with enhanced salinities.

[END FIGURE 2.32 HERE]

2.3.3.3 Sea-level

Both AR5 and SROCC concluded that global mean sea level (GMSL) has risen in response to past and recent increases in global mean surface temperature, from decadal to over millennial timescales (with *very high confidence* in SROCC). AR5 concluded with *high confidence* that the rate of GMSL rise since 1890 was larger than the mean rate during the previous two millennia. After weighing additional tide gauge and altimeter analyses since AR5, including new methods and bias-corrected data (Section 9.6.2), SROCC concluded that the rate of GMSL rise in the early years of the 20th century and satellite era (1993-1999) was slower than assessed during AR5 implying that uncertainty was larger than previously thought but also revealing a larger acceleration in the rate of GMSL rise over both periods. In this light, SROCC estimated a *very likely* GMSL rate of 1.6 (1.1-2.1) mm per year for 1902-2015 (with *high confidence* in an acceleration range from -0.002 to 0.019 mm yr⁻²); 1.53 (0.96-2.1) mm per year for 1901-1990; 2.06 (1.76-2.36) mm per year for 1970-2015; 3.07 (2.70-3.44) mm per year for 1993-2015 (acceleration range from -0.059 to 0.090 mm yr⁻²); and 3.5 (3.3-3.7) mm per year for 2005-2015.

During geological periods with global mean surface temperature warmer than present (Section 2.3.1), paleo reconstructions indicate substantially higher sea levels than present (Section 9.6.2.1). New analyses since AR5 and SROCC, however, have not substantially changed key conclusions and confidence levels (Section 9.6.2.1).

For the past two millennia, a new analysis (Kemp et al. 2018) based on proxy data (Figure 2.32), combined with a GMSL tide gauge reconstruction ((Hay et al., 2015)), indicates that rising GMSL rates were not larger than 0.5 mm per year for any 100-year interval prior to the 19th century (at least since –500 BP) but consistently increased, from 0.2 mm per year in the 19th century to 1.3 mm per year in 20th century (Figure 2.33), corroborating AR5’s conclusions for a transition from lower to higher GMSL rise rates since 1890.

[START FIGURE 2.32 HERE]

Figure 2.32: Global mean sea level timeseries from 500 Before Present (BP) to 2018, based on proxy data integrated with Hay et al.’s (2015) tide gauge reconstruction (Kemp et al., 2018; Kopp et al., 2016)), tide gauge only reconstructions and satellite altimeter products (Nerem et al. (2018), AVISO, ESA (Legeais et al. 2018), NASA and NOAA). Vertical dotted lines indicate the first year of a century. The pre-industrial period is 1750-1849, the early-industrial period is 1850-1899, and the sea level altimeter era starts in 1993 (inset). Period definitions are found in Cross-Chapter Box 1.3 All data were plotted as annual means except for proxy (10-year mean). For clarity, errorbars are only shown for proxy data and all data have been vertically shifted. Proxy data is relative to zero in 1850. Tide gauge data is aligned to the 1910-1990 mean of the proxy data. Altimeter data is aligned to the 1993-2000 mean of the proxy data. Units: mm.

[END FIGURE 2.32 HERE]

Since SROCC, an additional 20th century GMSL tide gauge reconstruction by Dangendorf et al. (submitted, 2019) is available, as an improved extension of Hay et al. (2015) (Section 9.6.2). In this effort, more altimeter products were extended to 2018 and bias-corrected for the 1.5 (0.4-3.4) mm per year drift (Figure 2.32), found through careful cross-calibration with tide gauge data (Watson et al., 2015). The drift was linked to a standalone satellite mission from 1993 to 1999 (WCRP Global Sea Level Budget Group, 2018) and subsequently confirmed in budget closure analyses (Chen et al. 2017; Dieng et al. 2017; Legeais et al. 2018).

GMSL rates from Dangendorf et al. for 1901-1990, 1901-2015 and 1993-2015 are consistently lower than assessed in SROCC whereas the rate for 1971-2015 is higher than in SROCC, albeit not statistically different in any of the aforementioned periods (Figure 2.33). Dangendorf et al.’s linear rates starting in 1901 are consistently lower than those of Church and White (2011) but both rates closely track each other when the starting year is 1971 for an overall 37-year period (1971-2007). Linear rates from different satellite products for 1993-2018 and 2006-2018 agree with each other within expected uncertainties. Modulation of the altimeter rates (acceleration or deceleration) for successive longer trend periods, is partly tied to interannual to decadal variability (Section 9.6.2). The 26-year altimeter GMSL rates for 1993-2018 are largely outside [placeholder for recomputed uncertainty/significance] the 26-year rate envelope for tide gauge reconstructions, from 1900 to 1970 (Figure 2.33).

[START FIGURE 2.33 HERE]

Figure 2.33: Linear rates from the global mean sea level timeseries in Figure 2.32. (Top panel). 100-year (diamond) and 26-year (circles) rate intervals. Observed rate ranges for the 20th century (1901-

2000) and satellite altimeter era (1993-2018) are highlighted by gray shading (inset) while individual rates by symbols with black countours. (Bottom panels). Linear rates for different start years (symbols) and end years (x-axis). Left panel: linear rates starting in 1901/2, 1971 and 2001. Right: linear rates starting in 1993 (satellite altimeter era) and 2006 (Argo-GRACE era). Vertical bars are linear rates estimated in AR5 (red), SROCC (blue) and current AR6 (gray). Linear rates were calculated using ordinary least squares. Units: mm yr⁻¹. [uncertainties for individual RATES will be included]

[END FIGURE 2.33 HERE]

Considering new evidence since AR5 and SROCC, and consistent with these previous assessments, we conclude with *high confidence* that the rate of Global Mean Sea Level rise is accelerating since the 19th century and that the rate of GMSL rise for the 20th century is higher than at any point in at least the past two millennia. The GMSL linear rate of increase is 1.6 (1.0-2.1) mm per year for 1901-2015 (with a *likely* acceleration from -0.002 - 0.019 mm yr⁻²); 1.5 (0.9-2.1) mm per year for 1901-1990; 2.1 (1.8 -2.4) mm per year for 1971-2015; 3.2 (2.8-3.6) mm per year for 1993-2018; and 4.0 (3.7 -4.4) mm per year for 2006-2018.

2.3.3.4 Ocean circulation

2.3.3.4.1 Oceanic Overturning Circulation

AR5 concluded that there was no evidence for a decline in the AMOC based on direct measurements. The deep branch of the AMOC is fed by three main processes: overflow across the Greenland-Scotland Ridge, entrainment of these overflow waters, and deep convection in the Labrador and Irminger Seas. In AR5, no trends were noted in the magnitude of the overflows across the Greenland-Scotland Ridge, and no comment was made about the entrainment of these water masses. AR5 found that the dense overflow waters freshened from the mid-1960s to the mid-1990s before a reversal of this trend with increasing salinities until the mid-2000s. AR5 concluded that estimates of the formation rate of Labrador Sea Water (LSW) decreased from 7.6–8.9 Sv in 1997–1999 to roughly 0.5 Sv in 2003–2005, and since 1997, only less dense LSW was formed compared to the high NAO years before. Beyond the Atlantic, AR5 concluded that the global volume of the Antarctic Bottom Water, the key water mass of the deep Antarctic-driven overturning circulation, decreased during the last two decades, making it more likely than not that the export rate of AABW from the Southern Ocean declined during this period. However, there is little evidence of circulation changes in the deep Pacific, as it is filled with deep waters originating in the North Atlantic and Southern Ocean on timescales over one thousand years.

Changes in the deep ocean overturning circulation have occurred in the past on millennial timescales, with dramatic consequences for regional and global climate (McManus et al., 2004; W. D. P., 2015), although these changes are generally inferred to be associated with glacial intervals of anomalously high input rates of freshwater from glacial melt. The AMOC has varied substantially during the last glacial period, with a dramatically weakened and shallower AMOC around the time of the LGM. There is evidence for AMOC changes associated with abrupt changes in past climate, including Dansgaard-Oeschger and Heinrich events (Lynch-Stieglitz, 2017). In contrast, after the final demise of the Laurentide ice sheet circa 8ka, the mean strength of AMOC has been remarkably stable throughout the Holocene epoch (Hoffmann et al., 2018). Proxy evidence from the near surface flow of the Gulf Stream suggests that AMOC may have slowed on the order of 10% during the LIA (Lund et al., 2006). Deepwater production in the Labrador Sea appears to have reinitiated in the early Holocene (Hillaire-Marcel et al., 2001) and persisted continuously since then, with a possible decline since the LIA to be currently at its weakest point in the past 1600 years (Thornalley et al., 2018).

Proxy evidence of ocean circulation support a recently weakened AMOC although the timing and

duration of this weakening is disputed. Surface temperatures in the subpolar gyre of the North Atlantic have not increased in line with the rest of the globe since the mid-20th Century. This ‘warming hole’ (Drijfhout et al., 2012) taken as a proxy for AMOC strength (Caesar et al., 2018; Rahmstorf et al., 2015) suggests the AMOC has been in decline over that period. Difficulties exist with the use of sea surface temperatures (SSTs) as a proxy for AMOC. For example, record cold sea-surface temperatures (SSTs) in the summer of 2015 were due to a combination of ocean re-emergence and extreme air-sea heat exchange rather than changes in the AMOC (Josey et al., 2018). Meridional sea level gradients in the Western North Atlantic have been proposed as a proxy for AMOC strength and as support for a recent weakening of the AMOC (Ezer, 2013; McCarthy et al., 2015). However, sea level responds to a variety of factors (e.g. atmospheric pressure and local winds) and the signature of the AMOC is not always clear (Woodworth et al., 2014; Piecuch et al., 2016; Piecuch and Ponte, 2015).

Direct observations of the AMOC since 2004 based on basinwide, continuous moored observations at the RAPID array at 26.5°N in the North Atlantic have shown that the AMOC is in a reduced state, with multi-year averages from 2008-2011 and 2012-2016 being weaker than the period 2004-2007 by 2.5 Sv and 3.0 Sv respectively (Smeed et al., 2014, 2018). Indirect observations of the AMOC based on satellite altimetry and Argo float data at 41°N in the North Atlantic also find a similar decline of 3 Sv per decade from 2004 to 2016, close to the trend of 2 Sv per decade from 2004 to 2016 in the RAPID data, though neither trend is statistically significant (Baringer et al., 2018). An apparent contradiction based on moored observations at the MOVE array at 16°N in the North Atlantic, which showed a strengthening lower branch of the AMOC over the same time period, was highlighted as a methodological difference (Frajka-Williams and Co-Authors, 2014). Estimates of the AMOC in density co-ordinates at the OVIDE hydrographic survey line between Portugal and Greenland noted a 2.5 ± 1.4 Sv decline in the period 1993-2010 superimposed on large decadal variability (Mercier et al., 2013). Additional moored observations at latitudes in the South Atlantic have been installed at 11°S (Hummels et al., 2015) and 34°S (Meinen et al., 2018) but high variability and short timeseries length hamper conclusions about trends.

Source waters for the upper cell of the AMOC have two origins: deep convection in the Labrador and Irminger Seas; and overflow, and the subsequent entrainment, of deep water across the Greenland-Scotland ridge. Overflow across the Greenland-Scotland Ridge has remained stable with no discernible trend since the mid-1990s (Hansen et al., 2016; Jochumsen et al., 2017). Deep convection has recently returned to the Labrador (Yashayaev and Loder, 2016) and Irminger Seas (de Jong et al., 2018; Jong and Steur, 2016) following an extended period with weak convection over the early 2000s. An associated strengthening of the AMOC has not been observed. The lack of relationship between deep convection in the Labrador Sea and the strength of the AMOC was reported in the first results from the OSNAP programme (Lozier et al., 2019), which built upon similar findings from the 53°N array (Zanntopp et al., 2017).

In summary, there is robust agreement from a limited number of direct observations for a rapid weakening of the AMOC beginning around 2005-2008 (high confidence) but records are of insufficient length to determine whether this is a significant trend or a manifestation of multi-decadal variability. There is agreement from proxy evidence for a longer-term decline in the AMOC, however, uncertainties in the robustness of the proxies and inconsistencies in the timing and duration of the associated AMOC decline means that confidence in a weakened AMOC on the basis of proxy data is low. There is no evidence for a long term weakening in the AMOC on the basis of observed source water variability.

2.3.3.4.2 Wind-driven ocean circulation

A strengthening and poleward shift of westerlies was reported with *medium confidence* in AR5. They further reported with *medium to high confidence* intensification of the North Pacific subpolar gyre, the South Pacific subtropical gyre, and the subtropical cells, along with the expansion of the North Pacific

subtropical gyre and a southward shift of the Antarctic Circumpolar Current (ACC) during the period from 1993. It was pointed out that these wind-driven changes are predominantly due to interannual-to-decadal variability.

Recent analyses demonstrate strengthening of the westerlies in the southern hemisphere, in association with a positive phase of the Southern Annular Mode (SAM) (Section 2.4.5.2). Observations in the Agulhas current (Beal and Elipot, 2016; Elipot and Beal, 2018) have noted a broadening of the current associated with the transfer of wind energy to eddies rather than mean flow. In the south Pacific, there has been little change noted in boundary currents near New Zealand (Fernandez et al., 2018) and the East Australian Current has not strengthened (Sloyan and O’Kane, 2015) nor penetrated farther south (Cetina-Heredia et al., 2014) in line with changing wind patterns. An increased strength of the Antarctic Circumpolar Current based on revised estimates of bottom velocity in Drake Passage lead to a revised estimate of 173 Sv, almost 30% higher than the canonical value of 134 Sv (Donohue et al., 2016). This is not associated with a strengthening of SH winds but with improved observational estimates of the barotropic flow.

In the northern hemisphere, neither the Kuroshio nor the Gulf Stream have notably strengthened. Based on satellite altimetry, the Kuroshio has weakened east of Taiwan (Wang et al., 2016e; Wang and Wu, 2018) while strengthening upstream of this. Differing trends of strengthening and weakening at different points along the Kuroshio are consistent with multi-decadal trends reported by (Hsin, 2015). The Gulf Stream has recently reversed a long term poleward migration (Bisagni et al., 2017) and the Gulf Stream Extension has become more eddying (Andres, 2016) and broader (McCarthy et al., 2018).

In summary, the response of wind-driven circulation to changing wind patterns is inconsistent and thus there is *low confidence*. There is little evidence for a predicted strengthening of western boundary currents in response to strengthening winds. There is some evidence that increased wind forcing is leading to increased eddying of WBCs.

2.3.4 Biosphere

2.3.4.1 Changes in the seasonal cycle of CO₂

SRCCCL (2019) noted an increase in the amplitude of the seasonal cycle of atmospheric CO₂ concentrations in the Northern Hemisphere (*robust evidence, high agreement*). The amplitude of the seasonal cycle is driven by carbon uptake and release in the biosphere; consequently, changes in the amplitude of the seasonal cycle imply changes in terrestrial ecosystem exchanges.

In situ observations document a rising amplitude of the seasonal cycle over the past half-century. Increases have been measured at both the Mauna Loa (3.2 +/- 0.07% dec⁻¹) and Barrow (6.0 +/- 0.05% dec⁻¹) surface observatories since the early 1960s (Figure 2.34; (Graven et al., 2013; Keeling et al., 1996). Likewise, airborne data poleward of 45°N exhibit an increase of 56 +/- 9.8% between field campaigns in 1958-1961 and 2009-2011 (Graven et al., 2013). Increases for the period 1980-2012 have also been documented at eight surface observatories north of 50°N (Piao et al., 2017), related primarily to a larger drawdown in June and July. In general, increases are larger in Arctic and boreal regions, indicative of changes in vegetation and carbon cycle dynamics in northern ecosystems (Forkel et al., 2016). As noted in SRCCCL (IPCC, 2019), the increase in the amplitude of the seasonal cycle is consistent with greater greening during the growing season (Section 2.3.4.5) and an increase in the length of the growing season (Section 2.3.4.6), both of which enhance gross primary production. Measurements of δ¹³C-CO₂ indicate that inter-annual variability in CO₂ growth rates (Figure 2.4.d) are related to relatively small variations in large two-way fluxes between the atmosphere and terrestrial biosphere (photosynthesis and respiration).

In summary, there is *high confidence* that the amplitude of the seasonal cycle of atmospheric CO₂ has increased by as much as 50% poleward of 45°N since the early 1960s.

[START FIGURE 2.34 HERE]

Figure 2.34: Changes in the amplitude of the seasonal cycle of atmospheric CO₂. Squares depict amplitudes at Barrow, and triangles depict amplitudes at Mauna Loa. The black line illustrates airborne data from a 1958-63 field campaign, and the magenta line illustrates airborne data from a 2009-11 field campaign.

[END FIGURE 2.34 HERE]

2.3.4.2 Ocean pH and deoxygenation

AR5 concluded with *high confidence* that the pH of the surface ocean waters decreased by 0.1 (or a 28% increase in acidity) since the preindustrial era, primarily as a result of ocean uptake of anthropogenic CO₂. It was *virtually certain* that the ocean is storing anthropogenic CO₂, about 118 ± 24 PgC from 1850 to 1994, and *very likely* that it had taken up approximately 34 PgC more anthropogenic CO₂ from 1994 to 2007. There was *medium confidence* that oxygen levels decreased in many ocean regions since 1960 and it was *likely* that the tropical oxygen minimum zones (OMZs) had expanded in recent decades.

There has been considerable work addressing past changes in surface ocean pH since AR5, largely associated with the emergence of the boron isotope proxy (Foster and Rae, 2016b). In particular, recent studies allowed comparing the rate of pH decrease associated with pronounced carbon cycle perturbations in the past with historic reconstructions and future predictions (Gutjahr et al., 2017b; Sosdian et al., 2018; Zeebe et al., 2016b). Similarly, the rejuvenation and further development of the marine sediment δ¹³C gradient method allowed quantifying past changes in oxygen levels in intermediate and deep waters (Gottschalk et al., 2016; Hoogakker et al., 2015, 2018a).

Throughout the last 65 million years there have been a number of time intervals when surface waters have transiently acidified (Figure 2.35). However, few intervals have sufficient chronological control to allow direct comparison with historic trends. The most notable exception relates to the PETM which involved a large pulse of volcanic CO₂ released into the ocean-atmosphere system within 5-20 kyrs (Kirtland Turner, 2018) and was associated with profound perturbations of the global carbon cycle. Estimates of the total amount of carbon released during the PETM vary from around 3,000 Pg C to more than 10,000 Pg C (Cui et al., 2011; Gutjahr et al., 2017c; Zeebe et al., 2009), although CO₂ release rates during the PETM were about an order of magnitude slower than today (i.e. 0.5-1.1 PgC yr⁻¹; (Panchuk et al., 2008; Zeebe et al., 2016c).

As a result, the PETM was associated with a negative surface ocean pH excursion *likely* ranging from 0.15 to 0.30 units (Babila et al., 2018; Gutjahr et al., 2017c; Penman et al., 2014). The degree of acidification *likely* occurred at one order of magnitude slower than the current rate of ocean acidification (Zeebe et al., 2016c).

Recent Earth System Model simulations as well as sedimentary data reveal widespread ocean deoxygenation during the PETM (Dickson et al., 2012; Winguth et al., 2012), with parts of the ocean potentially becoming sulfidic (Yao et al., 2018). Deoxygenation affected the surface ocean globally, due to vertical and lateral expansion of oxygen minimum zones (OMZs)(Zhou et al., 2014b) that resulted from warming and related changes in ocean stratification.

Surface ocean pH has gradually declined over the last 50 million years (Anagnostou et al., 2016), reaching similar to modern values in the last million years (Chalk et al., 2017; Martínez-Botí et al.,

2015; Seki et al., 2010) (Figure 2.35). There is *medium confidence* that surface pH values as low as today have not been experienced since the Mid-Pliocene Warm Period (MPWP), three million years ago (Figure 2.35). The Cenozoic, long-term trend in oxygenation is poorly documented, yet emerging studies suggest stable ocean redox conditions throughout the interval (Wang et al., 2019).

The Pleistocene glacial-interglacial cycles were associated with changes in surface ocean pH of the same order as historic pH change (0.12-0.16 pH units; (Chalk et al., 2017) but average rates of change were *likely* two orders of magnitude slower than modern. The last deglaciation is perhaps the best documented climatic transition in the recent past associated with a substantial atmospheric pCO₂ rise ranging from 190 to 265 ppm (equivalent to 160 Pg C) between 17.8 to 11.6 kyr ago (Marcott et al., 2014). Maximum rates of pH change during the last deglaciation reached -0.03 pH units per century at ~16.5 kyr (Hönisch et al., 2009; Martínez-Botí et al., 2015) as previously sequestered CO₂ was transferred from the ocean interior to the shallow subsurface ocean (Jaccard et al., 2016; Rae et al., 2018) (Figure 2.35). Concomitantly, marine sediments recorded an overall expansion of low-oxygen waters in the upper ocean as a result of warming as the Earth emerged from the last ice age and a reorganization of the global overturning circulation (Galbraith and Jaccard, 2015), with large swaths of the intermediate Indo-Pacific *very likely* becoming oxygen-depleted (Hoogakker et al., 2018b; Jaccard and Galbraith, 2012).

Present-day rates of ocean acidification, as detected in observations from repeated hydrographic sections and at ocean timeseries sites (Feely et al., 2012; Bates et al., 2014; Lauvset et al., 2015) at the surface and upper ocean, are now much higher than when averaged over the past 300 million years (Hönisch et al., 2012). All ocean basins are acidifying (Lauvset et al., 2015), but at different rates, depending on regional interactions of background ocean chemistry with circulation, temperature, carbon cycling and resident biota (Hurd et al., 2018). Some of the largest changes have occurred in the Arctic, where altered circulation patterns, and freshwater input from rivers and sea-ice melt have led to an acceleration in the rates of ocean acidification (Qi et al., 2017).

New analyses also substantiate an ongoing global-scale ocean deoxygenation over the past 50 years (Schmidtko et al., 2017) and Levin (2018). This decrease in oxygen – from 2% to 4% of the total seawater column inventory since 1960 (77 billion metric tons or $4.8 \pm 2.1 \times 10^{15}$ mol out of $227.4 \pm 1.1 \times 10^{15}$ mol, (Schmidtko et al., 2017), is strongly tied to the multidecadal increase in global ocean heat content, particularly after mid-1980s (Ito et al., 2017). Declines in solubility induced by ocean warming explains almost 15% of the deoxygenation trend (Schmidtko et al., 2017). The remaining 85% may be explained by reduced ventilation due to changes in ocean circulation and mixing, caused by stronger stratification (Breitburg et al., 2018), however, other factors may play a role (Levin, 2018).

Multidecadal rates in deoxygenation are highly variable across the water column and throughout the ocean basins SROCC Chapter 5 (IPCC, in prep), with the largest absolute losses in the North and Equatorial Pacific, South Atlantic and the Southern Ocean (Schmidtko et al., 2017). Oxygen losses in these regions along with the Arctic Ocean and the tropical oceans within OMZs cannot be explained by changes in seawater solubility. OMZs have expanded by about 4.5 million km² (Stramma et al., 2010) and the volume of water completely devoid of oxygen (anoxic seawater) has increased more than 400% (Schmidtko et al., 2017).

To conclude, all ocean basins are storing anthropogenic CO₂ and declining pH since the preindustrial era. Rapid deoxygenation is occurring across the Oceans with commensurate increases in oxygen minimum zones. It is *virtually certain* that the current rate of change in both Ocean pH and Ocean oxygenation is unprecedented in the available geological record.

[START FIGURE 2.35 HERE]

Figure 2.35: a) Low-latitude surface ocean pH over the last 65 million years, reconstructed using boron isotopes in foraminifera (red). Uncertainty is shown at 95% confidence as error bars or a shaded band. Data include: (Anagnostou et al., 2016; Gutjahr et al., 2017c; Sosdian et al., 2018). Also shown is the benthic $\delta^{18}\text{O}$ stack of Cramer et al. (2009; black – right axis). b) Low-latitude surface ocean pH over the last 3.5 million years reconstructed using boron isotopes in foraminifera (Sosdian et al., 2018). Uncertainty at 95% confidence is shown as blue shaded bands. The dotted horizontal line shows an acidification of 0.1 pH units relative to the Holocene value. The Pliocene (prior to 2.5 Ma) is the last interval when pH was as low as today for a sustained interval (hence medium confidence that modern pH is unusual in the last 3 million years).

[END FIGURE 2.35 HERE]

2.3.4.3 Ocean Phenology

AR5 WGII reported that 90% of observations of marine phenology were consistent with climate change. Spring advanced by 4.4 ± 0.7 days decade⁻¹ and summer events by 4.4 ± 1.1 days decade⁻¹ over the previous 50-year period. AR5 concluded that there was *robust evidence*, *high agreement*, and *high confidence* in ocean phenology changes.

Since AR5, data have continued to show that the timing of biological events in the ocean is advancing (Figure 2.36a, n=391). Overall 74.1% of all lifecycle events have moved earlier in time, significantly different from the null hypothesis of no impact of climate change of 50%, the phenology of marine life is thus *very likely* to have advanced earlier. There is also *high agreement* as all marine groups have moved earlier (Figure 2.36a), giving *high confidence* overall that there is a coherent response. There is *high confidence* that fish are occurring earlier (78.7%, n=164), including earlier peak abundance of fish larvae in upwelling systems (Asch, 2015), earlier appearance of fish in estuaries (Chevillot et al., 2017), earlier spawning migrations for anadromous fish such as salmon (Bal et al., 2017; Rubenstein et al., 2019), and earlier spawning of key commercial demersal (bottom-dwelling) species such as cod (McQueen and Marshall, 2017), and sole (Fincham et al., 2013). There is *high confidence* in the timing of phytoplankton blooms in spring and summer (Edwards & Richardson 2004; Guinder et al., 2010) and in earlier occurrence of peaks in zooplankton abundance (Chevillot et al., 2017; McGinty et al., 2011; Persson et al., 2012). In contrast, there is *low confidence* in earlier breeding seabirds (Sydeman et al., 2015) because of the *low agreement* in the direction of change (51.0%, Figure 2.36a. For example, many seabirds are breeding earlier (Byrd et al., 2008; Sydeman et al., 2009), while others in temperate and polar regions have been delayed, which has been linked to delayed sea ice breakup locally or limited prey resources (Barbraud and Weimerskirch, 2006; Chambers et al., 2014; WANLESS et al., 2009). Although the response of lifecycle events for many seabird species is variable in direction, there has usually been a more complex driver associated with climate that has been considered to be responsible (Sydeman et al., 2015).

The mean shift in timing across all marine groups is earlier by 3.22 ± 0.36 days decade⁻¹ (Figure 2.36b), significantly different from the null hypothesis of no net movement with climate change. Consequently, it is *likely* that most marine species are exhibiting earlier lifecycle events. There is *moderate agreement* amongst marine groups that they have moved earlier, giving *medium confidence* overall that there is a coherent response. Different marine groups are responding to climate change at different rates in terms of their phenology, with sufficient data to assess seabirds, zooplankton, and fish. There is *robust evidence* and *high agreement* in observation of fish (4.95 days decade⁻¹, n=113) and zooplankton (4.15 days decade⁻¹, n=52) in rapid shifts in phenology. There is *low confidence* that seabirds and marine mammals are exhibiting a net rate of movement close to zero. In particular, polar bears are delaying their time to enter their dens because of the warmer conditions (Escajeda et al., 2018; Hauser et al., 2017).

In summary, there is *high confidence* that the timing of lifecycle events in marine species has advanced over the past several decades.

[START FIGURE 2.36 HERE]

Figure 2.36: Observed responses to climate change by marine species, from time series >19 years in length (median 36 years). (a) The proportion consistent with climate change. (b) The shift in time (days per decade). Based on an update to the global database of (Poloczanska et al., 2013). Database contained 157 peer-reviewed papers published in the last 20 years, covering 100s of species, and including 391 time series >19 years in length. As in Poloczanska et al. (2013), time series were classified as being consistent (i.e., earlier timing with warming) or inconsistent with climate change (i.e., later timing), with non-responsive species being excluded.

[END FIGURE 2.36 HERE]

2.3.4.4 Terrestrial Ecosystem Changes

AR5 WGII concluded with *high confidence* that many terrestrial plant and animal species had experienced changes in abundance and had shifted their geographic ranges in recent decades in many regions. Since AR5, there have been further improvements in long-term ecological monitoring records and resurveys, dendroecological records of tree growth and mortality, and subfossil records from sedimentary archives. The gathering of thousands of individual site-level records into open community data resources (Babst et al., 2019; Brewer et al., 2012; Dornelas et al., 2018; Flantua et al., 2015; Uhen et al., 2013; Vincens et al., 2007; Williams et al., 2018) has greatly facilitated studies of changes in rates of ecosystem turnover and shifts in species distributions at continental to global scales.

From the LGM to present, terrestrial ecosystems worldwide experienced large changes in turnover. For example, temperate, high-latitude, and high-elevation regions exhibited high vegetation sensitivity to climate changes accompanying the last deglaciation (Nolan et al., 2018). In eastern North America and Europe, rates of ecosystem turnover were high during the last deglaciation, then were generally low during the early to middle Holocene (Finsinger et al., 2017; Seddon et al., 2015; Shuman, 2012; Williams et al., 2004). In both Europe and North America, rates of plant ecosystem turnover have accelerated over the last several centuries to millennia (Figure 2.37). In Europe, increases in turnover began nearly 6,000 years ago, while in North America, accelerated ecosystem turnover began within the last several centuries. In Europe, recent rates of ecosystem turnover exceed any observed over the last 10,000 years, while in North America rates are at or above any observed over the last 10,000 years.

[START FIGURE 2.37 HERE]

Figure 2.37: Rates of compositional turnover in Europe, eastern North America, and western North America. Derived from Neotoma pollen data and prior work (Finsinger et al., 2017). Squared chord distance (SCD) is a metric of compositional change (Overpeck et al., 1985).

[END FIGURE 2.37 HERE]

Over the last century, long-term ecological surveys encompassing mammals, birds, invertebrates, and plants show that ecosystem turnover is significantly increasing across a broad array of ecosystems (Dornelas et al., 2014). No globally significant trend in site-level species diversity is apparent; in other words, most local assemblages have experienced biodiversity change rather than systematic loss, despite global losses to biodiversity (Pimm et al., 2014). Rapid species turnover is also evident in undisturbed montane areas in Asia, Europe, the Americas, and the Indo-Pacific (Gibson-Reinemer et al., 2015). With rising community turnover, the novelty of contemporary communities relative to

historical baselines has increased (Hobbs et al., 2009; Radeloff et al., 2015), due to greater spatial homogenization, mixtures of exotic and native species, altered disturbance regimes, and legacies of current or historic land use (Goring et al., 2016; Olden and Rooney, 2006; Schulte et al., 2007; Thompson et al., 2013). In the Arctic, some ecosystems are reaching states not seen in 200,000 years (Axford et al., 2009).

In addition to changes in ecosystem turnover, there have been significant shifts in species distributions. During the Pliocene, boreal forest extended to the Arctic coastline and northeastern Greenland, with the treeline reaching 78°N (Figure 2.40) and tundra reduced in extent; temperate forests and grasslands also shifted poleward, and savannahs and woodlands expanded in Africa and Australia at the expense of deserts (Salzmann et al., 2008, 2013). During the last glacial maximum, tundra and steppe expanded, and forests were globally reduced in extent (Binney et al., 2017; Prentice et al., 2000), the treeline retreating southward to 62°N. As with turnover, rates of change in species distributions in Europe and eastern North America were high during the last deglaciation, then slowed during the Holocene (Giesecke et al., 2017; Ordonez and Williams, 2013; Williams et al., 2004). In eastern North America, leading edges expanded more quickly than trailing edges, leading to range expansion for most species (Ordonez and Williams, 2013). By the mid-Holocene, the treeline had shifted northward again to about 75°N (Binney et al., 2009; MacDonald et al., 2000; Williams et al., 2011) and North Africa experienced a widespread conversion from grasslands to desert (Hoelzmann et al., 1998; Jolly et al., 1998; Prentice et al., 2000).

Species range shifts during the 20th and early 21st century have been widely documented for many global land areas using a suite of long-term ecological records (Lenoir and Svenning, 2015; Pecl et al., 2017). Research has been most extensive for North America and western Eurasia, with relatively few studies for central Africa, eastern Asia, South America, Greenland, and Antarctica (Lenoir and Svenning, 2015). Consistent with increasing temperatures, species have usually moved poleward and upslope (Elmendorf et al., 2015; Evans and Brown, 2017; Harsch et al., 2009; Lenoir et al., 2008; Parmesan and Hanley, 2015). Notably, a large, quasi-global analysis (Chen et al., 2011) estimated that insect, bird, and plant species distributions had shifted by 17 kilometers per decade toward higher latitudes and 11 meters per decade toward higher elevations since the mid-20th century. Both the leading and trailing edges of species distributions have changed (Rumpf et al., 2018). Community assemblages have typically exhibited a consistent poleward/upslope pattern of change, but some individual species have diverged from this pattern because various biotic and abiotic factors (such as precipitation and land use) superceded the physiological effects of temperature (Gibson-Reinemer and Rahel, 2015; Ordonez et al., 2016; Scheffers et al., 2016; Vanderwal et al., 2013). Other species have generally advanced poleward but at rates that have not kept pace with the rates of climate change. Trees are one such example because of their long lifespan and slow maturity (Renwick and Rocca, 2015); in particular, advance is only evident at about half of the sites in a large global dataset of treeline dynamics for 1900-present (Harsch et al., 2009). Northward movement has been more extensive in areas with strong winter warming (Harsch et al., 2009), the current treeline reaching 73°N (Figure 2.40) in the Taimyr Peninsula region of central Siberia (MacDonald et al., 2008), roughly 2° poleward of its pre-industrial limit.

Consistent with species range shifts, analyses of in situ climate records indicate geographic changes in the distribution of eco-climate zones. For instance, in a global study for 1950-2010, (Chan and Wu, 2015) detected poleward shifts in temperate, continental, and polar climates (increases in the average elevation of the latter were also apparent). Within North America, Daly et al. (2012) very generally found zonal changes in winter plant hardiness regions since the 1970s over the central and eastern United States, with elevational changes also being important in the West. Likewise, McKenney et al. (2014) detected a clear northward migration in winter plant hardiness zones across western Canada since 1930, with somewhat lesser changes in the southeastern part of the country.

In summary, based on consistency between ecological and instrumental records, there is *high confidence* in observed changes in ecosystem composition and species distributions over much of the

globe during the past century, with species generally moving poleward and upslope. The accelerated rates of change during the late 20th and early 21st century are unusual in the context of the current Holocene interglacial (*medium confidence*).

2.3.4.5 Global Greening

AR5 WGII briefly discussed changes in global vegetation greenness derived from satellite proxies for photosynthetic activity. Observed trends varied in their strength and consistency; consequently, AR5 made no *confidence* statement regarding observed changes. Analyses since AR5 depict increases in vegetation activity and extent (i.e., greening) since the early 1980s (*medium evidence*; SRCCL). For example, Normalized Difference Vegetation Index (NDVI) increased globally through the early 2010s (Liu et al., 2015b), particularly in the Northern Hemisphere, with Pan et al. (2018) finding increases over about 70% of the earth's surface. Globally integrated Leaf Area Index (LAI) also rose through at least the early 2010s (Forzieri et al., 2017; Jiang et al., 2017; Xiao et al., 2017; Zhu et al., 2016) and likely through near-present, (Chen et al., 2019a) documenting a 34% areal increase from 2000-2017. Although less frequently analyzed for temporal trends, Fraction of Absorbed Photosynthetically Active Radiation (FAPAR) exhibits greening over many global land areas in the past two decades (Figure 2.38) (Forkel et al., 2014; Gobron, 2018; Keenan and Riley, 2018). Areas with extensive greening generally include northern North America, northern South America, Europe, southern and eastern Asia, and equatorial Africa. Some analyses also document changes in specific vegetation types, such as a 7% increase in global tree cover for 1982-2016 (Song et al., 2018) and an expansion of shrub extent in the Arctic tundra (Myers-Smith and Hik, 2018). The increased greening is largely consistent with CO₂ fertilization, nitrogen deposition, land-cover change, and temperature increases (Chen et al., 2019a; Zhu et al., 2016), the latter being particularly important at northern high latitudes (Keenan and Riley, 2018; Kong et al., 2017).

[START FIGURE 2.38 HERE]

Figure 2.38: Satellite-based trends in Fraction of Absorbed Photosynthetically Active Radiation (per decade) for 1998-2018 (updated from Gobron (2018)). White areas indicate trends that are not statistically significant ($p < 0.1$).

[END FIGURE 2.38 HERE]

From the mid-1990s to the mid-2000s, many studies document an increasing prevalence of decreases in vegetation (i.e., browning), especially in the northern hemisphere mid- to low-latitudes e.g. (Bhatt et al., 2017; Frost et al., 2014; Ju and Masek, 2016; Li et al., 2015; Liu et al., 2015b; Pan et al., 2018; Vickers et al., 2016; Zhou et al., 2014a). While a number of factors contribute to browning, such regions are generally associated with drought and wildfire (Yang et al., 2018; Zhou et al., 2014a), while over smaller areas, land-use changes such as agricultural activity and land abandonment are also important (Horion et al., 2016; Viglizzo et al., 2011).

Notably, there is a broad range of estimated changes in global terrestrial vegetation among the suite of recent analyses. Global-scale linear trends differ substantially across products even when the period of record and trend metric are held constant (Jiang et al., 2017). From an areal perspective, some individual studies provide ranges rather than single estimates to underscore uncertainty; for instance, Zhu et al. (2016) concluded that 25-50% of the global vegetated surface experienced greening from 1982-2009. A number of factors likely contribute to this large span in estimated changes. Remotely sensed vegetation products vary in their spatial and temporal completeness and are sensitive to contamination from atmospheric composition, clouds, snow cover, and anisotropy, as well as orbital

changes and sensor degradations (de Jong et al., 2012; Jiang et al., 2017; Pan et al., 2018; Xiao et al., 2017; Zhu et al., 2016). Ground measurements for calibration and validation are scarce before 2000 because of a paucity of suitable field campaigns (Xiao et al., 2017). Land use and disturbance regimes can mask large-scale signals (Franklin et al., 2016). In addition, there is a plethora of models for the detection of phenological metrics from satellite data as well as a variety of statistical techniques for analyzing historical changes (Wang et al., 2016b).

In summary, there is *medium confidence* that terrestrial vegetation across the globe has increased since the early 1980s.

2.3.4.6 Growing Season

AR5 WGII briefly discussed large-scale changes in the length of the growing season but made no *confidence* statement about observed trends. However, AR5 did conclude with *high confidence* that warming contributed to an overall spring advancement in the Northern Hemisphere. Since AR5 variations in the length of the growing season have been estimated using both in situ and satellite observations, the former providing a longer-term perspective and the latter having better geographic coverage.

In situ analyses depict increases in the length of the thermal growing season over much of the extratropical land surface since the mid-20th century. For example, (Donat et al., 2013a) detected a globally averaged rise of about 1.0 days per decade for 1951-2011, with statistically significant increases over much of the Northern Hemisphere. (Barichivich et al., 2013) found a 1.7 day per decade widening of the growing season poleward of 45°N over the same period, with slightly larger changes across Eurasia. Within North America, Kukul and Irmak (2018) identified growing season increases of about 1.3 days per decade in the United States for 1900-2014, with (McCabe et al., 2015) noting larger increases after 1980, while Pedlar et al. (2015) identified rises for nearly all ecozones in Canada from 1950-2010. Within Asia, (Xia et al., 2018) documented an increase in growing season length of at least 1.0 days per decade in China since 1960, and (Jung et al., 2015) identified an increase of several days per decade in South Korea since 1970, both related to an earlier start of the season. Generally speaking, phenological indicators are consistent with changes documented by instrumental data (Parmesan and Hanley, 2015). For instance, over recent decades, peak bloom dates for cherry blossoms in Kyoto, Japan have occurred progressively earlier in the growing season, as have grape harvest dates in Beaune, France (Figure 2.39).

[START FIGURE 2.39 HERE]

Figure 2.39: Phenological indicators of changes in growing season. The left panel depicts the time series of cherry blossom peak bloom dates in Kyoto, Japan (from Aono and Saito, 2010). The right panel depicts the time series of grape harvest dates in Beaune, France, from Labbe et al. (2019).

[END FIGURE 2.39 HERE]

Changes in the length of the photosynthetically active growing season (derived from NDVI) are also evident over many land areas since the early 1980s. In general, the strongest increases are in the mid- to high-latitudes of the Northern Hemisphere from the early 1980s, but with evidence of a reversal to a decline in season length starting in the early 2000s. (Park et al., 2016) identified an increase of about 2.6 days per decade poleward of 45°N for 1982-2014, centered over mid-latitude Eurasia and northeastern North America, but this linear metric masks a decrease in season length starting after 2000. (Garonna et al., 2016) found an overall linear trend of 2.2 days per decade and generally comparable patterns of change above 40°N for 1982-2012, including a lower trend in the most recent

decade. (Barichivich et al., 2013) detected a trend of 2.2 days per decade in the length of the growing season above 45°N, finding a good correspondence between satellite and in situ estimates of non-frozen days for 1988-2007 over Eurasia and North America. Several other large-scale analyses (e.g., (Liu et al., 2016; Wang et al., 2016b) also document general increases in growing season length poleward of 45°N from the early 1980s through the early 2000s, (Zhao et al., 2015) pointing out reduced trends after 2002. In addition, a number of studies capture changes at smaller areas, examples being Chen et al. (2016) for the Canadian Arctic, (Hogda et al., 2013) for Fennoscandia, Garonna (2014) for much of Europe, and Vrieling (2013) for much of subsaharan Africa. In many ecosystems, changes in growing season length result in changes of gross primary production (GPP) and net ecosystem exchange (NEE) (Keenan et al., 2014; Richardson et al., 2013).

The very general consistency between in situ and satellite estimates is noteworthy given that a variety of factors contribute to uncertainty in observed changes. For example, there is no universally accepted definition of growing season length across in situ analyses; some define the growing season as the period with mean daily temperatures above a certain threshold (e.g., 5°C) whereas others use the span between the date of the last spring freeze and the date of the first autumn frost. Spatial and temporal coverage can also be problematic with in situ studies (Donat et al., 2013a). For satellite analyses, two somewhat different sources of uncertainty are germane (Wang et al., 2016c): i) uncertainties related to the satellite datasets themselves (e.g., satellite drift, sensor differences, calibration uncertainties, atmospheric effects); and ii) uncertainties related to the methods for determining phenological metrics (e.g., start, end, and length of season).

In summary, based on multiple independent analyses of in situ, satellite, and phenological data, there is *high confidence* that the length of the growing season has increased over much of the extratropical Northern Hemisphere since at least the mid-20th century. There are no widely available and reliable sources of proxy information that permit a longer-term context for growing season length changes.

2.3.5 *Synthesis of evidence for past changes*

Section 2.3 has assessed the observational evidence basis for changes across the atmosphere, cryosphere, oceans and biosphere. For each component of the climate system key indicators have been considered starting, where applicable, from the long-term proxy record and coming up to the present day instrumental period. This synthesis can inform a final assessment of the evidence for change in the climate system and its unusualness or otherwise.

Evidence is ubiquitous that climate has varied across a broad range of timescales historically with very large swings in the last 2 million years between glacials and interglacials, including the current Holocene interglacial. Prior to this there is evidence for even greater variations still in the global climate. The climate system has repeatedly been both much warmer and much cooler than the present day climate in Earth's history with associated large-scale changes in multiple components of the system (Figure 2.40).

Over multiple glacial cycles (the last approximately 800ka) and then, with even more detail, in the Common Era, it is possible to consider both more aspects of the climate system and with increasing fidelity the rates of change. Solely for the last 150 years or so are direct observations available, and only in the past few decades have observational systems attained truly global monitoring capabilities. The direct observations point unequivocally to rapid change in very many components of the climate system since the pre-industrial. These are all internally consistent in pointing towards a world that has warmed rapidly.

The question of unusualness of present climate changes is key to understanding the importance of what we have directly observed. The climate system consists of many observable aspects that vary over a very broad range of different timescales. Some biogeochemical indicators of atmospheric

change vary almost instantaneously and are in states unseen in at least 800ka and *virtually certain* millions of years. Several relatively fast varying indicators such as Global Mean Surface Temperatures are now *more likely than not* in states that have not been seen in the Common Era and *about as likely as not* since the last interglacial 125ka. More slowly varying indicators such as sea-level, glaciers and ice sheets as well as various biospheric indicators are collectively in states unseen in Centuries to millenia. The rate of change in almost all the assessed components since the pre-industrial is highly unusual in the context of preceding slowly varying changes during the current postglacial period.

In summary, directly observed changes in key large-scale atmospheric, oceanic, cryospheric and biospheric indicators of climate are unequivocal. Many of the assessed indicators are now in states not seen for centuries, and some have exceeded conditions over many millennia. The climate system as a whole has shifted significantly and unusually rapidly since industrialization

[START FIGURE 2.40 HERE]

Figure 2.40: Pictorial representation of the changes in selected key indicators at various timeslices from the deep past to the present. Numbers are based upon the assessment carried out in the present chapter and other chapters of the report.

[END FIGURE 2.40 HERE]

2.4 Changes in modes of variability

Climate variations are not randomly distributed in space and time, but rather exhibit internally preferred modes of behaviour. The modes of variability denote the organized patterns of climate variability with coherent spatial structures whose amplitude, phase, location and roles within the climate system undergo important fluctuations over time. Variability of these modes spans a wide range of time scales and may modulate the observed climatic trends. Climate variability with interannual and longer timescales (Cross-Chapter Box 2.1) are assessed herein. Additional modes are considered as required in latter process based and regional chapters.

2.4.1 *El Niño-Southern Oscillation (ENSO)*

AR5 reported with *medium confidence* that ENSO-like variability existed, at least sporadically, through the warm period of the Pliocene. It was also found (*high confidence*) that ENSO has remained highly variable during the last 7000 years with no discernible orbital modulation. It was concluded in AR5 that large variability on interannual to decadal timescales, and differences between datasets, precluded robust conclusions on any changes in ENSO during the instrumental period. SROCC reported epochs of strong ENSO variability throughout the Holocene, with no evidence of a systematic trend in ENSO amplitude, but with some indication that the ENSO amplitude over 1979-2009 was greater than at any point in the period from 1590-1880. It was also reported that the frequency and intensity of El Niño events in the period from 1951-2000 was high relative to 1901-1950.

ENSO has been monitored through a range of indices, which are based on sea surface temperature (SST) anomalies or atmospheric pressure during the instrumental period, and various proxies on longer timescales. Differences between indices and thresholds used to define the occurrence of an El Niño or La Niña event mean that there is no commonly accepted list of events and their amplitude. A large number of studies have historically used SST estimates, but in-situ monitoring prior to the TAO array was often poor. Since AR5, there have been new versions of a number of SST data sets which have been widely used in ENSO assessments and these have altered apparent tropical Pacific SSTs in

substantial and important ways. In particular, the ERSSTv3 data set has been replaced by ERSSTv4 and ERSSTv5, addressing known SST biases in the equatorial Pacific which affected the mean state and amplitude of indices based on that data set (Huang et al., 2015). [We expect that there will be something to add here from HadSST4 once available]. Analyses using earlier versions of certain SST products thus need to be treated with a degree of caution.

2.4.1.1 Evidence prior to instrumental records

There is increasing geological evidence that ENSO-like variability existed, at least sporadically, during the warm period of the Pliocene. Manucharyan and Fedorov (2014) found that such variability has been present during epochs of millions of years, with proxy records indicating that this was the case even when cross-Pacific SST gradients were much weaker than present. Watanabe et al. (2011) found ENSO-like variability from coral proxies in the Philippines dated between 3.5 and 3.8 Ma. The extent to which cross-Pacific SST gradients weakened during the Pliocene warm period is uncertain, with earlier findings of a substantial weakening possibly being an artefact of proxy bias (Susskind et al., 2019).

Cobb et al. (2013) found a period of relatively weak ENSO activity in the mid-Holocene, approximately 6000 years ago, while being based on limited evidence. A number of subsequent studies (McGregor et al., 2013; Carré et al., 2014; Emile-Geay et al., 2016; Thompson et al., 2017; Tian et al., 2017; White et al., 2018) have since found similar or more pronounced weakening of ENSO at various times in the mid-Holocene, with the stronger decreases generally revealed by remotely teleconnected proxies rather than those close to the core region of ENSO activity. However, Karamperidou et al. (2015) find that weakening in ENSO-related variability in eastern Pacific proxies does not necessarily correspond to weakening in central Pacific proxies, potentially indicating an along-equatorial shift in ENSO activity rather than a weakening of ENSO during some periods. There is substantial disagreement between proxy records as to ENSO activity during the early Holocene (White et al., 2018b; Zhang et al., 2014a). Following the period of weak ENSO in the mid-Holocene, a number of studies find an increase in ENSO activity which commences between 3000 and 4000 years ago (Chen et al., 2016; Cobb et al., 2013; Emile-Geay et al., 2016; Koutavas and Joanides, 2012a; Thompson et al., 2017b; Zhang et al., 2014b).

Numerous studies (Hope et al., 2017; Ledru et al., 2013a; Li et al., 2013a; Liu et al., 2017b; McGregor et al., 2013a; Rustic et al., 2015) find substantial variability in ENSO activity on multi-decadal to centennial timescales over the last 500 to 1000 years, although due to a wide spread of proxy outcomes they differ in their detailed assessment in the pre-instrumental period (e.g. Dätwyler et al., 2019). Most also find that ENSO activity in recent decades was higher than in the few hundred years prior to the instrumental period. McGregor et al. (2013) found stronger ENSO variance for the 30-year period 1979-2009 compared to any 30-year period within the timespan 1590-1880. Koutavas and Joanides (2012), Ledru et al. (2013) and Thompson et al., (2017) identify a period ending 700 to 1000 years ago when ENSO activity was greater than in the following few hundred years, and more closely comparable to the 20th century behaviour.

2.4.1.2 Instrumental period records

There remains no clear evidence of any significant long-term trend in the east-west SST gradient across the equatorial Pacific Ocean during the instrumental period, with periods of gradients both above and below the long-term average on decadal timescales, associated with a predominance of La Niña or El Niño events respectively. Changes in the Walker Circulation are also complex and sensitive to choice of analysis approach, with some studies reporting weakening over the period since 1950 (e.g. Power & Kociuba, 2011a; Tokinaga et al., 2012a), and others finding strengthening over the last 30-40 years (e.g. Hu et al., 2013; L'Heureux et al., 2013). Based on estimations of changes in

mid-tropospheric pressure velocity from changes in observed cloud cover, Bellomo and Clement (2015) suggest a weakening and eastward shift of the Walker circulation during the period 1920-2010. Conversely, reanalysis products show a robust strengthening and a westward shift of the Walker circulation over the last three decades (Bayr et al., 2014; Ma and Zhou, 2016).

ENSO amplitude since 1950 has increased relative to the 1910-1950 period, as assessed by independent evidence provided by proxy records (e.g., Gergis & Fowler, 2009), the Southern Oscillation Index (SOI) (Braganza et al., 2009) and sea surface temperatures (e.g., Ohba, 2013; Yu & Kim, 2013). SOI data (Figure 2.41) and some newer SST-based studies show high ENSO magnitude in the period from the mid-late 19th century to about 1910 supporting the SOI. The frequency of El Niño and La Niña events is also subject to considerable decadal variability (e.g. Hu et al., 2013) but with no clear evidence of a long-term signal in the frequency of events.

[START FIGURE 2.41 HERE]

Figure 2.41: 30-year running variance of (left) reconstructed annual mean Niño 3.4 from a range of proxy-based studies (Stahle et al. 1998, Cook 2000, Mann et al. 2000, Evans et al. 2001, Evans et al. 2001, Cook 2008, Braganza et al. 2009, McGregor et al. 2010, Wilson et al. 2010, Li et al. 2011, Emile-Geay et al. 2013, Li et al. 2013, Tierney et al. 2015) (right) the June-November Southern Oscillation Index and April-March mean Niño 3.4 (from ERSSTv5, 1981–2010 base period).

[END FIGURE 2.41 HERE]

There has been a particular predominance of high-amplitude ENSO events since the 1970s. SROCC considered extreme El Niño and La Niña events, using definitions proposed by Cai et al. (2014, 2015), based on rainfall anomalies over the Niño 3 region of the eastern Pacific (El Niño) and the Niño 4 SST index (La Niña); subsequently alternative definitions have been proposed by Santoso et al., (2017). Under the Cai et al. (2014) definition, the El Niño events of 1982-83, 1997-98 and (marginally) 2015-16 are categorised as extreme. However, this definition cannot be applied to the period prior to the availability of satellite rainfall observations and hence cannot be used to assess pre-1980 El Niño events.

The El Niño events of 1982-83, 1997-98 and 2015-16 had the strongest anomalies in the Niño 3.4 SST index in the instrumental record. Their predominance was less evident from indices based on de-trended data such as the ONI (which still ranked them as the three strongest events since 1950, but only by a small margin), and in the SOI, which indicates comparable or stronger events in 1896 and 1905. Yu & Kim (2013)'s implementation of the ONI found a number of events with the ONI above 1.5°C between 1888 and 1905, then no such events until 1972, although uncertainties in SST observations reduce the *confidence* in ONI estimates prior to 1950. Giese & Ray (2011) also found a number of events between 1890 and 1920 in the SODA ocean reanalysis with comparable magnitude to the most significant recent events. There have also been a number of strong La Niña events (e.g. 1973-74, 1975-76 and 2010-11) with few clear analogues in the 1920-1970 period; the proxy-based analysis of McGregor et al. (2010) indicates that the mid-1970s La Niña period was also extreme in a multi-centennial context. Given the relatively large uncertainty in pre-1950 SST data in the tropical Pacific region, and evidence for several high-amplitude events in the late 19th and early 20th century, there is no clear evidence that the frequency of high-amplitude events since the 1970s constitutes a long-term trend which can be separated from multi-decadal variability.

2.4.1.3 Categorisation of ENSO events

AR5 noted the importance of distinguishing different types of ENSO events, in particular between El

Niño events focused in the eastern Pacific (‘Eastern Pacific’ (EP) or ‘classical’ events) and those focused in the Central Pacific (‘Central Pacific’ (CP) or ‘Modoki’ events), which have different typical teleconnections (e.g. Ashok et al., 2007; Capotondi et al., 2015; Ratnam et al., 2014; Timmermann et al., 2018). Banholzer & Donner (2014) found that CP events had a weaker influence on global mean surface temperature than did EP events. Kug & Ham (2011) concluded that central and eastern Pacific SST indices were more highly correlated with each other during La Niña than during El Niño events, indicating that La Niña events have less geographic diversity than El Niño events.

The distinction between CP and EP events is not necessarily clearly defined; Wiedermann et al. (2016) found that different methods are inconsistent in their classification of events, implying that no robust consensus classification into CP or EP is possible. Furthermore, Giese & Ray (2011) found that the central longitude of peak SST anomalies associated with individual El Niño events had a distribution not distinguishable from a Gaussian distribution peaked at 140°W, as opposed to a bimodal distribution as implied by the EP/CP paradigm. Instrumental data have historically been of limited value in assessing the classification of events before 1955, because of the almost total lack of SST observations in the western half of the Niño 4 region (5°N-5°S, 160°E-150°W) (L’Heureux et al., 2013); it is not yet known whether newer SST data set versions which benefit from improved data archives (Section 2.3.1.1.3) have reduced this uncertainty.

A number of studies, using a range of indicators, have found an increase in recent decades of the fraction of CP El Niño events, particularly after 2000 (Yu and Kim, 2013; Lübbecke and McPhaden, 2014; Pascolini-Campbell et al., 2015; Jiang and Zhu, 2018). Johnson (2013) found that the frequency of event types with a CP focus had increased (although not significantly) over the 1950-2011 period, accompanied by a significant increase in the frequency of La Niña events with a warm (as opposed to cool) western Pacific warm pool. A coral-based reconstruction starting in 1600 (Freund et al., submitted) found that the ratio of CP to EP events in the last 30 years was substantially higher than at any other time over the last 400 years.

2.4.1.4 ENSO Teleconnections

There is limited evidence of long-term changes in ENSO teleconnections. Some studies (e.g. Chiodi and Harrison (2015)) found that teleconnections over the most recent decades are broadly consistent with those over the last 100 years. Shi and Wang (2018) found that teleconnections with the Asian summer monsoon were generally stable since the 17th century during the developing phase of the monsoon, and showed substantial decadal variability, but no clear trend, during the decaying phase.

Teleconnections associated with ENSO are well-known to vary on decadal to multi-decadal timescales (e.g. Power et al., 1999; He et al., 2013; Lee and Ha, 2015; Ashcroft et al., 2016; Jin et al., 2016a), although Yun and Timmermann (2018) found that decadal variations in teleconnections between ENSO and the Indian monsoon were not distinguishable from a stochastic process. Many observed decadal changes in teleconnections are consistent with a shift to more central Pacific El Niño events (Evtushevsky et al., 2018; Yeh et al., 2018; Yu and Sun, 2018; Zhao and Wang, 2018). Impacts of the PDO (Kwon et al. 2013; Wang et al., 2014; Dong et al., 2018) and AMO/V (Kayano et al.) can also modulate ENSO teleconnections, and affect the frequency of CP versus EP events (Ashok et al., 2007). Variability in teleconnections can also occur on timescales longer than characteristic PDO timescales (e.g. Gallant et al., 2013).

2.4.1.5 Recent events and their implications for longer-term trends

A number of indicators reported in AR5 showed trends, most of which were not significant, consistent with a more La Niña-like mean state over the 1979-2012 period. Since 2012, there has been a very

strong El Niño event in 2015-16 (Newman et al., 2018), with the November-January ONI index (+2.6°C) being the highest in the post-1950 record. Its eastern Pacific signal was relatively weak compared to the 1982-83 and 1997-98 events, especially near the South American coast (Santoso et al., 2017b), although it still had Niño 3 index higher than Niño 4 (one definition used to distinguish EP from CP events). This was followed by an unusual event in early 2017, when large positive SST anomalies occurred along the South American coast. This event was accompanied by typical El Niño-like atmospheric responses in both North and South America, whilst the broader Pacific was in a neutral to weak La Niña state. (The March 2017 value of the Niño 1+2 index was +1.6°C, a level which has only previously been reached in the instrumental period during basin-wide El Niño events). The 2010-11 and 2011-12 La Niña events were also not fully incorporated in many studies published at the time of AR5. Most of the trends towards a La Niña-like mean state over the 1979-2012 period have weakened or reversed when assessed over 1980-2018.

2.4.1.6 Summary of ENSO understanding

In summary, there is no clear evidence for a recent sustained shift in ENSO or associated features such as the Walker Circulation, or in teleconnections associated with these, beyond the range of variability on decadal to millennial timescales. There is *medium confidence* that both ENSO amplitude and the frequency of high-magnitude events since 1950 is higher than over the pre-industrial period from 1850 as far back as 1400, but *low confidence* that it is outside the range of variability over periods prior to 1400. A high proportion of El Niño events in the last 20-30 years have been centred in the central, rather than eastern Pacific (with an associated westward shift in the Walker Circulation) but there is *low confidence* that this represents a long-term change.

2.4.2 Indian Ocean Basin and dipole modes

AR5 did not provide any assessment of the Indian Ocean Dipole (IOD) records based on paleo reconstructions. For the instrumental era AR5 reported statistically insignificant trends for the IOD over 1901–2012, 1950–2012 and 1979–2012. However, the Indian Ocean Basinwide (IOB) mode, as represented by the basin-scale averaged SST index depicted significant warming trends in all periods except 1979–2012.

The tropical Indian Ocean experiences uniform warming (cooling) during the decay phase of El Niño (La Niña) events, most pronounced during austral summer into autumn, giving rise to the IOB mode. Conversely, the IOD denotes the SST fluctuations whereby the eastern and western parts of the basin become alternately warmer and then colder. The IOB and IOD represent the first and second modes of the tropical Indian Ocean variability, respectively (Saji et al., 1999). Indian Ocean temperatures have historically exhibited considerable warming which is largely removed in the IOD index which is by definition the difference between two warming regions (Figure 2.42).

Paleo reconstructions based on proxy evidence from corals suggest an intensification of the IOD during the 20th century (Abram et al., 2008). Using Seychelles coral records from the mid-Holocene to present, (Zinke et al., 2014) identified changes in the bimodality of the seasonal cycle, with the lowest seasonal SST range in the mid-Holocene (6.2-5.2ka) and then again around 2ka, while the largest seasonal range occurred at 4.6ka and during the present (1990-2003). Reconstructions from fossil corals for the eastern Indian Ocean point to stronger IOD SST anomalies due to enhanced upwelling and cooling driven by a stronger monsoon with enhanced anomalous easterly winds in the eastern Indian Ocean during the mid-Holocene (Abram et al., 2007). Niedermeyer et al. (2014) suggest that the period 6.5-4.5ka was characterized by an anomalously positive IOD mean state represented by an east-west dipole in hydroclimate across the tropical Indian Ocean between East Africa and the western Pacific warm pool based on stable isotopes in terrestrial plant waxes.

In contrast, during the last deglaciation (16-8ka), including the Younger Dryas event, Wurtzel et al. (2018) contend that Indian Ocean hydroclimate does not mirror the zonal asymmetry observed in Indian Ocean SSTs, according to their analysis of a speleothem record from Sumatra, but instead reflects shifts in moisture transport pathways and sources. For the LGM (25-17ka), enhanced equatorial Indian Ocean productivity (during 25-17ka) in sediment records was also associated with strengthened westerly jets, in line with a shallower central-western Indian Ocean thermocline and stronger negative IOD events (Punyu et al., 2014).

[START FIGURE 2.42 HERE]

Figure 2.42: Climate indices for the Indian and Atlantic Oceans in different historical observational data sets, 1854–2018. Shown are the following indices from the top: Indian Ocean Basin (IOB) mode, Indian Ocean Dipole (IOD), Atlantic Multidecadal Oscillation/Variability (AMO/V), Atlantic meridional mode (AMM) and the Atlantic Niño (Atl3). The indices were derived from five different SST data sets: (1) the Objective Analyses of Sea-Surface Temperature and Marine Meteorological Variables for the 20th Century using ICOADS and the Kobe Collection, COBE v2 (Hirahara et al., 2014b); (2) the Extended Reconstructed Sea Surface Temperature, ERSST v5 (Huang et al., 2017); (3) the Hadley Centre Global Sea Ice and Sea Surface Temperature, HadISST v1 (Rayner et al., 2003); (4) the Kaplan Extended SST dataset, KAPLAN v2 (Kaplan et al., 1998); and (5) the satellite-derived Optimum Interpolation SST (OISST) (Reynolds et al., 2002).

[END FIGURE 2.42 HERE]

Using a coral $\delta^{18}\text{O}$ record from the Seychelles covering the period 1854-1994, (Du et al., 2014) showed a 3-7 yr dominant period associated with the IOB in response to ENSO forcing, with the Seychelles record maximally correlated at 5 months lag with Pacific conditions. They identified multi-decadal variability in the IOB, with more active IOB phases during 1870-1890, 1930-1955, and 1975-1992, while decadal variability in the IOB dominated during 1940-1975 (Du et al., 2014). Evidence for changes in IOB characteristics during earlier periods of interest (e.g., mid-Holocene, LGM) is limited.

The role of decadal to multi-decadal variability has recently emerged as an important consideration for the IOD. Decadal variations in SST featuring an IOD-like out-of-phase pattern between the western and eastern tropical Indian Ocean was found to be linked to the PDO (Krishnamurthy and Krishnamurthy, 2016) or IPO (Dong et al., 2016). In this link a combination of processes transmits the signal from the Pacific to the Indian Ocean through the atmospheric and oceanic bridges, leading to variations in the subsurface thermal structure in the Indian Ocean (Jin et al., 2018; Zhou et al., 2017). Decadal modulations of the mean thermocline depth in the eastern Indian Ocean provides pre-conditioning for the IOD events (Annamalai et al., 2005); with positive events in the 1960s and 1990s linked to relatively shallow eastern Indian Ocean thermocline and negative events in the 1970s and 1980s due to deeper thermocline (Ummenhofer et al., 2016).

Apparent changes in the frequency and/or magnitude of the IOD events in the second half of the 20th and into the 21st century are sensitive to choice of SST dataset. It has been suggested that the number of positive IOD events in recent decades has increased (Cai et al., 2009a). Recent IOD index values are skewed towards more frequent positive and fewer negative IOD events (Cai et al., 2009b), potentially due to recent shoaling of the thermocline in the eastern Indian Ocean (Cai et al., 2008). Although the seasonal evolution and type of ENSO (Section 2.4.1) may influence the character of the IOD (Fan et al., 2016; Guo et al., 2015; Zhang et al., 2015), the higher occurrence of the latter may be independent of the former (Sun et al., 2015).

To summarize, there is *low confidence* in any changes to long-term Indian Ocean Dipole variability. Understanding of the Indian Ocean Basin mode of variability trends during the instrumental period is also limited due to underlying Indian Ocean warming trends.

2.4.3 Atlantic Multidecadal Oscillation / Variability

AR5 reported no robust signals in AMO/V reconstructions based on paleo records arguing for low correlations between different AMO/V reconstructions prior to 1900. AR5 concluded that there have been no significant trends in the Atlantic Multidecadal Oscillation/Variability (AMO/V) index during the instrumental period. There was concluded to be difficulty in interpreting the AMO/V signal because of the long-term underlying SST warming trend.

The AR5 conclusions about large uncertainties in AMO/V paleo reconstructions have been further demonstrated in the last years. This is associated with the large uncertainties in the reconstructions of the occurrence of the AMO/V several hundred years prior to the instrumental period revealed by tree-rings (Gray et al., 2004), corals (Kilbourne et al., 2008), Greenland ice (Chylek et al., 2012), and multi-proxy reconstructions of low-latitude corals (Kilbourne et al., 2014; Svendsen et al., 2014).

Historical observations show that AMO/V is characterized by basin-wide warm and cool periods with an average SST difference of $\sim 0.4^{\circ}\text{C}$, but with larger differences in the North Atlantic subpolar gyre. A variety of indices have been used to define the AMO/V; the simplest being SST averaged over the North Atlantic Ocean, low-pass filtered (e.g., 11 year moving average) and with the linear trend removed (Enfield et al., 2001; Sutton and Hodson, 2005). The main difference among AMO/V indices is the approach used to remove the influence of background warming, which introduces some uncertainty in the strength of the recent phase of AMO/V (e.g., Ting et al., 2009). Furthermore, the AMO/V indices depend on the latitudinal span over which the SST is averaged (Gulev et al., 2013). All indices show warm periods occurred approximately between 1880 to 1900, 1940 to 1960, and from the mid-1990's to present, with cool periods in between (Figure 2.42). The pattern and strength of the AMO/V differs among the periods (e.g., Svendsen et al., 2014) and there are indications that there may have been a shift since 2005 toward a negative phase of the AMO/V (Robson et al., 2016).

Reconstructions based on observations going back to the mid-19th Century show that the North Atlantic Ocean drives turbulent fluxes variations closely related to AMO/V, with the atmosphere damping the SST variations (Gulev et al., 2013). The oceanic changes are linked with salinity and temperature variations over the upper 3000m of the North Atlantic (Keenlyside et al., 2015; Polyakov et al., 2005), and to sea level variations in the Western North Atlantic along the Gulf Stream passage (McCarthy et al., 2015). The winter NAO and related atmospheric blocking co-varied with the AMO/V, with warm phases of AMO/V associated with negative NAO and increased atmospheric blocking (Häkkinen et al., 2011; Omrani et al., 2014).

In summary, it is *very likely* that the AMO/V index has not exhibited any significant trend during the instrumental period (*high confidence*). However, instrumental records may not be long enough to distinguish any oscillatory tendency from trends. There is *low confidence* in paleo-AMO/V reconstructions due to a paucity of high-resolution records.

2.4.4 Pacific Decadal Variability

AR5 reported a statistically significant weakening trend of the Pacific Decadal Oscillation (PDO) from 1979-2012, although the trends were not statistically different from zero during the instrumental period as a whole. No consideration was given to the pre-instrumental PDO signals and the assessment of PDO during the instrumental period was not associated with any confidence or likelihood statement.

The Pacific decadal variability (PDV) is dominated by an SST pattern with anomalies in the eastern tropical Pacific and similar strength but opposite signed anomalies in the extra-tropical western north and south Pacific. The existence of Pacific decadal variability in the centuries prior to the instrumental period is evidenced by a variety proxy evidence, based on trees (Biondi et al., 2001), corals (Deng et al., 2013; Felis et al.; Linsley et al., 2015) and lake sediments (Lapointe et al., 2017). There is little agreement in the phasing of the various paleo-proxy indices prior to the instrumental record, and neither these nor the instrumental records provide evidence of a clearly defined spectral peak (Chen and Wallace, 2015; Henley, 2017; Newman et al., 2016a; Zhang et al., 2018b). Variability associated with PDV SSTs is commonly referred to as the Pacific Decadal Oscillation (PDO) (Mantua et al., 1997) or the Interdecadal Pacific Oscillation (IPO) (Henley et al., 2015; Power et al., 1999). The PDO is defined as the normalized first principal component of North Pacific (20-70°N) SST monthly anomalies with the global mean temperature signal removed (Mantua et al., 1997); while the IPO is defined as the normalized first principal component of low-pass filtered (i.e., 13-year cutoff) global monthly SST anomalies (Power et al., 1999). The two indices are closely related to each other and show coinciding transitions from warm to cold tropical Pacific conditions in the 1920's and 1970's, and from cold to warm conditions in the 1940's and late 1990's (Chen and Wallace, 2015; Newman et al., 2016a). The associated patterns are quite similar, but with the PDO pattern exhibiting stronger SST anomalies in the extra-tropical North Pacific than the IPO (Chen and Wallace, 2015). The strength and structure of the SST patterns also differ among the epochs (Newman et al., 2016a).

Instrumental observations are sparse prior to 1950, and thus the fidelity of the PDO index derived for the early decades of the 1900s is relatively low (Deng et al., 2013; Wen et al., 2014). This factor led to disagreements in the classification of the PDO phase among several indices, even during recent years with high data quality and availability. Nevertheless, the disagreements do not appear to contribute to teleconnection instability in any significant way (McAfee, 2017). Analysis of long-term time series of PDO shows the decadal nature of this mode of variability with no significant long-term trends but highlights a recent switch from positive to negative phase across all indicators.

In summary, the PDO is dominated by large decadal variability with persistence of positive and negative phases over the last 150 years with no overall trend (*high confidence*). Since AR5 a shift from positive to negative phase of the PDO has occurred (*high confidence*). There is *low confidence* in paleo-PDO reconstructions due to poor agreement among the various available estimates.

2.4.5 Annular modes (NAO, NAM, SAM)

2.4.5.1 Northern Annular Mode (NAM)

AR5 reported that the increase in the North Atlantic Oscillation (NAO), a mode of variability closely related to the NAM / Arctic Oscillation (AO) from the 1950s to the 1990s had been largely offset by recent changes, particularly during 2010 and 2013 (*high confidence*). Moreover, the strong positive phase of the NAO observed during the 20th century was not unusual based on NAO reconstructions spanning the last half millennium (*high confidence*).

Since AR5, new multi-millennial-timescale NAO reconstructions are available based on marine and lake sediments, speleotherms, tree rings and ice cores (Faust et al., 2016; Ortega et al., 2015; Sjolte et al., 2018). Past NAO variability over the last several millennia indicates that around the middle Holocene and the beginning of the LIA, the NAO changed from generally positive to variable, intermittently negative conditions (Olsen et al., 2012). Disagreement exists in terms of the NAO phase during the MCA (Faust et al., 2016; Olsen et al., 2012; Ortega et al., 2015; Trouet et al., 2009). Several instrumentally based NAO reconstructions extend back to the early industrial or pre-industrial era, most of them constructed from the difference in sea level pressure between two sites. Cornes et al. (2013) used the pressure gradient between Paris and London to estimate the state of the NAO back

to 1692, highlighting that the positive NAO values during the 1990s were not exceptional in the long-term context. An unprecedented persistent phase of positive values during boreal summer was observed over the 1770-1825 period, in line with the Luterbacher et al. (1999) NAO reconstruction. The NAO showed a persistence of positive values during the 1900-1930 period, particularly for the extended winter season (DJFM) (Cornes et al., 2013; Cropper et al., 2015). A declining trend in NAO index was observed from around 1920-1970, with a recovery thereafter to a period of consistently high values between 1970 and early 1990s, highlighting a large multidecadal variability that affects the assessment of NAO trends.

Based on observations and reanalysis datasets, multidecadal fluctuations were found for the Arctic Oscillation (AO) patterns: the Atlantic centre of the AO pattern remained unchanged throughout the period 1920-2010, whereas the Pacific centre of the AO was strong during 1920-1959 and 1986-2010 and weak during 1960-1985 (Gong et al., 2018). These multidecadal changes were also observed considering the centres of action of the NAO, with a northeast displacement in the period 1930-1950 and after 1980s, and a southwest displacement between 1950 and 1970 (Moore et al., 2013).

Time series for NAM/NAO/AO indices for the extended winter period (DJFM) derived from station data, gridded SLP, and reanalysis 1000hPa heights are dominated by multidecadal variations. Strong increases were identified during the most recent c.50 years, being associated with the mostly negative phases in the 1960s and a persistence of positive values in the late 1980s and the 1990s. Trends for this period are statistically significant only for the NAO CPC index (Barnston and Livezey, 1987), consistent with the results for the AO index as defined by Thompson and Wallace (2000) and the NAM index. This upward trend has considerably weakened for the 1980-2018 period due to the negative phase of NAO lasting from the late 1990s through the early 2000s. After 2000 all NAM/NAO/AO timeseries assessed here experience an increasing trend.

Boreal winter AO/NAM and NAO tend towards stronger variability since the 1990s, that is particularly evident in December NAO (Hanna et al., 2015), with similar larger variability during the latest decade also found in the December AO index (Overland and Wang, 2015). The 2009/2010 winter extremes in the NAO index as well as in the AO/NAM index are the lowest negative anomaly since at least the reanalysis era.

Being much more pronounced in the cold season, NAO experiences seasonality in long-term variability. During the 1970s the summer (July–August) NAO experienced positive anomalies. After that, there is a tendency towards a more negative summer NAO (Allan and Folland, 2017; Hanna et al., 2015), that is in association with evident interdecadal variations (Linderholm and Folland, 2017).

In summary, strong positive trends for the NAM/NAO/AO indices were observed since 1960, which have weakened since the 1980s (*high confidence*). The NAO variability in the instrumental record was *likely* not unusual in the millennial and multi-centennial context.

2.4.5.2 Southern Annular Modes (SAM)

AR5 found that it was *likely* that the SAM become more positive since the 1950s. It concluded that this increase was unusual in the context of the prior 400 years with *medium confidence*.

Using a variety of temperature-sensitive proxy records from Antarctica and South America, spanning the mid- to polar latitudes to reconstruct the SAM, Abram et al. (2014) concluded that the long-term mean of the SAM index of recent decades is at its highest value for at least the past 1,000 years. Abram et al. (2014) also identified a prolonged period with low SAM polarity during the period 1400-1700, which is also reflected in remaining SAM reconstructions (Dätwyler et al., 2017; Villalba et al., 2012), albeit less pronounced. A summer SAM reconstruction over the last millennium by Dätwyler et al. (2018) indicates that recent trends in the SAM during summer lie outside the 5-95% range of

pre-industrial natural variability. Using vegetation and fire activity records from Tasmania and southern South America, Fletcher et al. (2018) found the period 1000 to 1500 CE to be characterized by anomalously dry (i.e. positive SAM) conditions, while enhanced westerly wind activity was found for 0-1000 CE, as reflected in increased burning activity in Patagonia (Turney et al., 2016a) and tree records from Campbell and Auckland Islands (Turney et al., 2016b).

Over the last Century, various reanalysis and observation-based datasets (ERA-20C, HadSLP2r and 20CR) disagree in the representation of the SAM for 1900-1955, both in temporal variability and in amplitude (Lee et al., 2018), thus hinting at the impacts of limited underlying data. Over 1871-1950 the annual mean SAM index shows pronounced interannual and decadal time scale variability but no significant secular trends. In contrast after 1950 there is a clear trend towards positive values (Swart et al., 2015). The largest 30- and 50-year trends in the annual SAM index occurred at the end of the 20th century, indicating that the recent increase in the SAM is highly unusual in the context of the past three centuries (Yang and Xiao, 2018). This recent trend towards a positive phase of the SAM is consistent among different reanalyses and proxy records.

Considering the last 40 to 60 years, the seasonal trends in the SAM index based on HadSLP2r and 20CR reanalyses are positive for all seasons, being significant during Southern Hemisphere summer and autumn (Cerrone and Fusco, 2018; Swart et al., 2015). However, Marshall et al. (2013) described a reversal in the relationship between the SAM and East Antarctic temperatures during austral summer/autumn in the first decade of the 21st century, due to regional variability of the pressure pattern associated with the zonal wavenumber 3. Other studies have analysed the temporal variability in the interactions between the SAM and ENSO at different timescales (e.g. Clem et al., 2016; Yu et al., 2015). Reliable instrumental records used for the calculation of the SAM index extend back to 1957. Thus, trends are only calculated for the 1960-2018 and 1980-2018 periods, and are positive and significant, particularly for the latter period.

In summary, over the instrumental period there has been a robust positive trend in the SAM index, particularly since 1970 (*high confidence*). It is *likely* that the recent trend in the SAM is unprecedented in the past several centuries, although *medium confidence* arises due to the differences between reanalyses and proxy records before the 1950s.

2.4.6 Atlantic Meridional and Zonal Modes

AR5 reported no significant warming trend in the Atlantic Meridional Mode (AMM) during the twentieth century or shorter periods thereof. For the Atlantic Niño (Atl3), also referred as the zonal mode, the AR5 reported significant warming trends during the 1950–2012 period.

The Atl3 dominates climate variability in the tropical Atlantic which exhibits a zonal pattern similar to El Niño in the equatorial Pacific (Zebiak, 1993). The SST anomalies are associated with fluctuations of the southeasterly trade winds and changes in atmospheric convection and precipitation over the adjacent continental areas. In particular, the Atl3 has been linked to significant climate anomalies over West Africa (Rodríguez-Fonseca et al., 2015). The Atl3 peaks during the boreal summer and is typically associated with intensified upwelling and development of the cold water tongue governed by the ocean dynamics described by the Bjerknes feedback (Keenlyside and Latif, 2007; Lübbecke et al., 2018; Zebiak, 1993), intrinsic dynamics of deep equatorial jets (Brandt et al., 2011) and meridional temperature advection (Richter et al., 2012). An alternative hypothesis is that the SST anomalies are controlled by stochastic atmospheric driven heat flux changes (Nnamchi et al., 2015). The Atl3 mode according to a broad range of SST products (Figure 2.42) has been increasingly positive over the instrumental period as a whole.

The AMM exhibits an inter-hemispheric spatial structure of SST anomalies linked to cross-equatorial wind flow from the cooler to the warmer hemisphere on interannual to decadal timescales (Amaya et

al., 2017). This mode peaks in boreal summer and is governed by the Wind-Evaporation-SST (WES) feedback which drives anomalous meridional SST gradients. Because the tropical Atlantic has warmed substantially since the twentieth century (Deser et al., 2010; Tokinaga and Xie, 2011), the AMM index (the difference between SST north and south of the equator) does not exhibit significant trends. On the other hand, the basin-wide warming is reflected in significant long-term Atl3 upward trend, weakening to insignificant values after the late 1970s which may point to some low frequency modulation of the dynamically active cold tongue region (Servain et al., 2014).

In summary, it is *likely* that the Atl3 index has exhibited a significant positive trend since the beginning of instrumental period, but this trend has become weaker since the 1970s. For the AMM, no significant trend was identified.

2.5 Gaps in knowledge

- Significant uncertainties remain in reconstructions of natural drivers. For solar radiation, in particular prior to satellite measurements, the magnitude of total irradiance fluctuations, variability in spectral characteristics and upper atmosphere impacts of energetic particles are not well known. The radiative effects of large volcanic eruptions are highly uncertain. Further analysis of sulphate and sulphur isotope in ice cores, as well modelling, can provide further information on the location, season, and plume height of eruptions.
- Uncertainties remain in both the magnitude and rates of change of atmospheric CO₂ concentration through the last 500 million years. Better characterisation of rapid climate events, like the PETM, may provide additional context for climates warmer than now.
- The lack of knowledge of LLGHGs other than CO₂ prior to the last 800,000 years precludes longer-term assessment of their evolution. Offsets between different ice core records of CO₂ and N₂O hamper an accurate definition of timings and magnitudes of the GHG changes, such as the exact timing of the onset of the N₂O increase during the last 200 years. Acquiring glacier ice older than 1 Ma would extend understanding through the transition from 40,000 yr to 100,000 yr glacial-interglacial periodicity, which occurred during the Mid Pleistocene Transition (0.8-1.0 Ma).
- There is a need to better constrain the drivers of emission variations and budgets of LLGHGs. There is a large uncertainty in the relative contributions of biogenic and fossil-fuel related CH₄ emissions from different sources, and also the drivers for the increase in atmospheric burden starting in 2007. Improved use of modern instrumentation, including improved uncertainty estimation may help to close this gap.
- Sparse geographical coverage and limited time series of in-situ observations limit our confidence in large scale concentration changes of short-lived climate forcers, in particular prior to the satellite era. Due to the lack of global-scale observations of pre-industrial levels, current estimates of radiative forcing rely largely on models. Improvements in the consistency of satellite retrievals (e.g. tropospheric ozone) and careful re-assessment of observations (including ice-core data for specific components), targeted at model evaluation, can further reduce uncertainties. For quantification of effective forcings, it is imperative to better constrain cloud adjustments, requiring improved homogenisation of time series from ground-based networks and satellites.
- Significant uncertainty remains in the depiction of historical changes in global land cover, particularly in the pre-satellite era. In addition, hydrological and other feedbacks resulting from land use change have highly uncertain, but potentially large, impacts on the effective radiative forcing.
- A lack of robust observationally-based estimates of the differences between GMST and GSAT, and the present paucity of GSAT based estimates of long-term change serve to reduce confidence in estimation of GSAT changes. Early MAT observations are more prevalent than SST observations, so analyses of GSAT may also permit creation of more spatio-temporally complete 19th Century temperature estimates.

- 1 • There remain global-scale differences between available GMST analyses that often exceed their
2 quantified parametric uncertainty estimates. Differences are most substantial in the early data
3 sparse period when many non-standardised instruments and observing practices existed. Further
4 focus on the early period and, for marine data, the period around WWII, may yield additional
5 benefits. Improvements in such data may also improve understanding of modes of variability.
- 6 • The ability to place post-industrial warming in context of centennial variability over the Holocene
7 is limited due to the increase in uncertainty with time for large-scale temperature reconstructions.
8 Efforts to recover and analyse more proxy sources with decadal or better resolution that extend
9 through the Holocene, would reduce uncertainty.
- 10 • Many early instrumental data and other observations across a broad range of climate relevant
11 parameters currently exist in hard copy or image format only which, if digitized, could help to
12 improve our confidence in pre-industrial and early industrial era change across a broad suite of
13 indicators. Conservative estimates (Allan et al., 2011) suggest that globally as much data pre-
14 1950 as is currently digitised remains to be rescued.
- 15 • There remains uncertainty as to whether there has been long-term amplification of boundary layer
16 temperature changes throughout the tropical troposphere. There remains *low confidence* in
17 changes prior to the start of the 21st Century with the advent of GNSS-RO measurements because
18 of limited radiosonde observations, satellite observations which lack vertical resolution, and
19 biases in reanalysis data sets associated with the assimilation of aircraft observations.
- 20 • Assessment of many aspects of cryospheric change is limited by the paucity of available
21 observations for key aspects. For example, there exists limited observational data on sea ice and
22 snow thickness from the Arctic Basin for the winter season (Gerland et al., 2019) and Arctic
23 permafrost conditions. A significant gap in this context is the lack of reliable basin-scale snow
24 depth observations on sea ice (Webster et al., 2018).
- 25 • Understanding of changes in ocean circulation is hampered by short and sparse timeseries.
26 Confidence in the observed changes in circulation in response to climatic change would improve
27 with increased observations in both time and space. Reconstructed estimates of ocean circulation
28 offer a window on the past but more work is needed to increase confidence in these proxies.
- 29 • Although species range shifts have been widely documented for many global land
30 areas (particularly North America and western Eurasia), there have been relatively few studies
31 (and thus uncertainty is greater) for central Africa, eastern Asia, South America, Greenland, and
32 Antarctica.
- 33 • While there have *likely* been increases in terrestrial vegetation in many areas since the 1980s,
34 there is significant spread (and thus uncertainty) in the satellite-derived estimates of such changes,
35 and there is a paucity of in situ analyses for corroboration.
- 36 • Changes in growing season have been established in the extratropical Northern Hemisphere since
37 the mid-20th century; however, changes in other areas and periods have not been widely assessed,
38 and there are currently no widely available and reliable sources of proxy information that permit a
39 longer-term context.
- 40 • Assessment of pre-instrumental evidence for the Pacific Decadal Variability relies mostly on
41 paleoclimate reconstructions from the Northern Hemisphere, with *limited evidence* for the South
42 Pacific Ocean and, thus, changes of the Pacific Decadal Variability over the last millennium have
43 *low confidence*.
- 44 • The overall paucity of highly resolved proxy records that can be interpreted in terms of specific
45 climate variables, along with assumptions about the stationarity of spatial patterns of variability,
46 hampers the ability to draw robust conclusions about the behaviour of modes of variability prior
47 to the instrumental record, with increasing uncertainties further back in time.

Frequently Asked Questions

FAQ 2.1: The Earth's temperature has varied before. How is the current warming any different?

Summary: *Climate is always changing but the rate and global synchronicity of recent warming appears to be without precedent in geological history. By looking into the past, we gain a long-term perspective on the unusualness of recent climate changes. Among the most prominent climate indicators is the Earth's surface temperature. The recent warming has reversed the long-term natural cooling trend and global temperature is likely higher now than it has been for millennia.*

A rich variety of geological evidence shows that climate has changed throughout Earth's history. Through information gathered from tree rings, or ocean and lake sediments, for example, we can see there were times when the planet was cooler, and times when it was warmer. While climate can be described by many variables, the temperature of the planet's surface is a basic indicator of climate and is fundamental to understanding climate change.

Inferring temperature changes of the past is challenging. Our confidence in quantifying large-scale temperature reconstructions generally decreases back in time, especially beyond the last millennium when the number of temperature records calculated from tree rings declines sharply. Despite this uncertainty, scientists can still identify several differences between recent warming and that of the distant past.

Firstly, it's currently warming when natural drivers of climate are acting to cool the Earth. Temperature over the Northern Hemisphere, and possibly both hemispheres, was decreasing over thousands of years. The cooling trend was reversed by the end of the 19th Century, and warming prevailed throughout the 20th Century. During recent decades, temperatures have continued to rise despite a slight downturn in the output of the Sun, which is known to alter temperatures over short time frames (~10 years).

It's been a very long time since global temperature has been this high. The past 50 years are probably the warmest over the last 1400 years, and possibly much longer. Although there is ongoing discussion on whether the world was warmer before the long cooling trend (that begun ~6000 years ago), the last time global temperature was similar to today could well have been during the peak of the last interglaciation - about 125,000 years ago - when global temperatures might have been up to a degree higher than present warming and sea level was about 5 m higher than now.

It's warming quickly. Most of the rapid temperature shifts in Earth's climate history have occurred as ice sheets collapsed, giving way to interglacial periods like the present (Holocene Epoch). The shift from glacial to interglacial climate corresponded to a temperature change of about 5°C and took about 5000 years. Within this transition, the maximum temperature change during any thousand-year period was likely about 1.5°C. Compare this to the present temperature increase of 1°C in just 170 years. The shift from the mid-19th Century to date is likely the single most extreme temperature shift since the last deglaciation.

Finally, it's warming almost everywhere. When temperatures changed in the past, some regions warmed while others did not. For example, the North Atlantic region warmed more than other places during the Medieval Climate Anomaly. While the pattern of recent warming shows similarities to past warmings – continents have warmed more than the ocean, and the Arctic more than the tropics – its footprint is more global than has been reconstructed for any previous rapid global change.

Not only is warming unprecedented, it is occurring at a time when human activity creates stresses that compound the effects of warming. So much that some scientists coin this era to be the beginning of a new geological period: the Anthropocene, a time where human activity is the most dominant

influence on the world.

[START FAQ 2.1, FIGURE 1 HERE]

FAQ 2.1, Figure 1: Approximate average global surface temperature (y-axis) over the past two million years, combined from multiple sources and shown at five different time scales (x-axis). Temperature is relative to 1850-1900 (pre-industrial reference); 2,000,000 to 300,000 years ago and 200,000 to 30,000 years ago from marine oxygen isotopes; 20,000 to 10,000 years ago from multi-proxy records; 10,000 to 3,000 years ago from multi-proxy records; 2,000 to 300 years ago from multi-proxy records, mainly tree rings; 1800 to 2015 CE from instrumental temperature records.

[END FAQ 2.1, FIGURE 1 HERE]

FAQ 2.2: What is the evidence for climate change?

Summary: *The evidence for climate change rests on more than just increasing surface temperatures. Evidence abounds from a broad range of indicators that collectively leads to the inescapable conclusion that we are witnessing rapid changes to our global climate. We are seeing changes in the atmosphere and oceans, in our glaciers and ice-regions, and throughout the biosphere. Our scientific understanding depicts a coherent picture of a warming world from the depths of the oceans to high in the atmosphere and at all points in between that has occurred since the late 19th century.*

We have long observed our changing climate. From the earliest amateur scientists taking observations in the 16th and 17th Centuries to present we have seen a revolution in our ability to observe and diagnose our changing climate. We truly live in a golden age of earth observation capabilities. Today we can observe diverse aspects of our climate system from space, from aircraft and weather balloons, using a range of ground-based observing technologies, and using instruments that can measure to great depths in the oceans.

Observed changes across a diverse range of key indicators of the climate system clearly point to a climate that has warmed since the industrial revolution. Global mean land and ocean surface temperature has increased since the late 19th century, with each of the last four decades being warmer than any decade that preceded it. Change is not limited to the surface. The temperature of the lowest few kilometers of the atmosphere has risen, with associated increases in atmospheric water vapor and rainfall. Aspects of atmospheric circulation have evolved as well, including a widening of the tropical Hadley Circulation, an expansion in the subtropical dry zone and a poleward shift of mid-latitude storm tracks. Change has also transmitted to the depths of the global oceans. The heat content of the global oceans has increased since the 1970s, with more than 90% of the excess energy gained in the climate system being stored in the oceans. This warming has caused ocean water itself to expand, which in turn has contributed to the observed increase in global sea level in the past century.

Significant changes are also evident over the Cryosphere, the portion of the Earth where water is seasonally or continuously in solid form. There have been decreases in Arctic sea ice extent and thickness since the mid-1970s, with the Antarctic sea ice extent remaining quasi-stable. Spring snow cover in the Northern Hemisphere has fallen since the late-1970s. In addition, there is an observed warming of permafrost and evidence of thawing in many permafrost regions. The planet's two largest ice sheets (Greenland and Antarctica) are shrinking, as are the vast majority of glaciers worldwide, further contributing to the observed sea level rise. Confidence in our understanding of melting glaciers and ice sheets and increasing ocean temperatures is increased because when their contributions to sea-level rise are added together they match very well the directly observed sea-level changes. Such closure highlights robust understanding.

Many aspects of the biosphere are also changing. Over the last century, long-term ecological surveys encompassing mammals, birds, invertebrates, and plants show that rates of change are increasing across a broad array of ecosystems. A broad array of species have generally moved poleward and upslope. There have been general increases in vegetation activity and extent (i.e., global greening) since the early 1980s, and the length of the growing season has increased over much of the extratropical Northern Hemisphere since at least the mid-20th century. The global ocean has become more acidic as a result of increased carbon dioxide concentrations, and there have also been changes in marine ecosystems.

Change is ubiquitous across many components of the climate system. It has been observed using a very broad range of techniques and analysed independently by numerous groups around the world. The changes are consistent in pointing to a climate system that has undergone rapid warming since the industrial revolution.

1 **[START FAQ 2.2, FIGURE 1 HERE]**

2
3 **FAQ 2.2, Figure 1:** Independent analyses of many components of the climate system that would be expected to
4 change in a warming world exhibit trends consistent with warming.
5

6 **[START FAQ 2.1, FIGURE 1 HERE]**

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Figures

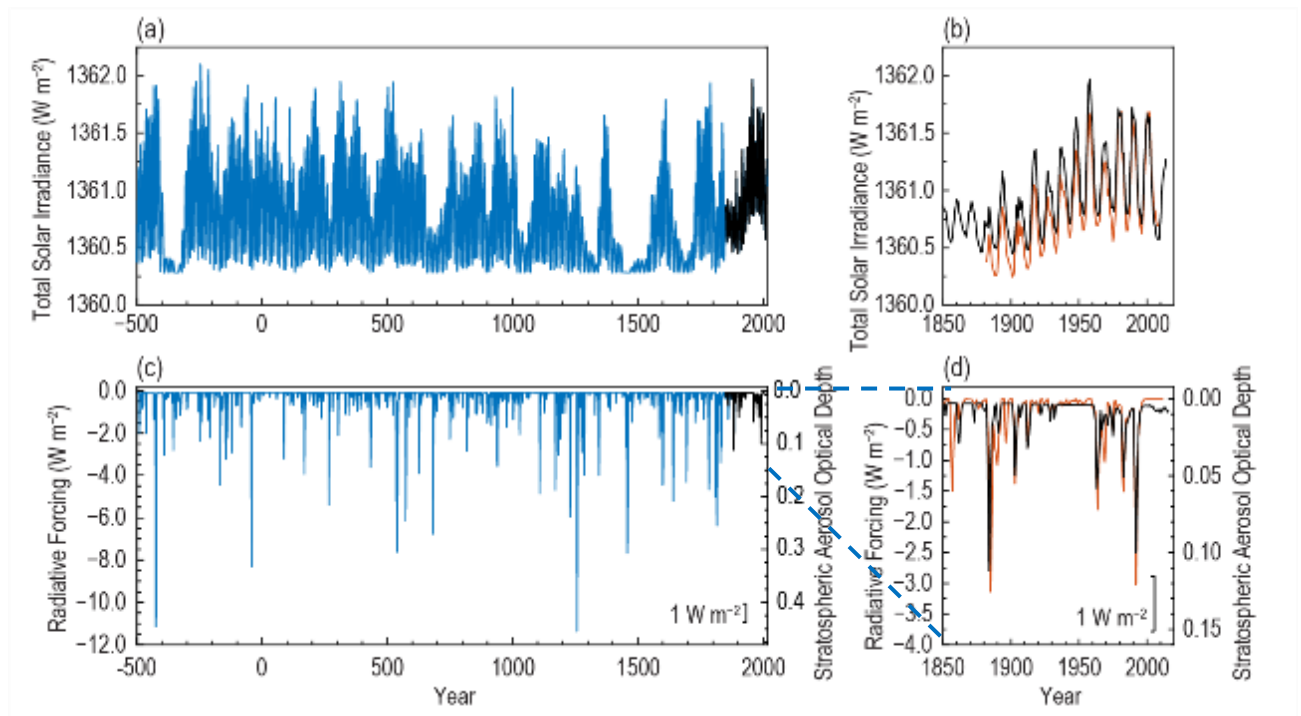


Figure 2.1: Time series of solar and volcanic forcing for two periods: (a, c) past 2500 years, and (b,d) past 165 years. (a) Total solar irradiance (TSI) reconstruction recommend for CMIP6/PMIP4 millennial experiments based on the radiocarbon dataset before 1850 (blue) scaled to the CMIP6 historical forcing after 1850 (black) (Jungclaus et al., 2017). (b) TSI time series (6-month running averages) from CMIP6 historical forcing as inferred from sunspot numbers (black) (Matthes et al., 2017), with CMIP5 forcing based on (Lean, 2000) and (Wang et al., 2005) for comparison (red). (c) Volcanic forcing represented as reconstructed stratospheric aerosol optical depth (SAOD) at 550 nm. Estimates covering 500 BCE to 1900 CE (blue) are from Toohey and Sigl (2017) and those for the period 1850-2015 (black) are from (Luo et al., 2018). SAOD was translated to global volcanic forcing based on the scaling of $RF = -25 \times SAOD$, as in AR5, which differs somewhat from the formulation for effective radiative forcing presented in Section 7.3.4.4. (d) SAOD reconstruction from CMIP6 (v 4) (black) (Luo et al., 2018), with CMIP5 forcing for comparison (red) (Sato et al., 1993). Note the change in y-axis range between panels c and d.

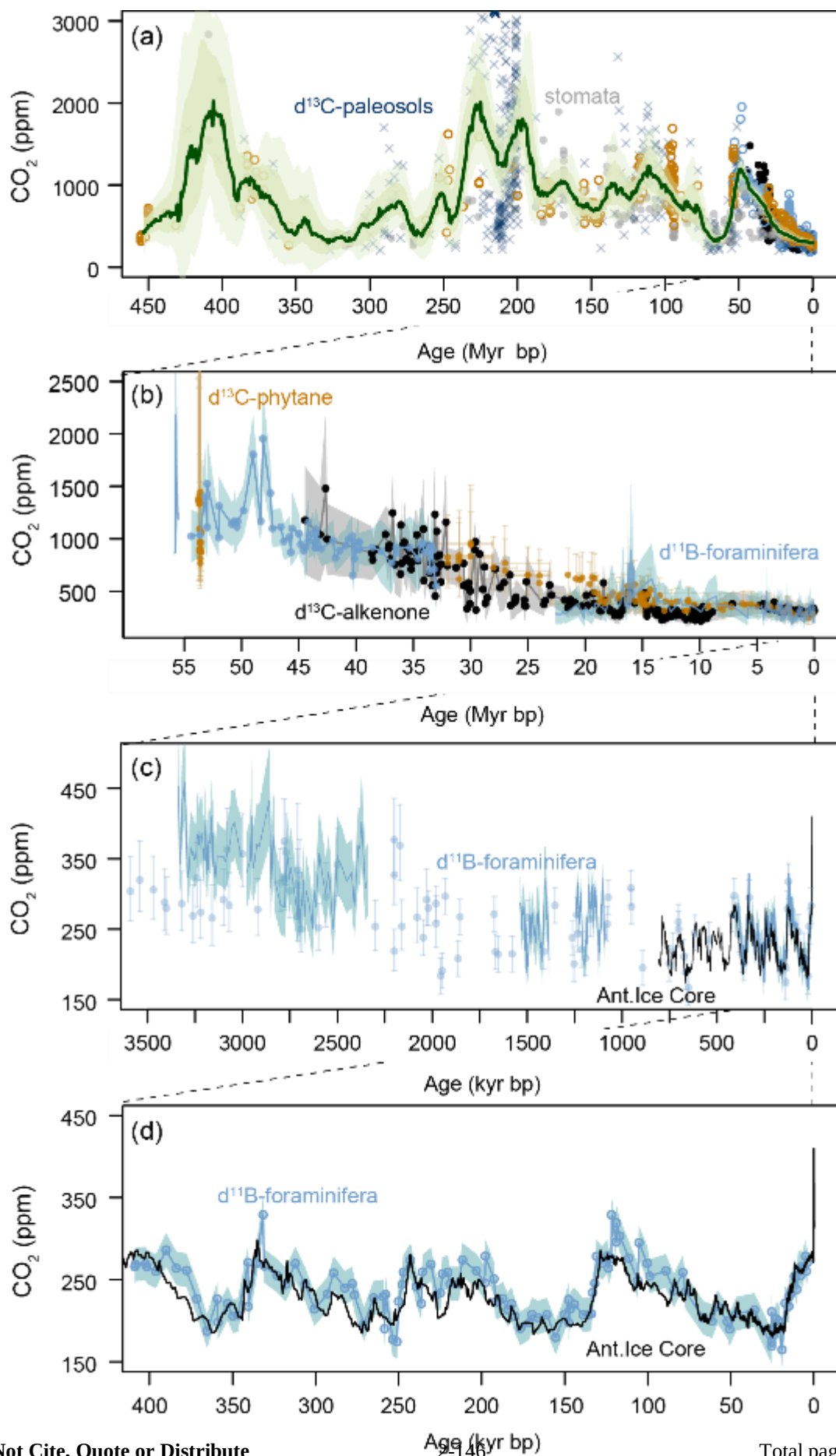


Figure 2.2: The evolution of atmospheric CO₂ through the last 450 million years reconstructed from ice core and marine sediment records. (a) 0 to 450 million years ago, data from (Foster et al., 2017) and references therein (and as detailed in subsequent panels). Stomatal based estimates in grey circles, paleosol carbonate $\delta^{13}\text{C}$ based estimates as blue crosses, $\delta^{13}\text{C}$ of phytane as brown circles. Dark green line is non-parametric LOESS fit (see Foster et al., 2017) through the data presented here, red band is uncertainty envelope at 68% uncertainty. See (Royer et al., 2001; Witkowski et al., 2018) for a description of the basis of these CO₂ proxies. (b) 0 to 58 million years ago, boron isotope based CO₂ in blue (Anagnostou et al., 2016, 2019; Sosdian et al., 2018), $\delta^{13}\text{C}$ of phytane in brown (Witkowski et al., 2018), $\delta^{13}\text{C}$ of alkenones in black (Pagani et al., 2005, 2011; Super et al., 2018). The basics of these CO₂ proxies are described in (Foster and Rae, 2016a; Pagani, 2002; Witkowski et al., 2018). (c) 0 to 3500 thousand years ago, ice core data shown in black (Bereiter et al., 2015) and boron isotope based CO₂ in blue (Chalk et al., 2017a; de la Vega et al., 2019; Hönisch et al., 2009; Martínez-Botí et al., 2015a; Sosdian et al., 2018) (d) 0 to 450 thousand years ago, ice core data shown in black (Bereiter et al., 2015) and boron isotope based CO₂ in blue (Chalk et al., 2017a; de la Vega et al., 2019).

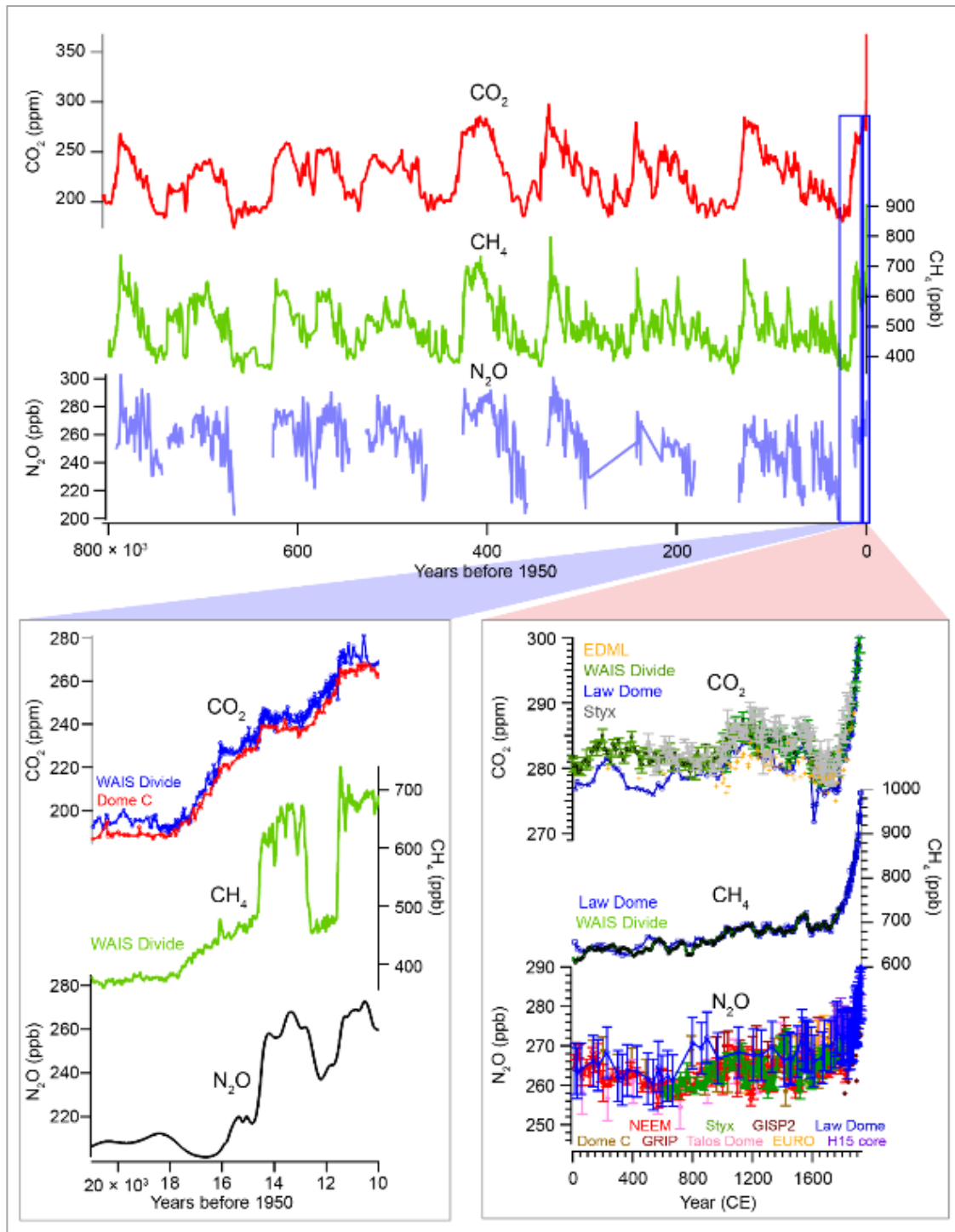


Figure 2.3: Atmospheric greenhouse gas concentrations from ice cores. a) Low-resolution records during the last 800ka (Bereiter et al., 2015; Loulergue et al., 2008; Lüthi et al., 2008; Schilt et al., 2010). b) High-resolution records during the last glacial termination (Köhler et al., 2017; Rhodes et al., 2015, 2017; WAIS Divide Project Members., 2015). c) Multiple high-resolution records over the common era. CO₂ data from (SIEGENTHALER et al., 2005; MacFarling Meure et al., 2006; Lüthi et al., 2008; Ahn et al., 2012; Ahn et al., in prep.). Antarctic ice core CH₄ records from (MacFarling Meure et al., 2006; Mitchell et al., 2013b); N₂O concentrations from Greenland and Antarctic ice cores (Machida et al., 1995; Sowers, 2001; Flückiger et al., 2002; MacFarling Meure et al., 2006; Schilt et al., 2010; Ryu et al., 2019b).

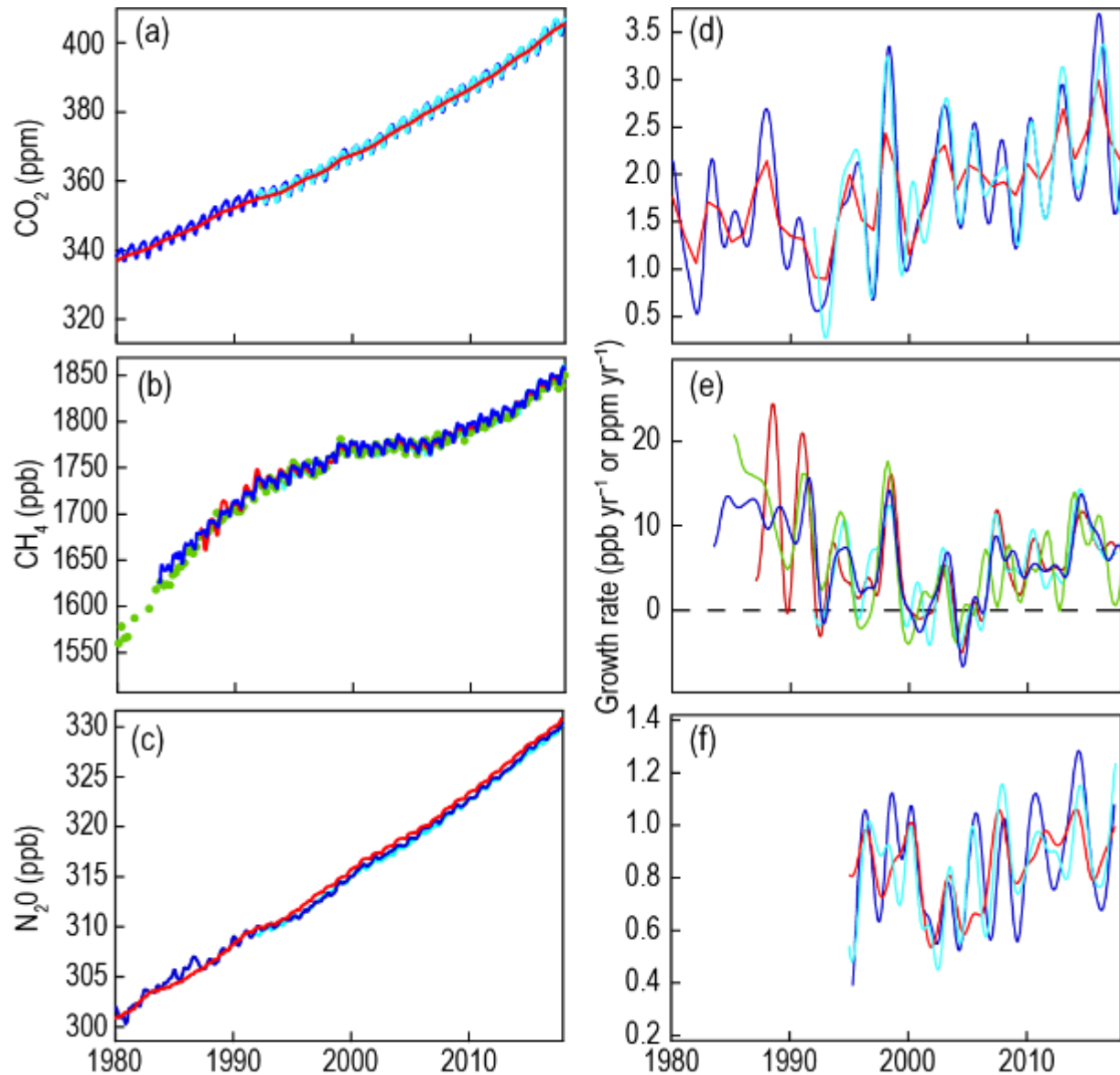


Figure 2.4: Globally averaged dry-air mole fractions of (a) CO₂ from Scripps Institution of Oceanography (SIO) at annual time resolution based on measurements from Mauna Loa, Hawaii and South Pole (orange), CSIRO at monthly time resolution (cyan), and NOAA/ESRL/GMD at quasi-weekly time resolution (blue). SIO values are deseasonalized. (b) CH₄ from NOAA (blue), AGAGE (red), CSIRO (cyan), and UCI (green); and (c) N₂O from NOAA (blue), AGAGE (red), and CSIRO (cyan). Annual growth rates, calculated as the time derivative of the deseasonalized global averages (Dlugokencky et al., 1994) for globally averaged atmospheric mole fractions are shown in corresponding right panels, (d) CO₂, (e) CH₄, and (f) N₂O using the same color code as in (a, b, c). Note that the annual growth rate of CO₂ was consistently below 2 ppm yr⁻¹ between 1960 and 1980 (not shown). Insufficient and noisy data prevent calculation of growth rates prior to 1995. For abbreviations see Table 2.2.

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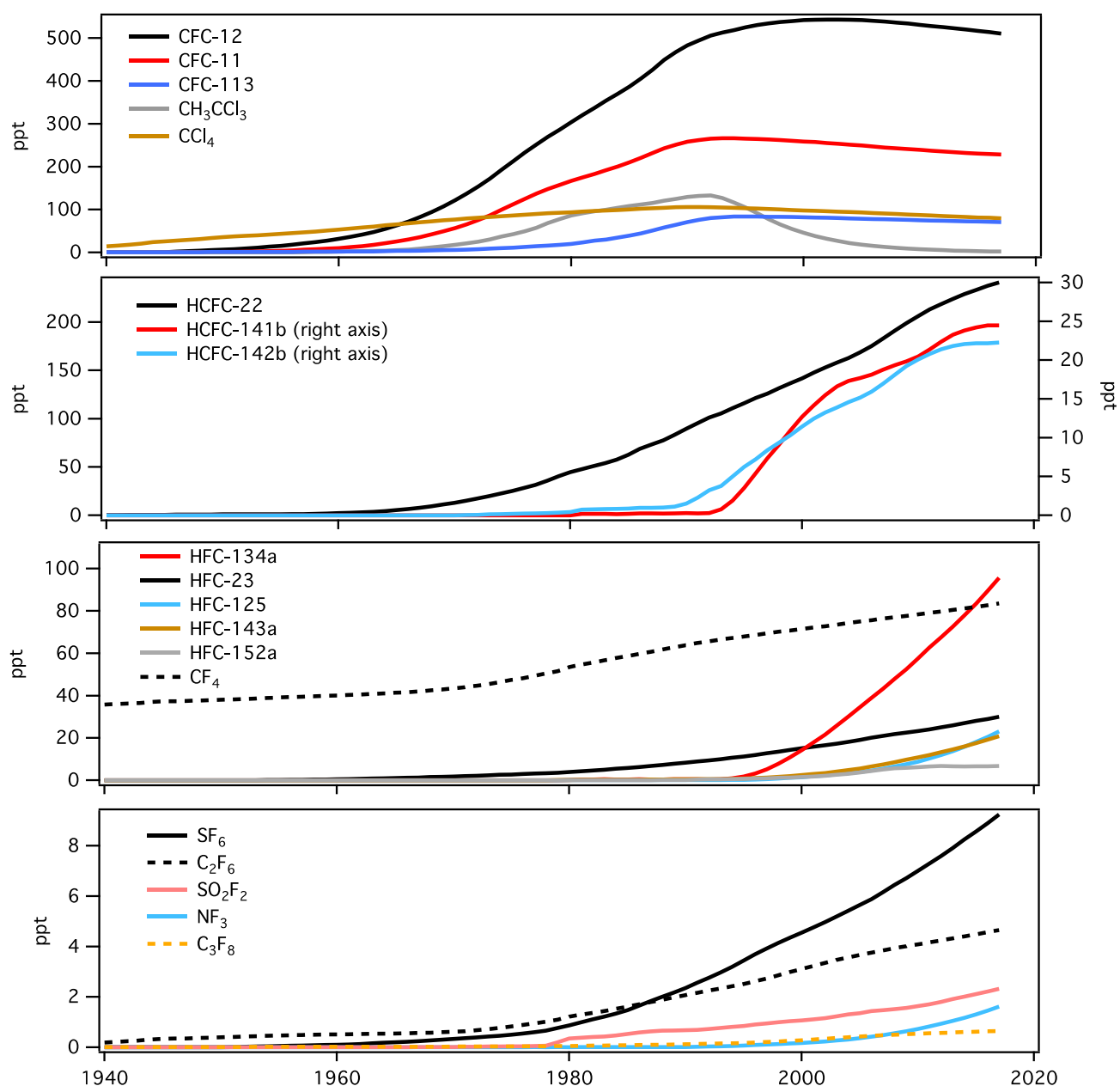
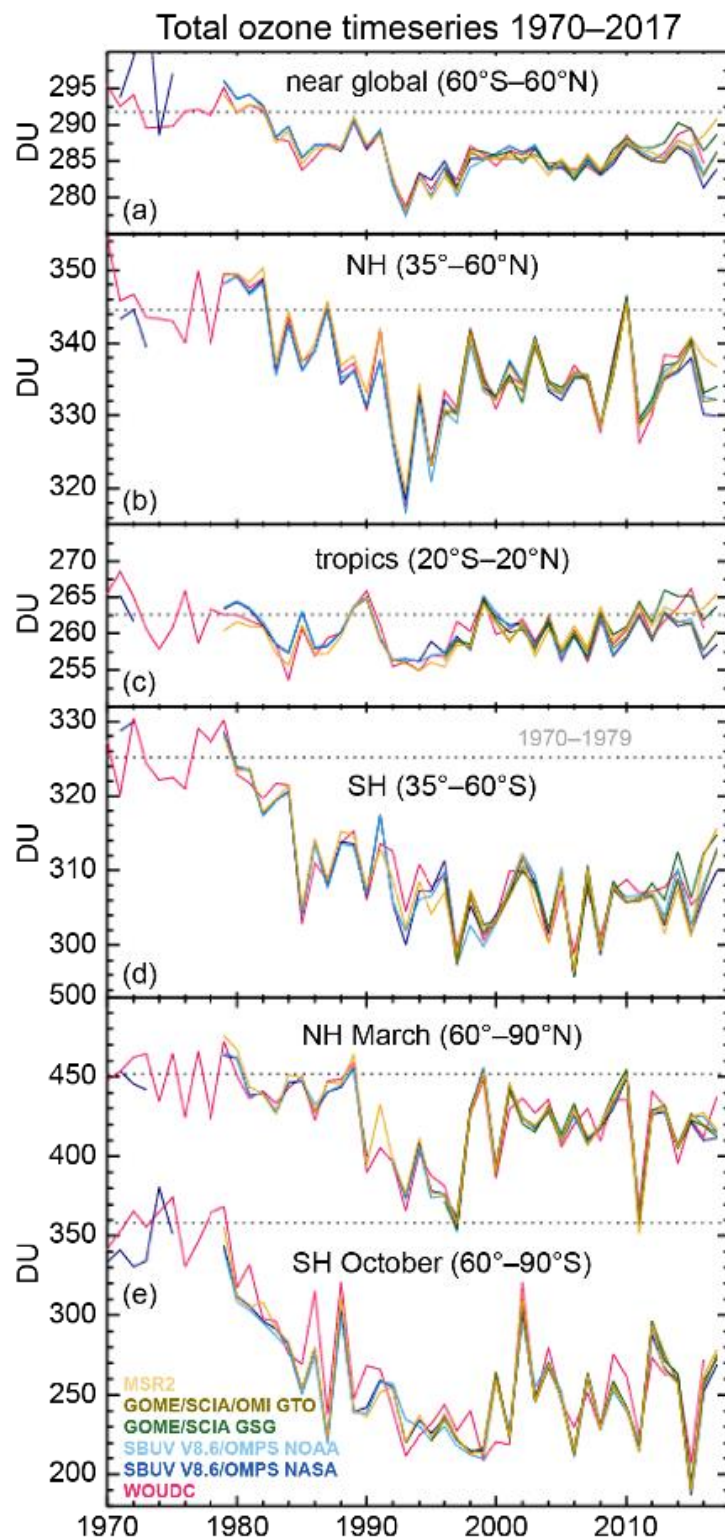


Figure 2.5: Global mean atmospheric mole fractions of select ozone-depleting substances and other greenhouse gases, based on the CMIP6 historical dataset (Meinshausen et al., 2017) and data from NOAA and AGAGE global networks. See Table 2.2 for information on lifetime and radiative effectiveness.



[Note: Panel (a) line SBUV V8.6/OMPS NASA is clipped at the top.
 Author contacted but not able to resolve in time for FOD submission.]

Figure 2.6: Time-series of annual mean total ozone (a good proxy for stratospheric ozone) from 1970–2017 in four zonal bands (a–d) and polar (60°–90°) total ozone in March (NH) and October (SH) (e). March (NH) and October (SH) are the months when polar ozone losses usually are largest. The values are given in Dobson units, where 100DU correspond to a 1mm thick slab of pure ozone gas at standard conditions. Adopted from Weber et al (2018).

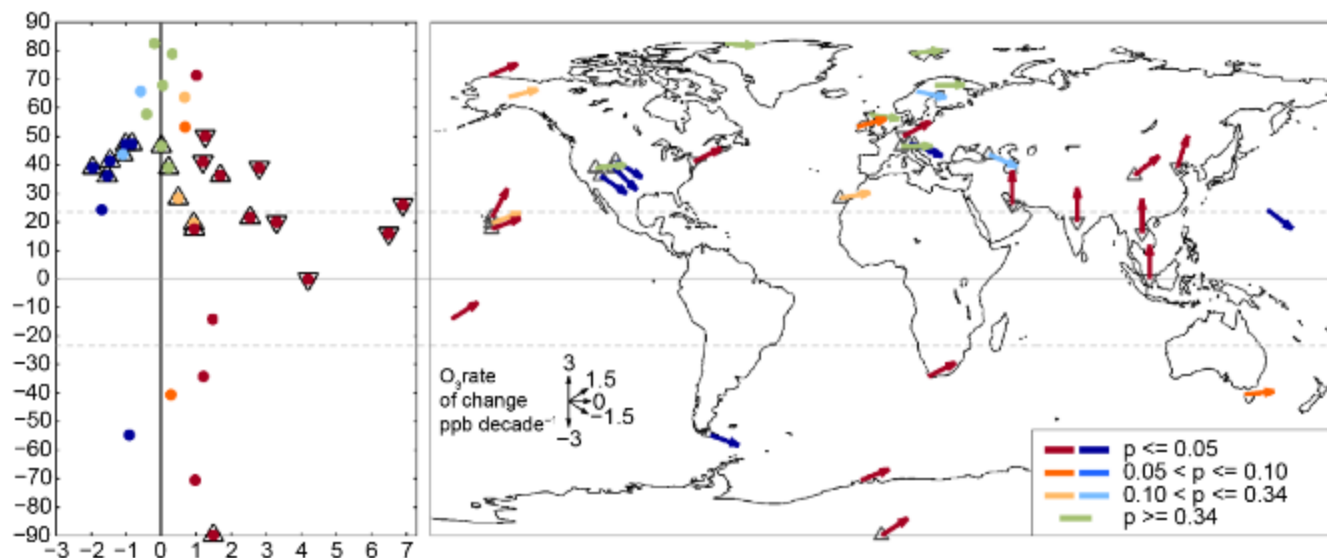


Figure 2.7: Decadal ozone trends based on monthly anomalies (relative to 1995-2010) at 28 remote surface sites, plus IAGOS aircraft trends at 650 hPa (∇) above Europe the northeastern USA, NE China, SE Asia, southern India, the Persian Gulf and Malaysia/Indonesia (as reported by (Gaudel et al., 2018). High elevation sites (>1500 m a.s.l.) are indicated by (Δ). Sites below 1500 m a.s.l. are indicated by unaccompanied circles or arrows. All surface trends end with the most recently available year, but begin in 1995. Mauna Loa, Hawaii trends are shown for dry mid-latitude (top arrow), all (middle arrow) and moist tropical (bottom arrow) air masses. [Note a known need to tidy up the title to the graphic]

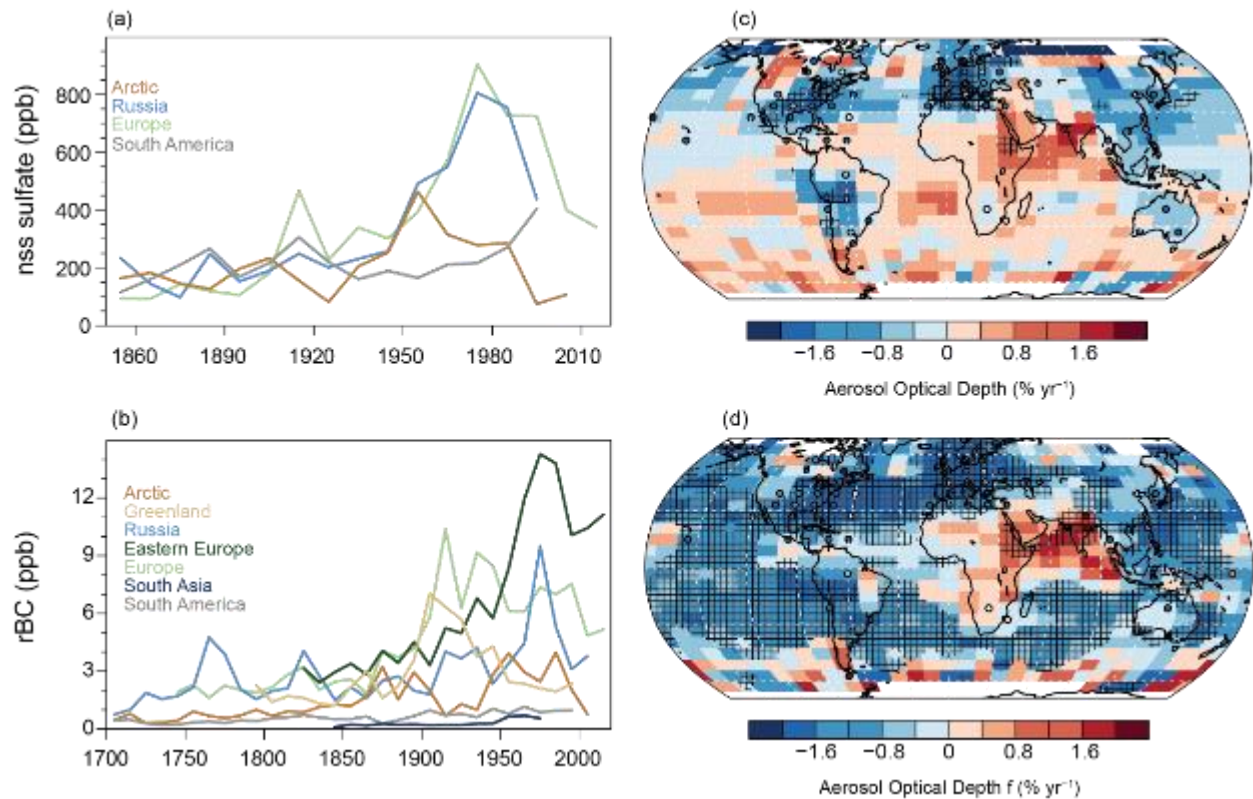


Figure 2.8: Ice core concentrations of (a) sulphate (ng g⁻¹) and (b) black carbon (ng g⁻¹) from Europe (reproduced from (Engardt et al., 2017; Sigl et al., 2018), Russia (reproduced from Oliver et al. (2006), Arctic (Svalbard, reproduced from Wendl et al., 2015 and Osmont et al., 2018), South America (ex-sulphate, corrected for mineral dust input, reproduced from Kellerhals et al. (2010), and BC from (Osmont et al., 2019), Eastern Europe (reproduced from Lim et al. 2017), and Greenland (stacked rBC record from Greenland including the four ice cores NEEM-2011-S1, D4, TUNU2013), and Summit2010 (Keegan et al., 2014; McConnell et al., 2007; Mernild et al., 2015; Sigl et al., 2013, 2015) reproduced from Osmont et al. (2019) and Sigl et al. (2018). Given are 10-year averages. Sulphate concentrations were not corrected for sea-salt input, which is negligible at the presented ice core locations. (c) geographical distribution of the linear trend in annual mean AOD retrieved from satellite data for the 2001 – 2017 period (% per year). The average trend from MODerate Imaging Spectroradiometer (MODIS; (Levy et al., 2010)) and Multi-Angle Imaging Spectroradiometer (MISR, (Garay et al., 2017)) is shown. Hatched areas show a trend that is significant at the 0.9 level (Kendall’s rank correlation test) according to both MISR and MODIS retrievals, with the two trends agreeing to within 50%. Superimposed are the trends in annual-mean AOD from AERONET surface sunphotometer network sites (Giles et al., 2019; Holben et al., 1998) for 2000 – 2018 where significant at the 0.9 level and where at least 10 years of data are available. (d) Trend in 2001-2016 AODf derived from MISR. Hatched areas show trends significant at the 0.9 level (Kendall’s rank correlation test), AERONET trends for 2000 – 2018 superimposed where statistically significant.

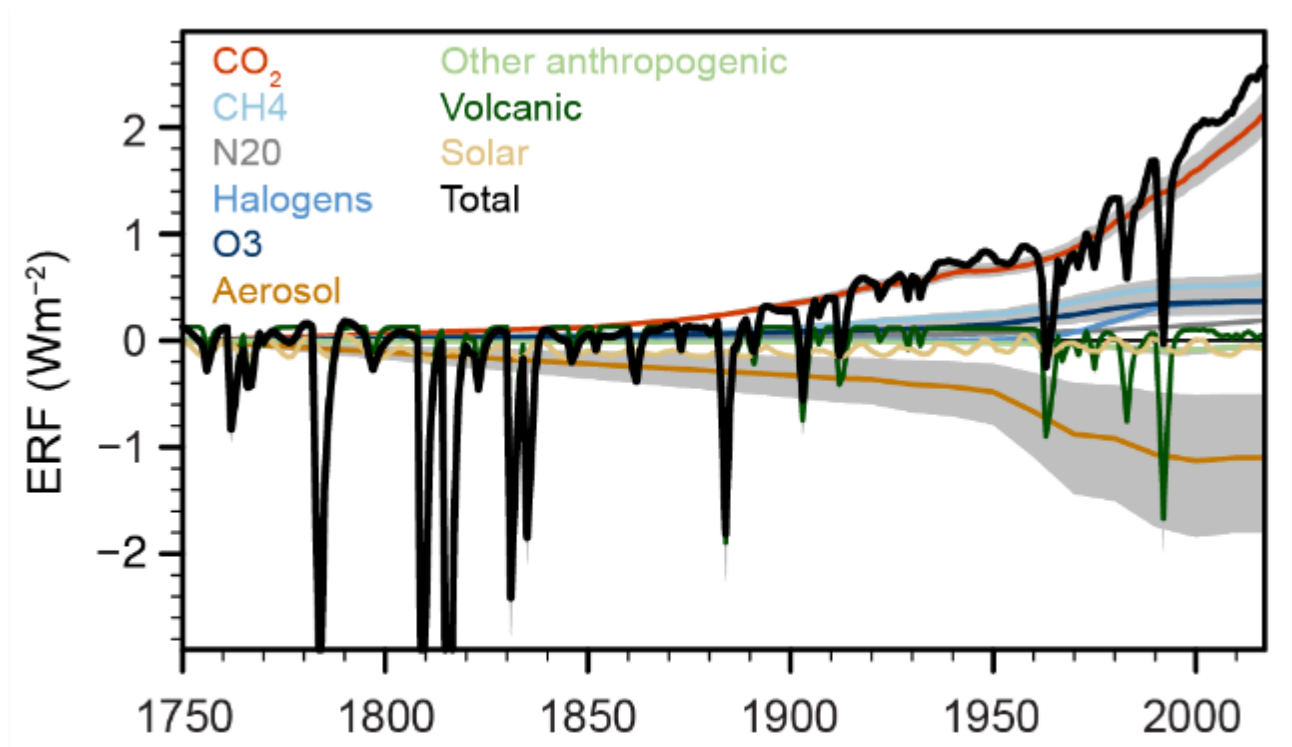


Figure 2.9: Changes in radiative forcing estimates related to the factors assessed in Section 2.2. effective radiative forcings are based upon the calculations described in Chapter 7. The global annual mean temporal evolution since 1850 is shown as the central assessment value. Gaseous changes to the atmospheric composition are shown for carbon dioxide (CO₂), methane (CH₄), nitrous oxide (N₂O), Halogens, and tropospheric ozone (O₃). Aerosol changes are the sum of the effective radiative forcing due to aerosol – radiation and aerosol – cloud interactions. Other anthropogenic forcings include stratospheric ozone, stratospheric water vapour, land use/land cover changes, black carbon deposition on snow, and contrails. The two other key forcings shown are variability in TSI and in volcanic forcing. Shading shows the 5-95% uncertainty range as assessed in Chapter 7 (Section 7.3). The sum of all forcings is shown as the total forcing.

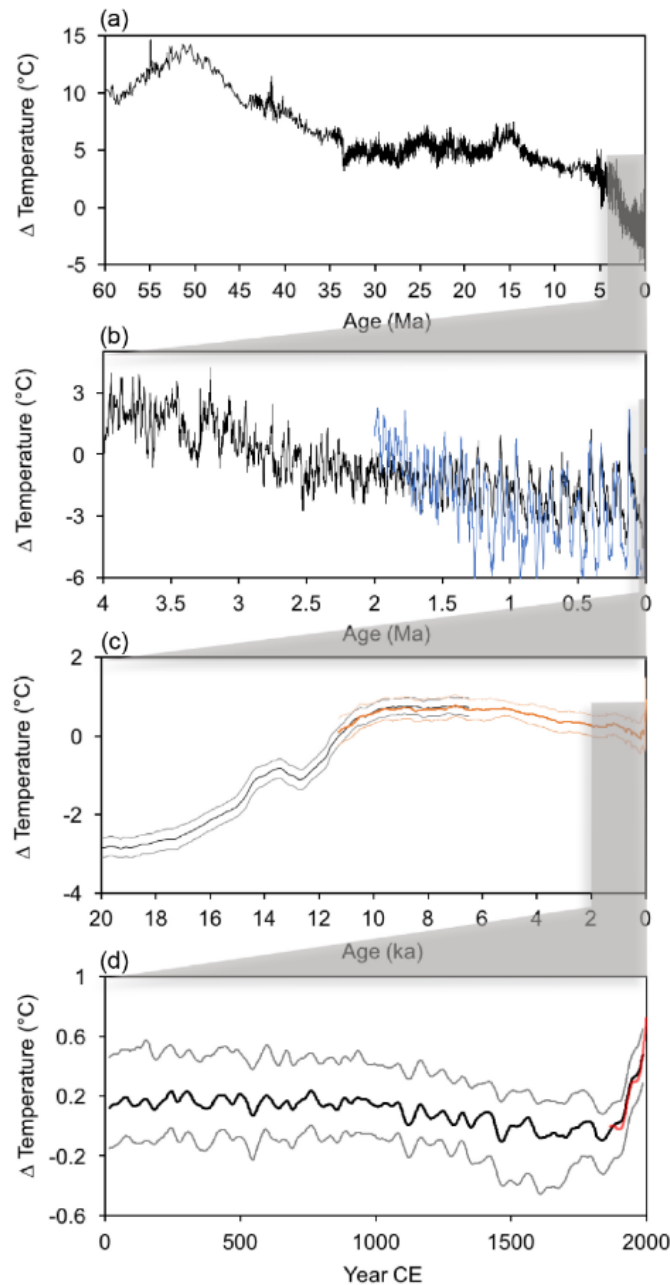


Figure 2.10: Global mean surface temperature reconstructed from various proxy records over four time scales. **(a)** 60Ma to present based on a scaling of deep-ocean temperature inferred from oxygen isotopes (grey line) (Hansen et al., 2013), with temperatures adjusted (Δ temperature = 0°C) relative to mean of last 1600 years (youngest data point). **(b)** 4Ma to present based on the same data in panel **a**, along with a composite of multi-proxy sea-surface temperature reconstructions (blue line) (Snyder, 2016) adjusted to the mean of the last 1000 years (youngest data point). Note that the temperature estimate for the last inter-glacial in these reconstructions ($+2^{\circ}\text{C}$ relative) is at the high end of the estimated range discussed in the text. **(c)** 20ka to present from the data syntheses of (Shakun et al., 2012) for the last deglaciation (black line) and (Marcott et al., 2013) for the Holocene (orange line), with 2.5% and 97.5% uncertainty bands. Temperatures were increased by 0.31°C from published values to adjust them to relative to pre-industrial (1850-1900), rather than relative to 1961-1990. **(d)** 2ka to present (aka, Common Era) relative to pre-industrial from a synthesis of seven different reconstruction methods (PAGES 2k Consortium, in revision) all based on the proxy data compilation of the (PAGES 2k Consortium et al., 2017), showing the median (solid line) and the 2.5th and 97.5th percentiles of the ensemble members (bands), smoothed with a 31-year Butterworth low-pass filter. Instrumental data (1850-2017) from (Cowtan and Way, 2014),

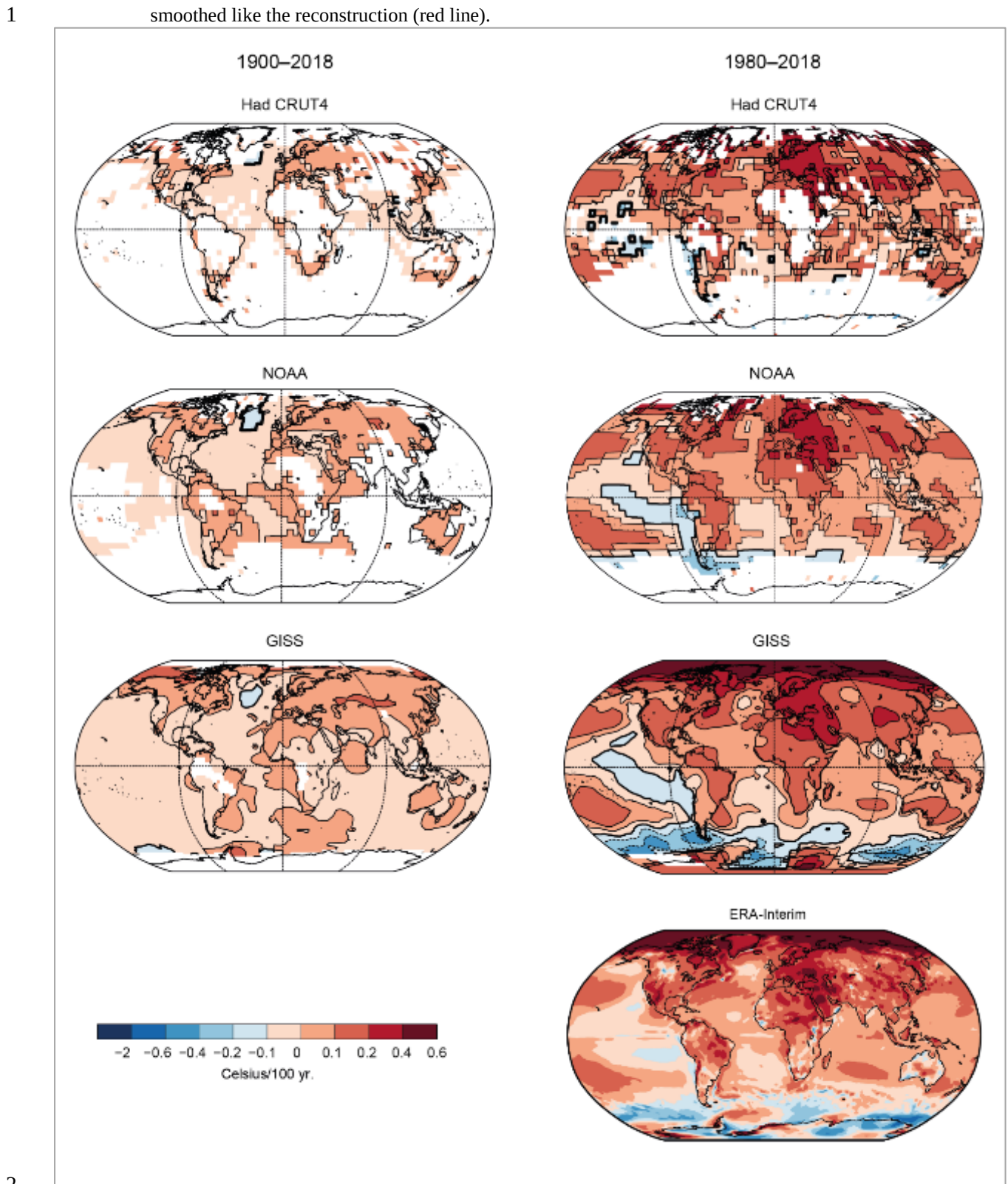


Figure 2.11: Trends in a range of global data sets over the 1900–2018 (left) and 1980–2018 (right) periods [significance overlay to be added in SOD], calculated using ordinary least squares, with time series (bottom left) of annual anomalies from a 1981–2010 base period. Trends are shown where data are available for all 12 months in at least 70% of years during the trend period, and at least 20% of years during the first and last decile of the trend period.

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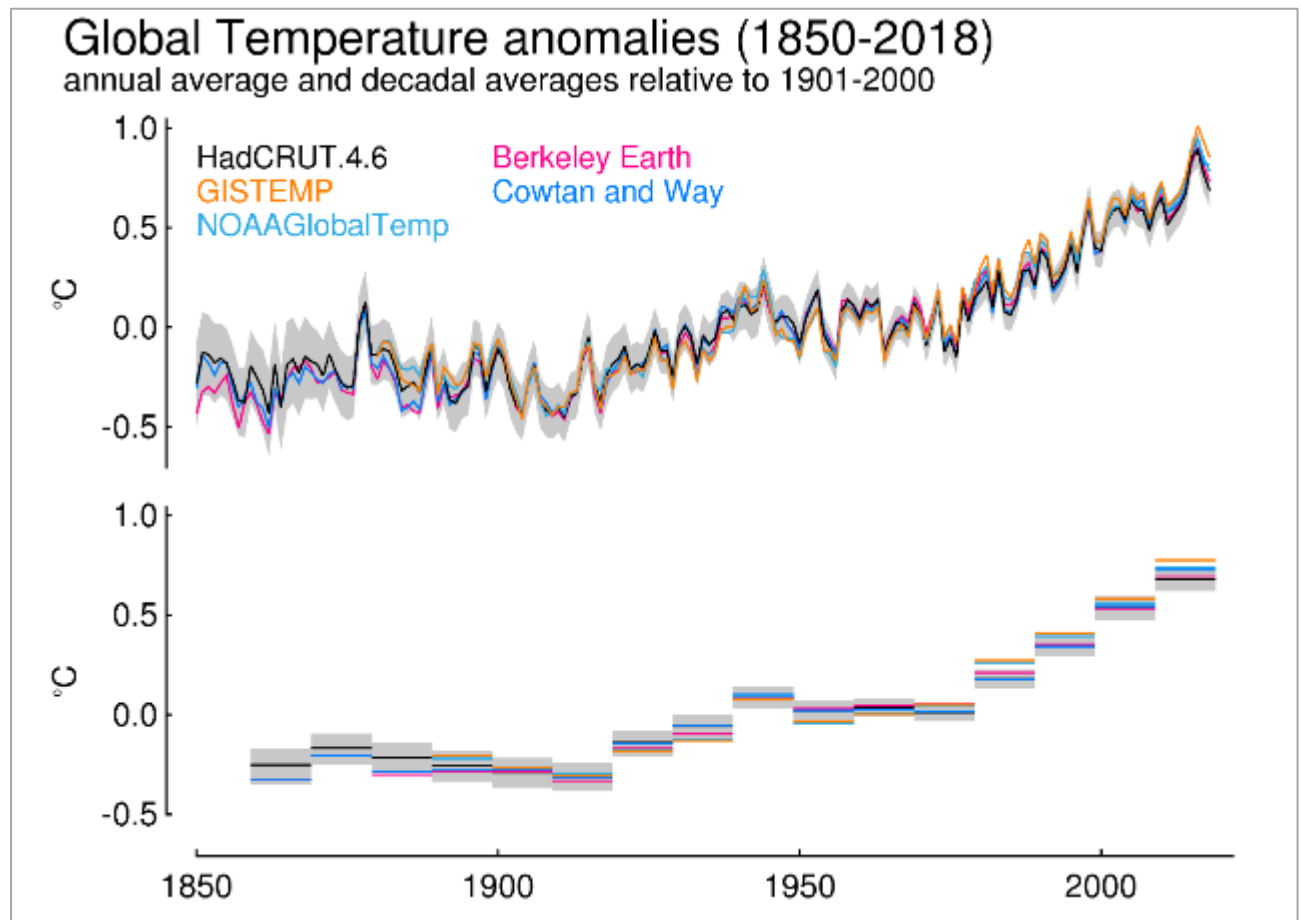
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Figure 2.12: Global mean timeseries at annual and decadal scales for the five conventional surface products from 1850 to date (as available). The HadCRUT4 estimates include uncertainties (grey shading) calculated per (Morice et al., 2012). All timeseries are rebased relative to a 1961-90 climatology.

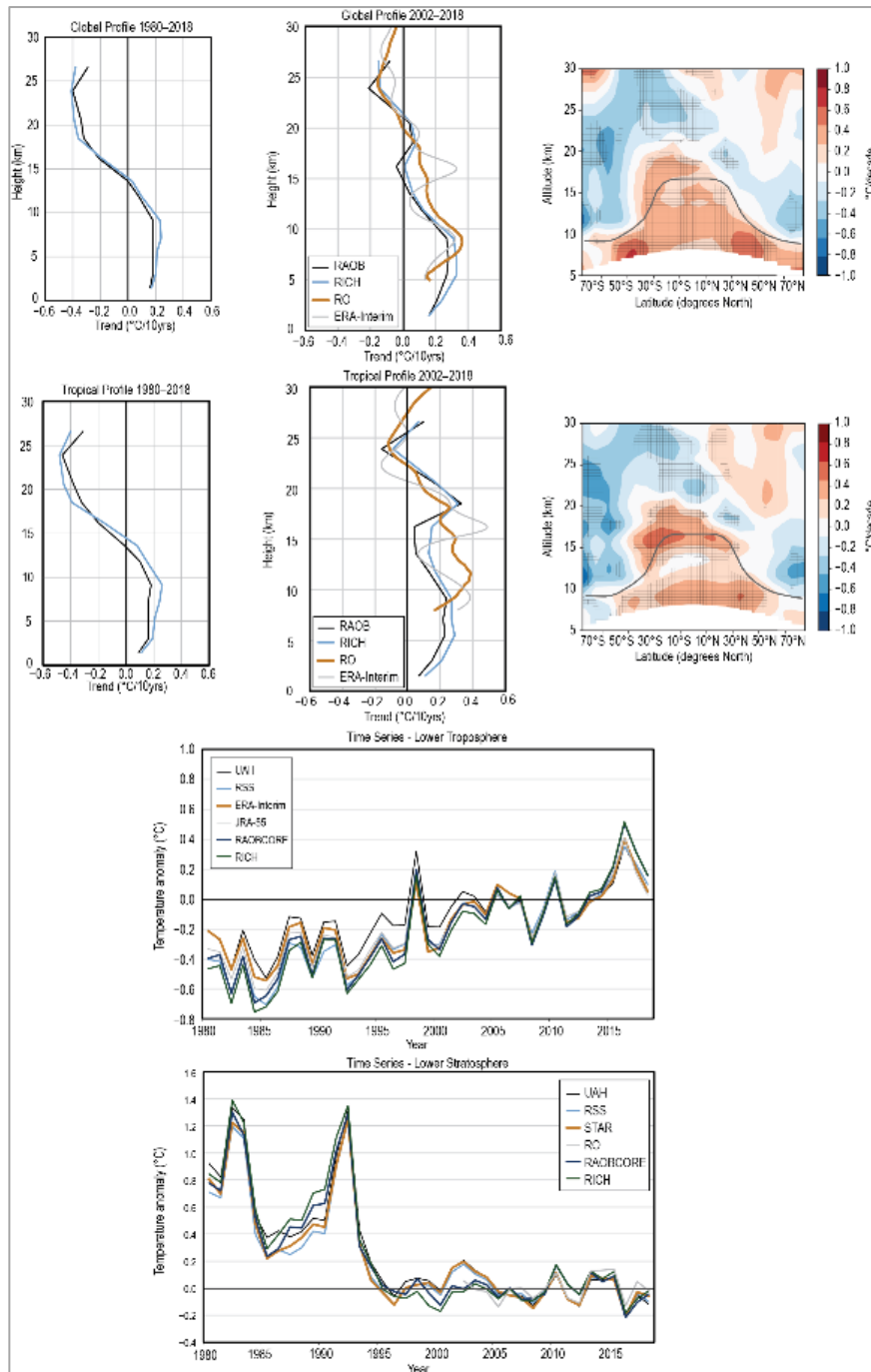


Figure 2.13: (Top left) Trends in temperature at various heights for 1980-2018 (left) and 2002-2018 (centre), from the RAOBCORE and RICH radiosonde data sets, the ERA-Interim reanalysis and, for 2002-2018, the Wegener Centre radio occultation (RO) data set (Angerer et al., 2017), for near-global (70°N to 70°S) (upper) and tropical (20°N to 20°S) (lower) regions. (Top right) Latitude/altitude cross-section of temperature trends for 2002-2018 (with ENSO and QBO influences removed by multiple regression) for the Wegener Centre radio occultation data set (upper) and the ERA-Interim reanalysis (lower). (Bottom) Time series of globally-averaged (upper) lower troposphere and (lower) lower stratosphere temperatures, using a 2002-2016 baseline to match the availability of radio occultation data. [Note known need to better label panels and improve description accordingly in SOD. Will likely also flip the order of lower two panels so stratosphere above troposphere for physical intuition]

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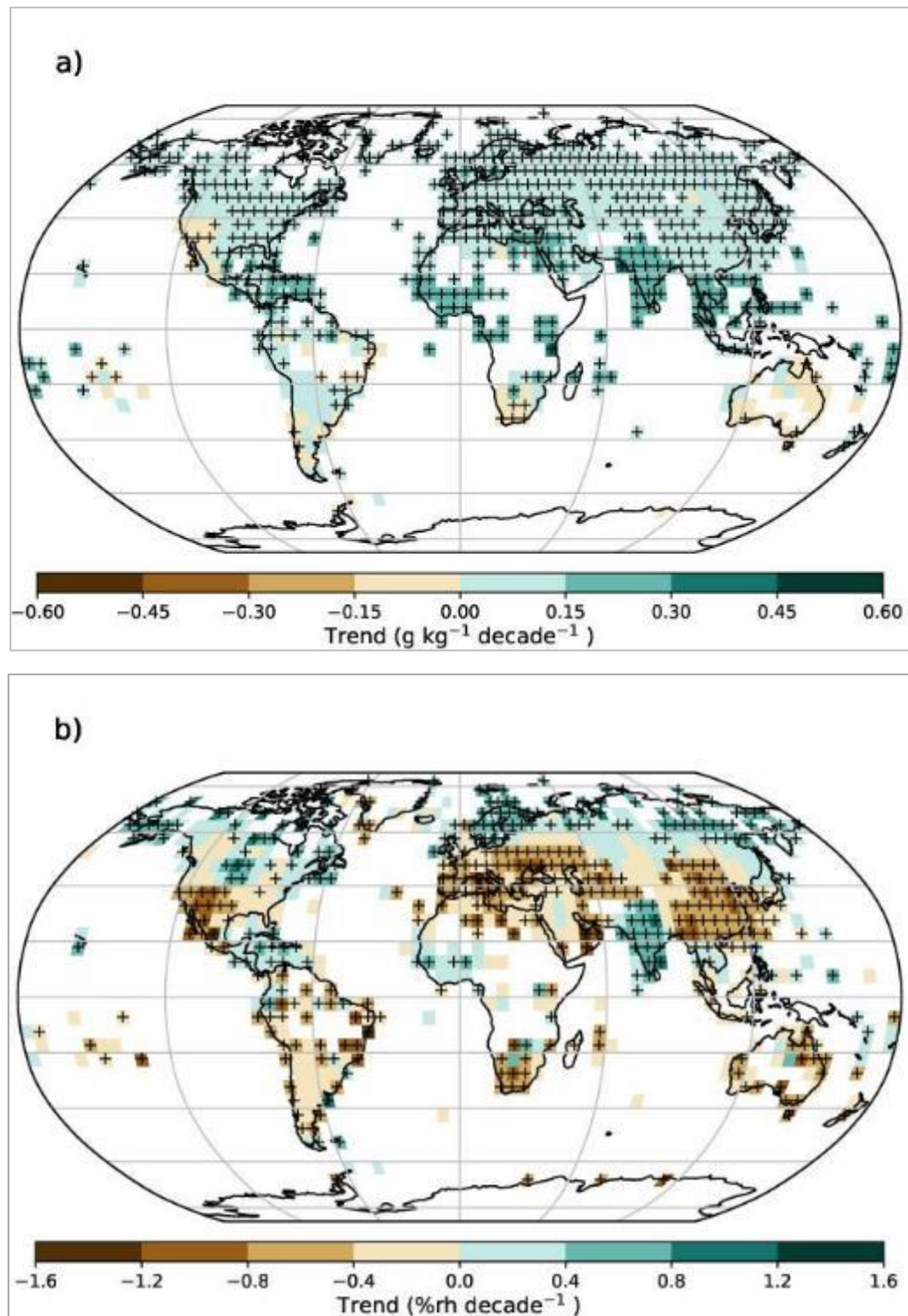


Figure 2.14: Decadal trends in a) surface specific humidity and b) surface relative humidity over 1973 to 2018 from HadISDH in situ monitoring product. Trends are fitted using the median of pairwise slopes. Trend significance is assigned to gridboxes (identified with crosses) where the 5th to 95th percent confidence intervals are both above or below zero.

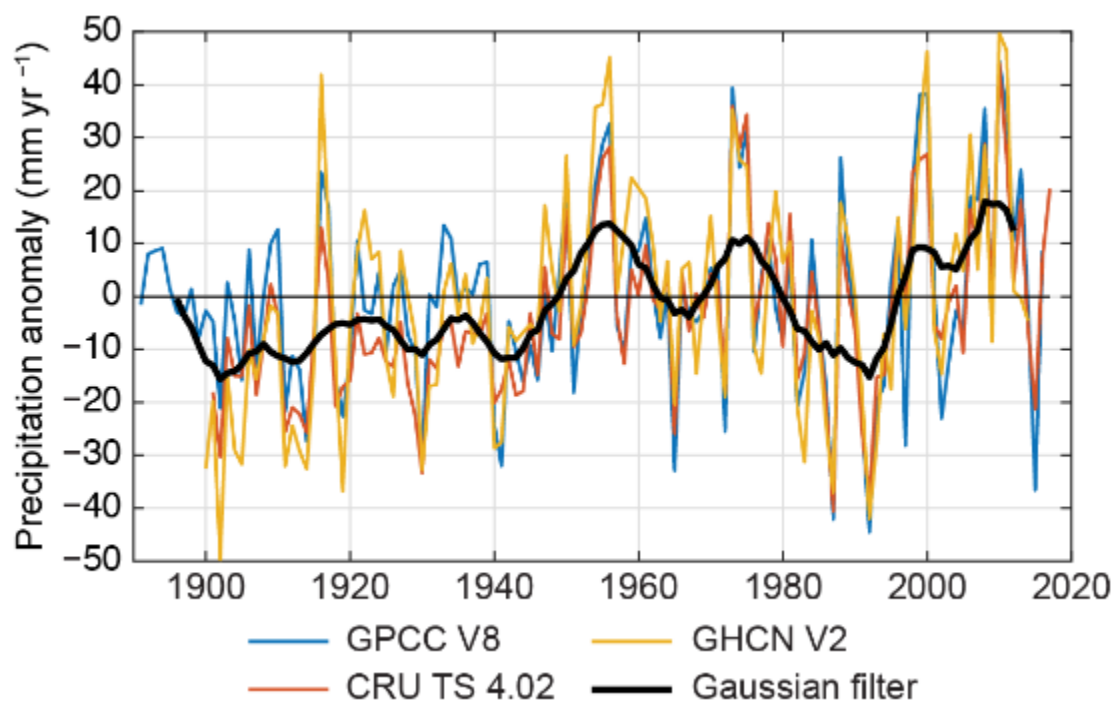


Figure 2.15: Annual precipitation anomalies averaged over global land areas from three global precipitation data sets (GPCC V8; CRU TS 4.02 and GHCN V2) relative to a 1981-2010 climatology. The resulting average of the annual precipitation anomalies from the three global precipitation data sets was smoothed with an 11-year Gaussian-weighted filter. The weights used, from the first through to the 11th years in the running window, are 1/24, 1/24, 1/12, 1/8, 1/8, 1/6, 1/8, 1/8, 1/12, 1/24, and 1/24 respectively.

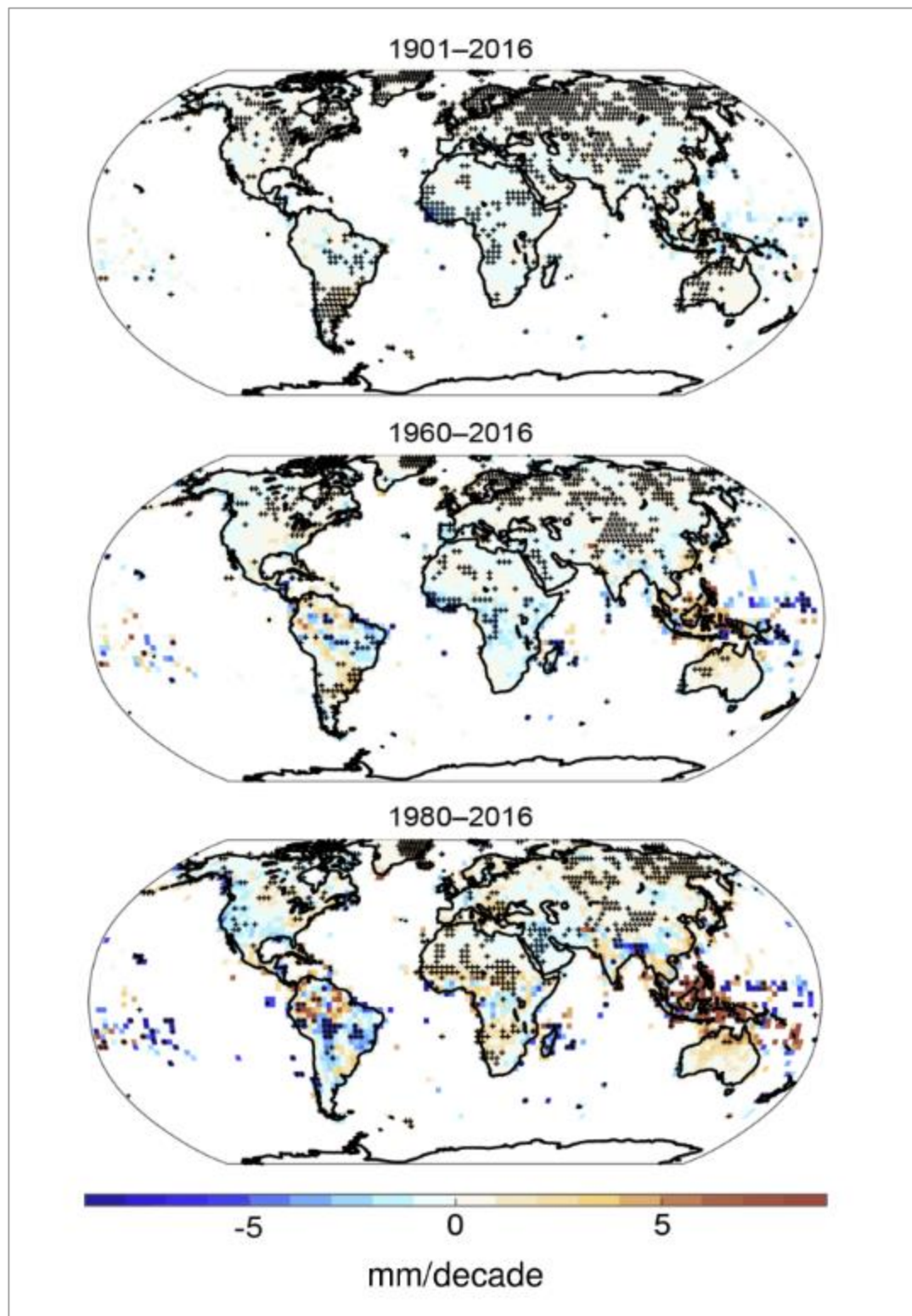


Figure 2.16: Spatial variability of observed GPCC precipitation change over different subperiods: a) 1901–2016, b) 1960–2016 and c) 1980–2016. Trends have been calculated using OLS regression and the confidence intervals in all plots have been corrected to account for serial correlation following the method of Santer et al. (2008).

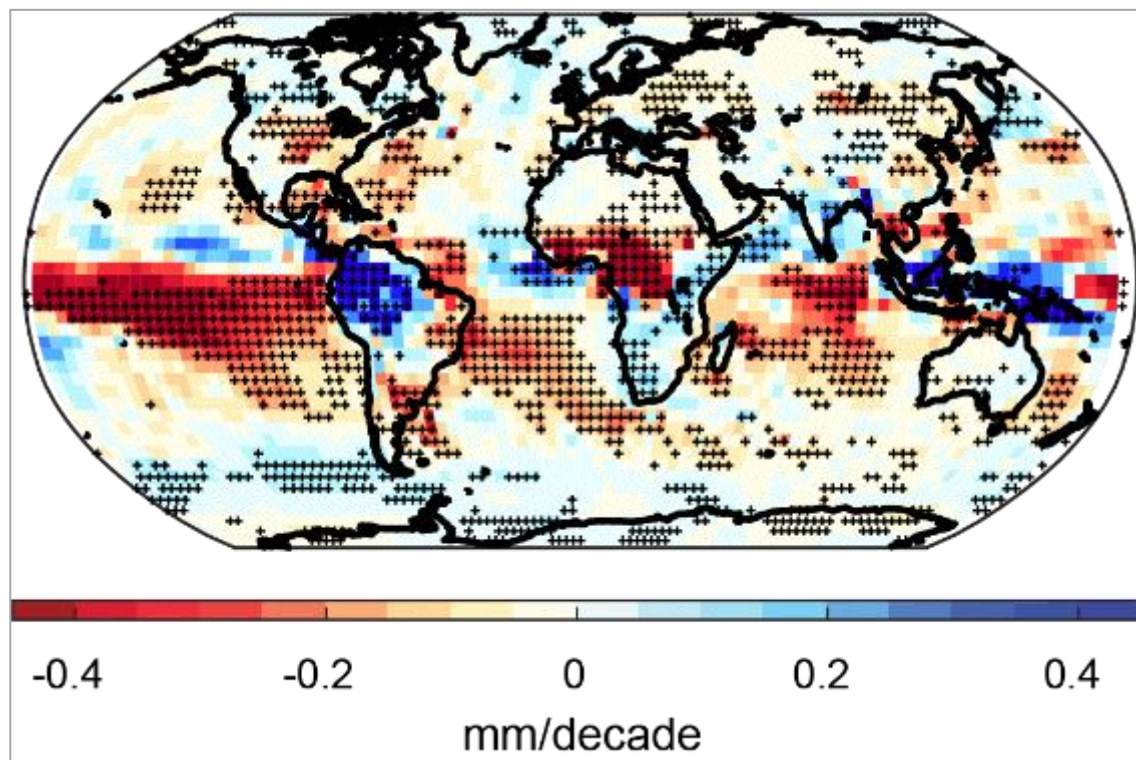


Figure 2.17: Linear trends in precipitation minus evaporation between 1980 and 2018 from the ERA-Interim reanalysis. Blue shading shows regions that have moistened [$\delta(P-E) > 0$] and red shading shows regions that have dried [$\delta(P-E) < 0$]. The stippling indicates regions where the trends are significant at the $p = 0.1$ level. Trends have been calculated using OLS regression and the confidence intervals in all plots have been corrected to account for serial correlation following the method of Santer et al. (2008).

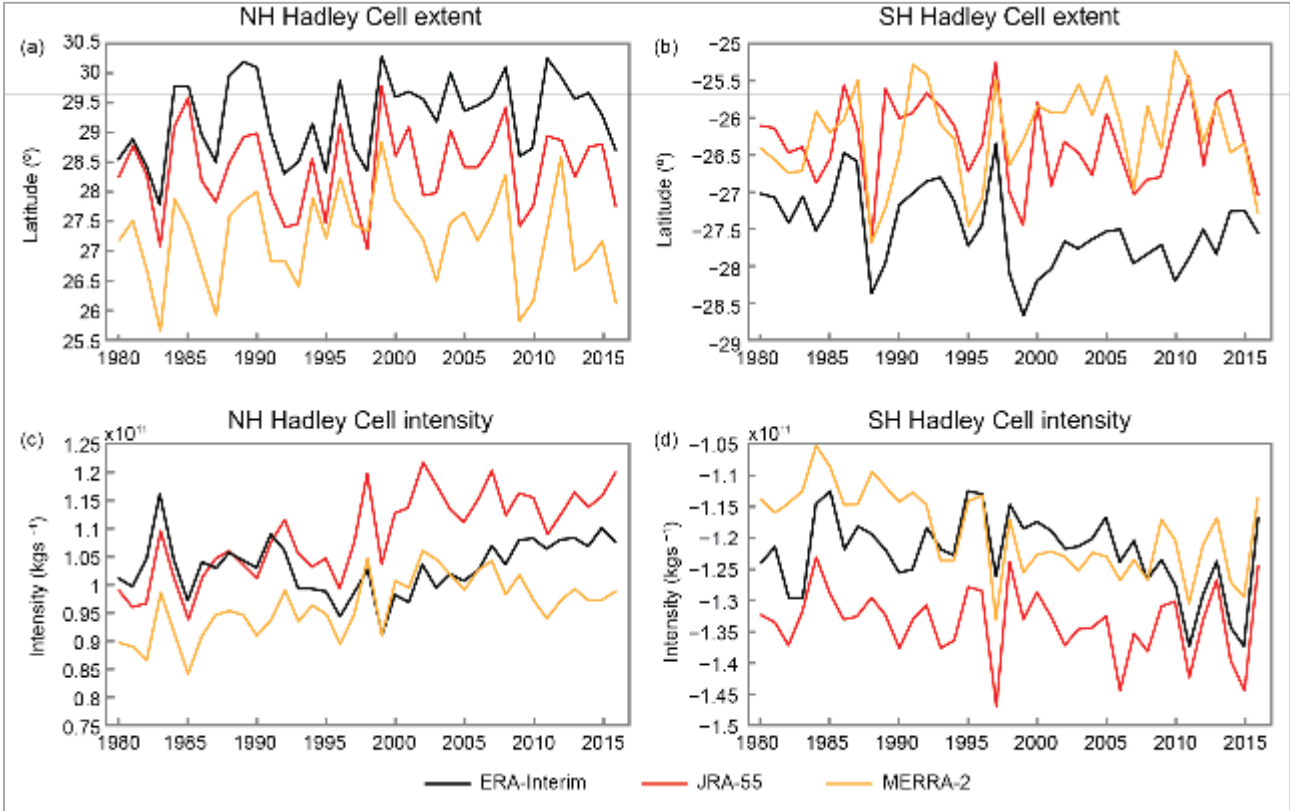


Figure 2.18: Time series of the annual mean (a) NH and (b) SH Hadley Cell extent and (c) NH and (d) SH Hadley Cell intensity computed with ERA-Interim, JRA-55 and MERRA-2 reanalysis. The edge of the HC is taken as the average position of the weakened overturning peak value between 800 and 400 hPa. HC intensity is taken as the vertically averaged maximum value of the meridional stream function between 900 and 200 hPa in each overturning cell.

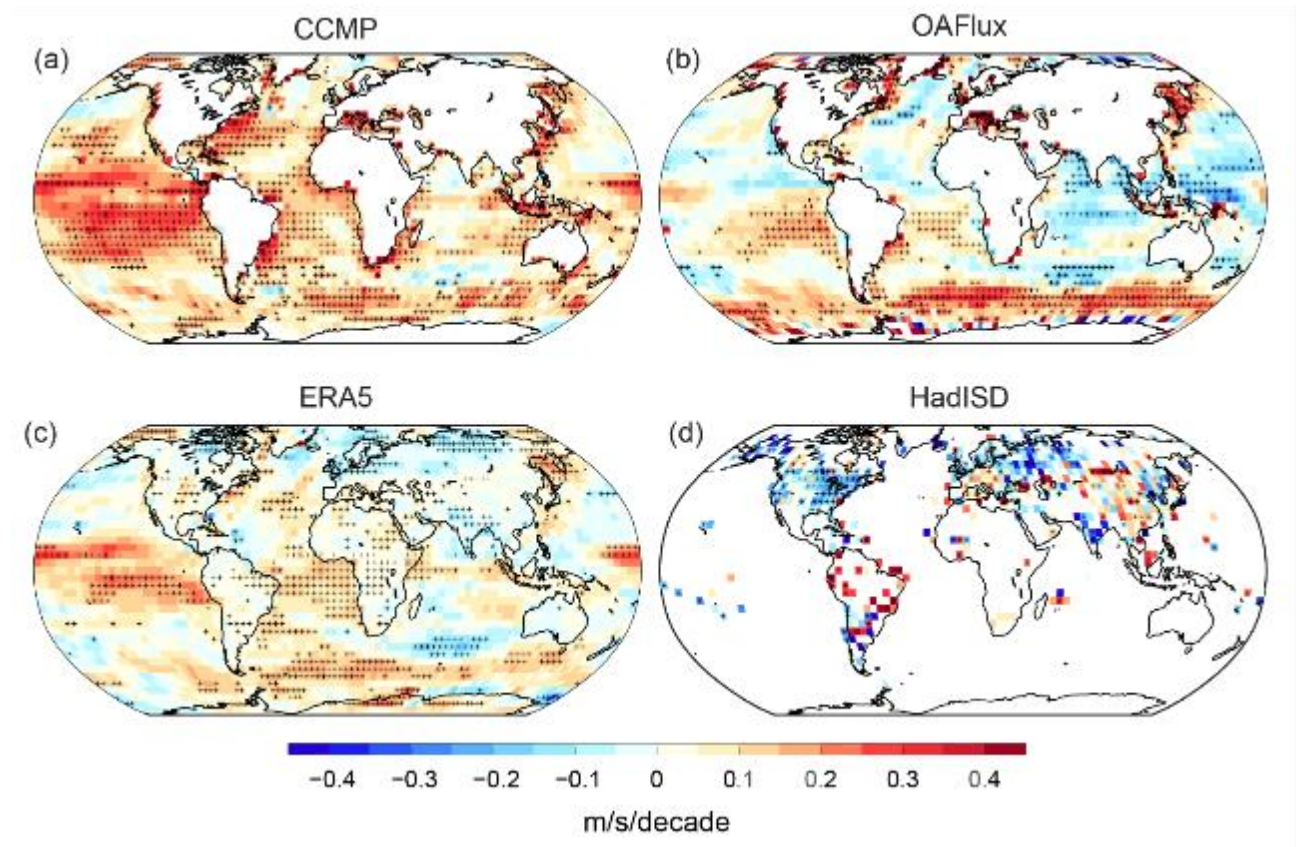


Figure 2.19: Trends in surface wind speed during 1988-2017. (a) Cross-Calibrated Multi-Platform wind product (Atlas et al., 2011); (b) wind speed from the Objectively Analyzed Air-Sea Heat Fluxes data set, release 3 (Yu et al., 2008); (c) ERA5, and (d) Station observation wind from the Integrated Surface Database (HadISD) v2.0.2.2017f (Smith et al., 2011). Trends for HadISD were computed only if at least 36 years had values and each year has at least 3 seasons of observations available. White areas indicate incomplete or missing data. The symbol of '+' indicates grid boxes where trends are significant (i.e. a trend of zero lies outside the 90% confidence interval) using OLS with AR(1) correction per Santer et al., 2008.

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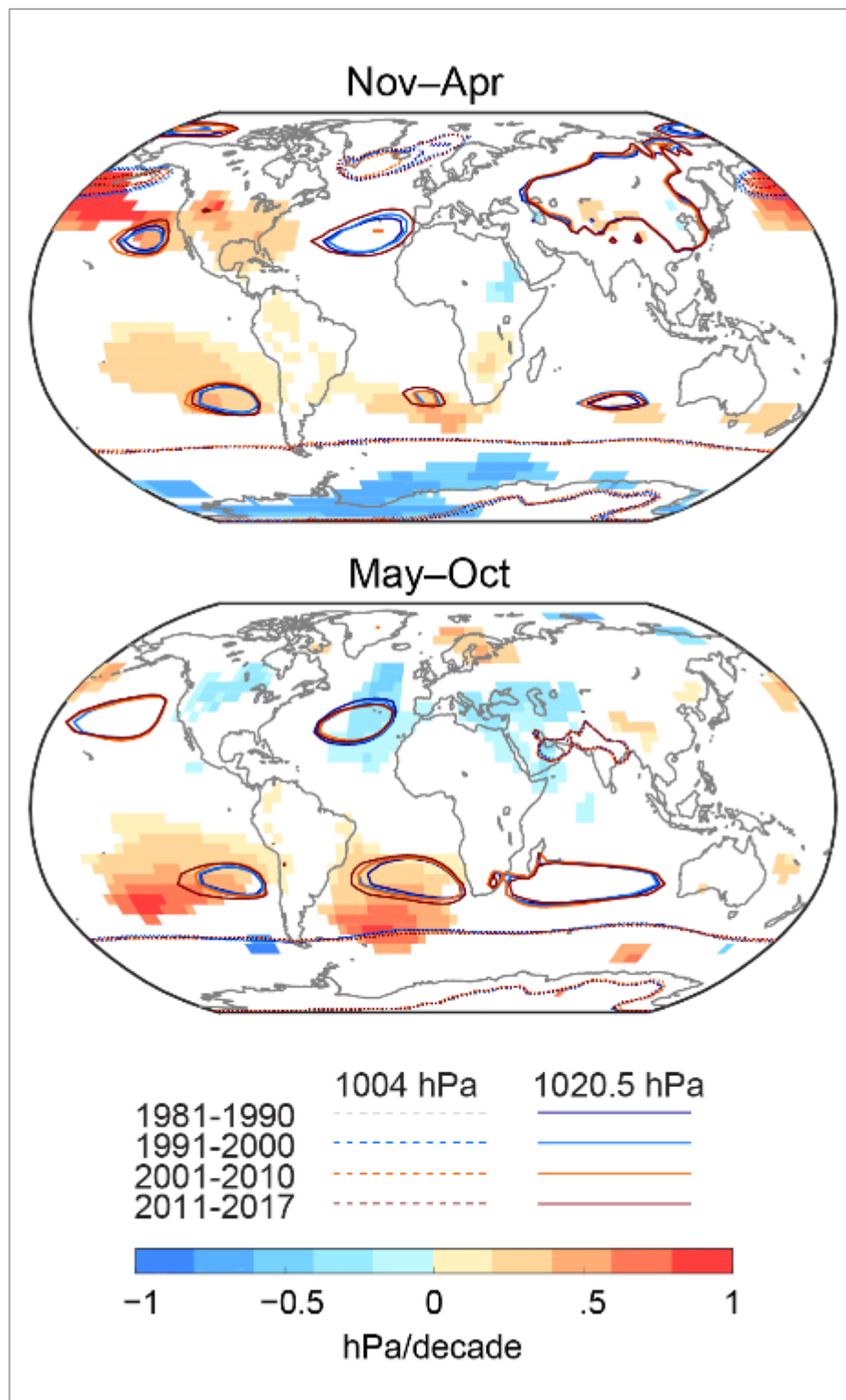


Figure 2.20: Changes in the ERA5 SLP since 1979. Colors denote linear trends estimated by the ordinary least squares fitting and only significant values (i.e. a trend of zero lies outside the 90% confidence interval) are plotted. Trend calculation is via OLS accounting for AR(1) effects as per Santer et al., 2008. Contour lines are locations of the decadal mean SLP for 1004hPa and 1020.5hPa.

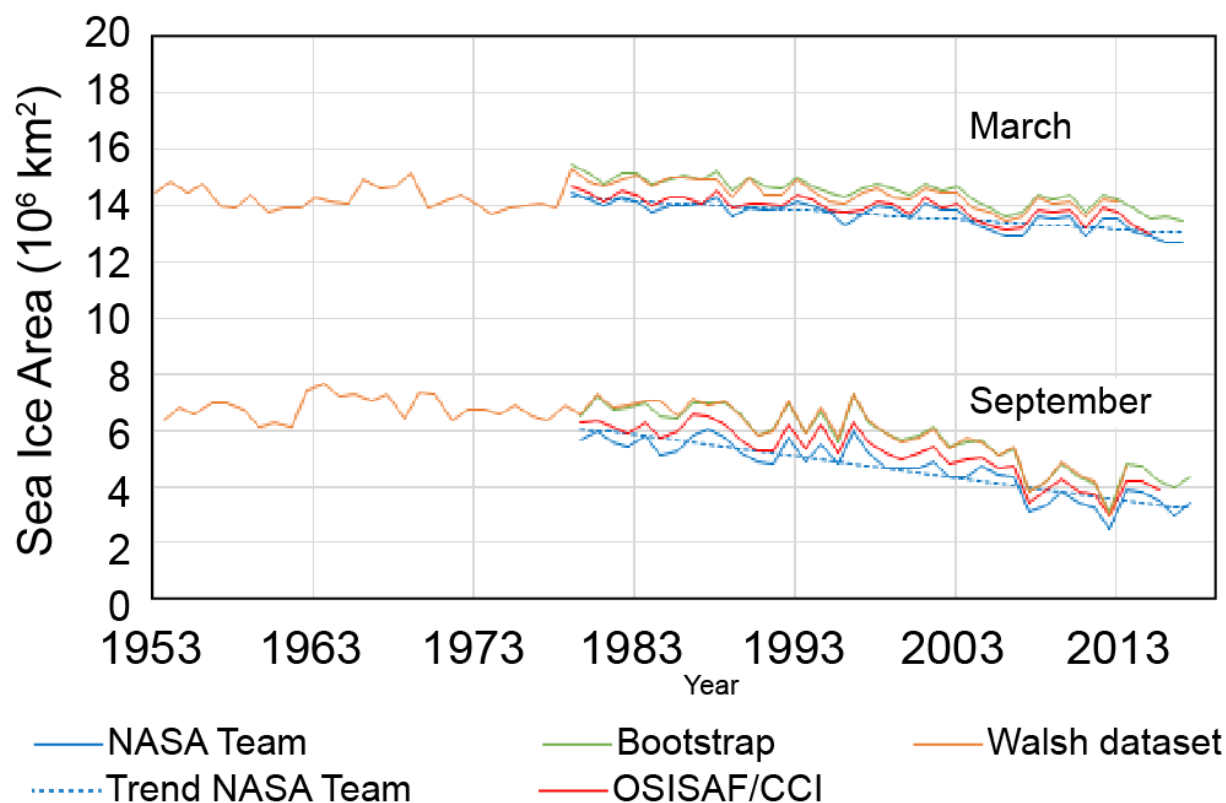


Figure 2.21: Arctic sea ice area for March and September since 1953 (Walsh dataset) and 1979 (passive microwave satellite era). Sea-ice area values have been calculated from raw sea-ice concentration fields provided by OSISAF/CCI, NASA Team and Bootstrap from NOAA CDR 3.0, and Walsh NSIDC G10010 data sets. Linear trend fits for NASA Team (dashed lines) are added for visual aid.

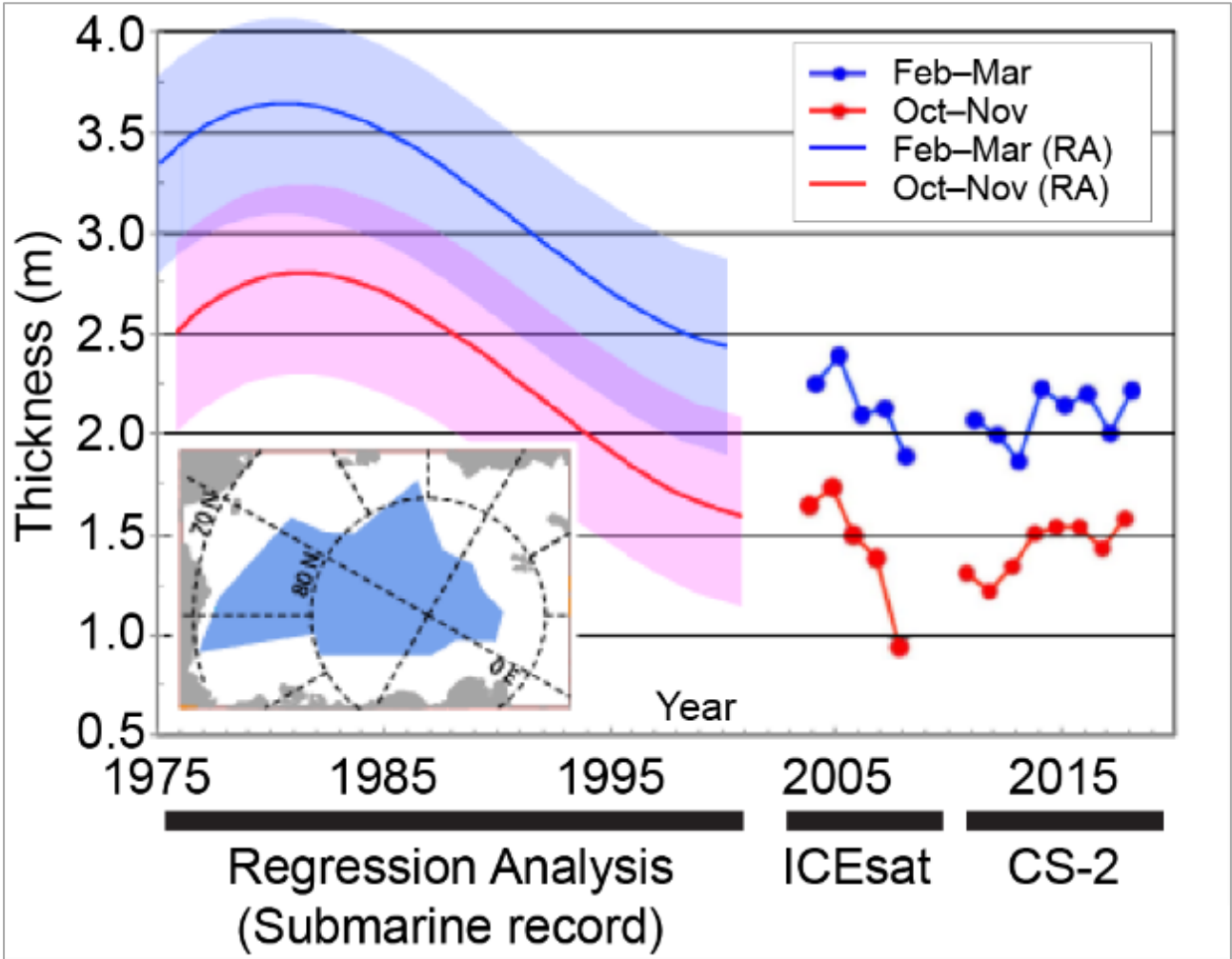


Figure 2.22: Arctic sea ice thickness changes (means) for winter and fall from submarine transects (Rothrock et al., 2008), ICESat (Kwok et al., 2009) and CryoSat-2 retrievals (Kwok and Cunningham, 2015). Shadings (blue and red) show expected residuals in the regression analysis. Inset shows the data release area of submarine data. [Placeholder from (Kwok and Kacimi, 2018). It is intended to update the figure for the SOD and add uncertainty ranges for satellite based ice thickness levels.]

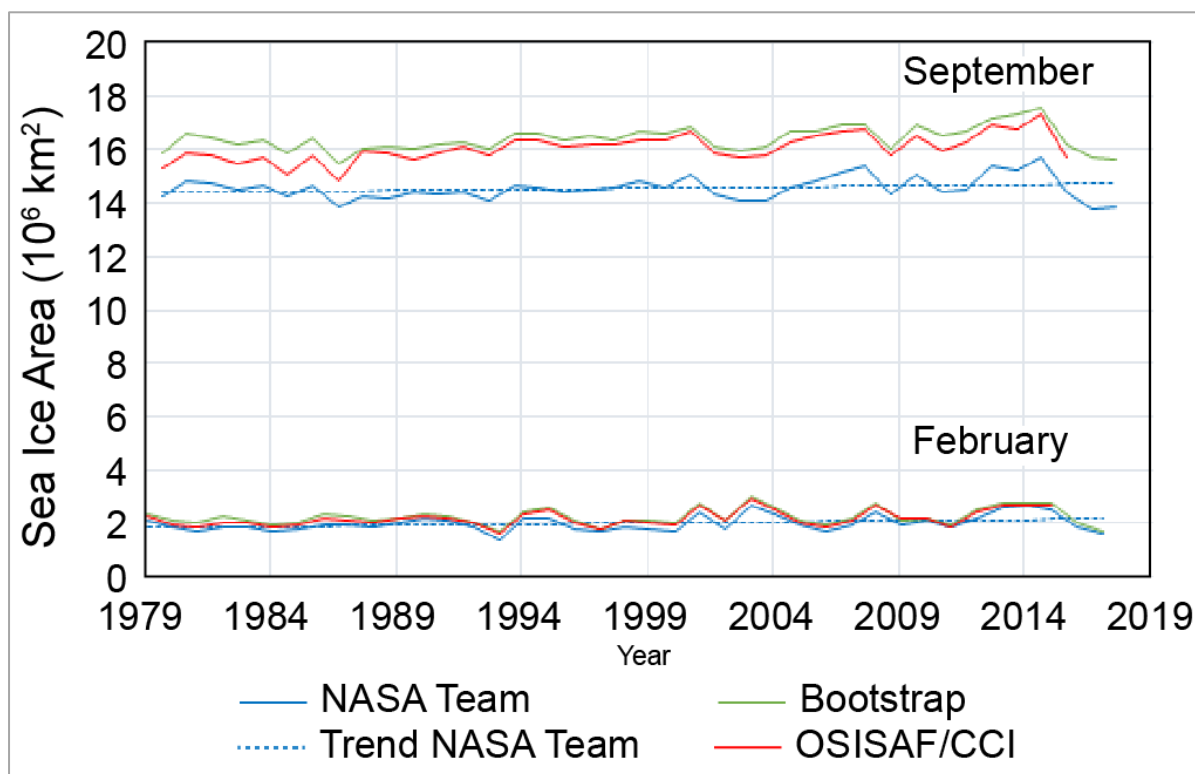


Figure 2.23: Antarctic sea ice area since 1979. Sea-ice area values have been calculated from raw sea-ice concentration fields provided by OSISAF/CCI, NASA Team and Bootstrap from NOAA CDR 3.0. Linear trend fit for the NASA Team (dashed lines) is added for visual aid.

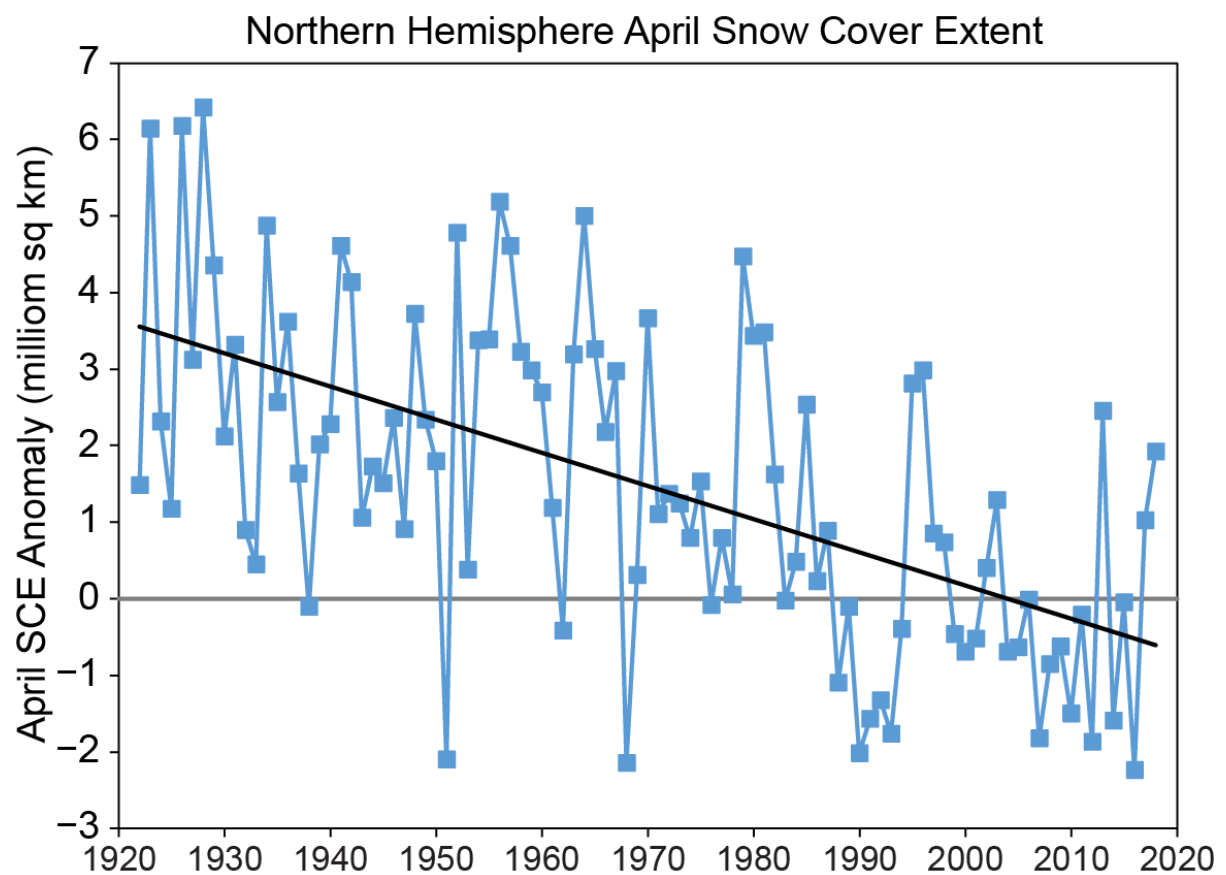


Figure 2.24: April SCE anomaly (departure from 1995-2014 mean) and linear least squares trend over the entire period 1921-2018 for the Northern Hemisphere. Data are from NOAA (<https://www.ncdc.noaa.gov/snow-and-ice/extent/>) for 1967-2018 with the earlier part of record based on in situ data (Brown, 2002 <https://nsidc.org/data/g02131>).

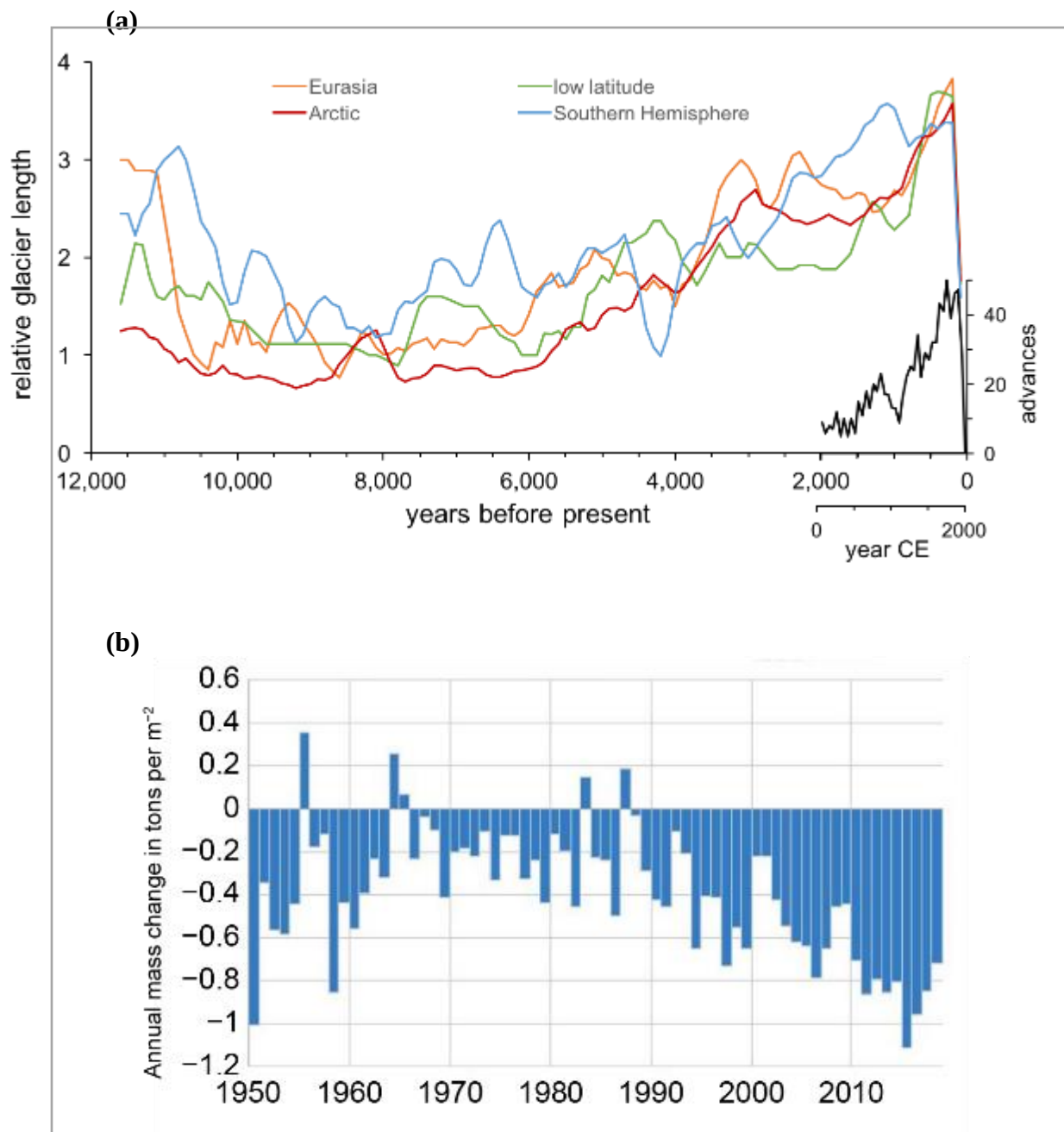
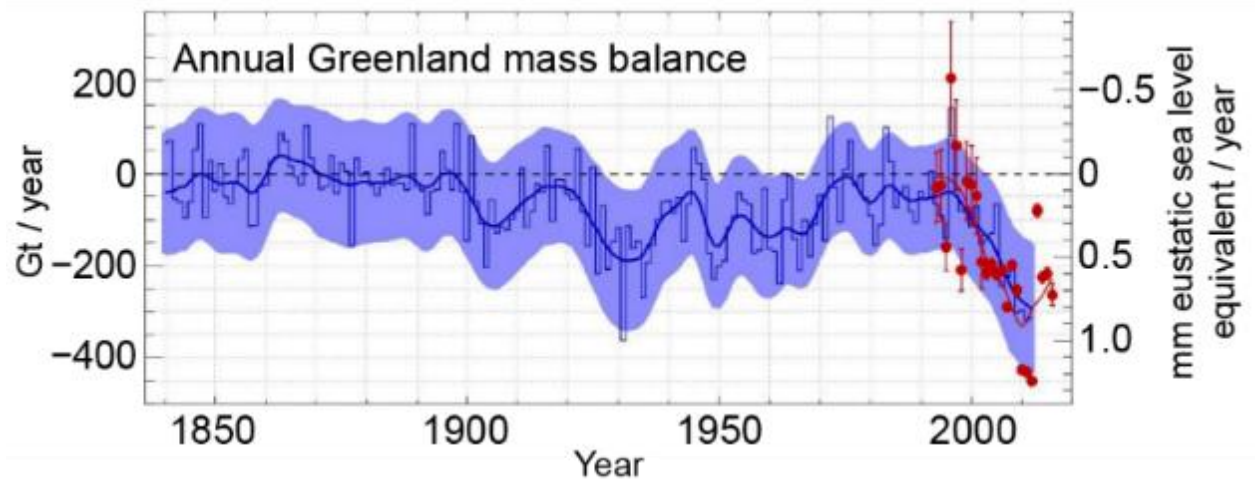


Figure 2.25: (a) Mountain glacier relative extent over the Holocene. Mean qualitative status of mountain glacier length summarized from a recent global review (Solomina et al., 2015). The relative length of glaciers studied around the world was averaged for four major regions in 100-year intervals and smoothed with a 300-year running mean, where 4 = Holocene maximum position, 3 = larger than modern (approximately the past 30 years) and advancing, 2 = glaciers present in the study area to some unknown extent, 1 = larger than modern and retreating, 0 = near the present extent. Intervals lacking information were not included in the mean values (set to null). Inset shows the total number of glacier advances reported in a recent global compilation of mountain glacier fluctuations during the past 2000 years, summed in 50-year intervals (Solomina et al., 2016). (b) Annual mass balance for a set of global reference glaciers with more than 30 years of continuous glaciological measurements. Annual mass change values are given on the vertical axis in meter water equivalent (m w.e.) which corresponds to tons per square meter ($t\ m^{-2}$). Source: WGMS (2017, updated and earlier reports; <https://wgms.ch/>). Global values are calculated using only one single value (averaged) for each of 19 mountain regions in order to avoid a bias to well observed regions.

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Figure 2.26: Greenland ice sheet total mass balance (gigatons per year, left axis) since the end of the Little Ice Age, combining a dataset spanning from 1840 to 2010 (based on a combination of in situ and simulated climate data reconstruction of surface mass balance and parameterized marine ice loss; Box and Colgan 2013), and 1993-2016 (mainly based on GRACE satellite data; WCRP 2018). Shown are annual levels and Gaussian running average filter for the first dataset, and annual levels and error estimates for the second. The right axis shows the equivalent eustatic sea level change in mm.

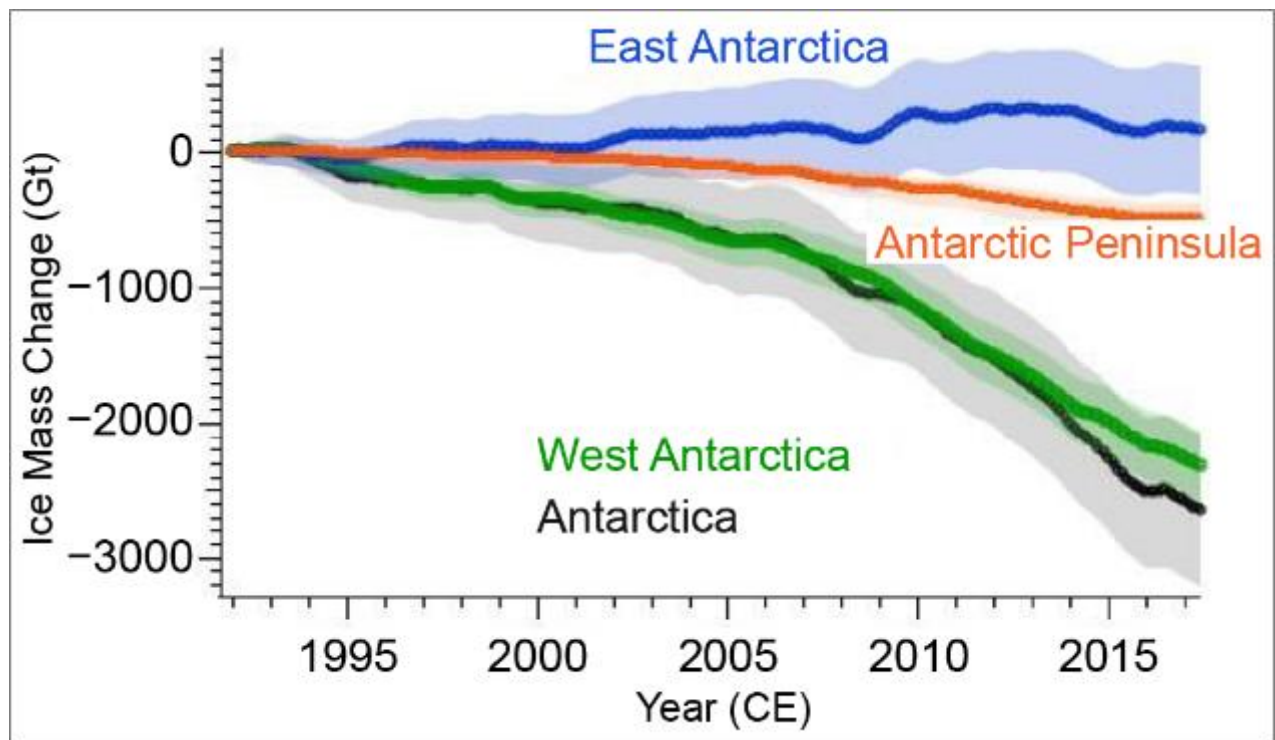


Figure 2.27: Cumulative Antarctic Ice Sheet mass change (gigatons) from satellite-based measurements (The IMBIE Team, 2018). The cumulative ice-sheet mass changes (lines) are determined from the integral of monthly measurement-class averages for each part of the Antarctic ice sheet (see colored labels). The estimated 1σ uncertainties for the respective cumulative changes are shaded.

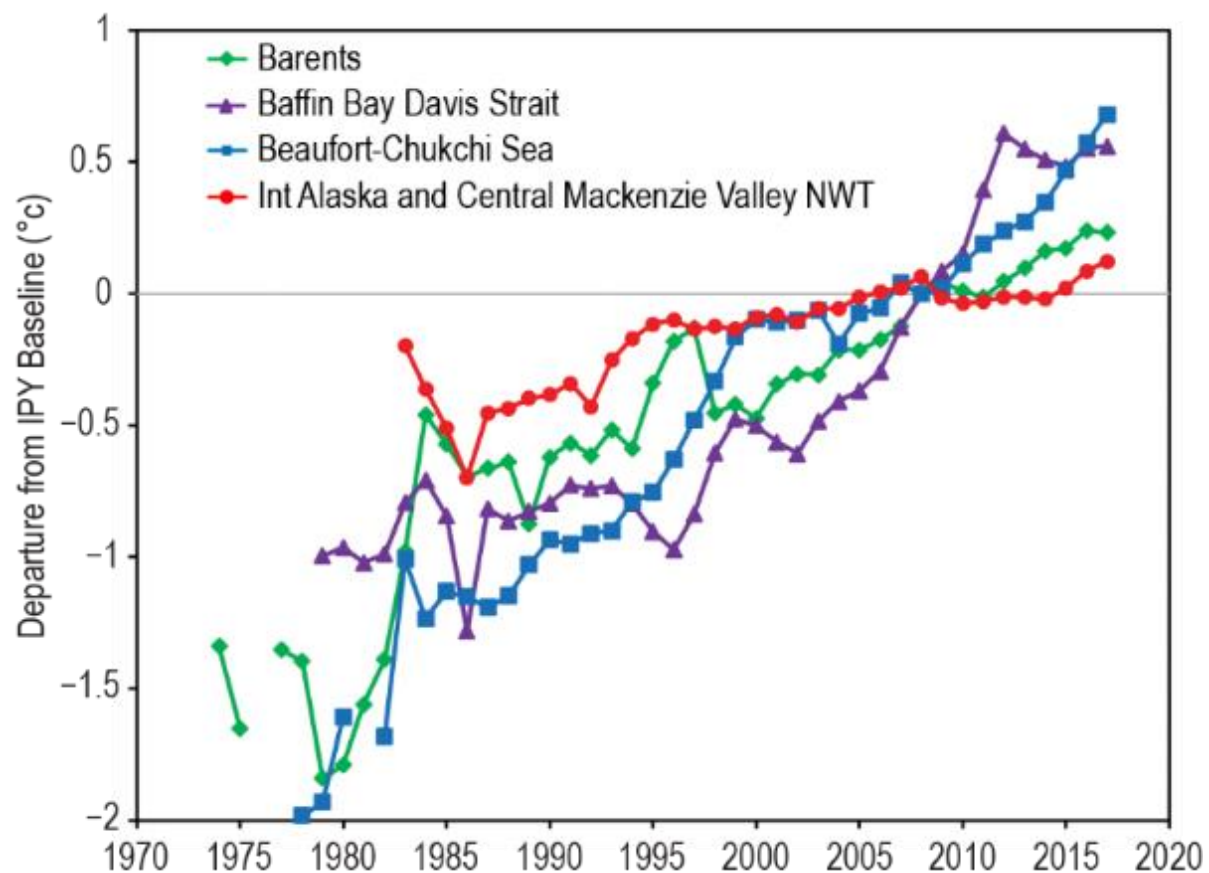


Figure 2.28: Average departures of permafrost temperature (measured near the depth where seasonal variation is negligible) from a baseline established during International Polar Year (IPY 2007-09). Regions are those described in (Romanovsky et al., 2017; 2018). Barents: continuous-discontinuous permafrost Nordic region and Russia/Siberia 1974-2017 (11 sites with 6 starting 1998 or later). Baffin Bay Davis Strait: Continuous permafrost high and eastern Canadian Arctic 1978-2017 (7 sites with 4 initiated 2008). Beaufort-Chukchi Sea: continuous permafrost northern Alaska, Northwest Territories and eastern Siberia 1978-2017 (9 sites 4 initiated 2003 or later). Interior Alaska & Central Mackenzie Valley: discontinuous permafrost 1983-2017 (10 sites, 1 initiated 2001).

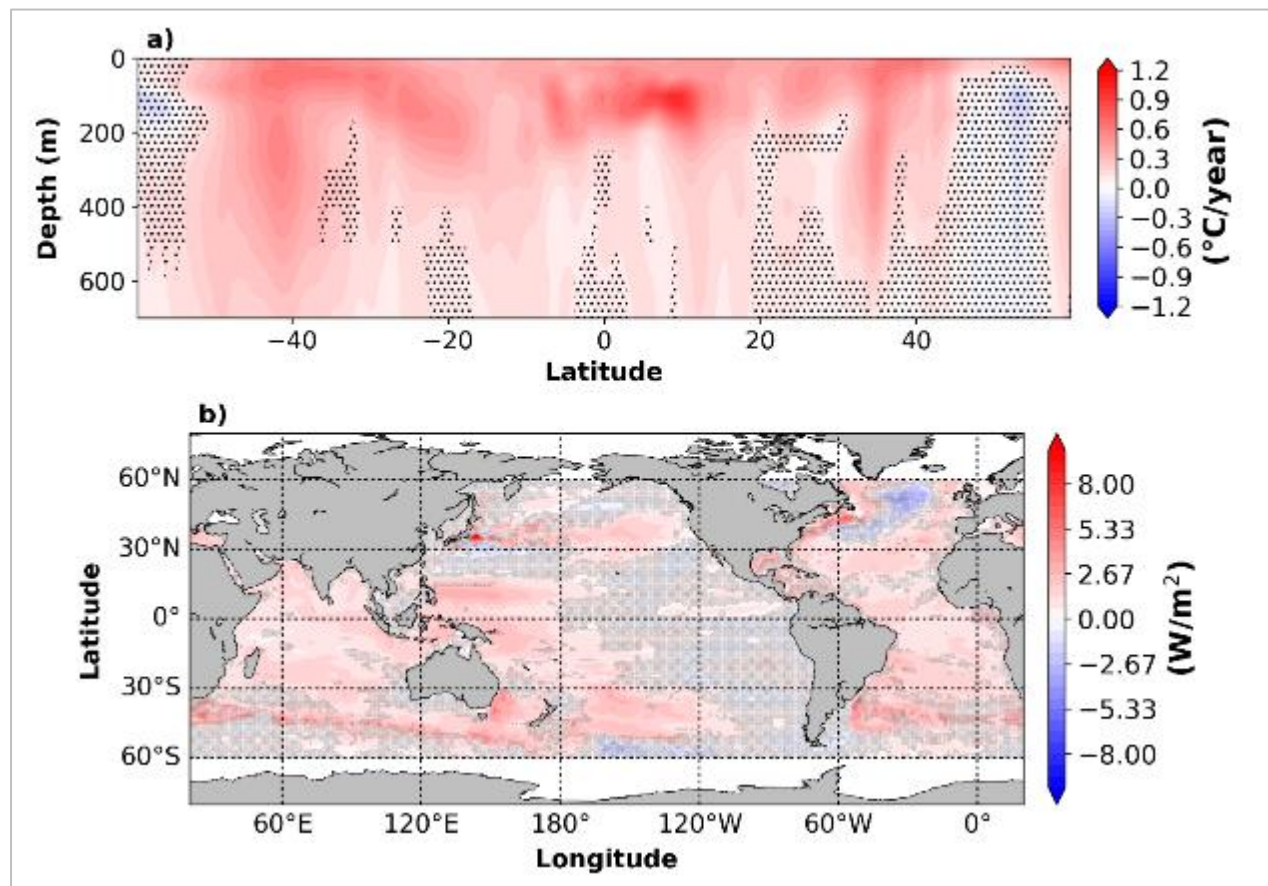


Figure 2.29: Rate of change over the period 1993-2017 a) of global zonal mean ocean subsurface temperature anomalies as a function of latitude, b) of ocean heat content (0-700m) anomalies relative to the 1993-2014 reference period based on the ensemble mean of the Copernicus Marine Service multi-product approach (3 global ocean reanalyses: GLORYS, CGLOR, OSRS5; 2 in situ based reprocessed products: CORA, ARMOR-3D, von Schuckmann et al., 2018; <http://marine.copernicus.eu/science-learning/ocean-monitoring-indicators/catalogue/>). Black dots indicate areas where the noise (ensemble standard deviation of trends) exceeds the signal (ensemble mean of trend), indicating areas of most robust signatures from the multi-product approach.

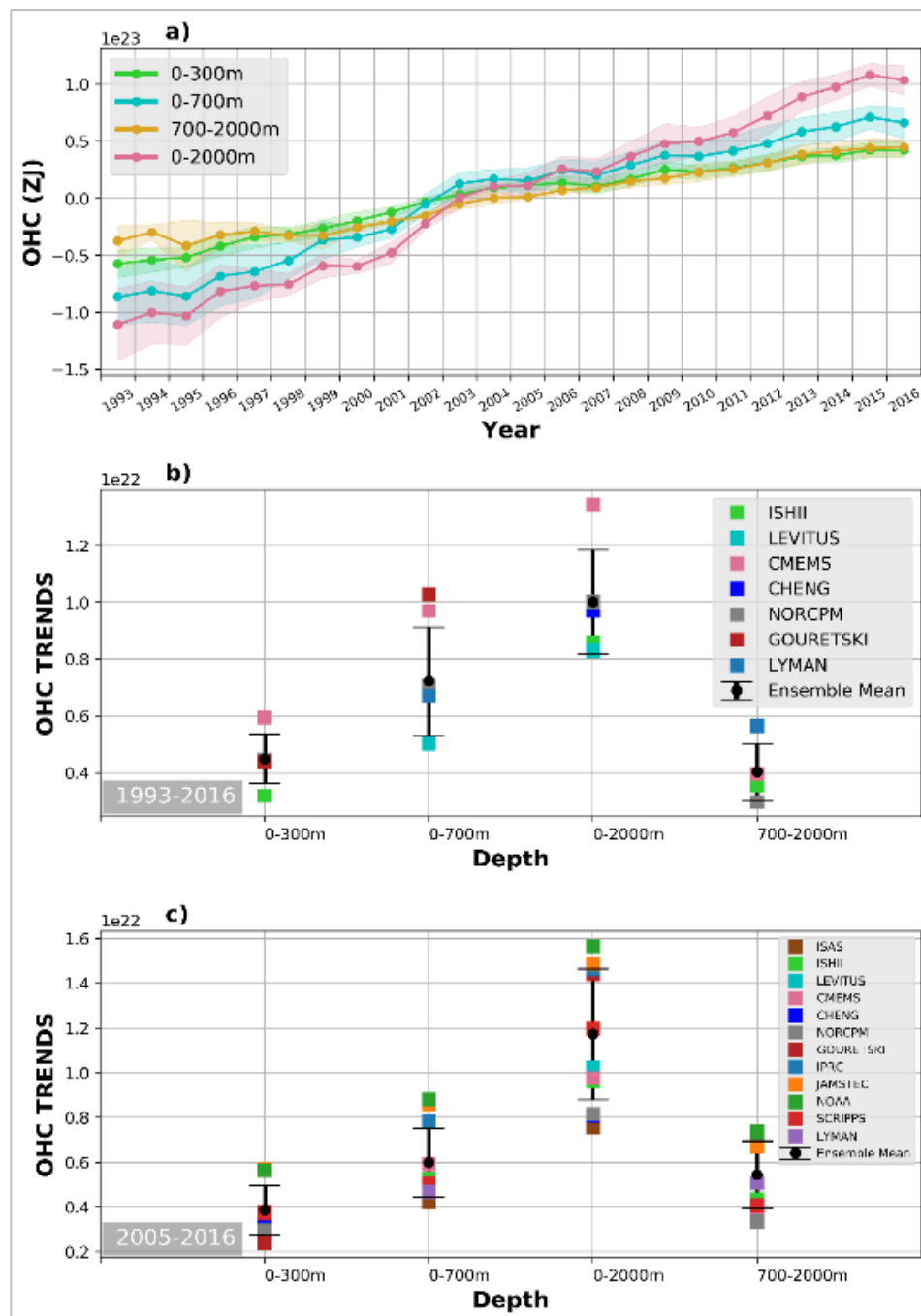


Figure 2.30: a) Ensemble (see legend of panel b)) mean of yearly averaged near-global (60°S-60°S latitude) ocean heat content integrated over different depth layers (green: 0-300m; blue: 0-700m; yellow: 700-2000m; red: 0-2000m) over the period 1993-2016, together with the ensemble spread (shaded areas). b) Rate of change for near global ocean heat content over the period 1993-2016 as derived from different data sources, which are listed in the figure panel, and integrated over different depth layers. c) Same as b), but over the time period 2005-2016, which allows the assessment of a larger variety of data product (see legend). Adopted from von Schuckmann et al., 2019. Data sources linked to legend abbreviations include: Ishii: Ishii et al., 2009; Levitus: Levitus et al., 2012; CMEMS: von Schuckmann et al., 2018; CHENG: Cheng et al., 2017a; NORCPM: Counillon et al., 2014; GOURETSKI: Gouretski and Koltermann (2007); LYMAN: Lyman and Johnson, 2014; ISAS: Gaillard et al., 2015; IPRC: Yoshinari et al., 2006; JAMSTEC: Hosoda et al., 2008; NOAA: Levitus et al., 2012; Scripps: Roemmich and Gilson, 2009).

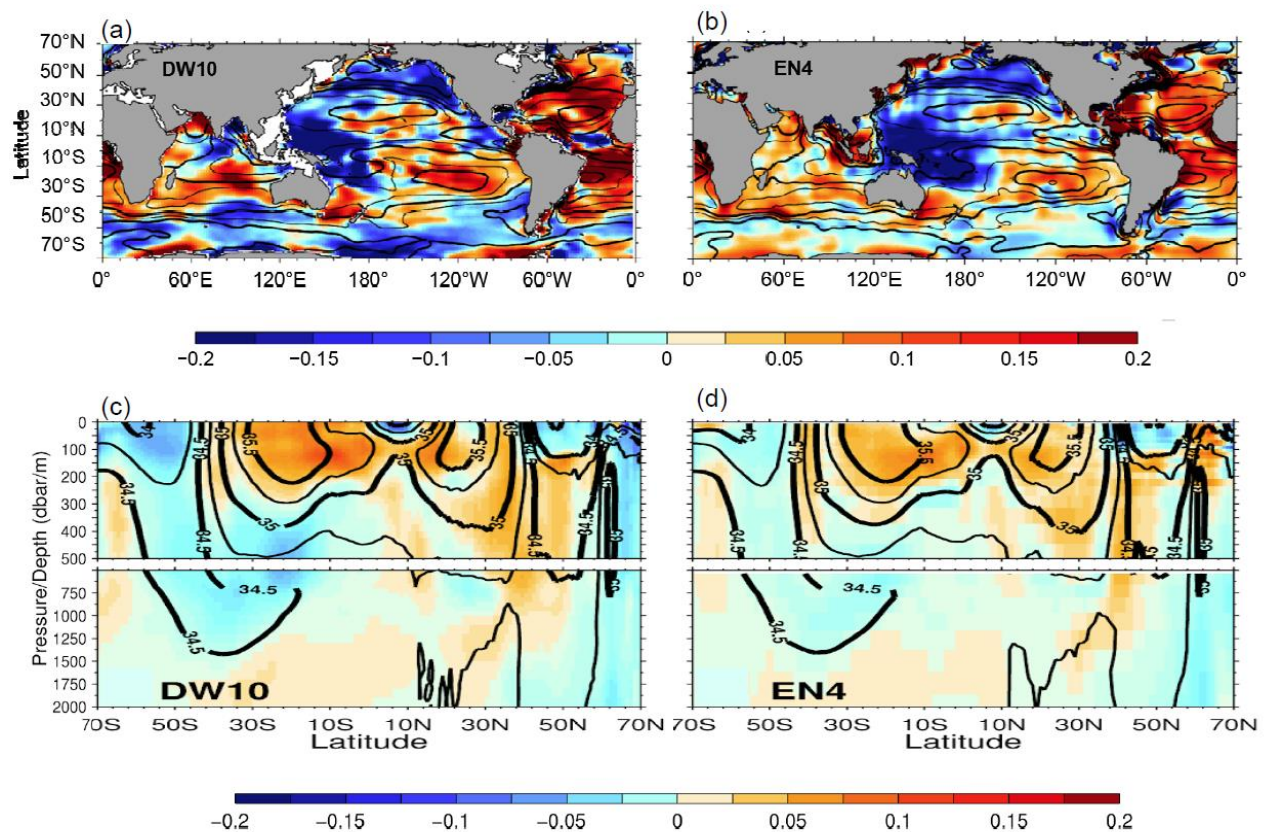


Figure 2.31: Long-term estimates of global near-surface salinity (SSS) change after (A) Durack and Wijffels (2010; analysis period 1950–2008), (B) Good et al. (2014; analysis period 1950–2012). Black contours show the associated climatological mean SSS for the analysis period. (C) and (D) Three long-term estimates of global zonal mean subsurface salinity changes after (A) Durack and Wijffels (2010) and (B) Good et al. (2013). Both estimates in (A)–(D) are scaled to represent equivalent magnitude changes over a 50-year period (Practical Salinity Scale 1978 [PSS-78], 50 yr⁻¹). Black contours in maps (A, B) and cross-sections (C, D) show the associated climatological mean subsurface salinity for the analysis period from each corresponding climatological assessment. In all panels blue denotes freshening regions and red denotes regions with enhanced salinities.

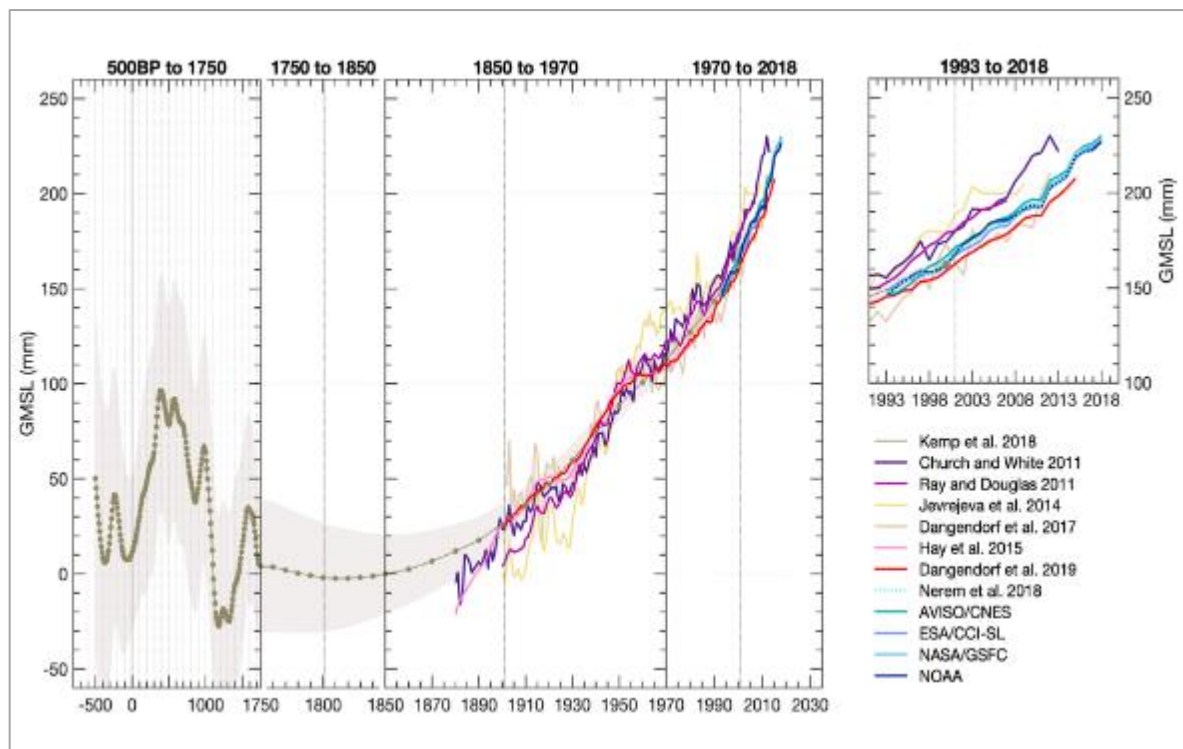


Figure 2.32: Global mean sea level timeseries from 500 Before Present (BP) to 2018, based on proxy data integrated with Hay et al.'s (2015) tide gauge reconstruction (Kemp et al., 2018; Kopp et al., 2016)), tide gauge only reconstructions and satellite altimeter products (Nerem et al. (2018), AVISO, ESA (Legeais et al. 2018), NASA and NOAA). Vertical dotted lines indicate the first year of a century. The pre-industrial period is 1750-1849, the early-industrial period is 1850-1899, and the sea level altimeter era starts in 1993 (inset). Period definitions are found in cross-chapter Box 1.4. All data were plotted as annual means except for proxy (10-year mean). For clarity, errorbars are only shown for proxy data and all data have been vertically shifted. Proxy data is relative to zero in 1850. Tide gauge data is aligned to the 1910-1990 mean of the proxy data. Altimeter data is aligned to the 1993-2000 mean of the proxy data. Units: mm.

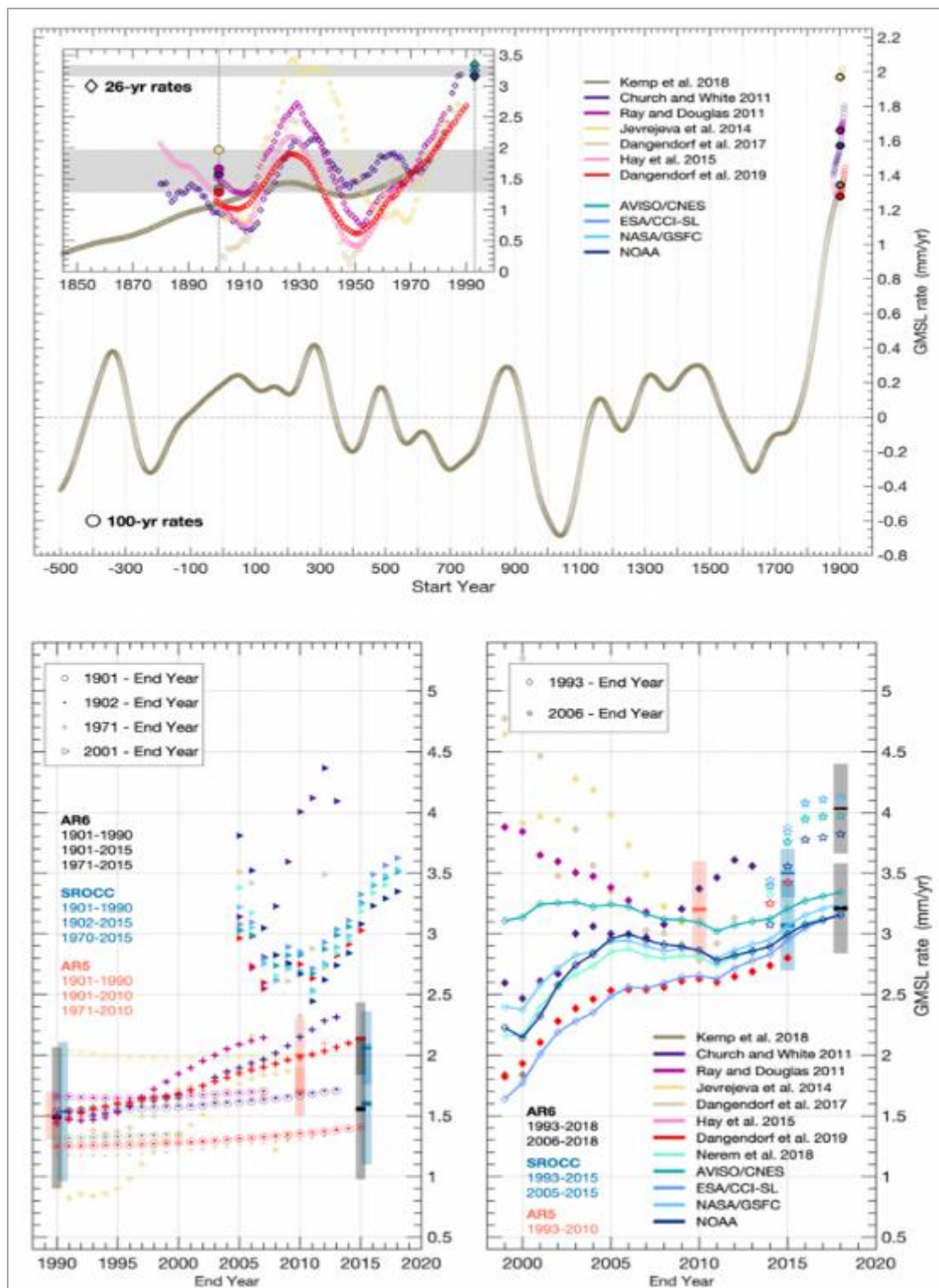


Figure 2.33: Linear rates from the global mean sea level timeseries in Figure 2.32. (Top panel). 100-year (diamond) and 26-year (circles) rate intervals. Observed rate ranges for the 20th century (1901–2000) and satellite altimeter era (1993–2018) are highlighted by gray shading (inset) while individual rates by symbols with black contours. (Bottom panels). Linear rates for different start years (symbols) and end years (x-axis). Left panel: linear rates starting in 1901/2, 1971 and 2001. Right: linear rates starting in 1993 (satellite altimeter era) and 2006 (Argo–GRACE era). Vertical bars are linear rates estimated in AR5 (red), SROCC (blue) and current AR6 (gray). Linear rates were calculated using ordinary least squares. Units: mm yr⁻¹. [uncertainties for individual RATES will be included]

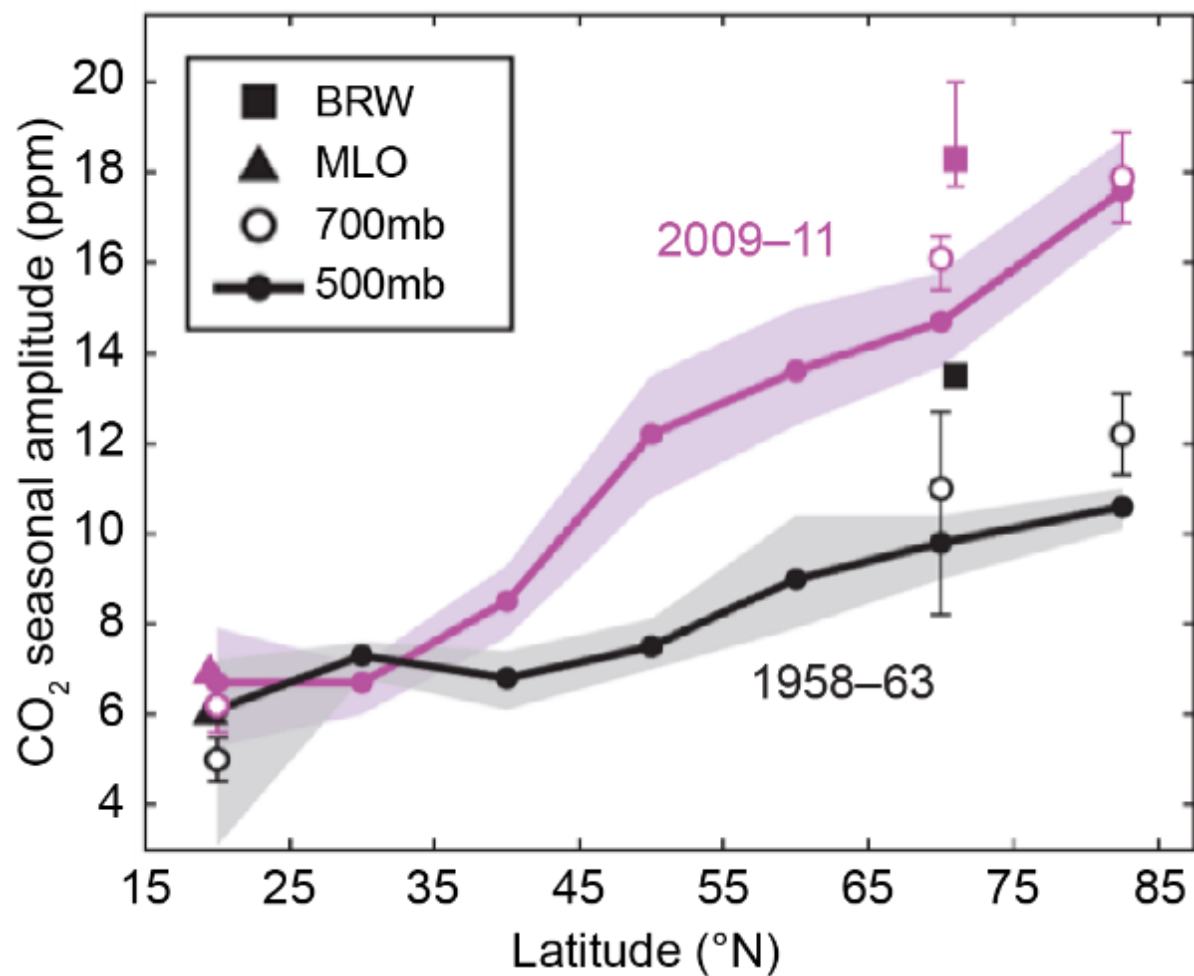


Figure 2.34: Changes in the amplitude of the seasonal cycle of atmospheric CO₂. Squares depict amplitudes at Barrow, and triangles depict amplitudes at Mauna Loa. The black line illustrates airborne data from a 1958-63 field campaign, and the magenta line illustrates airborne data from a 2009-11 field campaign.

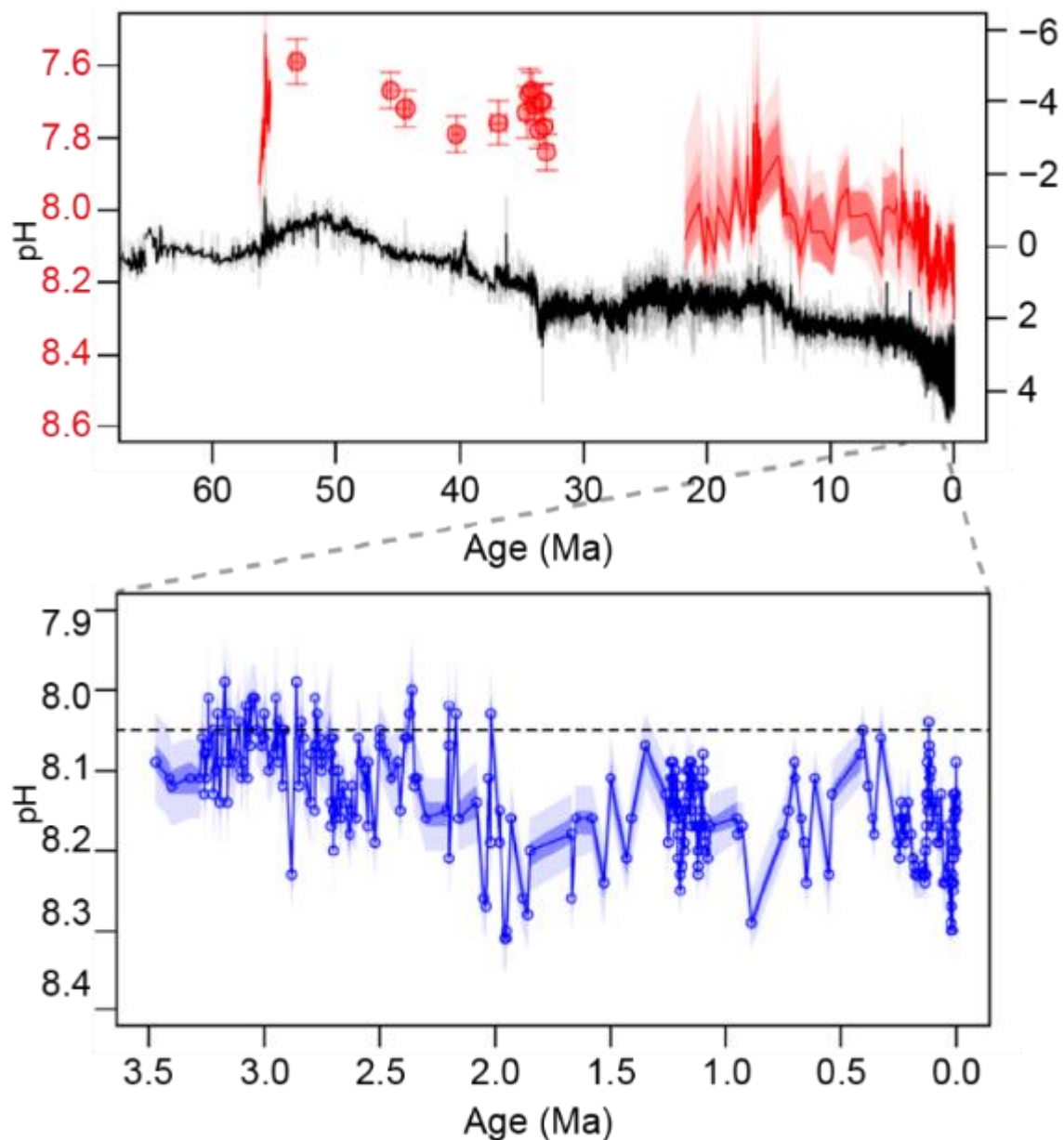


Figure 2.35: a) Low-latitude surface ocean pH over the last 65 million years, reconstructed using boron isotopes in foraminifera (red). Uncertainty is shown at 95% confidence as error bars or a shaded band. Data include: Gutjahr et al. (2017); Anagnostou et al. (2016) and Sossian et al. (2018). Also shown is the benthic $\delta^{18}\text{O}$ stack of Cramer et al. (2009; black – right axis). b) Low-latitude surface ocean pH over the last 3.5 million years reconstructed using boron isotopes in foraminifera (Sossian et al. 2018). Uncertainty at 95% confidence is shown as blue shaded bands. The dotted horizontal line shows an acidification of 0.1 pH units relative to the Holocene value. The Pliocene (prior to 2.5 Ma) is the last interval when pH was as low as today for a sustained interval (hence medium confidence that modern pH is unusual in the last 3 million years).

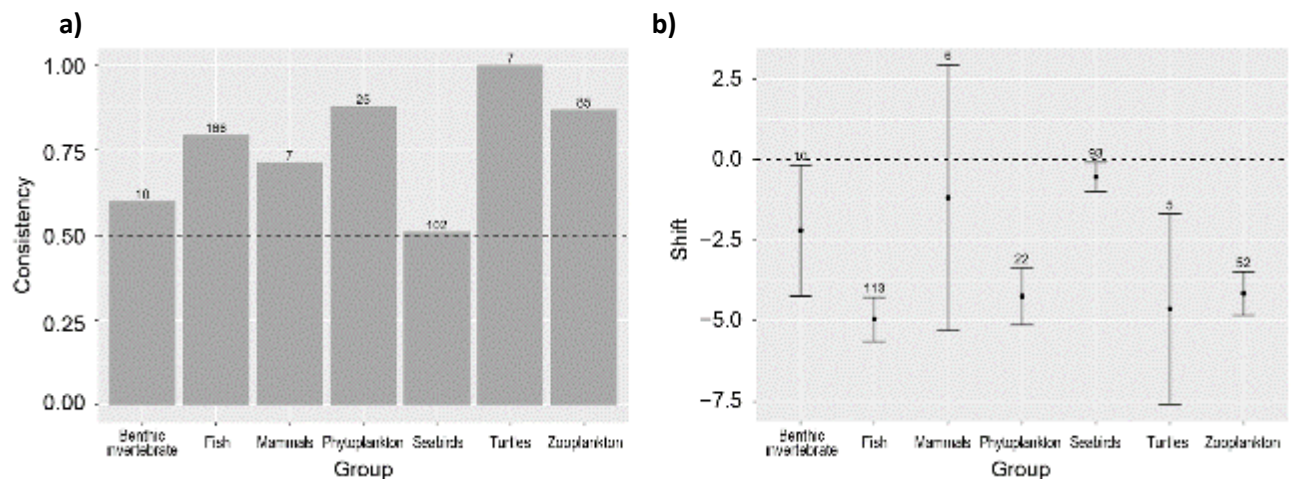


Figure 2.36: Observed responses to climate change by marine species, from time series >19 years in length (median 36 years). (a) The proportion consistent with climate change. (b) The shift in time (days per decade). Based on an update to the global database of (Poloczanska et al., 2013). Database contained 157 peer-reviewed papers published in the last 20 years, covering 100s of species, and including 391 time series >19 years in length. As in Poloczanska et al. (2013), time series were classified as being consistent (i.e., earlier timing with warming) or inconsistent with climate change (i.e., later timing), with non-responsive species being excluded.

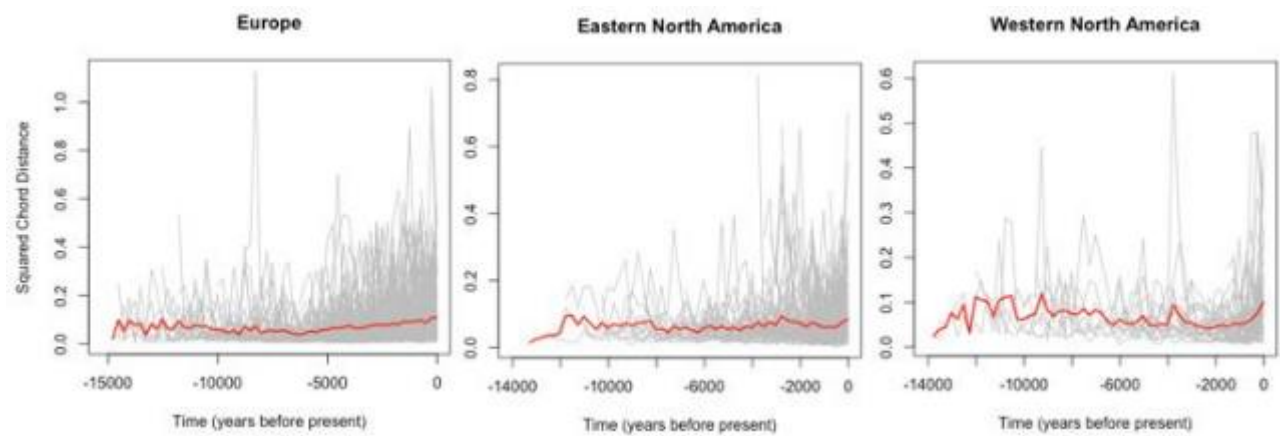


Figure 2.37: Rates of compositional turnover in Europe, eastern North America, and western North America. Derived from Neotoma pollen data and prior work (Finsinger et al., 2017; Shuman et al., 2009). Squared chord distance (SCD) is a metric of compositional change (Overpeck et al., 1985).

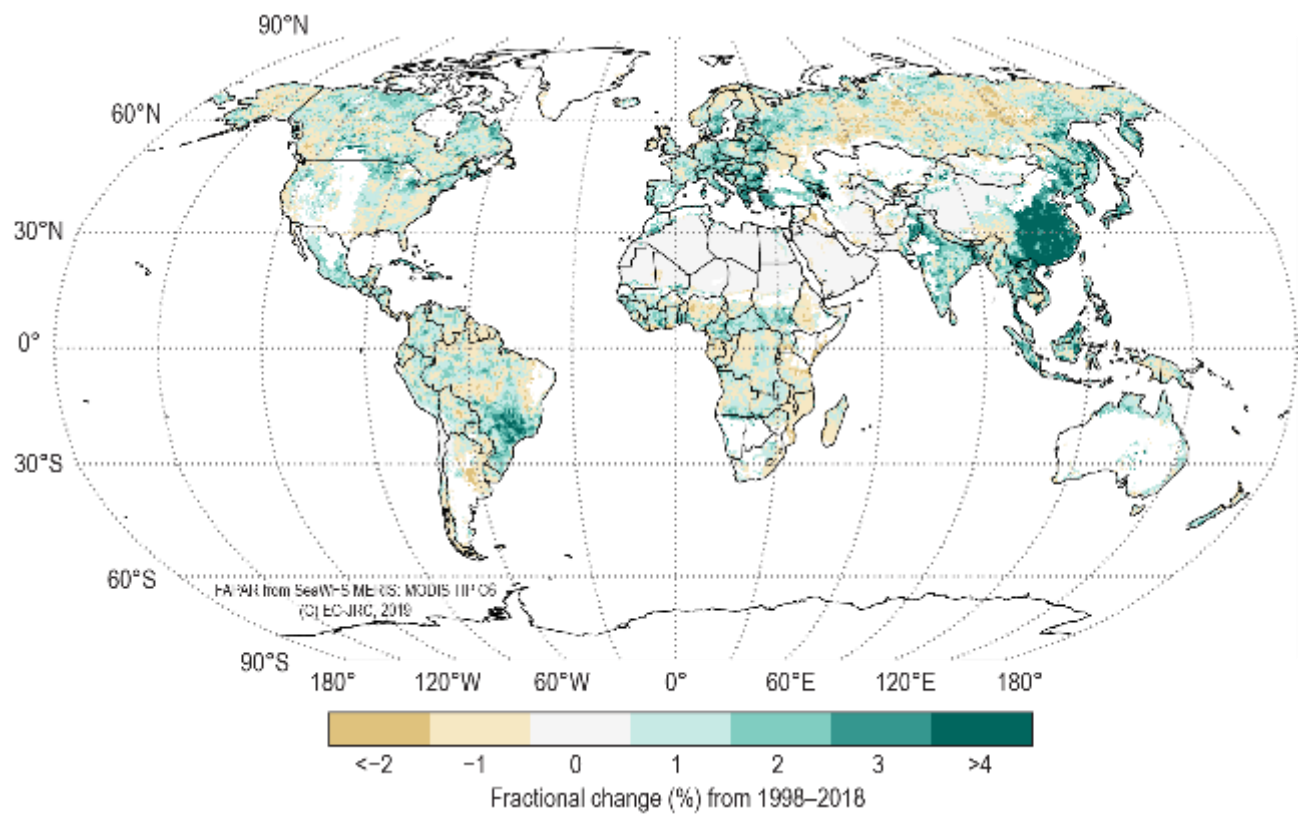


Figure 2.38: Satellite-based trends in Fraction of Absorbed Photosynthetically Active Radiation (per decade) for 1998-2018 (updated from Gobron, 2017). White areas indicate trends that are not statistically significant ($p < 0.1$).

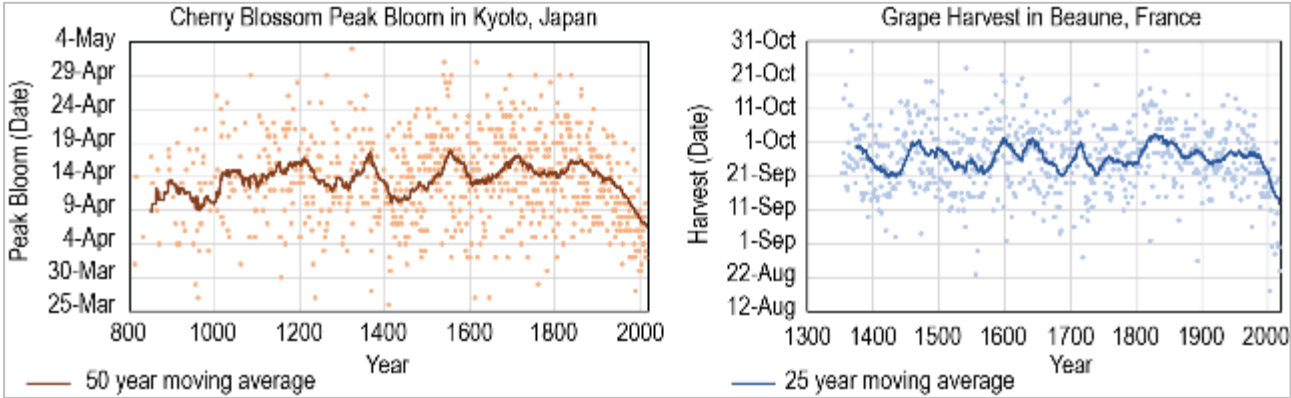


Figure 2.39: Phenological indicators of changes in growing season. The left panel depicts the time series of cherry blossom peak bloom dates in Kyoto, Japan (from Aono and Saito, 2010). The right panel depicts the time series of grape harvest dates in Beaune, France, from Labbe et al. (2019).

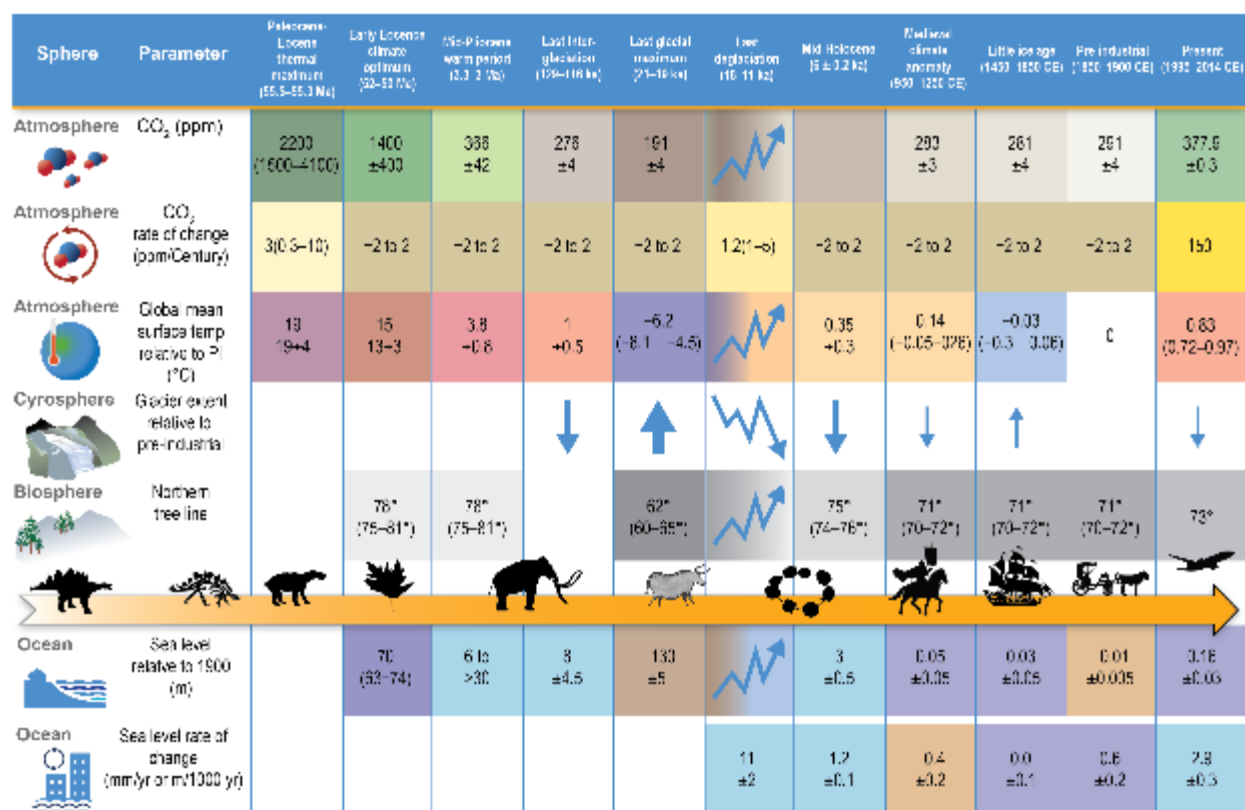


Figure 2.40: Pictorial representation of the changes in selected key indicators at various timeslices from the deep past to the present. Numbers are based upon the assessment carried out in the present chapter and other chapters of the report.

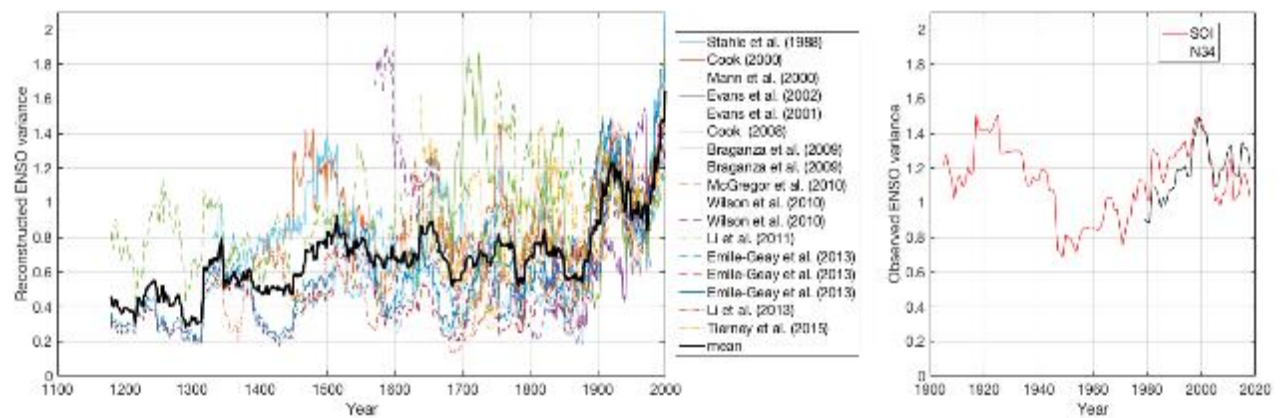


Figure 2.41: 30-year running variance of (left) reconstructed annual mean Niño 3.4 from a range of proxy-based studies (Stahle et al. 1998, Cook 2000, Mann et al. 2000, Evans et al. 2001, Evans et al. 2001, Cook 2008, Braganza et al. 2009, McGregor et al. 2010, Wilson et al. 2010, Li et al. 2011, Emile-Geay et al. 2013, Li et al. 2013, Tierney et al. 2015) (right) the June–November Southern Oscillation Index and April–March mean Niño 3.4 (from ERSSTv5, 1981–2010 base period).

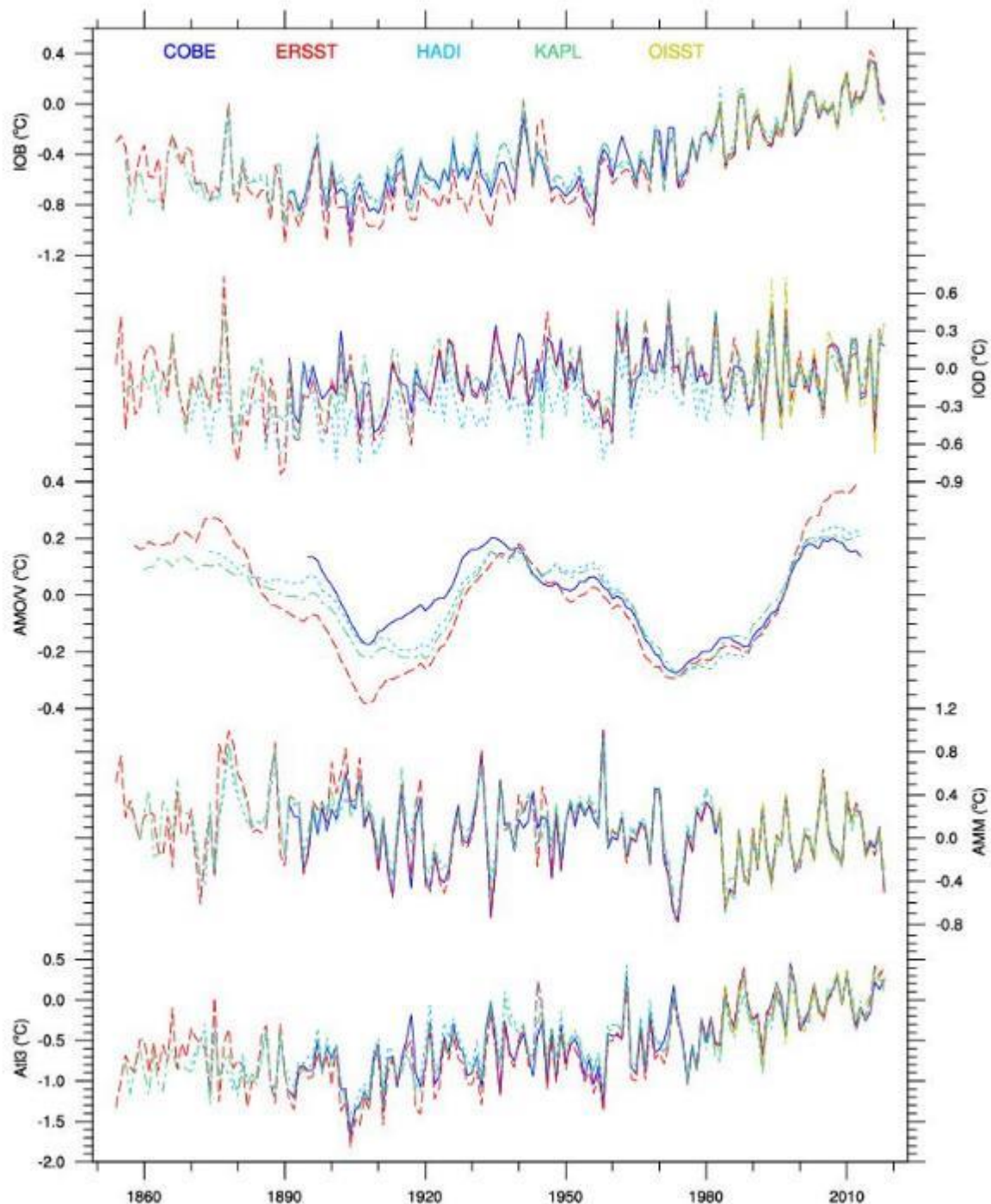
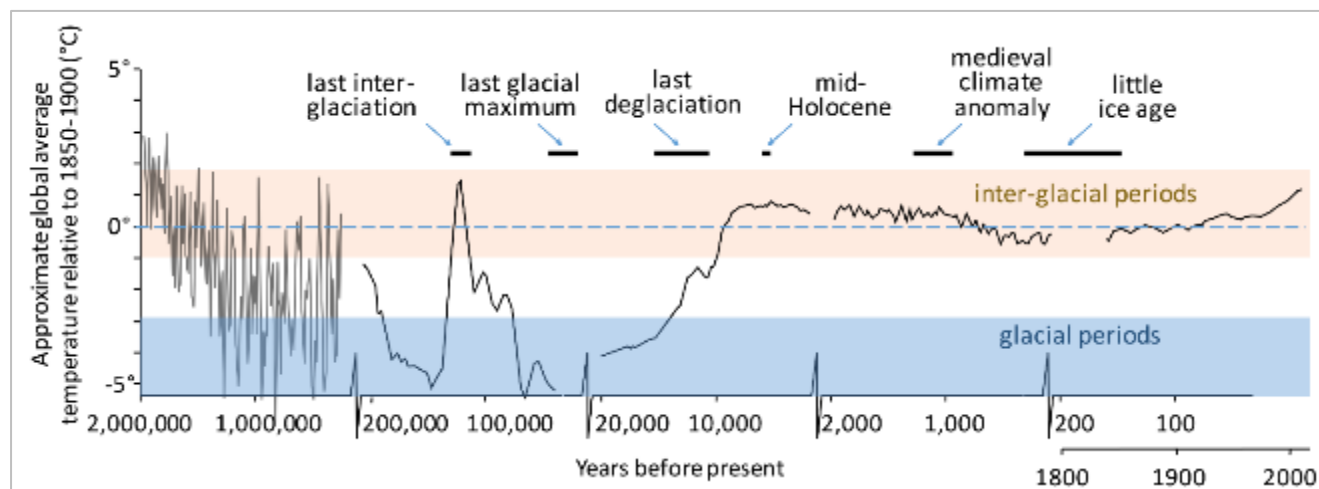


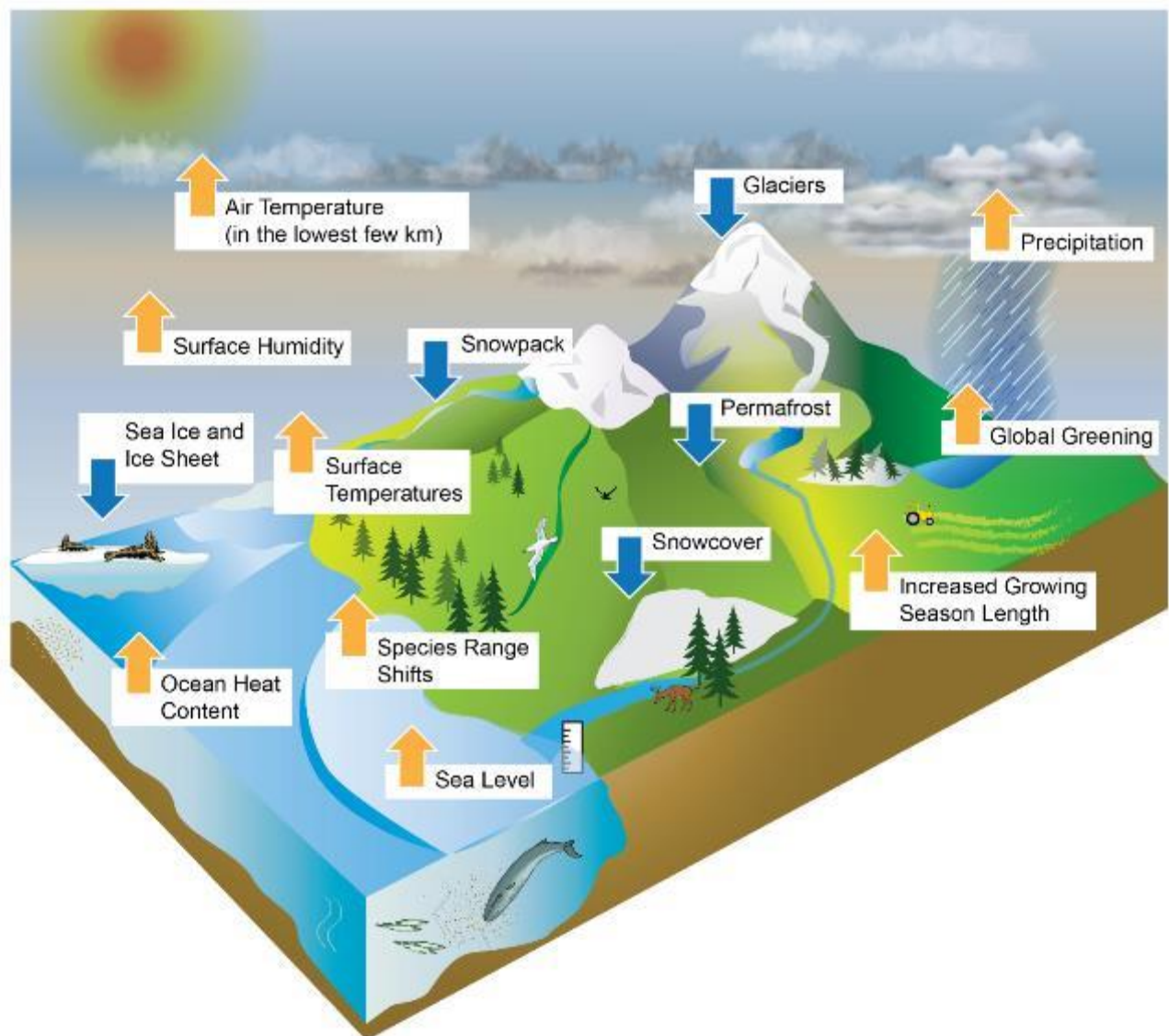
Figure 2.42: Climate indices for the Indian and Atlantic Oceans in different historical observational data sets, 1854–2018. Shown are the following indices from the top: Indian Ocean Basin (IOB) mode, Indian Ocean Dipole (IOD), Atlantic Multidecadal Oscillation/Variability (AMO/V), Atlantic meridional mode (AMM) and the Atlantic Niño (Atl3). The indices were derived from five different SST data sets: (1) the Objective Analyses of Sea-Surface Temperature and Marine Meteorological Variables for the 20th Century using ICOADS and the Kobe Collection, COBE v2 (Hirahara et al., 2014); (2) the Extended Reconstructed Sea Surface Temperature, ERSST v5 (Huang et al., 2017); (3) the Hadley Centre Global Sea Ice and Sea Surface Temperature, HadISST v1 (Rayner et al., 2003); (4) the Kaplan Extended SST dataset, KAPLAN v2 (Kaplan et al., 1998); and (5) the satellite-derived Optimum Interpolation SST (OISST) (Reynolds et al., 2002).

FAQs



FAQ 2.1, Figure 1: Approximate average global surface temperature (y-axis) over the past two million years, combined from multiple sources and shown at five different time scales (x-axis). Temperature is relative to 1850-1900 (pre-industrial reference); 2,000,000 to 300,000 years ago and 200,000 to 30,000 years ago from marine oxygen isotopes; 20,000 to 10,000 years ago from multi-proxy records; 10,000 to 3,000 years ago from multi-proxy records; 2,000 to 300 years ago from multi-proxy records, mainly tree rings; 1800 to 2015 CE from instrumental temperature records.

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FAQ 2.2, Figure 1: Independent analyses of many components of the climate system that would be expected to change in a warming world exhibit trends consistent with warming.

2.SM Supplementary Material Chapter 2

Table 2 SM.1: Ozone trends at remote surface locations worldwide, and in the free troposphere, as reported in the peer-reviewed literature. Measurements were made using modern UV-absorption methods, ozonesondes or satellite instruments. Historical data sets (1896-1975) discussed by Tarasick et al. (2019) are not shown. In most cases the ozone time series are longer than 20 years and the trend is reported for the longest period available. When possible, trends are also reported for the recent period beginning in approximately 1995. When available, trends are reported with 95% confidence intervals. When p-values are less than 0.05 trends are presented in bold font; when p-values are in the range 0.05-0.10 trends are presented in italics. When available, trends are reported in units of percent and one of the following: ppbv for surface (Sfc) and free troposphere (FT); DU for tropospheric column (TC); Tg for the total mass (TM) of ozone in a latitude band. Note: while Mauna Loa and South Pole are high elevation surface stations, their data are considered to be representative of the lower free troposphere.

Location and length of full ozone record	ozone increase over the full record % and % decade ⁻¹	ozone increase since approx. 1995 % and % decade ⁻¹	ozone increase over the full record ppbv and ppbv decade ⁻¹	ozone increase since approx. 1995 ppbv and ppbv decade ⁻¹	Reference and notes
Northern Polar Region					
Sfc: Alert, Canada 82.5° N, -62.5° E, 210 m 1992-2012	4 % 2 % decade ⁻¹	-1 % -0.5 % decade ⁻¹	1.3 ppbv 0.7 ± 1.0 ppbv decade ⁻¹	-0.3 ppbv -0.2 ± 1.2 ppbv decade ⁻¹	Cooper et al. [2019],
Sfc: Zeppelin, Svalbard 78.9° N, 11.9° E, 474 m 1989-2017	5 % 2 % decade ⁻¹	2 % 1 % decade ⁻¹	1.8 ppbv 0.6 ± 0.7 ppbv decade ⁻¹	0.7 ppbv 0.3 ± 1.0 ppbv decade ⁻¹	Cooper et al. [2019],
Sfc: Barrow, Alaska 71.3° N, -156.6° E, 11 m 1973-2018	13 % 3 % decade ⁻¹	7 % 3 % decade ⁻¹	2.8 ppbv 0.6 ± 0.4 ppbv decade ⁻¹	2.3 ppbv 1.0 ± 0.9 ppbv decade ⁻¹	Cooper et al. [2019],
Sfc: Esrange, Sweden 67.9° N, 21.1° E, 524 m 1991-2017	7 % 3 % decade ⁻¹	0 % 0 % decade ⁻¹	2.0 ppbv 0.8 ± 0.7 ppbv decade ⁻¹	0.1 ppbv 0 ± 0.8 ppbv decade ⁻¹	Cooper et al. [2019],
FT: Canada, 1966-2013	Insignificant changes	Insignificant changes	Insignificant changes	Insignificant changes	Tarasick et al. [2016]. Three ozonesonde sites
Northern mid-latitudes					
Sfc: Tustervatn, Norway 65.8° N, 13.9° E, 439 m	9 % 3 % decade ⁻¹	-4 % -2 % decade ⁻¹	2.6 ppbv 0.9 ± 0.8 ppbv decade ⁻¹	-1.3 ppbv -0.6 ± 0.7 ppbv decade ⁻¹	Cooper et al. [2019],

1989-2017					
Sfc: Denali, Alaska 63.7° N, -149.0° E, 663 m 1987-2017	8 % 3 % decade ⁻¹	4 % 2 % decade ⁻¹	2.2 ppbv 0.7 ± 0.6 ppbv decade ⁻¹	1.5 ppbv 0.7 ± 0.9 ppbv decade ⁻¹	Cooper et al. [2019],
Sfc: Strath Vaich Dam UK, 57.7° N, -4.8° E, 266 m, 1987-2017	2 % 1 % decade ⁻¹	-3 % -1 % decade ⁻¹	0.7 ppbv 0.2 ± 0.7 ppbv decade ⁻¹	-0.9 ppbv -0.4 ± 1.0 ppbv decade ⁻¹	Cooper et al. [2019],
Sfc: Mace Head, Ireland 53.3° N, -9.9° E, 8 m 1988-2017	8 % 3 % decade ⁻¹	4 % 2 % decade ⁻¹	3.1 ppbv 1.1 ± 0.5 ppbv decade ⁻¹	1.5 ppbv 0.7 ± 0.8 ppbv decade ⁻¹	Cooper et al. [2019],
Sfc: Zugspitze, Germany 47.4° N, 11.0° E, 2800 m 1978-2017	26 % 7 % decade ⁻¹	-4 % -2 % decade ⁻¹	9.1 ppbv 2.3 ± 0.7 ppbv decade ⁻¹	-1.8 ppbv -0.8 ± 0.6 ppbv decade ⁻¹	Cooper et al. [2019],
Sfc: Sonnblick, Austria 47.1° N, 13.0° E, 3106 m 1990-2016	0 % 0 % decade ⁻¹	-5 % -2 % decade ⁻¹	0.0 ppbv -0.0 ± 0.6 ppbv decade ⁻¹	-2.1 ppbv -1.0 ± 0.7 ppbv decade ⁻¹	Cooper et al. [2019],
Sfc: Jungfrauoch, CH 46.5° N, 8.0° E, 3580 m 1986-2017	5 % 2 % decade ⁻¹	0 % 0 % decade ⁻¹	3.3 ppbv 1.1 ± 0.6 ppbv decade ⁻¹	0.0 ppbv 0.0 ± 0.6 ppbv decade ⁻¹	Cooper et al. [2019],
Sfc: Kislovodsk, Russia 43.7° N, 42.7° E, 2096 m 1989-2013	-17 % -7 % decade ⁻¹	-5 % -3 % decade ⁻¹	-8.2 ppbv -3.4 ± 1.5 ppbv decade ⁻¹	-2.0 ppbv -1.1 ± 1.6 ppbv decade ⁻¹	Cooper et al. [2019],
Sfc: Centennial, WY 41.4°N, -106°E, 3175 m 1989-2017	-3 % -1 % decade ⁻¹	-8 % -4 % decade ⁻¹	-1.8 ppbv -0.6 ± 0.9 ppbv decade ⁻¹	-3.2 ppbv -1.5 ± 1.2 ppbv decade ⁻¹	Cooper et al. [2019],
Sfc: Great Basin, NV 39.0°N, -114.°E, 2058 m 1993-2017	3 % 1 % decade ⁻¹	2 % 1 % decade ⁻¹	1.3 ppbv 0.5 ± 0.9 ppbv decade ⁻¹	0.5 ppbv 0.2 ± 1.0 ppbv decade ⁻¹	Cooper et al. [2019],
Sfc: Gothic, Colorado 39.0° N, -107° E, 2915 m 1989-2017	-5 % -2 % decade ⁻¹	-10 % -5 % decade ⁻¹	-2.7 ppbv -1.0 ± 0.6 ppbv decade ⁻¹	-4.3 ppbv -1.9 ± 0.8 ppbv decade ⁻¹	Cooper et al. [2019],
Sfc: Grand Canyon, AZ 36.1°N, -112°E, 2073 m 1989-2017	-2 % -1 % decade ⁻¹	-9 % -4 % decade ⁻¹	-1.0 ppbv -0.3 ± 0.7 ppbv decade ⁻¹	-3.4 ppbv -1.5 ± 0.8 ppbv decade ⁻¹	Cooper et al. [2019],
Sfc: Mt. Waliguan, China 36.3° N, 100.9°E, 3810 m 1995-2016	9 % 4 % decade ⁻¹	8 % 4 % decade ⁻¹	4.0 ppbv 1.8 ± 0.5 ppbv decade ⁻¹	3.6 ppbv 1.7 ± 0.5 ppbv decade ⁻¹	Cooper et al. [2019],

Sfc: Izaña, Canary Is. 28.3° N, -16.5° E, 2367 m 1987-2017	8 % 3 % decade ⁻¹	3 % 1 % decade ⁻¹	3.9 ppbv 1.3 ± 0.6 ppbv decade ⁻¹	1.1 ppbv 0.5 ± 0.9 ppbv decade ⁻¹	Cooper et al. [2019],
Sfc: Minamitorishima, N. Pacific Ocean 24.3° N, 154° E, 8 m 1994-2017	-10 % -4 % decade ⁻¹	-9 % -4 % decade ⁻¹	-4.0 ppbv -1.7 ± 1.1 ppbv decade ⁻¹	-3.7 ppbv -1.7 ± 1.2 ppbv decade ⁻¹	Cooper et al. [2019],
650 hPa: IAGOS Europe 1994-2016	3 %, 2 % decade ⁻¹	3 %, 2 % decade ⁻¹	1.5 ppbv, 0.9 ppbv decade ⁻¹	1.5 ppbv, 0.9 ppbv decade ⁻¹	Gaudel et al. [2019], Table S2, commercial aircraft observations at 650 hPa
650 hPa: IAGOS northeast USA, 1994- 2016	4 %, 2 % decade ⁻¹	4 %, 2 % decade ⁻¹	1.9 ppbv, 1.2 ppbv decade ⁻¹	1.9 ppbv, 1.2 ppbv decade ⁻¹	Gaudel et al. [2019], Table S2, commercial aircraft observations at 650 hPa
650 hPa: IAGOS China/S. Korea, 1994- 2016	8 %, 5 % decade ⁻¹	8 %, 5 % decade ⁻¹	4.5 ppbv, 2.8 ppbv decade ⁻¹	4.5 ppbv, 2.8 ppbv decade ⁻¹	Gaudel et al. [2019], Table S2, commercial aircraft observations at 650 hPa
650 hPa: IAGOS Persian Gulf, 1998- 2016	15 %, 12 % decade ⁻¹	15 %, 12 % decade ⁻¹	8.9 ppbv, 6.9 ppbv decade ⁻¹	8.9 ppbv, 6.9 ppbv decade ⁻¹	Gaudel et al. [2019], Table S2, commercial aircraft observations at 650 hPa
FT: Tsukuba, Japan, 700-500 hPa, 1971- 2010	12.0 ± 7.6 %, 3.0 ± 1.9 % decade ⁻¹		6.4 ± 4.0 ppbv 1.6 ± 1.0 ppbv decade ⁻¹		Oltmans et al. [2013], ozonesondes
FT: Tsukuba, Japan, 500-300 hPa, 1971- 2010	21.6 ± 12.0 %, 5.4 ± 3.0 % decade ⁻¹		13.6 ± 8.0 ppbv 3.4 ± 2.0 ppbv decade ⁻¹		Oltmans et al. [2013], ozonesondes
FT: Sapporo, Japan, 700-500 hPa, 1971- 2010	17.6 ± 8.0 %, 4.4 ± 2.0 % decade ⁻¹		9.2 ± 4.0 ppbv 2.3 ± 1.0 ppbv decade ⁻¹		Oltmans et al. [2013], ozonesondes
FT: Sapporo, Japan, 500-300 hPa, 1971- 2010	9.6 ± 12.0 %, 2.4 ± 3.0 % decade ⁻¹		6.4 ± 8.0 ppbv 1.6 ± 2.0 ppbv decade ⁻¹		Oltmans et al. [2013], ozonesondes
FT: Wallops Island, Virginia, 700-500 hPa, 1971-2010	6.4 ± 6.4 %, 1.6 ± 1.6 % decade ⁻¹		3.6 ± 4.0 ppbv 0.9 ± 1.0 ppbv decade ⁻¹		Oltmans et al. [2013], ozonesondes

FT: Wallops Island, Virginia, 500-300 hPa, 1971-2010	$-0.4 \pm 10.4 \%$, $-0.1 \pm 2.6 \%$ decade ⁻¹		-0.0 ± 7.2 ppbv -0.0 ± 1.8 ppbv decade ⁻¹		Oltmans et al. [2013], ozonesondes
FT: Hohenpeissenberg, Germany, 700-500 hPa, 1971-2010	$19.6 \pm 4.8 \%$, $4.9 \pm 1.2 \%$ decade ⁻¹		10.4 ± 2.4 ppbv 2.6 ± 0.6 ppbv decade ⁻¹		Oltmans et al. [2013], ozonesondes
FT: Hohenpeissenberg, Germany, 500-300 hPa, 1971-2010	$21.6 \pm 6.0 \%$, $5.4 \pm 1.5 \%$ decade ⁻¹		13.6 ± 4.0 ppbv 3.4 ± 1.0 ppbv decade ⁻¹		Oltmans et al. [2013], ozonesondes
FT: Uccle, Belgium, 1969-2014	9-14 % 2-3 % decade ⁻¹	5-14 %, 3-8 % decade ⁻¹			Van Malderen et al. [2016], corrected ozonesondes, 0-10 km below tropopause
FT: Canada, 1966-2013	Insignificant changes	Insignificant changes	Insignificant changes	Insignificant changes	Tarasick et al. [2016]. Three ozonesonde sites
FT: IAGOS Europe 1994-2016	4 %, 3 % decade ⁻¹	4 %, 3 % decade ⁻¹	3 ppbv, 2 ppbv decade ⁻¹	3 ppbv, 2 ppbv decade ⁻¹	Gaudel et al. [2019], commercial aircraft profiles, based on Table S2, 700-300 hPa
FT: IAGOS northeast USA, 1994-2016	6 %, 4 % decade ⁻¹	6 %, 4 % decade ⁻¹	4 ppbv, 2 ppbv decade ⁻¹	4 ppbv, 2 ppbv decade ⁻¹	Gaudel et al. [2019], commercial aircraft profiles, based on Table S2, 700-300 hPa
FT: IAGOS China/S. Korea, 1994-2016	7 %, 4 % decade ⁻¹	7 %, 4 % decade ⁻¹	4 ppbv, 2 ppbv decade ⁻¹	4 ppbv, 2 ppbv decade ⁻¹	Gaudel et al. [2019], commercial aircraft profiles, based on Table S2, 700-300 hPa
FT: IAGOS Persian Gulf, 1998-2016	13 %, 10 % decade ⁻¹	13 %, 10 % decade ⁻¹	8 ppbv, 6 ppbv decade ⁻¹	8 ppbv, 6 ppbv decade ⁻¹	Gaudel et al. [2019], commercial aircraft profiles, based on Table S2, 700-300 hPa
FT: IAGOS upper troposphere above 6 regions, 1994-2013	7-12 %, 4-7 % decade ⁻¹	7-12 %, 4-7 % decade ⁻¹	4.5-8.1 ppbv, 2.5-4.5 ppbv decade ⁻¹	4.5-8.1 ppbv, 2.5-4.5 ppbv decade ⁻¹	Cohen et al. [2018], trend values span NE USA, Europe, NE China, SE Asia, India, Persian Gulf
FT: IAGOS upper troposphere above N. Atlantic, 1994-2013			5 ppbv, 3 ppbv decade ⁻¹	5 ppbv, 3 ppbv decade ⁻¹	Cohen et al. [2018], the only region with no significant trend was the North Atlantic Ocean

FT: western North America, 3-8 km, April-May, 1995-2014	11 ± 7 %, 5 ± 4 % decade ⁻¹	11 ± 7 %, 5 ± 4 % decade ⁻¹	6.2 ± 4.2 ppbv, 3.1 ± 2.1 ppbv decade ⁻¹	6.2 ± 4.2 ppbv, 3.1 ± 2.1 ppbv decade ⁻¹	Lin et al. [2015], data are a composite of all available ozonesondes and aircraft obs.
TC: 30°-60°N latitude band, 1998-2012	8 ± 2 %, 6 ± 1 % decade ⁻¹	8 ± 2 %, 6 ± 1 % decade ⁻¹	2.4 ± 0.6 DU, 1.7 ± 0.4 DU decade ⁻¹	2.4 ± 0.6 DU, 1.7 ± 0.4 DU decade ⁻¹	Gaudel et al. [2018], based on the TOST ozonesonde product
Northern Tropics					
FT: Mauna Loa, 3.4 km above sea level, all data, 1973-2018	17 %, 4 % decade ⁻¹	6 %, 3 % decade ⁻¹	5.9 ppbv, 1.3 ± 0.5 ppbv decade ⁻¹	2.2 ppbv, 0.9 ± 1.4 ppbv decade ⁻¹	Cooper et al. [2019], all data, restricted to nighttime observations.
FT: Mauna Loa, dry air, 1974-2018	24 %, 5 % decade ⁻¹	13 %, 6 % decade ⁻¹	9.1 ppbv, 2.1 ± 0.4 ppbv decade ⁻¹	5.6 ppbv, 2.5 ± 1.1 ppbv decade ⁻¹	Cooper et al. [2019], dry air masses at Mauna Loa have a mid-latitude bias
FT: Mauna Loa, moist air, 1974-2018	15 %, 3 % decade ⁻¹	7 %, 3 % decade ⁻¹	4.6 ppbv, 1.1 ± 0.3 ppbv decade ⁻¹	2.1 ppbv, 0.9 ± 0.9 ppbv decade ⁻¹	Cooper et al. [2019], moist air masses at Mauna Loa have a tropical bias
650 hPa: IAGOS India, 1994-2016	11 %, 7 % decade ⁻¹	11 %, 7 % decade ⁻¹	5.2 ppbv, 3.3 ppbv decade ⁻¹	5.2 ppbv, 3.3 ppbv decade ⁻¹	Gaudel et al. [2019], Table S2, commercial aircraft observations at 650 hPa
650 hPa: IAGOS SE Asia, 1994-2016	26 %, 16 % decade ⁻¹	26 %, 16 % decade ⁻¹	10.4 ppbv, 6.5 ppbv decade ⁻¹	10.4 ppbv, 6.5 ppbv decade ⁻¹	Gaudel et al. [2019], Table S2, commercial aircraft observations at 650 hPa
FT: IAGOS India, 1994-2016	13 %, 8 % decade ⁻¹	13 %, 8 % decade ⁻¹	7 ppbv, 4 ppbv decade ⁻¹	7 ppbv, 4 ppbv decade ⁻¹	Gaudel et al. [2019], commercial aircraft profiles, based on Table S2, 700-300 hPa
FT: IAGOS SE Asia, 1994-2016	21 %, 13 % decade ⁻¹	21 %, 13 % decade ⁻¹	10 ppbv, 6 ppbv decade ⁻¹	10 ppbv, 6 ppbv decade ⁻¹	Gaudel et al. [2019], commercial aircraft profiles, based on Table S2, 700-300 hPa
TC: 0°-30°N latitude band, 1998-2012	14 ± 4 %, 10 ± 3 % decade ⁻¹	14 ± 4 %, 10 ± 3 % decade ⁻¹	4.3 ± 1.3 DU, 3.1 ± 0.9 DU decade ⁻¹	4.3 ± 1.3 DU, 3.1 ± 0.9 DU decade ⁻¹	Gaudel et al. [2018], based on the TOST ozonesonde product

TM: 0° – 30°N latitude band, 1979-2016	10 ± 6 %, 3 ± 2 % decade ⁻¹		8.0 ± 4.6 Tg, 2.1 ± 1.2 Tg decade ⁻¹		Ziemke et al. [2019], a composite of the TOMS (1979-2005) and OMI/MLS (2005-2016) satellite products
Equatorial region, spanning the Northern and Southern Hemispheres					
650 hPa: IAGOS Malaysia/Indonesia, 1994-2016	25 %, 15 % decade ⁻¹	25 %, 15 % decade ⁻¹	7.2 ppbv, 4.2 ppbv decade ⁻¹	7.2 ppbv, 4.2 ppbv decade ⁻¹	Gaudel et al. [2019], Table S2, commercial aircraft observations at 650 hPa
FT: IAGOS Malaysia/Indonesia, 1995-2016	12 %, 7 % decade ⁻¹	12 %, 7 % decade ⁻¹	4 ppbv 2 ppbv decade ⁻¹	4 ppbv 2 ppbv decade ⁻¹	Gaudel et al. [2019], commercial aircraft profiles, based on Table S2, 700-300 hPa
TC: 15° S – 15° N, 1995-2015	-2 ± 6 %, -1 ± 3 % decade ⁻¹	-2 ± 6 %, -1 ± 3 % decade ⁻¹	-0.4 ± 1.2 DU, -0.2 ± 0.6 DU decade ⁻¹	-0.4 ± 1.2 DU, -0.2 ± 0.6 DU decade ⁻¹	Leventidou et al. [2018], satellite detected trends from a series of satellite products. Several regions within this zone had significant ozone increases.
TC: 20° S – 20° N, 1995-2015	7 ± 1 %, 4 ± 1 % decade ⁻¹	7 ± 1 %, 4 ± 1 % decade ⁻¹	1.4 ± 0.2 DU, 0.7 ± 0.1 DU decade ⁻¹	1.4 ± 0.2 DU, 0.7 ± 0.1 DU decade ⁻¹	Heue et al. [2016], satellite detected trends from a series of satellite products.
Southern tropics					
Sfc: American Samoa South Pacific Ocean -14.3° N, -171.6° E, 42 m 1975-2015	7 % 2 % decade ⁻¹	37 % 19 % decade ⁻¹	0.9 ppbv 0.2 ± 0.2 ppbv decade ⁻¹	2.9 ppbv 1.5 ± 0.7 ppbv decade ⁻¹	Cooper et al. [2019],
FT: Réunion Island, South Indian Ocean, 1992-2011	Annual trends not reported. Winter trends are 35-45 % decade ⁻¹				Thompson et al. [2014], SHADOZ ozonesondes; uncertainty estimates were not reported
FT: Irene, South Africa, 1990-2007	Annual trends not reported. Winter trends are ~25 % decade ⁻¹				Thompson et al. [2014], SHADOZ ozonesondes; uncertainty estimates were not reported
TC: 0°-30°S latitude band, 1998-2012	26 ± 4 %, 18 ± 3 % decade ⁻¹	26 ± 4 %, 18 ± 3 % decade ⁻¹	6.7 ± 1.1 DU, 4.8 ± 0.8 DU decade ⁻¹	6.7 ± 1.1 DU, 4.8 ± 0.8 DU decade ⁻¹	Gaudel et al. [2018], based on the TOST

					ozonesonde product
TM: 30° S – 0° latitude band, 1979-2016	6 ± 5 %, 2 ± 1 % decade ⁻¹		4.4 ± 4.2 Tg 1.2 ± 1.1 Tg decade ⁻¹		Ziemke et al. [2019], a composite of the TOMS (1979-2005) and OMI/MLS (2005-2016) satellite products
Southern mid-latitudes					
Sfc: Cape Point, SA -34.4° N, 18.5° E, 230 m 1983-2017	37 % 11 % decade ⁻¹	20 % 9 % decade ⁻¹	5.5 ppbv 1.6 ± 0.3 ppbv decade ⁻¹	2.7 ppbv 1.2 ± 0.6 ppbv decade ⁻¹	Cooper et al. [2019],
Sfc: Cape Grim, AU -40.7° N, 144.7° E, 94 m 1982-2017	10 % 3 % decade ⁻¹	3 % 1 % decade ⁻¹	1.6 ppbv 0.4 ± 0.2 ppbv decade ⁻¹	0.6 ppbv 0.3 ± 0.3 ppbv decade ⁻¹	Cooper et al. [2019],
Sfc: Ushuaia, Argentina -54.8° N, -68.3° E, 18 m 1994-2017	-13 % -6 % decade ⁻¹	-14 % -6 % decade ⁻¹	-2.1 ppbv -0.9 ± 0.5 ppbv decade ⁻¹	-2.0 ppbv -0.9 ± 0.5 ppbv decade ⁻¹	Cooper et al. [2019],
FT: Lauder, New Zealand, 3-6 km, 1987-2014			0.6 ± 0.2 ppbv decade ⁻¹		Zeng et al. [2017], ozonesondes Uncertainty is +/- one standard error. Ask Kai-Lan if this can be converted to 2-sigma
FT: Lauder, New Zealand, 6-9 km, 1987-2014			-0.9 ± 0.5 ppbv decade ⁻¹		Zeng et al. [2017], ozonesondes
FT: Lauder, New Zealand, 9-12 km, 1987-2014			-5.5 ± 2.7 ppbv decade ⁻¹		Zeng et al. [2017], ozonesondes
TC: 30°-60°S latitude band, 1998-2012	2 ± 2 %, 2 ± 1 % decade ⁻¹	2 ± 2 %, 2 ± 1 % decade ⁻¹	0.6 ± 0.4 DU, 0.4 ± 0.3 DU decade ⁻¹	0.6 ± 0.4 DU, 0.4 ± 0.3 DU decade ⁻¹	Gaudel et al. [2018], based on the TOST ozonesonde product
Southern polar region					
Sfc: Neumayer, Antarctica -70.7° N, -8.3° E, 42 m 1995-2017	13 % 6 % decade ⁻¹	13 % 6 % decade ⁻¹	2.1 ppbv 1.0 ± 0.4 ppbv decade ⁻¹	2.1 ppbv 1.0 ± 0.4 ppbv decade ⁻¹	Cooper et al. [2019],
FT: South Pole, 2.8 km above sea level, 1975-2018	4 %, 1 % decade ⁻¹	14 % 6 % decade ⁻¹	1.1 ppbv, 0.3 ± 0.4 ppbv decade ⁻¹	3.4 ppbv, 1.5 ± 0.6 ppbv decade ⁻¹	Cooper et al. [2019], Trends based on 1975-2018 data, with 1961-63 historical data being similar

					to present day values
FT: South Pole, 700-500 hPa, 1986-2010	$4.6 \pm 3.8 \%$, $1.9 \pm 1.6 \%$ decade ⁻¹		1.2 ± 1.0 ppbv, 0.5 ± 0.4 ppbv decade ⁻¹		Oltmans et al. [2013], ozonesondes
FT: South Pole, 500-300 hPa, 1986-2010	$7.2 \pm 8.3 \%$, $3.0 \pm 3.4 \%$ decade ⁻¹		2.9 ± 3.4 ppbv, 1.2 ± 1.4 ppbv decade ⁻¹		Oltmans et al. [2013], ozonesondes
FT: Syowa, Antarctica, 700-500 hPa, 1971-2010	$7.6 \pm 6.4 \%$, $1.9 \pm 1.6 \%$ decade ⁻¹		2.0 ± 1.6 ppbv 0.5 ± 0.4 ppbv decade ⁻¹		Oltmans et al. [2013], ozonesondes
FT: Syowa, Antarctica, 500-300 hPa, 1971-2010	$-10 \pm 10 \%$, $-2.5 \pm 2.5 \%$ decade ⁻¹		-3.9 ± 3.9 ppbv -1.0 ± 1.0 ppbv decade ⁻¹		Oltmans et al. [2013], ozonesondes

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