# Frequently Asked Questions

FAQ

# Frequently Asked Questions

These Frequently Asked Questions have been extracted from the chapters of the underlying report and are compiled here. When referencing specific FAQs, please reference the corresponding chapter in the report from where the FAQ originated (e.g., FAQ 3.1 is part of Chapter 3).

# Table of Contents

## Frequently Asked Questions



FAQ

Frequently Asked Questions

## FAQ 1.1 | If Understanding of the Climate System Has Increased, Why Hasn't the Range of Temperature Projections Been Reduced?

*The models used to calculate the IPCC's temperature projections agree on the direction of future global change, but the projected size of those changes cannot be precisely predicted. Future greenhouse gas (GHG) emission rates could take any one of many possible trajectories, and some underlying physical processes are not yet completely understood, making them difficult to model. Those uncertainties, combined with natural year-to-year climate variability, produce an 'uncertainty range' in temperature projections.*

*The uncertainty range around projected GHG and aerosol precursor emissions (which depend on projections of future social and economic conditions) cannot be materially reduced. Nevertheless, improved understanding and climate models—along with observational constraints—may reduce the uncertainty range around some factors that influence the climate's response to those emission changes. The complexity of the climate system, however, makes this a slow process. (FAQ1.1, Figure 1)*

Climate science has made many important advances since the last IPCC assessment report, thanks to improvements in measurements and data analysis in the cryosphere, atmosphere, land, biosphere and ocean systems. Scientists also have better understanding and tools to model the role of clouds, sea ice, aerosols, small-scale ocean mixing, the carbon cycle and other processes. More observations mean that models can now be evaluated more thoroughly, and projections can be better constrained. For example, as models and observational analysis have improved, projections of sea level rise have become more accurate, balancing the current sea level rise budget.

Despite these advances, there is still a range in plausible projections for future global and regional climate—what scientists call an 'uncertainty range'. These uncertainty ranges are specific to the variable being considered (precipitation vs. temperature, for instance) and the spatial and temporal extent (such as regional vs. global averages). Uncertainties in climate projections arise from natural variability and uncertainty around the rate of future emissions and the climate's response to them. They can also occur because representations of some known processes are as yet unrefined, and because some processes are not included in the models.

There are fundamental limits to just how precisely annual temperatures can be projected, because of the chaotic nature of the climate system. Furthermore, decadal-scale projections are sensitive to prevailing conditions—such as the temperature of the deep ocean—that are less well known. Some natural variability over decades arises from interactions between the ocean, atmosphere, land, biosphere and cryosphere, and is also linked to phenomena such as the El Niño-Southern Oscillation (ENSO) and the North Atlantic Oscillation (see Box 2.5 for details on patterns and indices of climate variability).

Volcanic eruptions and variations in the sun's output also contribute to natural variability, although they are externally forced and explainable. This natural variability can be viewed as part of the 'noise' in the climate record, which provides the backdrop against which the 'signal' of anthropogenic climate change is detected.

Natural variability has a greater influence on uncertainty at regional and local scales than it does over continental or global scales. It is inherent in the Earth system, and more knowledge will not eliminate the uncertainties it brings. However, some progress is possible—particularly for projections up to a few years ahead—which exploit advances in knowledge of, for instance, the cryosphere or ocean state and processes. This is an area of active research. When climate variables are averaged over decadal timescales or longer, the relative importance of internal variability diminishes, making the long-term signals more evident (FAQ1.1, Figure 1). This long-term perspective is consistent with a common definition of climate as an average over 30 years.

A second source of uncertainty stems from the many possible trajectories that future emission rates of GHGs and aerosol precursors might take, and from future trends in land use. Nevertheless, climate projections rely on input from these variables. So to obtain these estimates, scientists consider a number of alternative scenarios for future human society, in terms of population, economic and technological change, and political choices. They then estimate the likely emissions under each scenario. The IPCC informs policymaking, therefore climate projections for different emissions scenarios can be useful as they show the possible climatic consequences of different policy choices. These scenarios are intended to be compatible with the full range of emissions scenarios described in the current scientific literature, with or without climate policy. As such, they are designed to sample uncertainty in future scenarios. *(continued on next page)*

*FAQ 1.1 (continued)*

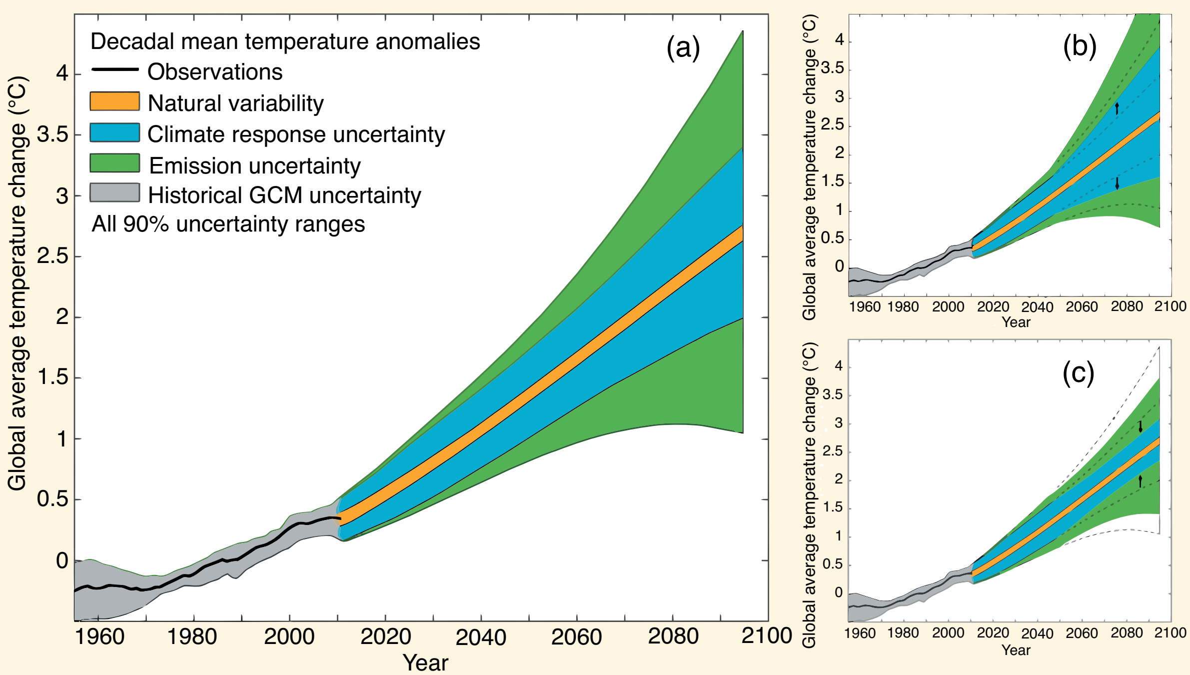
Projections for the next few years and decades are sensitive to emissions of short-lived compounds such as aerosols and methane. More distant projections, however, are more sensitive to alternative scenarios around long-lived GHG emissions. These scenario-dependent uncertainties will not be reduced by improvements in climate science, and will become the dominant uncertainty in projections over longer timescales (e.g., 2100) (FAQ 1.1, Figure 1).

The final contribution to the uncertainty range comes from our imperfect knowledge of how the climate will respond to future anthropogenic emissions and land use change. Scientists principally use computer-based global climate models to estimate this response. A few dozen global climate models have been developed by different groups of scientists around the world. All models are built on the same physical principles, but some approximations are needed because the climate system is so complex. Different groups choose slightly different approximations to represent specific processes in the atmosphere, such as clouds. These choices produce differences in climate projections from different models. This contribution to the uncertainty range is described as 'response uncertainty' or 'model uncertainty'.

The complexity of the Earth system means that future climate could follow many different scenarios, yet still be consistent with current understanding and models. As observational records lengthen and models improve, researchers should be able, within the limitations of the range of natural variability, to narrow that range in probable temperature in the next few decades (FAQ 1.1, Figure 1). It is also possible to use information about the current state of the oceans and cryosphere to produce better projections up to a few years ahead.

As science improves, new geophysical processes can be added to climate models, and representations of those already included can be improved. These developments can appear to increase model-derived estimates of climate response uncertainty, but such increases merely reflect the quantification of previously unmeasured sources of uncertainty (FAQ1.1, Figure 1). As more and more important processes are added, the influence of unquantified processes lessens, and there can be more confidence in the projections.






**FAQ 1.1, Figure 1 |** Schematic diagram showing the relative importance of different uncertainties, and their evolution in time. (a) Decadal mean surface temperature change (°C) from the historical record (black line), with climate model estimates of uncertainty for historical period (grey), along with future climate projections and uncertainty. Values are normalised by means from 1961 to 1980. Natural variability (orange) derives from model interannual variability, and is assumed constant with time. Emission uncertainty (green) is estimated as the model mean difference in projections from different scenarios. Climate response uncertainty (blue-solid) is based on climate model spread, along with added uncertainties from the carbon cycle, as well as rough estimates of additional uncertainty from poorly modelled processes. Based on Hawkins and Sutton (2011) and Huntingford et al. (2009). (b) Climate response uncertainty can appear to increase when a new process is discovered to be relevant, but such increases reflect a quantification of previously unmeasured uncertainty, or (c) can decrease with additional model improvements and observational constraints. The given uncertainty range of 90% means that the temperature is estimated to be in that range, with a probability of 90%.

Frequently Asked Questions

## FAQ 2.1 | How Do We Know the World Has Warmed?

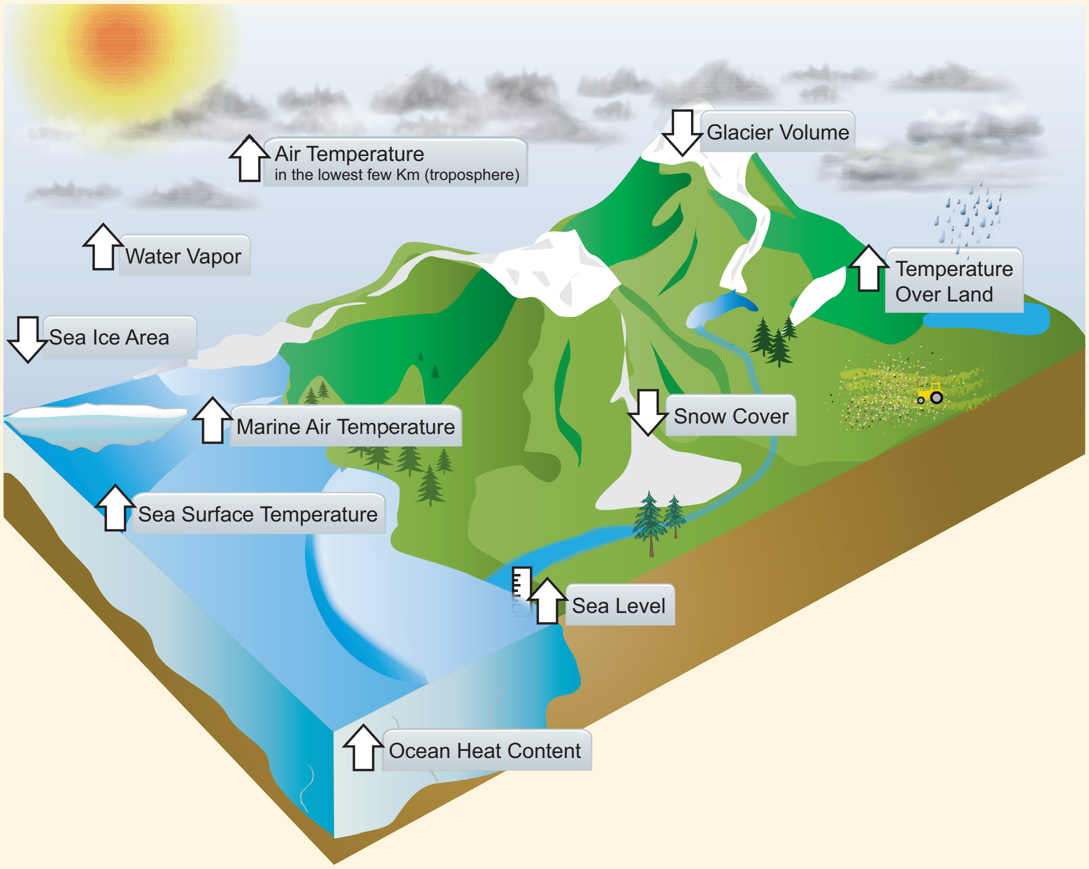
*Evidence for a warming world comes from multiple independent climate indicators, from high up in the atmosphere to the depths of the oceans. They include changes in surface, atmospheric and oceanic temperatures; glaciers; snow cover; sea ice; sea level and atmospheric water vapour. Scientists from all over the world have independently verified this evidence many times. That the world has warmed since the 19th century is unequivocal.*

Discussion about climate warming often centres on potential residual biases in temperature records from land-based weather stations. These records are very important, but they only represent one indicator of changes in the climate system. Broader evidence for a warming world comes from a wide range of independent physically consistent measurements of many other, strongly interlinked, elements of the climate system (FAQ 2.1, Figure 1).

A rise in global average surface temperatures is the best-known indicator of climate change. Although each year and even decade is not always warmer than the last, global surface temperatures have warmed substantially since 1900.

Warming land temperatures correspond closely with the observed warming trend over the oceans. Warming oceanic air temperatures, measured from aboard ships, and temperatures of the sea surface itself also coincide, as borne out by many independent analyses.

The atmosphere and ocean are both fluid bodies, so warming at the surface should also be seen in the lower atmosphere, and deeper down into the upper oceans, and observations confirm that this is indeed the case. Analyses of measurements made by weather balloon radiosondes and satellites consistently show warming of the troposphere, the active weather layer of the atmosphere. More than 90% of the excess energy absorbed by the climate system since at least the 1970s has been stored in the oceans as can be seen from global records of ocean heat content going back to the 1950s. *(continued on next page)*




**FAQ 2.1, 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 (arrow direction denotes the sign of the change), as shown in FAQ 2.1, Figure 2.

*FAQ 2.1 (continued)*

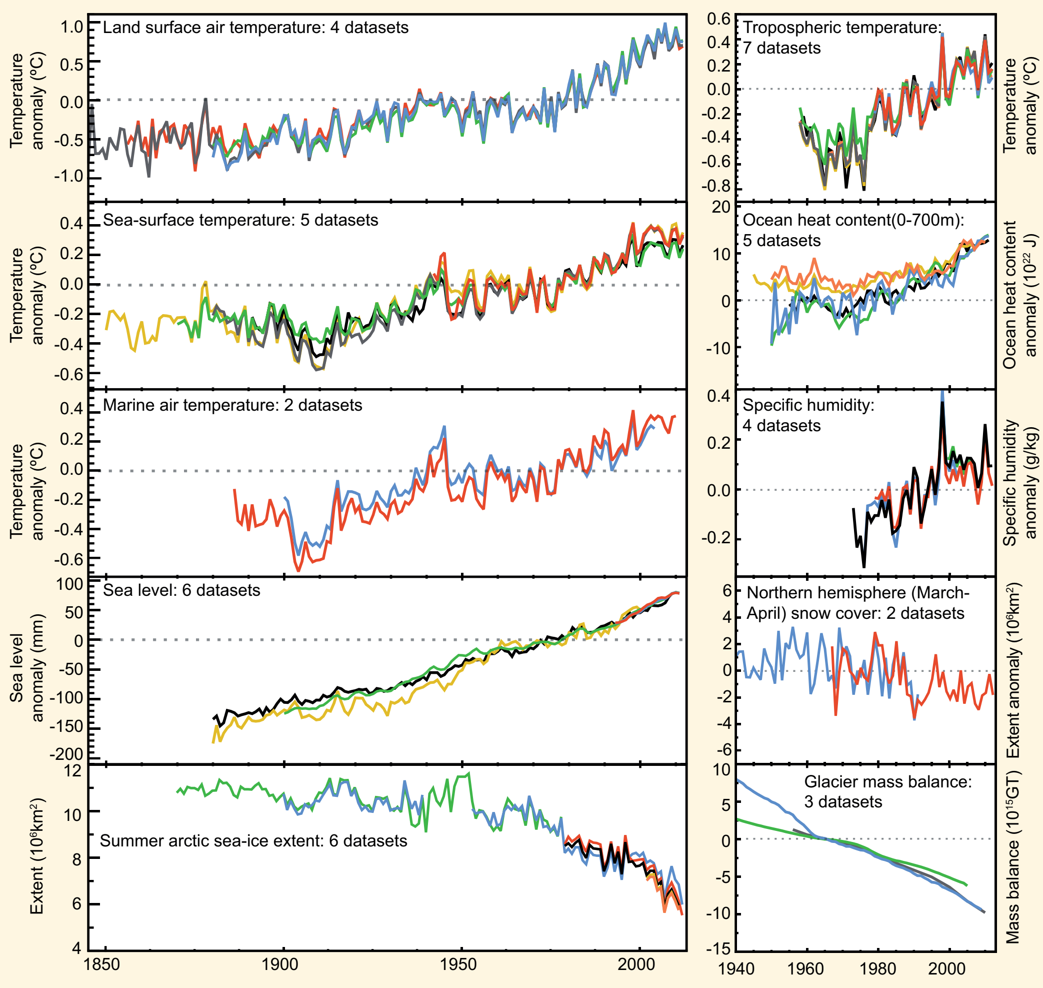
As the oceans warm, the water itself expands. This expansion is one of the main drivers of the independently observed rise in sea levels over the past century. Melting of glaciers and ice sheets also contribute, as do changes in storage and usage of water on land.

A warmer world is also a moister one, because warmer air can hold more water vapour. Global analyses show that specific humidity, which measures the amount of water vapour in the atmosphere, has increased over both the land and the oceans.

The frozen parts of the planet—known collectively as the cryosphere—affect, and are affected by, local changes in temperature. The amount of ice contained in glaciers globally has been declining every year for more than 20 years, and the lost mass contributes, in part, to the observed rise in sea level. Snow cover is sensitive to changes in temperature, particularly during the spring, when snow starts to melt. Spring snow cover has shrunk across the NH since the 1950s. Substantial losses in Arctic sea ice have been observed since satellite records began, particularly at the time of the mimimum extent, which occurs in September at the end of the annual melt season. By contrast, the increase in Antarctic sea ice has been smaller.

Individually, any single analysis might be unconvincing, but analysis of these different indicators and independent data sets has led many independent research groups to *all* reach the same conclusion. From the deep oceans to the top of the troposphere, the evidence of warmer air and oceans, of melting ice and rising seas all points unequivocally to one thing: the world has warmed since the late 19th century (FAQ 2.1, Figure 2).






**FAQ 2.1, Figure 2 |** Multiple independent indicators of a changing global climate. Each line represents an independently derived estimate of change in the climate element. In each panel all data sets have been normalized to a common period of record. A full detailing of which source data sets go into which panel is given in the Supplementary Material 2.SM.5.

Frequently Asked Questions

## FAQ 2.2 | Have There Been Any Changes in Climate Extremes?

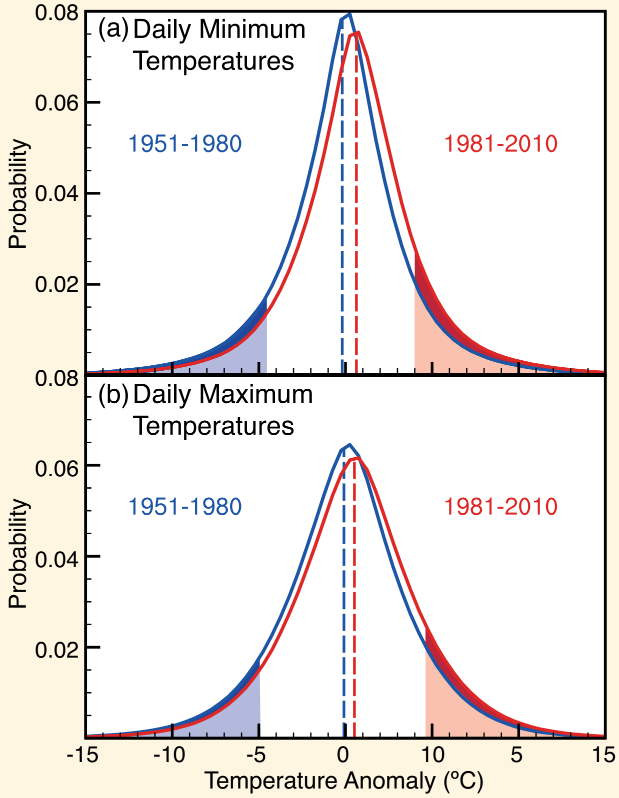
*There is strong evidence that warming has lead to changes in temperature extremes—including heat waves—since the mid-20th century. Increases in heavy precipitation have probably also occurred over this time, but vary by region. However, for other extremes, such as tropical cyclone frequency, we are less certain, except in some limited regions, that there have been discernable changes over the observed record.*

From heat waves to cold snaps or droughts to flooding rains, recording and analysing climate extremes poses unique challenges, not just because these events are rare, but also because they invariably happen in conjunction with disruptive conditions. Furthermore, there is no consistent definition in the scientific literature of what constitutes an extreme climatic event, and this complicates comparative global assessments.

Although, in an absolute sense, an extreme climate event will vary from place to place—a hot day in the tropics, for instance, may be a different temperature to a hot day in the mid-latitudes—international efforts to monitor extremes have highlighted some significant global changes.

For example, using consistent definitions for cold (<10th percentile) and warm (>90th percentile) days and nights it is found that warm days and nights have increased and cold days and nights have decreased for most regions of the globe; a few exceptions being central and eastern North America, and southern South America but mostly only related to daytime temperatures. Those changes are generally most apparent in minimum temperature extremes, for example, warm nights. Data limitations make it difficult to establish a causal link to increases in average temperatures, but FAQ 2.2, Figure 1 indicates that daily global temperature extremes have indeed changed. Whether these changes are simply associated with the average of daily temperatures increasing (the dashed lines in FAQ 2.2, Figure 1) or whether other changes in the distribution of daytime and nighttime temperatures have occurred is still under debate.

Warm spells or heat waves, that is, periods containing consecutive extremely hot days or nights, have also been assessed, but there are fewer studies of heat wave characteristics than those that compare changes in merely warm days or nights. Most global land areas with available data have experienced more heat waves since the middle of the 20th century. One exception is the south-eastern USA, where heat wave frequency and duration measures generally show decreases. This has been associated with a so-called 'warming hole' in this region, where precipitation has also increased and may be related to interactions between the land and the atmosphere and long-term variations in the Atlantic and Pacific Oceans. However, for large regions, particularly in Africa and South America, information on changes in heatwaves is limited.




**FAQ 2.2, Figure 1 |** Distribution of (a) daily minimum and (b) daily maximum temperature anomalies relative to a 1961–1990 climatology for two periods: 1951–1980 (blue) and 1981–2010 (red) using the HadGHCND data set. The shaded blue and red areas represent the coldest 10% and warmest 10% respectively of (a) nights and (b) days during the 1951–1980 period. The darker shading indicates by how much the number of the coldest days and nights has reduced (dark blue) and by how much the number of the warmest days and nights has increased (dark red) during the 1981–2010 period compared to the 1951–1980 period.

For regions such as Europe, where historical temperature reconstructions exist going back several hundreds of years, indications are that some areas have experienced a disproportionate number of extreme heat waves in recent decades. *(continued on next page)*

*FAQ 2.2 (continued)*

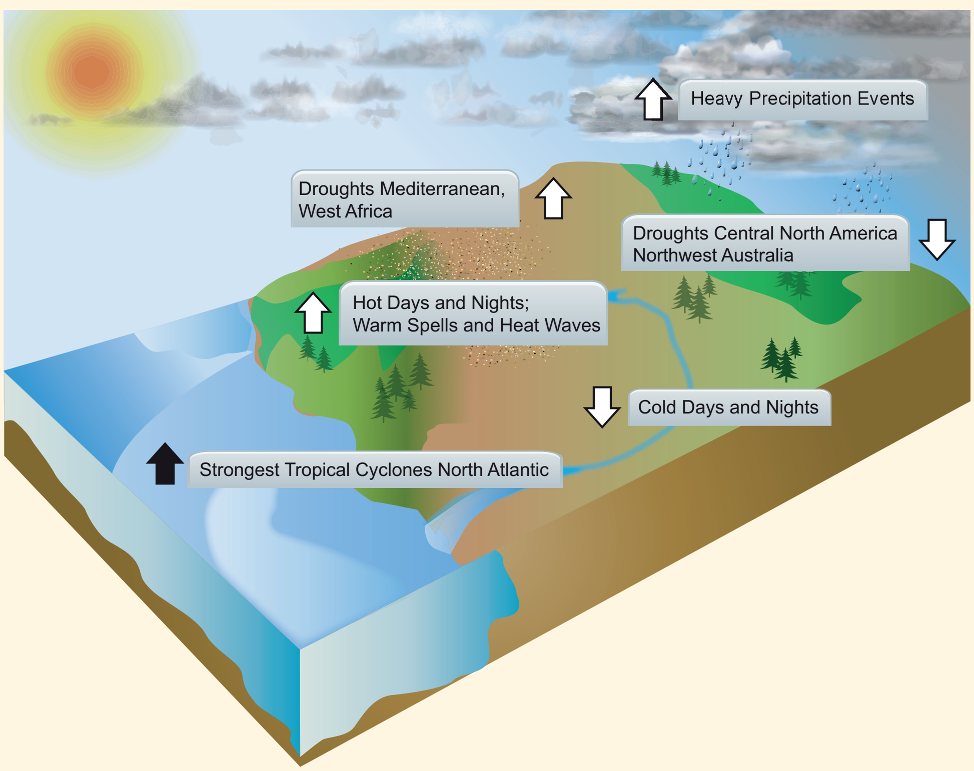
Changes in extremes for other climate variables are generally less coherent than those observed for temperature, owing to data limitations and inconsistencies between studies, regions and/or seasons. However, increases in precipitation extremes, for example, are consistent with a warmer climate. Analyses of land areas with sufficient data indicate increases in the frequency and intensity of extreme precipitation events in recent decades, but results vary strongly between regions and seasons. For instance, evidence is most compelling for increases in heavy precipitation in North America, Central America and Europe, but in some other regions—such as southern Australia and western Asia—there is evidence of decreases. Likewise, drought studies do not agree on the sign of the global trend, with regional inconsistencies in trends also dependent on how droughts are defined. However, indications exist that droughts have increased in some regions (e.g., the Mediterranean) and decreased in others (e.g., central North America) since the middle of the 20th century.

Considering other extremes, such as tropical cyclones, the latest assessments show that due to problems with past observing capabilities, it is difficult to make conclusive statements about long-term trends. There is very strong evidence, however, that storm activity has increased in the North Atlantic since the 1970s.

Over periods of a century or more, evidence suggests slight decreases in the frequency of tropical cyclones making landfall in the North Atlantic and the South Pacific, once uncertainties in observing methods have been considered. Little evidence exists of any longer-term trend in other ocean basins. For extratropical cyclones, a poleward shift is evident in both hemispheres over the past 50 years, with further but limited evidence of a decrease in wind storm frequency at mid-latitudes. Several studies suggest an increase in intensity, but data sampling issues hamper these assessments.



FAQ 2.2, Figure 2 summarizes some of the observed changes in climate extremes. Overall, the most robust global changes in climate extremes are seen in measures of daily temperature, including to some extent, heat waves. Precipitation extremes also appear to be increasing, but there is large spatial variability, and observed trends in droughts are still uncertain except in a few regions. While robust increases have been seen in tropical cyclone frequency and activity in the North Atlantic since the 1970s, the reasons for this are still being debated. There is limited evidence of changes in extremes associated with other climate variables since the mid-20th century.




**FAQ 2.2, Figure 2 |** Trends in the frequency (or intensity) of various climate extremes (arrow direction denotes the sign of the change) since the middle of the 20th century (except for North Atlantic storms where the period covered is from the 1970s).

Frequently Asked Questions

## FAQ 3.1 | Is the Ocean Warming?

*Yes, the ocean is warming over many regions, depth ranges and time periods, although neither everywhere nor constantly. The signature of warming emerges most clearly when considering global, or even ocean basin, averages over time spans of a decade or more.*

Ocean temperature at any given location can vary greatly with the seasons. It can also fluctuate substantially from year to year—or even decade to decade—because of variations in ocean currents and the exchange of heat between ocean and atmosphere.

Ocean temperatures have been recorded for centuries, but it was not until around 1971 that measurements were sufficiently comprehensive to estimate the average global temperature of the upper several hundred meters of the ocean confidently for any given year. In fact, before the international Argo temperature/salinity profiling float array first achieved worldwide coverage in 2005, the global average upper ocean temperature for any given year was sensitive to the methodology used to estimate it.

Global mean upper ocean temperatures have increased over decadal time scales from 1971 to 2010. Despite large uncertainty in most yearly means, this warming is a robust result. In the upper 75 m of the ocean, the global average warming trend has been 0.11 [0.09 to 0.13]°C per decade over this time. That trend generally lessens from the surface to mid-depth, reducing to about 0.04°C per decade by 200 m, and to less than 0.02°C per decade by 500 m.

Temperature anomalies enter the subsurface ocean by paths in addition to mixing from above (FAQ3.1, Figure 1). Colder—hence denser—waters from high latitudes can sink from the surface, then spread toward the equator beneath warmer, lighter, waters at lower latitudes. At a few locations—the northern North Atlantic Ocean and the Southern Ocean around Antarctica—ocean water is cooled so much that it sinks to great depths, even to the sea floor. This water then spreads out to fill much of the rest of the deep ocean. As ocean surface waters warm, these sinking waters also warm with time, increasing temperatures in the ocean interior much more quickly than would downward mixing of surface heating alone.

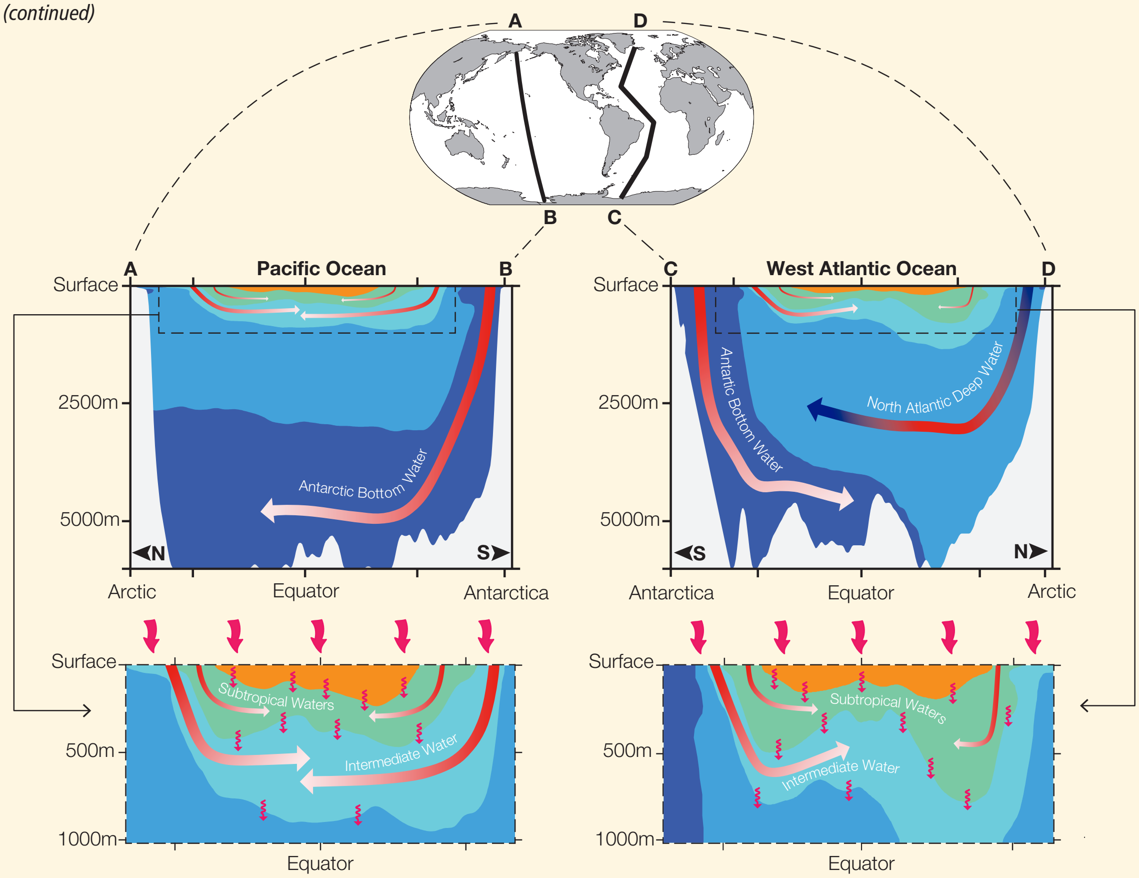
In the North Atlantic, the temperature of these deep waters varies from decade to decade—sometimes warming, sometimes cooling—depending on prevailing winter atmospheric patterns. Around Antarctica, bottom waters have warmed detectably from about 1992–2005, perhaps due to the strengthening and southward shift of westerly winds around the Southern Ocean over the last several decades. This warming signal in the deepest coldest bottom waters of the world ocean is detectable, although it weakens northward in the Indian, Atlantic and Pacific Oceans. Deep warming rates are generally less pronounced than ocean surface rates (around 0.03ºC per decade since the 1990s in the deep and bottom waters around Antarctica, and smaller in many other locations). However, they occur over a large volume, so deep ocean warming contributes significantly to the total increase in ocean heat.

Estimates of historical changes in global average ocean temperature have become more accurate over the past several years, largely thanks to the recognition, and reduction, of systematic measurement errors. By carefully comparing less accurate measurements with sparser, more accurate ones at adjacent locations and similar times, scientists have reduced some spurious instrumental biases in the historical record. These improvements revealed that the global average ocean temperature has increased much more steadily from year to year than was reported prior to 2008. Nevertheless, the global average warming rate may not be uniform in time. In some years, the ocean appears to warm faster than average; in others, the warming rate seems to slow.

The ocean's large mass and high heat capacity allow it to store huge amounts of energy—more than 1000 times that in the atmosphere for an equivalent increase in temperature. The Earth is absorbing more heat than it is emitting back into space, and nearly all this excess heat is entering the oceans and being stored there. The ocean has absorbed about 93% of the combined heat stored by warmed air, sea, and land, and melted ice between 1971 and 2010.

The ocean's huge heat capacity and slow circulation lend it significant thermal inertia. It takes about a decade for near-surface ocean temperatures to adjust in response to climate forcing (Section 12.5), such as changes in greenhouse gas concentrations. Thus, if greenhouse gas concentrations could be held at present levels into the future, increases in the Earth's surface temperature would begin to slow within about a decade. However, deep ocean temperature would continue to warm for centuries to millennia (Section 12.5), and thus sea levels would continue to rise for centuries to millennia as well (Section 13.5). *(continued on next page)*

*FAQ 3.1 (continued)*




**FAQ 3.1, Figure 1 |** Ocean heat uptake pathways. The ocean is stratified, with the coldest, densest water in the deep ocean (upper panels: use map at top for orientation). Cold Antarctic Bottom Water (dark blue) sinks around Antarctica then spreads northward along the ocean floor into the central Pacific (upper left panel: red arrows fading to white indicate stronger warming of the bottom water most recently in contact with the ocean surface) and western Atlantic oceans (upper right panel), as well as the Indian Ocean (not shown). Less cold, hence lighter, North Atlantic Deep Water (lighter blue) sinks in the northern North Atlantic Ocean (upper right panel: red and blue arrow in the deep water indicates decadal warming and cooling), then spreads south above the Antarctic Bottom Water. Similarly, in the upper ocean (lower left panel shows Pacific Ocean detail, lower right panel the Atlantic), cool Intermediate Waters (cyan) sink in sub-polar regions (red arrows fading to white indicating warming with time), before spreading toward the equator under warmer Subtropical Waters (green), which in turn sink (red arrows fading to white indicate stronger warming of the intermediate and subtropical waters most recently in contact with the surface) and spread toward the equator under tropical waters, the warmest and lightest (orange) in all three oceans. Excess heat or cold entering at the ocean surface (top curvy red arrows) also mixes slowly downward (sub-surface wavy red arrows).

FAQ

Frequently Asked Questions

## FAQ 3.2 | Is There Evidence for Changes in the Earth's Water Cycle?

*The Earth's water cycle involves evaporation and precipitation of moisture at the Earth's surface. Changes in the atmosphere's water vapour content provide strong evidence that the water cycle is already responding to a warming climate. Further evidence comes from changes in the distribution of ocean salinity, which, due to a lack of long-term observations of rain and evaporation over the global oceans, has become an important proxy rain gauge.*

The water cycle is expected to intensify in a warmer climate, because warmer air can be moister: the atmosphere can hold about 7% more water vapour for each degree Celsius of warming. Observations since the 1970s show increases in surface and lower atmospheric water vapour (FAQ 3.2, Figure 1a), at a rate consistent with observed warming. Moreover, evaporation and precipitation are projected to intensify in a warmer climate.

Recorded changes in ocean salinity in the last 50 years support that projection. Seawater contains both salt and fresh water, and its salinity is a function of the weight of dissolved salts it contains. Because the total amount of salt—which comes from the weathering of rocks—does not change over human time scales, seawater's salinity can only be altered—over days or centuries—by the addition or removal of fresh water.

The atmosphere connects the ocean's regions of net fresh water loss to those of fresh water gain by moving evaporated water vapour from one place to another. The distribution of salinity at the ocean surface largely reflects the spatial pattern of evaporation minus precipitation, runoff from land, and sea ice processes. There is some shifting of the patterns relative to each other, because of the ocean's currents.

Subtropical waters are highly saline, because evaporation exceeds rainfall, whereas seawater at high latitudes and in the tropics—where more rain falls than evaporates—is less so (FAQ 3.2, Figure 1b, d). The Atlantic, the saltiest ocean basin, loses more freshwater through evaporation than it gains from precipitation, while the Pacific is nearly neutral (i.e., precipitation gain nearly balances evaporation loss), and the Southern Ocean (region around Antarctica) is dominated by precipitation.

Changes in surface salinity and in the upper ocean have reinforced the mean salinity pattern. The evaporation-dominated subtropical regions have become saltier, while the precipitation-dominated subpolar and tropical regions have become fresher. When changes over the top 500 m are considered, the evaporation-dominated Atlantic has become saltier, while the nearly neutral Pacific and precipitation-dominated Southern Ocean have become fresher (FAQ 3.2, Figure 1c).

Observing changes in precipitation and evaporation directly and globally is difficult, because most of the exchange of fresh water between the atmosphere and the surface happens over the 70% of the Earth's surface covered by ocean. Long-term precipitation records are available only from over the land, and there are no long-term measurements of evaporation.

Land-based observations show precipitation increases in some regions, and decreases in others, making it difficult to construct a globally integrated picture. Land-based observations have shown more extreme rainfall events, and more flooding associated with earlier snow melt at high northern latitudes, but there is strong regionality in the trends. Land-based observations are so far insufficient to provide evidence of changes in drought.

Ocean salinity, on the other hand, acts as a sensitive and effective rain gauge over the ocean. It naturally reflects and smoothes out the difference between water gained by the ocean from precipitation, and water lost by the ocean through evaporation, both of which are very patchy and episodic. Ocean salinity is also affected by water runoff from the continents, and by the melting and freezing of sea ice or floating glacial ice. Fresh water added by melting ice on land will change global-averaged salinity, but changes to date are too small to observe.

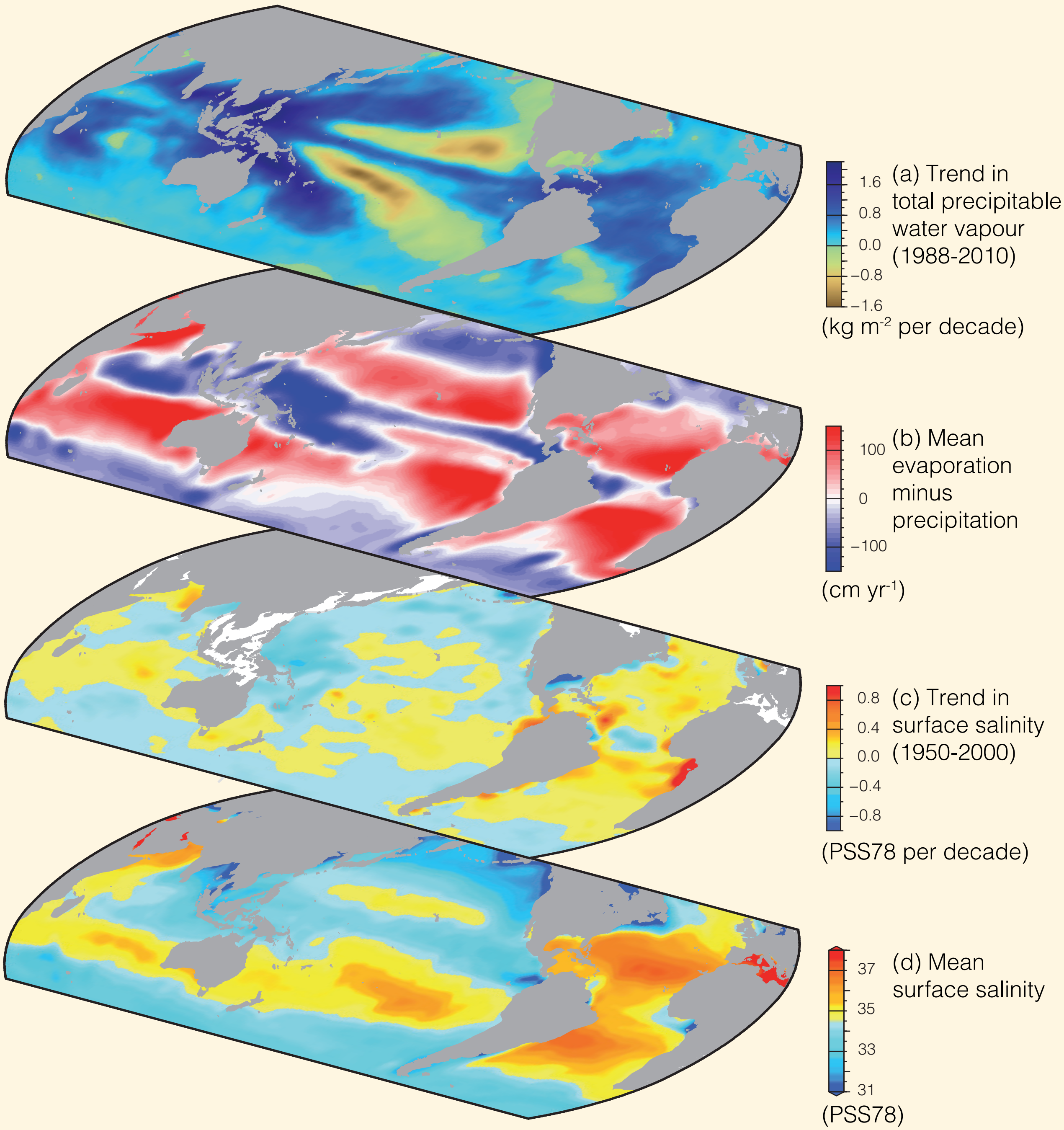
Data from the past 50 years show widespread salinity changes in the upper ocean, which are indicative of systematic changes in precipitation and runoff minus evaporation, as illustrated in FAQ 3.2, Figure 1.

FAQ 3.2 is based on observations reported in Chapters 2 and 3, and on model analyses in Chapters 9 and 12.

*(continued on next page)*

FAQ

*FAQ 3.2 (continued)*




**FAQ 3.2, Figure 1 |** Changes in sea surface salinity are related to the atmospheric patterns of evaporation minus precipitation (E – P) and trends in total precipitable water: (a) Linear trend (1988–2010) in total precipitable water (water vapor integrated from the Earth's surface up through the entire atmosphere) (kg $m^{-2}$ per decade) from satellite observations (Special Sensor Microwave Imager) (after Wentz et al., 2007) (blues: wetter; yellows: drier). (b) The 1979–2005 climatological mean net E –P (cm $yr^{-1}$) from meteorological reanalysis (National Centers for Environmental Prediction/National Center for Atmospheric Research; Kalnay et al., 1996) (reds: net evaporation; blues: net precipitation). (c) Trend (1950–2000) in surface salinity (PSS78 per 50 years) (after Durack and Wijffels, 2010) (blues freshening; yellows-reds saltier). (d) The climatological-mean surface salinity (PSS78) (blues: <35; yellows–reds: >35).

Frequently Asked Questions

## FAQ 3.3 | How Does Anthropogenic Ocean Acidification Relate to Climate Change?

*Both anthropogenic climate change and anthropogenic ocean acidification are caused by increasing carbon dioxide concentrations in the atmosphere. Rising levels of carbon dioxide ($CO_2$), along with other greenhouse gases, indirectly alter the climate system by trapping heat as it is reflected back from the Earth's surface. Anthropogenic ocean acidification is a direct consequence of rising $CO_2$ concentrations as seawater currently absorbs about 30% of the anthropogenic $CO_2$ from the atmosphere.*

Ocean acidification refers to a reduction in pH over an extended period, typically decades or longer, caused primarily by the uptake of $CO_2$ from the atmosphere. pH is a dimensionless measure of acidity. Ocean acidification describes the direction of pH change rather than the end point; that is, ocean pH is decreasing but is not expected to become acidic (pH < 7). Ocean acidification can also be caused by other chemical additions or subtractions from the oceans that are natural (e.g., increased volcanic activity, methane hydrate releases, long-term changes in net respiration) or human-induced (e.g., release of nitrogen and sulphur compounds into the atmosphere). Anthropogenic ocean acidification refers to the component of pH reduction that is caused by human activity.

Since about 1750, the release of $CO_2$ from industrial and agricultural activities has resulted in global average atmospheric $CO_2$ concentrations that have increased from 278 to 390.5 ppm in 2011. The atmospheric concentration of $CO_2$ is now higher than experienced on the Earth for at least the last 800,000 years and is expected to continue to rise because of our dependence on fossil fuels for energy. To date, the oceans have absorbed approximately 155 ± 30 PgC from the atmosphere, which corresponds to roughly one-fourth of the total amount of $CO_2$ emitted (555 ± 85 PgC) by human activities since preindustrial times. This natural process of absorption has significantly reduced the greenhouse gas levels in the atmosphere and minimized some of the impacts of global warming. However, the ocean's uptake of $CO_2$ is having a significant impact on the chemistry of seawater. The average pH of ocean surface waters has already fallen by about 0.1 units, from about 8.2 to 8.1 since the beginning of the Industrial Revolution. Estimates of projected future atmospheric and oceanic $CO_2$ concentrations indicate that, by the end of this century, the average surface ocean pH could be 0.2 to 0.4 lower than it is today. The pH scale is logarithmic, so a change of 1 unit corresponds to a 10-fold change in hydrogen ion concentration.

When atmospheric $CO_2$ exchanges across the air–sea interface it reacts with seawater through a series of four chemical reactions that increase the concentrations of the carbon species: dissolved carbon dioxide ($CO_{2(aq)}$), carbonic acid ($H_2CO_3$) and bicarbonate ($HCO_3^-$):

$$CO_{2(atmos)} \rightleftarrows CO_{2(aq)} \quad (1)$$

$$CO_{2(aq)} + H_2O \rightleftarrows H_2CO_3 \quad (2)$$

$$H_2CO_3 \rightleftarrows H^+ + HCO_3^- \quad (3)$$

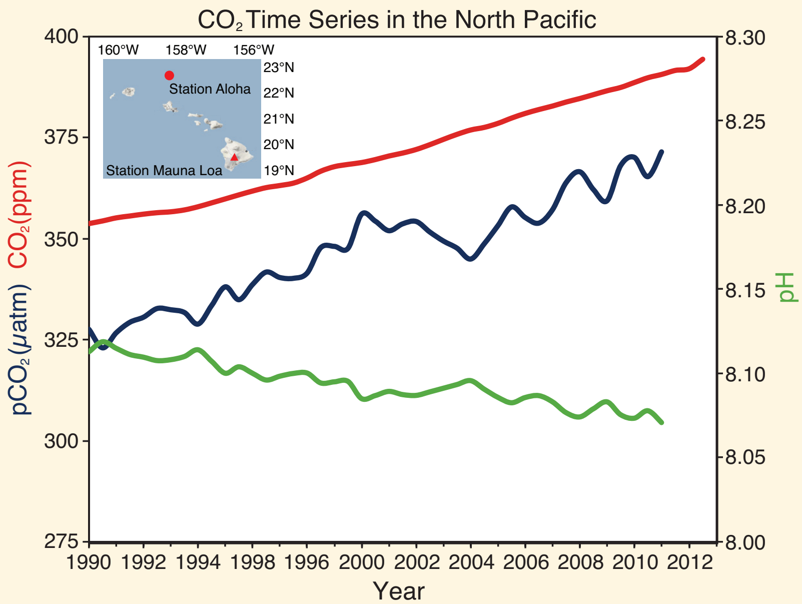
$$HCO_3^- \rightleftarrows H^+ + CO_3^{2-} \quad (4)$$

Hydrogen ions ($H^+$) are produced by these reactions. This increase in the ocean's hydrogen ion concentration corresponds to a reduction in pH, or an increase in acidity. Under normal seawater conditions, more than 99.99% of the hydrogen ions that are produced will combine with carbonate ion ($CO_3^{2-}$ ) to produce additional $HCO_3^-$. Thus, the addition of anthropogenic $CO_2$ into the oceans lowers the pH and consumes carbonate ion. These reactions are fully reversible and the basic thermodynamics of these reactions in seawater are well known, such that at a pH of approximately 8.1 approximately 90% the carbon is in the form of bicarbonate ion, 9% in the form of carbonate ion, and only about 1% of the carbon is in the form of dissolved $CO_2$. Results from laboratory, field, and modeling studies, as well as evidence from the geological record, clearly indicate that marine ecosystems are highly susceptible to the increases in oceanic $CO_2$ and the corresponding decreases in pH and carbonate ion.

Climate change and anthropogenic ocean acidification do not act independently. Although the $CO_2$ that is taken up by the ocean does not contribute to greenhouse warming, ocean warming reduces the solubility of carbon dioxide in seawater; and thus reduces the amount of $CO_2$ the oceans can absorb from the atmosphere. For example, under doubled preindustrial $CO_2$ concentrations and a 2°C temperature increase, seawater absorbs about 10% less $CO_2$ (10% less total carbon, $C_T$) than it would with no temperature increase (compare columns 4 and 6 in Table 1), but the pH remains almost unchanged. Thus, a warmer ocean has less capacity to remove $CO_2$ from the atmosphere, yet still experiences ocean acidification. The reason for this is that bicarbonate is converted to carbonate in a warmer ocean, releasing a hydrogen ion thus stabilizing the pH. *(continued on next page)*

FAQ

*FAQ 3.3 (continued)*




**FAQ 3.3, Figure 1 |** A smoothed time series of atmospheric $CO_2$ mole fraction (in ppm) at the atmospheric Mauna Loa Observatory (top red line), surface ocean partial pressure of $CO_2$ ($pCO_2$; middle blue line) and surface ocean pH (bottom green line) at Station ALOHA in the subtropical North Pacific north of Hawaii for the period from1990–2011 (after Doney et al., 2009; data from Dore et al., 2009). The results indicate that the surface ocean $pCO_2$ trend is generally consistent with the atmospheric increase but is more variable due to large-scale interannual variability of oceanic processes.

**FAQ 3.3, Table 1 |** Oceanic pH and carbon system parameter changes in surface water for a $CO_2$ doubling from the preindustrial atmosphere without and with a 2°C warming[a].

| Parameter | Pre-industrial (280 ppmv) 20°C | 2 × Pre-industrial (560 ppmv) 20°C | (% change relative to pre-industrial) | 2 × Pre-industrial (560 ppmv) 22°C | (% change relative to pre-industrial) |
|---|---|---|---|---|---|
| **pH** | 8.1714 | 7.9202 | – | 7.9207 | – |
| **$H^+$ (mol kg$^{-1}$)** | $6.739e^{-9}$ | $1.202e^{-8}$ | (78.4) | $1.200e^{-8}$ | (78.1) |
| **$CO_{2(aq)}$ (μmol kg$^{-1}$)** | 9.10 | 18.10 | (98.9) | 17.2 | (89.0) |
| **$HCO_3^-$ (μmol kg$^{-1}$)** | 1723.4 | 1932.8 | (12.15) | 1910.4 | (10.9) |
| **$CO_3^{2-}$ (μmol kg$^{-1}$)** | 228.3 | 143.6 | (-37.1) | 152.9 | (–33.0) |
| **$C_T$ (μmol kg$^{-1}$)** | 1960.8 | 2094.5 | (6.82) | 2080.5 | (6.10) |

Notes:

[a] $CO_{2(aq)}$ = dissolved $CO_2$, $H_2CO_3$ = carbonic acid, $HCO_3^-$ = bicarbonate, $CO_3^{2-}$ = carbonate, $C_T$ = total carbon = $CO_{2(aq)} + HCO_3^- + CO_3^{2-}$).

Frequently Asked Questions

## FAQ 4.1 | How Is Sea Ice Changing in the Arctic and Antarctic?

*The sea ice covers on the Arctic Ocean and on the Southern Ocean around Antarctica have quite different characteristics, and are showing different changes with time. Over the past 34 years (1979–2012), there has been a downward trend of 3.8% per decade in the annual average extent of sea ice in the Arctic. The average winter thickness of Arctic Ocean sea ice has thinned by approximately 1.8 m between 1978 and 2008, and the total volume (mass) of Arctic sea ice has decreased at all times of year. The more rapid decrease in the extent of sea ice at the summer minimum is a consequence of these trends. In contrast, over the same 34-year period, the total extent of Antarctic sea ice shows a small increase of 1.5% per decade, but there are strong regional differences in the changes around the Antarctic. Measurements of Antarctic sea ice thickness are too few to be able to judge whether its total volume (mass) is decreasing, steady, or increasing.*

A large part of the total Arctic sea ice cover lies above 60°N (FAQ 4.1, Figure 1) and is surrounded by land to the south with openings to the Canadian Arctic Archipelago, and the Bering, Barents and Greenland seas. Some of the ice within the Arctic Basin survives for several seasons, growing in thickness by freezing of seawater at the base and by deformation (ridging and rafting). Seasonal sea ice grows to only ~2 m in thickness but sea ice that is more than 1 year old (perennial ice) can be several metres thicker. Arctic sea ice drifts within the basin, driven by wind and ocean currents: the mean drift pattern is dominated by a clockwise circulation pattern in the western Arctic and a Transpolar Drift Stream that transports Siberian sea ice across the Arctic and exports it from the basin through the Fram Strait.

Satellites with the capability to distinguish ice and open water have provided a picture of the sea ice cover changes. Since 1979, the annual average extent of ice in the Arctic has decreased by 3.8% per decade. The decline in extent at the end of summer (in late September) has been even greater at 11% per decade, reaching a record minimum in 2012. The decadal average extent of the September minimum Arctic ice cover has decreased for each decade since satellite records began. Submarine and satellite records suggest that the thickness of Arctic ice, and hence the total volume, is also decreasing. Changes in the relative amounts of perennial and seasonal ice are contributing to the reduction in ice volume. Over the 34-year record, approximately 17% of this type of sea ice per decade has been lost to melt and export out of the basin since 1979 and 40% since 1999. Although the area of Arctic sea ice coverage can fluctuate from year to year because of variable seasonal production, the proportion of thick perennial ice, and the total sea ice volume, can recover only slowly.

Unlike the Arctic, the sea ice cover around Antarctica is constrained to latitudes north of 78°S because of the presence of the continental land mass. The Antarctic sea ice cover is largely seasonal, with an average thickness of only ~1 m at the time of maximum extent in September. Only a small fraction of the ice cover survives the summer minimum in February, and very little Antarctic sea ice is more than 2 years old. The ice edge is exposed to the open ocean and the snowfall rate over Antarctic sea ice is higher than in the Arctic. When the snow load from snowfall is sufficient to depress the ice surface below sea level, seawater infiltrates the base of the snow pack and snow-ice is formed when the resultant slush freezes. Consequently, snow-to-ice conversion (as well as basal freezing as in the Arctic) contributes to the seasonal growth in ice thickness and total ice volume in the Antarctic. Snow-ice formation is sensitive to changes in precipitation and thus changes in regional climate. The consequence of changes in precipitation on Antarctic sea ice thickness and volume remains a focus for research.

Unconstrained by land boundaries, the latitudinal extent of the Antarctic sea ice cover is highly variable. Near the Antarctic coast, sea ice drift is predominantly from east to west, but further north, it is from west to east and highly divergent. Distinct clockwise circulation patterns that transport ice northward can be found in the Weddell and Ross seas, while the circulation is more variable around East Antarctica. The northward extent of the sea ice cover is controlled in part by the divergent drift that is conducive in winter months to new ice formation in persistent open water areas (polynyas) along the coastlines. These zones of ice formation result in saltier and thus denser ocean water and become one of the primary sources of the deepest water found in the global oceans.

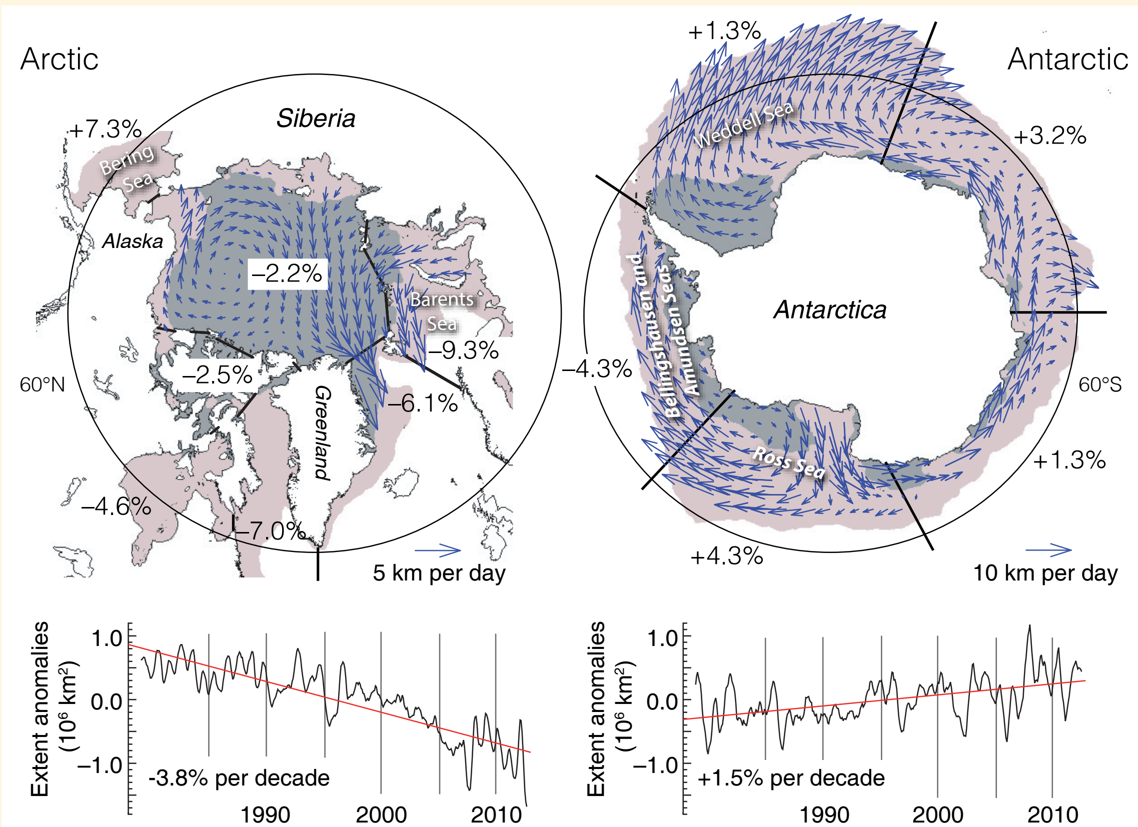
Over the same 34-year satellite record, the annual extent of sea ice in the Antarctic increased at about 1.5% per decade. However, there are regional differences in trends, with decreases seen in the Bellingshausen and Amundsen seas, but a larger increase in sea ice extent in the Ross Sea that dominates the overall trend. Whether the smaller overall increase in Antarctic sea ice extent is meaningful as an indicator of climate is uncertain because the extent

*(continued on next page)*

*FAQ 4.1 (continued)*

varies so much from year to year and from place to place around the continent. Results from a recent study suggest that these contrasting trends in ice coverage may be due to trends in regional wind speed and patterns. Without better ice thickness and ice volume estimates, it is difficult to characterize how Antarctic sea ice cover is responding to changing climate, or which climate parameters are most influential.

There are large differences in the physical environment and processes that affect the state of Arctic and Antarctic sea ice cover and contribute to their dissimilar responses to climate change. The long, and unbroken, record of satellite observations have provided a clear picture of the decline of the Arctic sea ice cover, but available evidence precludes us from making robust statements about overall changes in Antarctic sea ice and their causes.




**FAQ 4.1, Figure 1 |** The mean circulation pattern of sea ice and the decadal trends (%) in annual anomalies in ice extent (i.e., after removal of the seasonal cycle), in different sectors of the Arctic and Antarctic. Arrows show the average direction and magnitude of ice drift. The average sea ice cover for the period 1979 through 2012, from satellite observations, at maximum (minimum) extent is shown as orange (grey) shading.

Frequently Asked Questions
## FAQ 4.2 | Are Glaciers in Mountain Regions Disappearing?

*In many mountain ranges around the world, glaciers are disappearing in response to the atmospheric temperature increases of past decades. Disappearing glaciers have been reported in the Canadian Arctic and Rocky Mountains; the Andes; Patagonia; the European Alps; the Tien Shan; tropical mountains in South America, Africa and Asia and elsewhere. In these regions, more than 600 glaciers have disappeared over the past decades. Even if there is no further warming, many more glaciers will disappear. It is also likely that some mountain ranges will lose most, if not all, of their glaciers.*

In all mountain regions where glaciers exist today, glacier volume has decreased considerably over the past 150 years. Over that time, many small glaciers have disappeared. With some local exceptions, glacier shrinkage (area and volume reduction) was globally widespread already and particularly strong during the 1940s and since the 1980s. However, there were also phases of relative stability during the 1890s, 1920s and 1970s, as indicated by long-term measurements of length changes and by modelling of mass balance. Conventional *in situ* measurements—and increasingly, airborne and satellite measurements—offer robust evidence in most glacierized regions that the rate of reduction in glacier area was higher over the past two decades than previously, and that glaciers continue to shrink. In a few regions, however, individual glaciers are behaving differently and have advanced while most others were in retreat (e.g., on the coasts of New Zealand, Norway and Southern Patagonia (Chile), or in the Karakoram range in Asia). In general, these advances are the result of special topographic and/or climate conditions (e.g., increased precipitation).

It can take several decades for a glacier to adjust its extent to an instantaneous change in climate, so most glaciers are currently larger than they would be if they were in balance with current climate. Because the time required for the adjustment increases with glacier size, larger glaciers will continue to shrink over the next few decades, even if temperatures stabilise. Smaller glaciers will also continue to shrink, but they will adjust their extent faster and many will ultimately disappear entirely.

Many factors influence the future development of each glacier, and whether it will disappear: for instance, its size, slope, elevation range, distribution of area with elevation, and its surface characteristics (e.g., the amount of debris cover). These factors vary substantially from region to region, and also between neighbouring glaciers. External factors, such as the surrounding topography and the climatic regime, are also important for future glacier evolution. Over shorter time scales (one or two decades), each glacier responds to climate change individually and differently in detail.

Over periods longer than about 50 years, the response is more coherent and less dependent on local environmental details, which means that long-term trends in glacier development can be well modelled. Such models are built on an understanding of basic physical principles. For example, an increase in local mean air temperature, with no change in precipitation, will cause an upward shift of the equilibrium line altitude (ELA; see Glossary) by about 150 m for each degree Celsius of atmospheric warming. Such an upward shift and its consequences for glaciers of different size and elevation range are illustrated in FAQ 4.2, Figure 1.

Initially, all glaciers have an accumulation area (white) above and an ablation area (light blue) below the ELA (FAQ 4.2, Figure 1a). As the ELA shifts upwards, the accumulation area shrinks and the ablation area expands, thus increasing the area over which ice is lost through melt (FAQ 4.2, Figure 1b). This imbalance results in an overall loss of ice. After several years, the glacier front retreats, and the ablation area shrinks until the glacier has adjusted its extent to the new climate (FAQ 4.2, Figure 1c). Where climate change is sufficiently strong to raise the ELA permanently above the glacier's highest point (FAQ 4.2, Figure 1b, right) the glacier will eventually disappear entirely (FAQ 4.2, Figure 1c, right). Higher glaciers, which retain their accumulation areas, will shrink but not disappear (FAQ 4.2, Figure 1c, left and middle). A large valley glacier might lose much of its tongue, probably leaving a lake in its place (FAQ 4.2, Figure 1c, left). Besides air temperature, changes in the quantity and seasonality of precipitation influence the shift of the ELA as well. Glacier dynamics (e.g., flow speed) also plays a role, but is not considered in this simplified scheme.

Many observations have confirmed that different glacier types do respond differently to recent climate change. For example, the flat, low-lying tongues of large valley glaciers (such as in Alaska, Canada or the Alps) currently show the strongest mass losses, largely independent of aspect, shading or debris cover. This type of glacier is slow in

*(continued on next page)*

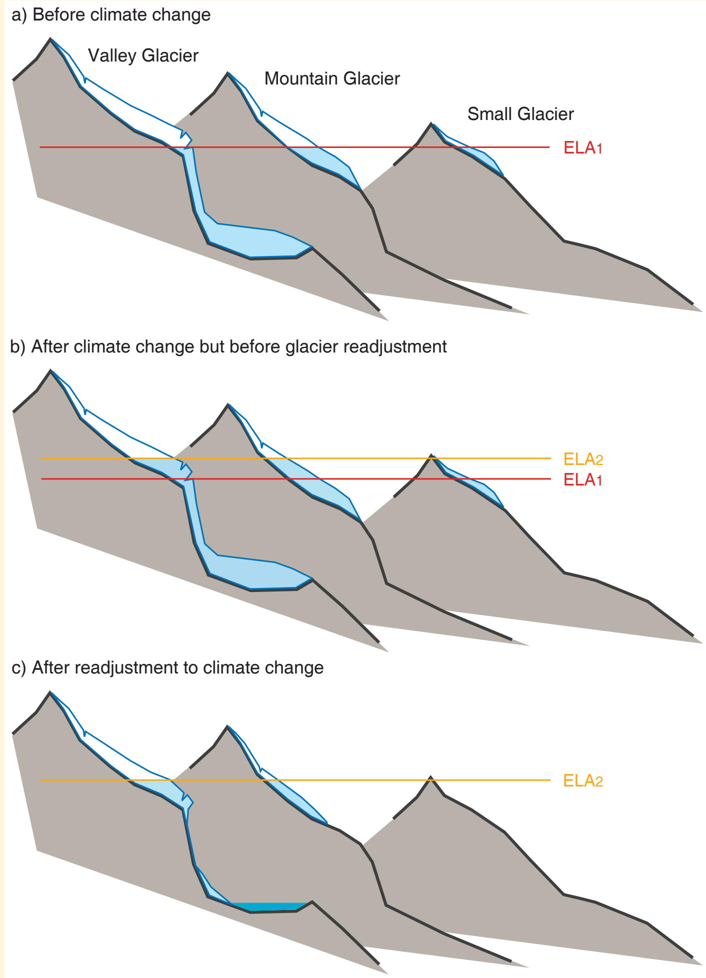
FAQ

*FAQ 4.2 (continued)*

adjusting its extent to new climatic conditions and reacts mainly by thinning without substantial terminus retreat. In contrast, smaller mountain glaciers, with fairly constant slopes, adjust more quickly to the new climate by changing the size of their ablation area more rapidly (FAQ 4.2, Figure 1c, middle).

The long-term response of most glacier types can be determined very well with the approach illustrated in FAQ 4.2, Figure 1. However, modelling short-term glacier response, or the long-term response of more complex glacier types (e.g., those that are heavily debris-covered, fed by avalanche snow, have a disconnected accumulation area, are of surging type, or calve into water), is difficult. These cases require detailed knowledge of other glacier characteristics, such as mass balance, ice thickness distribution, and internal hydraulics. For the majority of glaciers worldwide, such data are unavailable, and their response to climate change can thus only be approximated with the simplified scheme shown in FAQ 4.2, Figure 1.

The Karakoram–Himalaya mountain range, for instance, has a large variety of glacier types and climatic conditions, and glacier characteristics are still only poorly known. This makes determining their future evolution particularly uncertain. However, gaps in knowledge are expected to decrease substantially in coming years, thanks to increased use of satellite data (e.g., to compile glacier inventories or derive flow velocities) and extension of the ground-based measurement network.




**FAQ 4.2, Figure 1 |** Schematic of three types of glaciers located at different elevations, and their response to an upward shift of the equilibrium line altitude (ELA). (a) For a given climate, the ELA has a specific altitude (ELA1), and all glaciers have a specific size. (b) Due to a temperature increase, the ELA shifts upwards to a new altitude (ELA2), initially resulting in reduced accumulation and larger ablation areas for all glaciers. (c) After glacier size has adjusted to the new ELA, the valley glacier (left) has lost its tongue and the small glacier (right) has disappeared entirely.

In summary, the fate of glaciers will be variable, depending on both their specific characteristics and future climate conditions. More glaciers will disappear; others will lose most of their low-lying portions and others might not change substantially. Where the ELA is already above the highest elevation on a particular glacier, that glacier is destined to disappear entirely unless climate cools. Similarly, all glaciers will disappear in those regions where the ELA rises above their highest elevation in the future.

Frequently Asked Questions
## FAQ 5.1 | Is the Sun a Major Driver of Recent Changes in Climate?

*Total solar irradiance (TSI, Chapter 8) is a measure of the total energy received from the sun at the top of the atmosphere. It varies over a wide range of time scales, from billions of years to just a few days, though variations have been relatively small over the past 140 years. Changes in solar irradiance are an important driver of climate variability (Chapter 1; Figure 1.1) along with volcanic emissions and anthropogenic factors. As such, they help explain the observed change in global surface temperatures during the instrumental period (FAQ 5.1, Figure 1; Chapter 10) and over the last millennium. While solar variability may have had a discernible contribution to changes in global surface temperature in the early 20th century, it cannot explain the observed increase since TSI started to be measured directly by satellites in the late 1970s (Chapters 8, 10).*

The Sun's core is a massive nuclear fusion reactor that converts hydrogen into helium. This process produces energy that radiates throughout the solar system as electromagnetic radiation. The amount of energy striking the top of Earth's atmosphere varies depending on the generation and emission of electromagnetic energy by the Sun and on the Earth's orbital path around the Sun.

Satellite-based instruments have directly measured TSI since 1978, and indicate that on average, ~1361 W $m^{-2}$ reaches the top of the Earth's atmosphere. Parts of the Earth's surface and air pollution and clouds in the atmosphere act as a mirror and reflect about 30% of this power back into space. Higher levels of TSI are recorded when the Sun is more active. Irradiance variations follow the roughly 11-year sunspot cycle: during the last cycles, TSI values fluctuated by an average of around 0.1%.

For pre-satellite times, TSI variations have to be estimated from sunspot numbers (back to 1610), or from radioisotopes that are formed in the atmosphere, and archived in polar ice and tree rings. Distinct 50- to 100-year periods of very low solar activity—such as the Maunder Minimum between 1645 and 1715—are commonly referred to as grand solar minima. Most estimates of TSI changes between the Maunder Minimum and the present day are in the order of 0.1%, similar to the amplitude of the 11-year variability.

How can solar variability help explain the observed global surface temperature record back to 1870? To answer this question, it is important to understand that other climate drivers are involved, each producing characteristic patterns of regional climate responses. However, it is the combination of them all that causes the observed climate change. Solar variability and volcanic eruptions are natural factors. Anthropogenic (human-produced) factors, on the other hand, include changes in the concentrations of greenhouse gases, and emissions of visible air pollution (aerosols) and other substances from human activities. 'Internal variability' refers to fluctuations within the climate system, for example, due to weather variability or phenomena like the El Niño-Southern Oscillation.

The relative contributions of these natural and anthropogenic factors change with time. FAQ 5.1, Figure 1 illustrates those contributions based on a very simple calculation, in which the mean global surface temperature variation represents the sum of four components linearly related to solar, volcanic, and anthropogenic forcing, and to internal variability. Global surface temperature has increased by approximately 0.8°C from 1870 to 2010 (FAQ 5.1, Figure 1a). However, this increase has not been uniform: at times, factors that cool the Earth's surface—volcanic eruptions, reduced solar activity, most anthropogenic aerosol emissions—have outweighed those factors that warm it, such as greenhouse gases, and the variability generated within the climate system has caused further fluctuations unrelated to external influences.

The solar contribution to the record of global surface temperature change is dominated by the 11-year solar cycle, which can explain global temperature fluctuations up to approximately 0.1°C between minima and maxima (FAQ 5.1, Figure 1b). A long-term increasing trend in solar activity in the early 20th century may have augmented the warming recorded during this interval, together with internal variability, greenhouse gas increases and a hiatus in volcanism. However, it cannot explain the observed increase since the late 1970s, and there was even a slight decreasing trend of TSI from 1986 to 2008 (Chapters 8 and 10).

Volcanic eruptions contribute to global surface temperature change by episodically injecting aerosols into the atmosphere, which cool the Earth's surface (FAQ 5.1, Figure 1c). Large volcanic eruptions, such as the eruption of Mt. Pinatubo in 1991, can cool the surface by around 0.1°C to 0.3°C for up to three years. *(continued on next page)*

FAQ

*FAQ 5.1 (continued)*

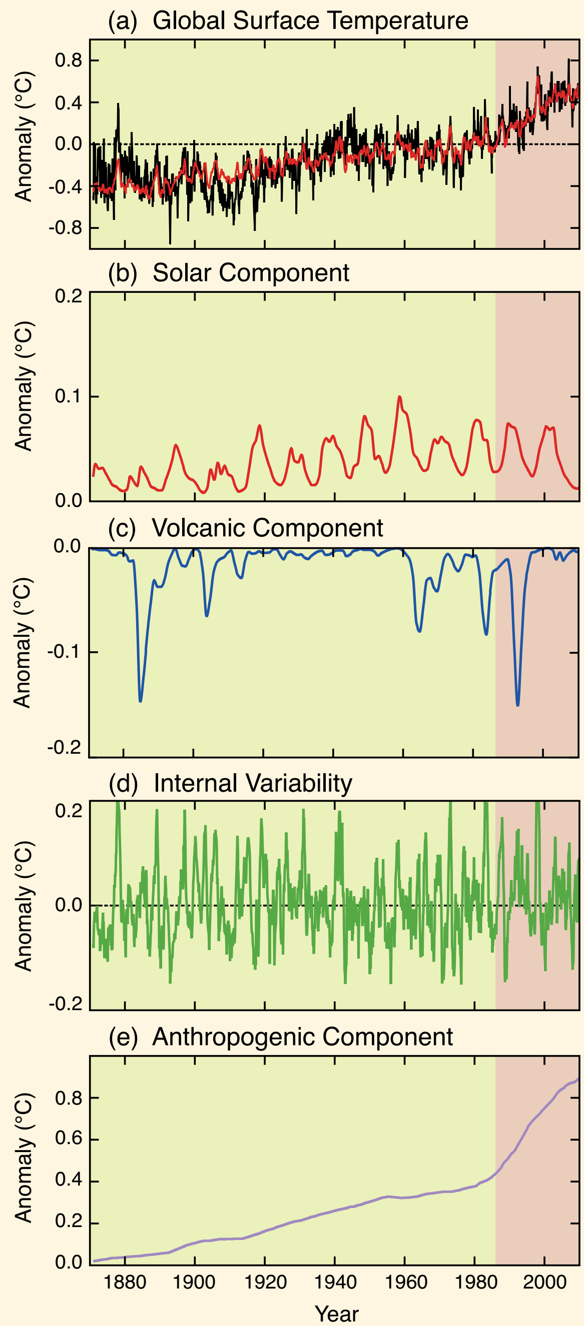
The most important component of internal climate variability is the El Niño Southern Oscillation, which has a major effect on year-to-year variations of tropical and global mean temperature (FAQ 5.1, Figure 1d). Relatively high annual temperatures have been encountered during El Niño events, such as in 1997–1998.

The variability of observed global surface temperatures from 1870 to 2010 (Figure 1a) reflects the combined influences of natural (solar, volcanic, internal; FAQ 5.1, Figure 1b–d) factors, superimposed on the multi-decadal warming trend from anthropogenic factors (FAQ 5.1, Figure 1e).

Prior to 1870, when anthropogenic emissions of greenhouse gases and aerosols were smaller, changes in solar and volcanic activity and internal variability played a more important role, although the specific contributions of these individual factors to global surface temperatures are less certain. Solar minima lasting several decades have often been associated with cold conditions. However, these periods are often also affected by volcanic eruptions, making it difficult to quantify the solar contribution.

At the regional scale, changes in solar activity have been related to changes in surface climate and atmospheric circulation in the Indo-Pacific, Northern Asia and North Atlantic areas. The mechanisms that amplify the regional effects of the relatively small fluctuations of TSI in the roughly 11-year solar cycle involve dynamical interactions between the upper and the lower atmosphere, or between the ocean sea surface temperature and atmosphere, and have little effect on global mean temperatures (see Box 10.2).

Finally, a decrease in solar activity during the past solar minimum a few years ago (FAQ 5.1, Figure 1b) raises the question of its future influence on climate. Despite uncertainties in future solar activity, there is *high confidence* that the effects of solar activity within the range of grand solar maxima and minima will be much smaller than the changes due to anthropogenic effects.




**FAQ 5.1, Figure 1 |** Global surface temperature anomalies from 1870 to 2010, and the natural (solar, volcanic, and internal) and anthropogenic factors that influence them. (a) Global surface temperature record (1870–2010) relative to the average global surface temperature for 1961–1990 (black line). A model of global surface temperature change (a: red line) produced using the sum of the impacts on temperature of natural (b, c, d) and anthropogenic factors (e). (b) Estimated temperature response to solar forcing. (c) Estimated temperature response to volcanic eruptions. (d) Estimated temperature variability due to internal variability, here related to the El Niño-Southern Oscillation. (e) Estimated temperature response to anthropogenic forcing, consisting of a warming component from greenhouse gases, and a cooling component from most aerosols.

FAQ

Frequently Asked Questions

## FAQ 5.2 | How Unusual is the Current Sea Level Rate of Change?

*The rate of mean global sea level change—averaging 1.7 ± 0.2 mm yr$^{-1}$ for the entire 20th century and between 2.8 and 3.6 mm yr$^{-1}$ since 1993 (Chapter 13)—is unusual in the context of centennial-scale variations of the last two millennia. However, much more rapid rates of sea level change occurred during past periods of rapid ice sheet disintegration, such as transitions between glacial and interglacial periods. Exceptional tectonic effects can also drive very rapid local sea level changes, with local rates exceeding the current global rates of change.*

'Sea level' is commonly thought of as the point where the ocean meets the land. Earth scientists define sea level as a measure of the position of the sea surface relative to the land, both of which may be moving relative to the center of the Earth. A measure of sea level therefore reflects a combination of geophysical and climate factors. Geophysical factors affecting sea level include land subsidence or uplift and glacial isostatic adjustments—the earth–ocean system's response to changes in mass distribution on the Earth, specifically ocean water and land ice.

Climate influences include variations in ocean temperatures, which cause sea water to expand or contract, changes in the volume of glaciers and ice sheets, and shifts in ocean currents. Local and regional changes in these climate and geophysical factors produce significant deviations from the global estimate of the mean rate of sea level change. For example, *local* sea level is falling at a rate approaching 10 mm yr$^{-1}$ along the northern Swedish coast (Gulf of Bothnia), due to ongoing uplift caused by continental ice that melted after the last glacial period. In contrast, *local* sea level rose at a rate of ~20 mm yr$^{-1}$ from 1960 to 2005 south of Bangkok, mainly in response to subsidence due to ground water extraction.

For the past ~150 years, sea level change has been recorded at tide gauge stations, and for the past ~20 years, with satellite altimeters. Results of these two data sets are consistent for the overlapping period. The globally averaged rate of sea level rise of ~1.7 ± 0.2 mm yr$^{-1}$ over the 20th century—and about twice that over the past two decades—may seem small compared with observations of wave and tidal oscillations around the globe that can be orders of magnitude larger. However, if these rates persist over long time intervals, the magnitude carries important consequences for heavily populated, low-lying coastal regions, where even a small increase in sea level can inundate large land areas.

Prior to the instrumental period, local rates of sea level change are estimated from indirect measures recorded in sedimentary, fossil and archaeological archives. These proxy records are spatially limited and reflect both local and global conditions. Reconstruction of a global signal is strengthened, though, when individual proxy records from widely different environmental settings converge on a common signal. It is important to note that geologic archives—particularly those before about 20,000 years ago—most commonly only capture millennial-scale changes in sea level. Estimates of century-scale rates of sea level change are therefore based on millennial-scale information, but it must be recognised that such data do not necessarily preclude more rapid rates of century-scale changes in sea level.

Sea level reconstructions for the last two millennia offer an opportunity to use proxy records to overlap with, and extend beyond, the instrumental period. A recent example comes from salt-marsh deposits on the Atlantic Coast of the United States, combined with sea level reconstructions based on tide-gauge data and model predictions, to document an average rate of sea level change since the late 19th century of 2.1 ± 0.2 mm yr$^{-1}$. This century-long rise exceeds any other century-scale change rate in the entire 2000-year record for this same section of coast.

On longer time scales, much larger rates and amplitudes of sea level changes have sometimes been encountered. Glacial–interglacial climate cycles over the past 500,000 years resulted in global sea level changes of up to about 120 to 140 m. Much of this sea level change occurred in 10,000 to 15,000 years, during the transition from a full glacial period to an interglacial period, at average rates of 10 to 15 mm yr$^{-1}$. These high rates are only sustainable when the Earth is emerging from periods of extreme glaciation, when large ice sheets contact the oceans. For example, during the transition from the last glacial maximum (about 21,000 years ago) to the present interglacial (Holocene, last 11,650 years), fossil coral reef deposits indicate that global sea level rose abruptly by 14 to 18 m in less than 500 years. This event is known as Meltwater Pulse 1A, in which the rate of sea level rise reached more than 40 mm yr$^{-1}$.

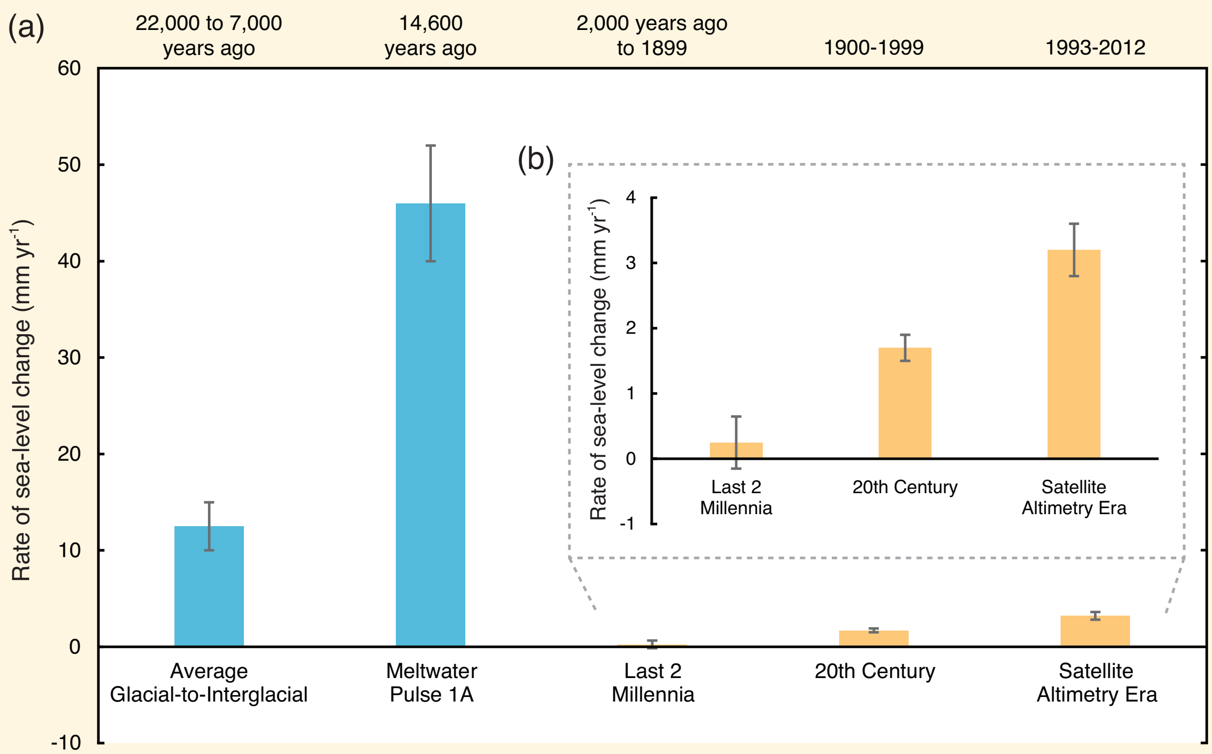
These examples from longer time scales indicate rates of sea level change greater than observed today, but it should be remembered that they all occurred in special circumstances: at times of transition from full glacial to interglacial condition; at locations where the long-term after-effects of these transitions are still occurring; at locations of

*(continued on next page)*

*FAQ 5.2 (continued)*

major tectonic upheavals or in major deltas, where subsidence due to sediment compaction—sometimes amplified by ground-fluid extraction—dominates.

The instrumental and geologic record support the conclusion that the current rate of mean global sea level change is unusual relative to that observed and/or estimated over the last two millennia. Higher rates have been observed in the geological record, especially during times of transition between glacial and interglacial periods.




**FAQ 5.2, Figure 1 |** (a) Estimates of the average rate of global mean sea level change (in mm yr$^{-1}$) for five selected time intervals: last glacial-to-interglacial transition; Meltwater Pulse 1A; last 2 millennia; 20th century; satellite altimetry era (1993–2012). Blue columns denote time intervals of transition from a glacial to an interglacial period, whereas orange columns denote the current interglacial period. Black bars indicate the range of likely values of the average rate of global mean sea level change. Note the overall higher rates of global mean sea level change characteristic of times of transition between glacial and interglacial periods. (b) Expanded view of the rate of global mean sea level change during three time intervals of the present interglacial.

FAQ

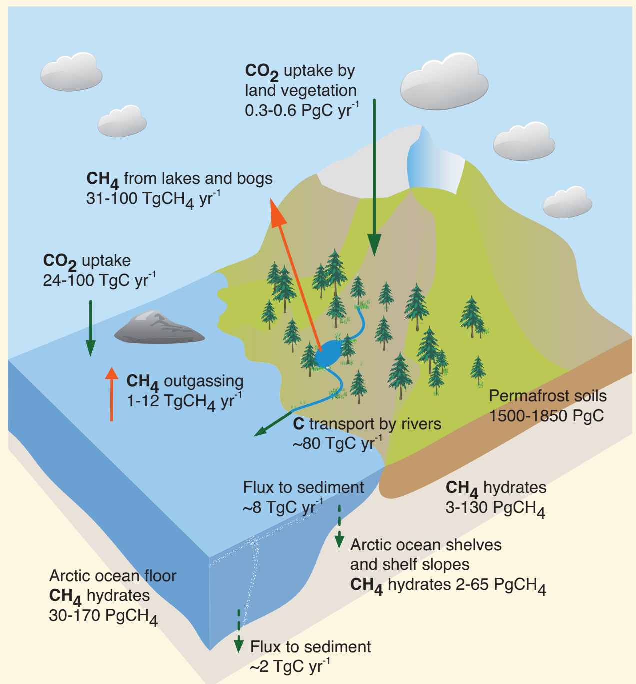
Frequently Asked Questions

## FAQ 6.1 | Could Rapid Release of Methane and Carbon Dioxide from Thawing Permafrost or Ocean Warming Substantially Increase Warming?

*Permafrost is permanently frozen ground, mainly found in the high latitudes of the Arctic. Permafrost, including the sub-sea permafrost on the shallow shelves of the Arctic Ocean, contains old organic carbon deposits. Some are relicts from the last glaciation, and hold at least twice the amount of carbon currently present in the atmosphere as carbon dioxide ($CO_2$). Should a sizeable fraction of this carbon be released as methane and $CO_2$, it would increase atmospheric concentrations, which would lead to higher atmospheric temperatures. That in turn would cause yet more methane and $CO_2$ to be released, creating a positive feedback, which would further amplify global warming.*

*The Arctic domain presently represents a net sink of $CO_2$—sequestering around 0.4 ± 0.4 PgC $yr^{-1}$ in growing vegetation representing about 10% of the current global land sink. It is also a modest source of methane ($CH_4$): between 15 and 50 Tg($CH_4$) $yr^{-1}$ are emitted mostly from seasonally unfrozen wetlands corresponding to about 10% of the global wetland methane source. There is no clear evidence yet that thawing contributes significantly to the current global budgets of these two greenhouse gases. However, under sustained Arctic warming, modelling studies and expert judgments indicate with medium agreement that a potential combined release totalling up to 350 PgC as $CO_2$ equivalent could occur by the year 2100.*

Permafrost soils on land, and in ocean shelves, contain large pools of organic carbon, which must be thawed and decomposed by microbes before it can be released—mostly as $CO_2$. Where oxygen is limited, as in waterlogged soils, some microbes also produce methane.




**FAQ 6.1, Figure 1 |** A simplified graph of current major carbon pools and flows in the Arctic domain, including permafrost on land, continental shelves and ocean. (Adapted from McGuire et al., 2009; and Tarnocai et al., 2009.) TgC = $10^{12}$ gC, and PgC = $10^{15}$ gC.

On land, permafrost is overlain by a surface 'active layer', which thaws during summer and forms part of the tundra ecosystem. If spring and summer temperatures become warmer on average, the active layer will thicken, making more organic carbon available for microbial decomposition. However, warmer summers would also result in greater uptake of carbon dioxide by Arctic vegetation through photosynthesis. That means the net Arctic carbon balance is a delicate one between enhanced uptake and enhanced release of carbon.

Hydrological conditions during the summer thaw are also important. The melting of bodies of excess ground ice may create standing water conditions in pools and lakes, where lack of oxygen will induce methane production. The complexity of Arctic landscapes under climate warming means we have *low confidence* in which of these different processes might dominate on a regional scale. Heat diffusion and permafrost melting takes time—in fact, the deeper Arctic permafrost can be seen as a relict of the last glaciation, which is still slowly eroding—so any significant loss of permafrost soil carbon will happen over long time scales.

Given enough oxygen, decomposition of organic matter in soil is accompanied by the release of heat by microbes (similar to compost), which, during summer, might stimulate further permafrost thaw. Depending on carbon and ice content of the permafrost, and the hydrological regime, this mechanism could, under warming, trigger relatively fast local permafrost degradation. *(continued on next page)*

*FAQ 6.1 (continued)*

Modelling studies of permafrost dynamics and greenhouse gas emissions indicate a relatively slow positive feedback, on time scales of hundreds of years. Until the year 2100, up to 250 PgC could be released as $CO_2$, and up to 5 Pg as $CH_4$. Given methane's stronger greenhouse warming potential, that corresponds to a further 100 PgC of equivalent $CO_2$ released until the year 2100. These amounts are similar in magnitude to other biogeochemical feedbacks, for example, the additional $CO_2$ released by the global warming of terrestrial soils. However, current models do not include the full complexity of Arctic processes that occur when permafrost thaws, such as the formation of lakes and ponds.

Methane hydrates are another form of frozen carbon, occurring in deep permafrost soils, ocean shelves, shelf slopes and deeper ocean bottom sediments. They consist of methane and water molecule clusters, which are only stable in a specific window of low temperatures and high pressures. On land and in the ocean, most of these hydrates originate from marine or terrestrial biogenic carbon, decomposed in the absence of oxygen and trapped in an aquatic environment under suitable temperature–pressure conditions.

Any warming of permafrost soils, ocean waters and sediments and/or changes in pressure could destabilise those hydrates, releasing their $CH_4$ to the ocean. During larger, more sporadic releases, a fraction of that $CH_4$ might also be outgassed to the atmosphere. There is a large pool of these hydrates: in the Arctic alone, the amount of $CH_4$ stored as hydrates could be more than 10 times greater than the $CH_4$ presently in the global atmosphere.

Like permafrost thawing, liberating hydrates on land is a slow process, taking decades to centuries. The deeper ocean regions and bottom sediments will take still longer—between centuries and millennia to warm enough to destabilise the hydrates within them. Furthermore, methane released in deeper waters has to reach the surface and atmosphere before it can become climatically active, but most is expected to be consumed by microorganisms before it gets there. Only the $CH_4$ from hydrates in shallow shelves, such as in the Arctic Ocean north of Eastern Siberia, may actually reach the atmosphere to have a climate impact.

Several recent studies have documented locally significant $CH_4$ emissions over the Arctic Siberian shelf and from Siberian lakes. How much of this $CH_4$ originates from decomposing organic carbon or from destabilizing hydrates is not known. There is also no evidence available to determine whether these sources have been stimulated by recent regional warming, or whether they have always existed—it may be possible that these $CH_4$ seepages have been present since the last deglaciation. In any event, these sources make a very small contribution to the global $CH_4$ budget—less than 5%. This is also confirmed by atmospheric methane concentration observations, which do not show any substantial increases over the Arctic.

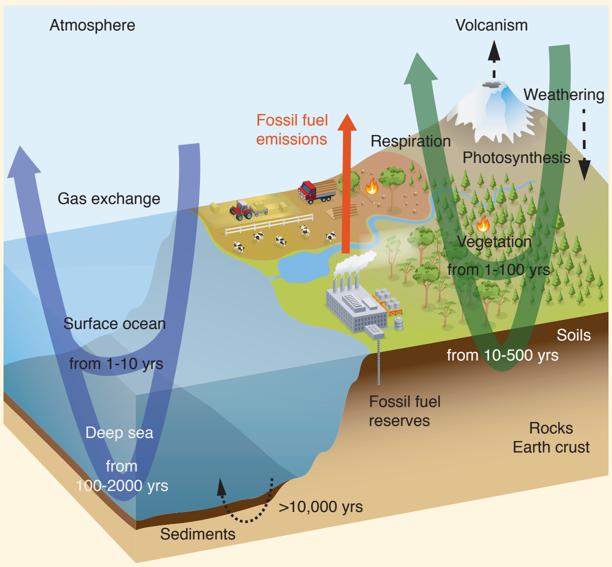
However modelling studies and expert judgment indicate that $CH_4$ and $CO_2$ emissions will increase under Arctic warming, and that they will provide a positive climate feedback. Over centuries, this feedback will be moderate: of a magnitude similar to other climate–terrestrial ecosystem feedbacks. Over millennia and longer, however, $CO_2$ and $CH_4$ releases from permafrost and shelves/shelf slopes are much more important, because of the large carbon and methane hydrate pools involved.

Frequently Asked Questions

## FAQ 6.2 | What Happens to Carbon Dioxide After It Is Emitted into the Atmosphere?

*Carbon dioxide ($CO_2$), after it is emitted into the atmosphere, is firstly rapidly distributed between atmosphere, the upper ocean and vegetation. Subsequently, the carbon continues to be moved between the different reservoirs of the global carbon cycle, such as soils, the deeper ocean and rocks. Some of these exchanges occur very slowly. Depending on the amount of $CO_2$ released, between 15% and 40% will remain in the atmosphere for up to 2000 years, after which a new balance is established between the atmosphere, the land biosphere and the ocean. Geological processes will take anywhere from tens to hundreds of thousands of years—perhaps longer—to redistribute the carbon further among the geological reservoirs. Higher atmospheric $CO_2$ concentrations, and associated climate impacts of present emissions, will, therefore, persist for a very long time into the future.*

$CO_2$ is a largely non-reactive gas, which is rapidly mixed throughout the entire troposphere in less than a year. Unlike reactive chemical compounds in the atmosphere that are removed and broken down by sink processes, such as methane, carbon is instead redistributed among the different reservoirs of the global carbon cycle and ultimately recycled back to the atmosphere on a multitude of time scales. FAQ 6.2, Figure 1 shows a simplified diagram of the global carbon cycle. The open arrows indicate typical timeframes for carbon atoms to be transferred through the different reservoirs.




**FAQ 6.2, Figure 1 |** Simplified schematic of the global carbon cycle showing the typical turnover time scales for carbon transfers through the major reservoirs.

Before the Industrial Era, the global carbon cycle was roughly balanced. This can be inferred from ice core measurements, which show a near constant atmospheric concentration of $CO_2$ over the last several thousand years prior to the Industrial Era. Anthropogenic emissions of carbon dioxide into the atmosphere, however, have disturbed that equilibrium. As global $CO_2$ concentrations rise, the exchange processes between $CO_2$ and the surface ocean and vegetation are altered, as are subsequent exchanges within and among the carbon reservoirs on land, in the ocean and eventually, the Earth crust. In this way, the added carbon is redistributed by the global carbon cycle, until the exchanges of carbon between the different carbon reservoirs have reached a new, approximate balance.

Over the ocean, $CO_2$ molecules pass through the air-sea interface by gas exchange. In seawater, $CO_2$ interacts with water molecules to form carbonic acid, which reacts very quickly with the large reservoir of dissolved inorganic carbon—bicarbonate and carbonate ions—in the ocean. Currents and the formation of sinking dense waters transport the carbon between the surface and deeper layers of the ocean. The marine biota also redistribute carbon: marine organisms grow organic tissue and calcareous shells in surface waters, which, after their death, sink to deeper waters, where they are returned to the dissolved inorganic carbon reservoir by dissolution and microbial decomposition. A small fraction reaches the sea floor, and is incorporated into the sediments.

The extra carbon from anthropogenic emissions has the effect of increasing the atmospheric partial pressure of $CO_2$, which in turn increases the air-to-sea exchange of $CO_2$ molecules. In the surface ocean, the carbonate chemistry quickly accommodates that extra $CO_2$. As a result, shallow surface ocean waters reach balance with the atmosphere within 1 or 2 years. Movement of the carbon from the surface into the middle depths and deeper waters takes longer—between decades and many centuries. On still longer time scales, acidification by the invading $CO_2$ dissolves carbonate sediments on the sea floor, which further enhances ocean uptake. However, current understanding suggests that, unless substantial ocean circulation changes occur, plankton growth remains roughly unchanged because it is limited mostly by environmental factors, such as nutrients and light, and not by the availability of inorganic carbon it does not contribute significantly to the ocean uptake of anthropogenic $CO_2$. *(continued on next page)*

*FAQ 6.2 (continued)*

On land, vegetation absorbs $CO_2$ by photosynthesis and converts it into organic matter. A fraction of this carbon is immediately returned to the atmosphere as $CO_2$ by plant respiration. Plants use the remainder for growth. Dead plant material is incorporated into soils, eventually to be decomposed by microorganisms and then respired back into the atmosphere as $CO_2$. In addition, carbon in vegetation and soils is also converted back into $CO_2$ by fires, insects, herbivores, as well as by harvest of plants and subsequent consumption by livestock or humans. Some organic carbon is furthermore carried into the ocean by streams and rivers.

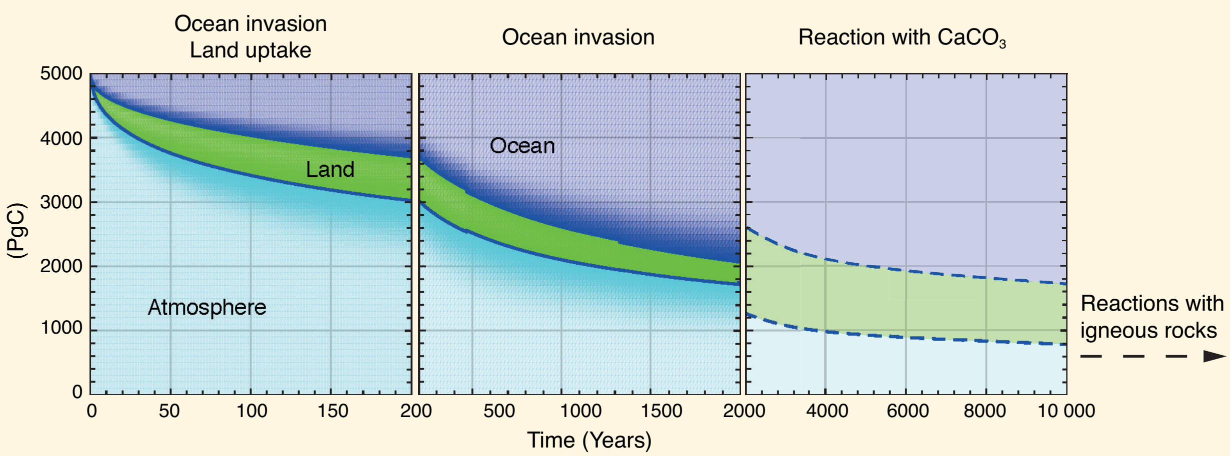
An increase in atmospheric $CO_2$ stimulates photosynthesis, and thus carbon uptake. In addition, elevated $CO_2$ concentrations help plants in dry areas to use ground water more efficiently. This in turn increases the biomass in vegetation and soils and so fosters a carbon sink on land. The magnitude of this sink, however, also depends critically on other factors, such as water and nutrient availability.

Coupled carbon-cycle climate models indicate that less carbon is taken up by the ocean and land as the climate warms constituting a positive climate feedback. Many different factors contribute to this effect: warmer seawater, for instance, has a lower $CO_2$ solubility, so altered chemical carbon reactions result in less oceanic uptake of excess atmospheric $CO_2$. On land, higher temperatures foster longer seasonal growth periods in temperate and higher latitudes, but also faster respiration of soil carbon.



The time it takes to reach a new carbon distribution balance depends on the transfer times of carbon through the different reservoirs, and takes place over a multitude of time scales. Carbon is first exchanged among the 'fast' carbon reservoirs, such as the atmosphere, surface ocean, land vegetation and soils, over time scales up to a few thousand years. Over longer time scales, very slow secondary geological processes—dissolution of carbonate sediments and sediment burial into the Earth's crust—become important.

FAQ 6.2, Figure 2 illustrates the decay of a large excess amount of $CO_2$ (5000 PgC, or about 10 times the cumulative $CO_2$ emitted so far since the beginning of the industrial Era) emitted into the atmosphere, and how it is redistributed among land and the ocean over time. During the first 200 years, the ocean and land take up similar amounts of carbon. On longer time scales, the ocean uptake dominates mainly because of its larger reservoir size (~38,000 PgC) as compared to land (~4000 PgC) and atmosphere (589 PgC prior to the Industrial Era). Because of ocean chemistry the size of the initial input is important: higher emissions imply that a larger fraction of $CO_2$ will remain in the atmosphere. After 2000 years, the atmosphere will still contain between 15% and 40% of those initial $CO_2$ emissions. A further reduction by carbonate sediment dissolution, and reactions with igneous rocks, such as silicate weathering and sediment burial, will take anything from tens to hundreds of thousands of years, or even longer.




**FAQ 6.2, Figure 2 |** Decay of a $CO_2$ excess amount of 5000 PgC emitted at time zero into the atmosphere, and its subsequent redistribution into land and ocean as a function of time, computed by coupled carbon-cycle climate models. The sizes of the colour bands indicate the carbon uptake by the respective reservoir. The first two panels show the multi-model mean from a model intercomparison project (Joos et al., 2013). The last panel shows the longer term redistribution including ocean dissolution of carbonaceous sediments as computed with an Earth System Model of Intermediate Complexity (after Archer et al., 2009b).

Frequently Asked Questions

## FAQ 7.1 | How Do Clouds Affect Climate and Climate Change?

*Clouds strongly affect the current climate, but observations alone cannot yet tell us how they will affect a future, warmer climate. Comprehensive prediction of changes in cloudiness requires a global climate model. Such models simulate cloud fields that roughly resemble those observed, but important errors and uncertainties remain. Different climate models produce different projections of how clouds will change in a warmer climate. Based on all available evidence, it seems likely that the net cloud–climate feedback amplifies global warming. If so, the strength of this amplification remains uncertain.*

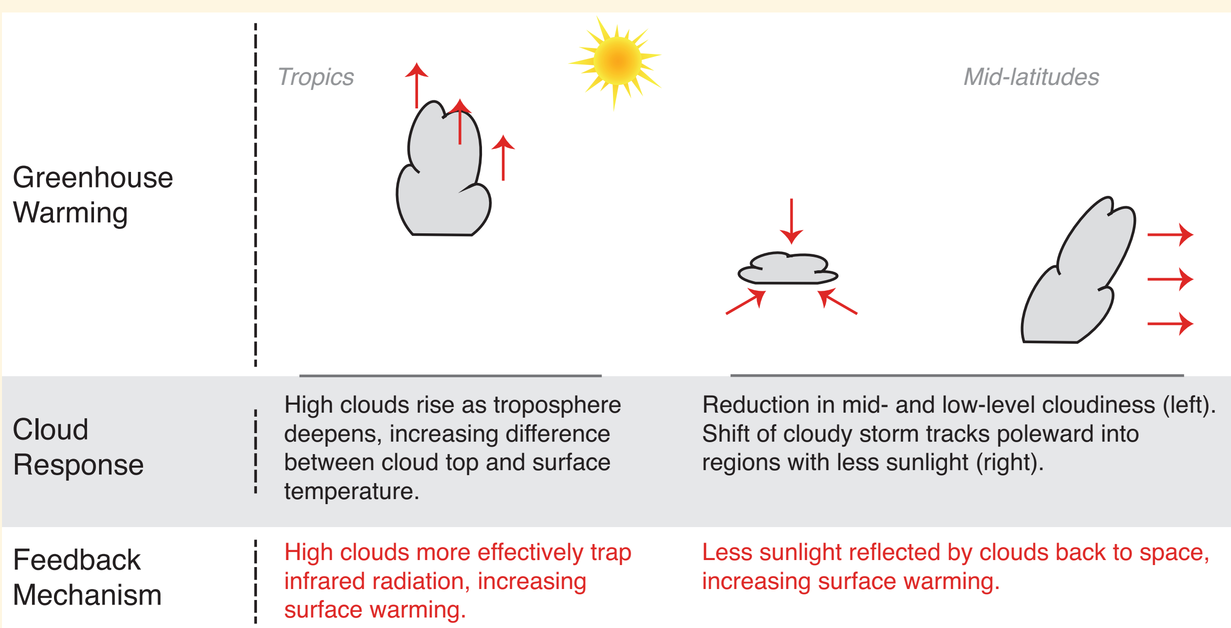
Since the 1970s, scientists have recognized the critical importance of clouds for the climate system, and for climate change. Clouds affect the climate system in a variety of ways. They produce precipitation (rain and snow) that is necessary for most life on land. They warm the atmosphere as water vapour condenses. Although some of the condensed water re-evaporates, the precipitation that reaches the surface represents a net warming of the air. Clouds strongly affect the flows of both sunlight (warming the planet) and infrared light (cooling the planet as it is radiated to space) through the atmosphere. Finally, clouds contain powerful updraughts that can rapidly carry air from near the surface to great heights. The updraughts carry energy, moisture, momentum, trace gases, and aerosol particles. For decades, climate scientists have been using both observations and models to study how clouds change with the daily weather, with the seasonal cycle, and with year-to-year changes such as those associated with El Niño.

All cloud processes have the potential to change as the climate state changes. Cloud feedbacks are of intense interest in the context of climate change. Any change in a cloud process that is caused by climate change—and in turn influences climate—represents a cloud–climate feedback. Because clouds interact so strongly with both sunlight and infrared light, small changes in cloudiness can have a potent effect on the climate system.

Many possible types of cloud–climate feedbacks have been suggested, involving changes in cloud amount, cloud-top height and/or cloud reflectivity (see FAQ7.1, Figure 1). The literature shows consistently that high clouds amplify global warming as they interact with infrared light emitted by the atmosphere and surface. There is more uncertainty, however, about the feedbacks associated with low-altitude clouds, and about cloud feedbacks associated with amount and reflectivity in general.

Thick high clouds efficiently reflect sunlight, and both thick and thin high clouds strongly reduce the amount of infrared light that the atmosphere and surface emit to space. The compensation between these two effects makes

*(continued on next page)*




**FAQ 7.1, Figure 1 |** Schematic of important cloud feedback mechanisms.

*FAQ 7.1 (continued)*

the surface temperature somewhat less sensitive to changes in high cloud amount than to changes in low cloud amount. This compensation could be disturbed if there were a systematic shift from thick high cloud to thin cirrus cloud or vice versa; while this possibility cannot be ruled out, it is not currently supported by any evidence. On the other hand, changes in the altitude of high clouds (for a given high-cloud amount) can strongly affect surface temperature. An upward shift in high clouds reduces the infrared light that the surface and atmosphere emit to space, but has little effect on the reflected sunlight. There is strong evidence of such a shift in a warmer climate. This amplifies global warming by preventing some of the additional infrared light emitted by the atmosphere and surface from leaving the climate system.

Low clouds reflect a lot of sunlight back to space but, for a given state of the atmosphere and surface, they have only a weak effect on the infrared light that is emitted to space by the Earth. As a result, they have a net cooling effect on the present climate; to a lesser extent, the same holds for mid-level clouds. In a future climate warmed by increasing greenhouse gases, most IPCC-assessed climate models simulate a decrease in low and mid-level cloud amount, which would increase the absorption of sunlight and so tend to increase the warming. The extent of this decrease is quite model-dependent, however.

There are also other ways that clouds may change in a warmer climate. Changes in wind patterns and storm tracks could affect the regional and seasonal patterns of cloudiness and precipitation. Some studies suggest that the signal of one such trend seen in climate models—a poleward migration of the clouds associated with mid-latitude storm tracks—is already detectable in the observational record. By shifting clouds into regions receiving less sunlight, this could also amplify global warming. More clouds may be made of liquid drops, which are small but numerous and reflect more sunlight back to space than a cloud composed of the same mass of larger ice crystals. Thin cirrus cloud, which exerts a net warming effect and is very hard for climate models to simulate, could change in ways not simulated by models although there is no evidence for this. Other processes may be regionally important, for example, interactions between clouds and the surface can change over the ocean where sea ice melts, and over land where plant transpiration is reduced.

There is as yet no broadly accepted way to infer global cloud feedbacks from observations of long-term cloud trends or shorter-time scale variability. Nevertheless, all the models used for the current assessment (and the preceding two IPCC assessments) produce net cloud feedbacks that either enhance anthropogenic greenhouse warming or have little overall effect. Feedbacks are not 'put into' the models, but emerge from the functioning of the clouds in the simulated atmosphere and their effects on the flows and transformations of energy in the climate system. The differences in the strengths of the cloud feedbacks produced by the various models largely account for the different sensitivities of the models to changes in greenhouse gas concentrations.

Frequently Asked Questions

## FAQ 7.2 | How Do Aerosols Affect Climate and Climate Change?

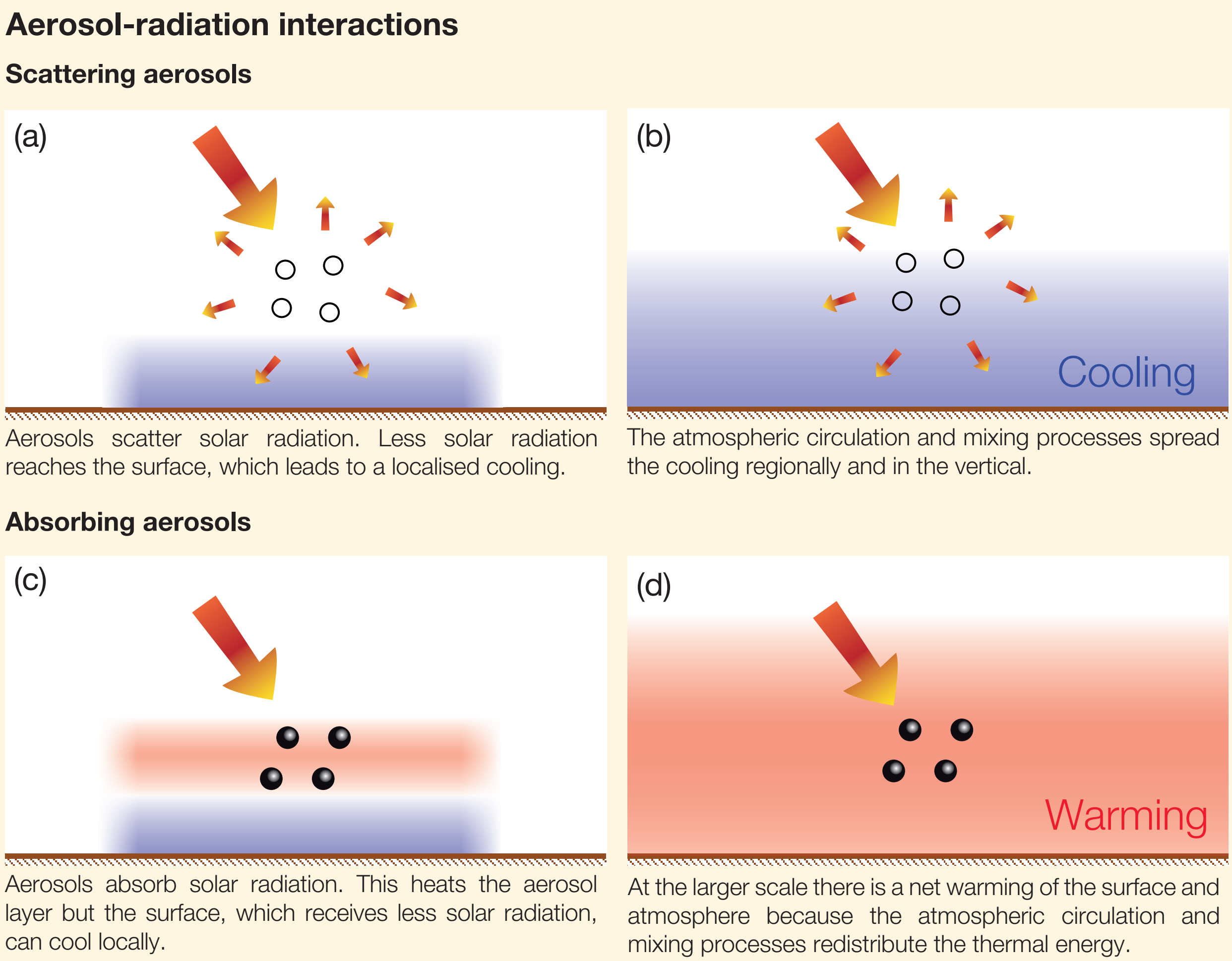
*Atmospheric aerosols are composed of small liquid or solid particles suspended in the atmosphere, other than larger cloud and precipitation particles. They come from natural and anthropogenic sources, and can affect the climate in multiple and complex ways through their interactions with radiation and clouds. Overall, models and observations indicate that anthropogenic aerosols have exerted a cooling influence on the Earth since pre-industrial times, which has masked some of the global mean warming from greenhouse gases that would have occurred in their absence. The projected decrease in emissions of anthropogenic aerosols in the future, in response to air quality policies, would eventually unmask this warming.*

Atmospheric aerosols have a typical lifetime of one day to two weeks in the troposphere, and about one year in the stratosphere. They vary greatly in size, chemical composition and shape. Some aerosols, such as dust and sea spray, are mostly or entirely of natural origin, while other aerosols, such as sulphates and smoke, come from both natural and anthropogenic sources.

Aerosols affect climate in many ways. First, they scatter and absorb sunlight, which modifies the Earth's radiative balance (see FAQ.7.2, Figure 1). Aerosol scattering generally makes the planet more reflective, and tends to cool the climate, while aerosol absorption has the opposite effect, and tends to warm the climate system. The balance between cooling and warming depends on aerosol properties and environmental conditions. Many observational studies have quantified local radiative effects from anthropogenic and natural aerosols, but determining their

*(continued on next page)*





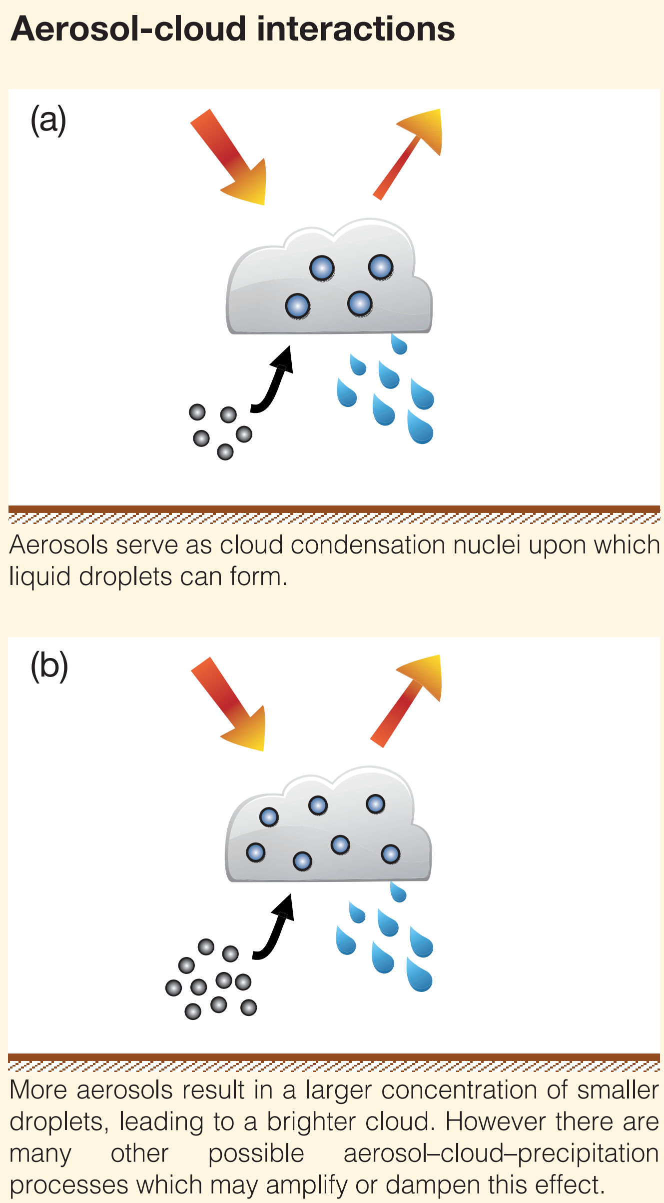

**FAQ 7.2, Figure 1 |** Overview of interactions between aerosols and solar radiation and their impact on climate. The left panels show the instantaneous radiative effects of aerosols, while the right panels show their overall impact after the climate system has responded to their radiative effects.

*FAQ 7.2 (continued)*

global impact requires satellite data and models. One of the remaining uncertainties comes from black carbon, an absorbing aerosol that not only is more difficult to measure than scattering aerosols, but also induces a complicated cloud response. Most studies agree, however, that the overall radiative effect from anthropogenic aerosols is to cool the planet.

Aerosols also serve as condensation and ice nucleation sites, on which cloud droplets and ice particles can form (see FAQ.7.2, Figure 2). When influenced by more aerosol particles, clouds of liquid water droplets tend to have more, but smaller droplets, which causes these clouds to reflect more solar radiation. There are however many other pathways for aerosol–cloud interactions, particularly in ice—or mixed liquid and ice—clouds, where phase changes between liquid and ice water are sensitive to aerosol concentrations and properties. The initial view that an increase in aerosol concentration will also increase the amount of low clouds has been challenged because a number of counteracting processes come into play. Quantifying the overall impact of aerosols on cloud amounts and properties is understandably difficult. Available studies, based on climate models and satellite observations, generally indicate that the net effect of anthropogenic aerosols on clouds is to cool the climate system.

Because aerosols are distributed unevenly in the atmosphere, they can heat and cool the climate system in patterns that can drive changes in the weather. These effects are complex, and hard to simulate with current models, but several studies suggest significant effects on precipitation in certain regions.




**FAQ 7.2, Figure 2 |** Overview of aerosol–cloud interactions and their impact on climate. Panels (a) and (b) represent a clean and a polluted low-level cloud, respectively.

Because of their short lifetime, the abundance of aerosols—and their climate effects—have varied over time, in rough concert with anthropogenic emissions of aerosols and their precursors in the gas phase such as sulphur dioxide ($SO_2$) and some volatile organic compounds. Because anthropogenic aerosol emissions have increased substantially over the industrial period, this has counteracted some of the warming that would otherwise have occurred from increased concentrations of well mixed greenhouse gases. Aerosols from large volcanic eruptions that enter the stratosphere, such as those of El Chichón and Pinatubo, have also caused cooling periods that typically last a year or two.

Over the last two decades, anthropogenic aerosol emissions have decreased in some developed countries, but increased in many developing countries. The impact of aerosols on the global mean surface temperature over this particular period is therefore thought to be small. It is projected, however, that emissions of anthropogenic aerosols will ultimately decrease in response to air quality policies, which would suppress their cooling influence on the Earth's surface, thus leading to increased warming.

Frequently Asked Questions

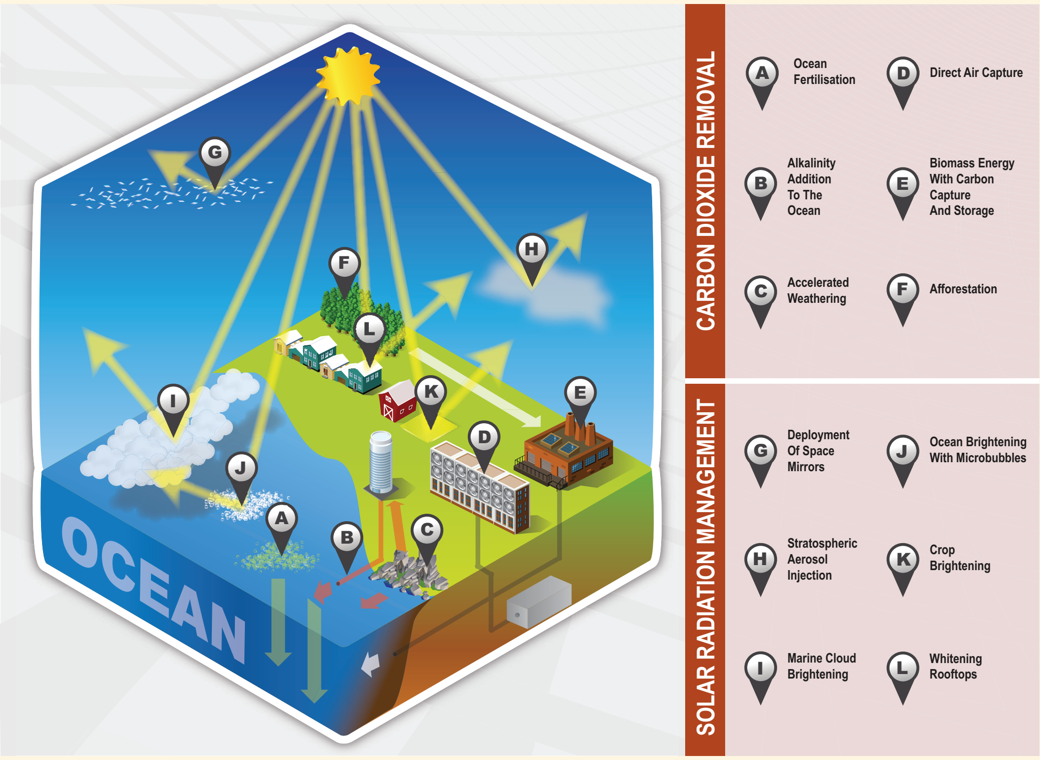
## FAQ 7.3 | Could Geoengineering Counteract Climate Change and What Side Effects Might Occur?

*Geoengineering—also called climate engineering—is defined as a broad set of methods and technologies that aim to deliberately alter the climate system in order to alleviate impacts of climate change. Two distinct categories of geoengineering methods are usually considered: Solar Radiation Management (SRM, assessed in Section 7.7) aims to offset the warming from anthropogenic greenhouse gases by making the planet more reflective while Carbon Dioxide Removal (CDR, assessed in Section 6.5) aims at reducing the atmospheric $CO_2$ concentration. The two categories operate on different physical principles and on different time scales. Models suggest that if SRM methods were realizable they would be effective in countering increasing temperatures, and would be less, but still, effective in countering some other climate changes. SRM would not counter all effects of climate change, and all proposed geoengineering methods also carry risks and side effects. Additional consequences cannot yet be anticipated as the level of scientific understanding about both SRM and CDR is low. There are also many (political, ethical, and practical) issues involving geoengineering that are beyond the scope of this report.*

### Carbon Dioxide Removal Methods

CDR methods aim at removing $CO_2$ from the atmosphere by deliberately modifying carbon cycle processes, or by industrial (e.g., chemical) approaches. The carbon withdrawn from the atmosphere would then be stored in land, ocean or in geological reservoirs. Some CDR methods rely on biological processes, such as large-scale afforestation/reforestation, carbon sequestration in soils through biochar, bioenergy with carbon capture and storage (BECCS) and ocean fertilization. Others would rely on geological processes, such as accelerated weathering of silicate and carbonate rocks—on land or in the ocean (see FAQ.7.3, Figure 1). The $CO_2$ removed from the atmosphere would

(continued on next page)




**FAQ 7.3, Figure 1 |** Overview of some proposed geoengineering methods as they have been suggested. Carbon Dioxide Removal methods (see Section 6.5 for details): (A) nutrients are added to the ocean (ocean fertilization), which increases oceanic productivity in the surface ocean and transports a fraction of the resulting biogenic carbon downward; (B) alkalinity from solid minerals is added to the ocean, which causes more atmospheric $CO_2$ to dissolve in the ocean; (C) the weathering rate of silicate rocks is increased, and the dissolved carbonate minerals are transported to the ocean; (D) atmospheric $CO_2$ is captured chemically, and stored either underground or in the ocean; (E) biomass is burned at an electric power plant with carbon capture, and the captured $CO_2$ is stored either underground or in the ocean; and (F) $CO_2$ is captured through afforestation and reforestation to be stored in land ecosystems. Solar Radiation Management methods (see Section 7.7 for details): (G) reflectors are placed in space to reflect solar radiation; (H) aerosols are injected in the stratosphere; (I) marine clouds are seeded in order to be made more reflective; (J) microbubbles are produced at the ocean surface to make it more reflective; (K) more reflective crops are grown; and (L) roofs and other built structures are whitened.

*FAQ 7.3 (continued)*

then be stored in organic form in land reservoirs, or in inorganic form in oceanic and geological reservoirs, where it would have to be stored for at least hundreds of years for CDR to be effective.

CDR methods would reduce the radiative forcing of $CO_2$ inasmuch as they are effective at removing $CO_2$ from the atmosphere and keeping the removed carbon away from the atmosphere. Some methods would also reduce ocean acidification (see FAQ 3.2), but other methods involving oceanic storage might instead increase ocean acidification if the carbon is sequestered as dissolved $CO_2$. A major uncertainty related to the effectiveness of CDR methods is the storage capacity and the permanence of stored carbon. Permanent carbon removal and storage by CDR would decrease climate warming in the long term. However, non-permanent storage strategies would allow $CO_2$ to return back to the atmosphere where it would once again contribute to warming. An intentional removal of $CO_2$ by CDR methods will be partially offset by the response of the oceanic and terrestrial carbon reservoirs if the $CO_2$ atmospheric concentration is reduced. This is because some oceanic and terrestrial carbon reservoirs will outgas to the atmosphere the anthropogenic $CO_2$ that had previously been stored. To completely offset past anthropogenic $CO_2$ emissions, CDR techniques would therefore need to remove not just the $CO_2$ that has accumulated in the atmosphere since pre-industrial times, but also the anthropogenic carbon previously taken up by the terrestrial biosphere and the ocean.

Biological and most chemical weathering CDR methods cannot be scaled up indefinitely and are necessarily limited by various physical or environmental constraints such as competing demands for land. Assuming a maximum CDR sequestration rate of 200 PgC per century from a combination of CDR methods, it would take about one and half centuries to remove the $CO_2$ emitted in the last 50 years, making it difficult—even for a suite of additive CDR methods—to mitigate climate change rapidly. Direct air capture methods could in principle operate much more rapidly, but may be limited by large-scale implementation, including energy use and environmental constraints.



CDR could also have climatic and environmental side effects. For instance, enhanced vegetation productivity may increase emissions of $N_2O$, which is a more potent greenhouse gas than $CO_2$. A large-scale increase in vegetation coverage, for instance through afforestation or energy crops, could alter surface characteristics, such as surface reflectivity and turbulent fluxes. Some modelling studies have shown that afforestation in seasonally snow-covered boreal regions could in fact accelerate global warming, whereas afforestation in the tropics may be more effective at slowing global warming. Ocean-based CDR methods that rely on biological production (i.e., ocean fertilization) would have numerous side effects on ocean ecosystems, ocean acidity and may produce emissions of non-$CO_2$ greenhouse gases.

**Solar Radiation Management Methods**

The globally averaged surface temperature of the planet is strongly influenced by the amount of sunlight absorbed by the Earth's atmosphere and surface, which warms the planet, and by the existence of the greenhouse effect, the process by which greenhouse gases and clouds affect the way energy is eventually radiated back to space. An increase in the greenhouse effect leads to a surface temperature rise until a new equilibrium is found. If less incoming sunlight is absorbed because the planet has been made more reflective, or if energy can be emitted to space more effectively because the greenhouse effect is reduced, the average global surface temperature will be reduced.

Suggested geoengineering methods that aim at managing the Earth's incoming and outgoing energy flows are based on this fundamental physical principle. Most of these methods propose to either reduce sunlight reaching the Earth or increase the reflectivity of the planet by making the atmosphere, clouds or the surface brighter (see FAQ 7.3, Figure 1). Another technique proposes to suppress high-level clouds called cirrus, as these clouds have a strong greenhouse effect. Basic physics tells us that if any of these methods change energy flows as expected, then the planet will cool. The picture is complicated, however, because of the many and complex physical processes which govern the interactions between the flow of energy, the atmospheric circulation, weather and the resulting climate.

While the globally averaged surface temperature of the planet will respond to a change in the amount of sunlight reaching the surface or a change in the greenhouse effect, the temperature at any given location and time is influenced by many other factors and the amount of cooling from SRM will not in general equal the amount of warming caused by greenhouse gases. For example, SRM will change heating rates only during daytime, but increasing greenhouse gases can change temperatures during both day and night. This inexact compensation can influence

*(continued on next page)*

*FAQ 7.3 (continued)*

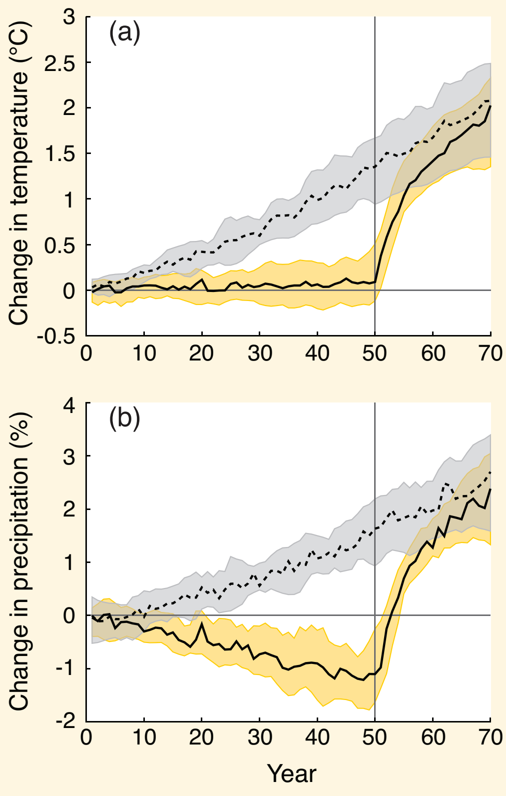
the diurnal cycle of surface temperature, even if the average surface temperature is unchanged. As another example, model calculations suggest that a uniform decrease in sunlight reaching the surface might offset global mean $CO_2$-induced warming, but some regions will cool less than others. Models suggest that if anthropogenic greenhouse warming were completely compensated by stratospheric aerosols, then polar regions would be left with a small residual warming, while tropical regions would become a little cooler than in pre-industrial times.

SRM could theoretically counteract anthropogenic climate change rapidly, cooling the Earth to pre-industrial levels within one or two decades. This is known from climate models but also from the climate records of large volcanic eruptions. The well-observed eruption of Mt Pinatubo in 1991 caused a temporary increase in stratospheric aerosols and a rapid decrease in surface temperature of about 0.5°C.

Climate consists of many factors besides surface temperature. Consequences for other climate features, such as rainfall, soil moisture, river flow, snowpack and sea ice, and ecosystems may also be important. Both models and theory show that compensating an increased greenhouse effect with SRM to stabilize surface temperature would somewhat lower the globally averaged rainfall (see FAQ 7.3, Figure 2 for an idealized model result), and there also could be regional changes. Such imprecise compensation in regional and global climate patterns makes it improbable that SRM will produce a future climate that is 'just like' the one we experience today, or have experienced in the past. However, available climate models indicate that a geoengineered climate with SRM and high atmospheric $CO_2$ levels would be generally closer to 20th century climate than a future climate with elevated $CO_2$ concentrations and no SRM.

SRM techniques would probably have other side effects. For example, theory, observation and models suggest that stratospheric sulphate aerosols from volcanic eruptions and natural emissions deplete stratospheric ozone, especially while chlorine from chlorofluorocarbon emissions resides in the atmosphere. Stratospheric aerosols introduced for SRM are expected to have the same effect. Ozone depletion would increase the amount of ultraviolet light reaching the surface damaging terrestrial and marine ecosystems. Stratospheric aerosols would also increase the ratio of direct to diffuse sunlight reaching the surface, which generally increases plant productivity. There has also been some concern that sulphate aerosol SRM would increase acid rain, but model studies suggest that acid rain is probably not a major concern since the rate of acid rain production from stratospheric aerosol SRM would be much smaller than values currently produced by pollution sources. SRM will also not address the ocean acidification associated with increasing atmospheric $CO_2$ concentrations and its impacts on marine ecosystems.

Without conventional mitigation efforts or potential CDR methods, high $CO_2$ concentrations from anthropogenic emissions will persist in the atmosphere for as long as a thousand years, and SRM would have to be maintained as long as $CO_2$ concentrations were high. Stopping SRM while $CO_2$ concentrations are still high would lead to a very rapid warming over one or two decades (see FAQ7.3, Figure 2), severely stressing ecosystem and human adaptation.




**FAQ 7.3, Figure 2 |** Change in globally averaged (a) surface temperature (°C) and (b) precipitation (%) in two idealized experiments. Solid lines are for simulations using Solar Radiation Management (SRM) to balance a 1% $yr^{-1}$ increase in $CO_2$ concentration until year 50, after which SRM is stopped. Dashed lines are for simulations with a 1% $yr^{-1}$ increase in $CO_2$ concentration and no SRM. The yellow and grey envelopes show the 25th to 75th percentiles from eight different models.

If SRM were used to avoid some consequences of increasing $CO_2$ concentrations, the risks, side effects and shortcomings would clearly increase as the scale of SRM increase. Approaches have been proposed to use a time-limited amount of SRM along with aggressive strategies for reducing $CO_2$ concentrations to help avoid transitions across climate thresholds or tipping points that would be unavoidable otherwise; assessment of such approaches would require a very careful risk benefit analysis that goes much beyond this Report.

Frequently Asked Questions

## FAQ 8.1 | How Important Is Water Vapour to Climate Change?

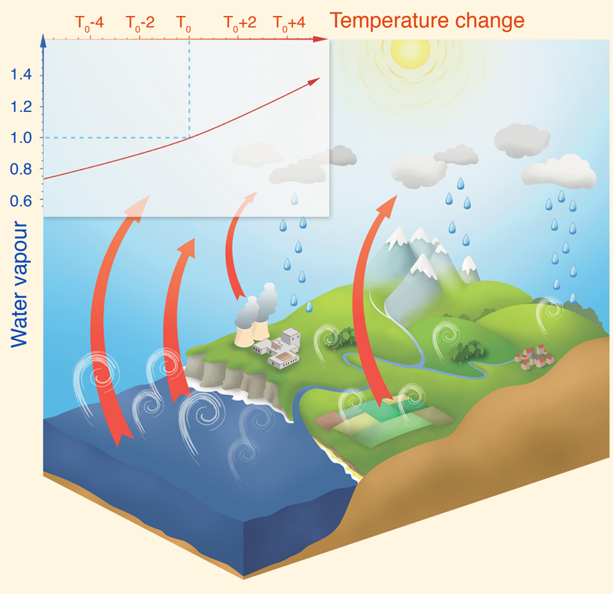
*As the largest contributor to the natural greenhouse effect, water vapour plays an essential role in the Earth's climate. However, the amount of water vapour in the atmosphere is controlled mostly by air temperature, rather than by emissions. For that reason, scientists consider it a feedback agent, rather than a forcing to climate change. Anthropogenic emissions of water vapour through irrigation or power plant cooling have a negligible impact on the global climate.*

Water vapour is the primary greenhouse gas in the Earth's atmosphere. The contribution of water vapour to the natural greenhouse effect relative to that of carbon dioxide ($CO_2$) depends on the accounting method, but can be considered to be approximately two to three times greater. Additional water vapour is injected into the atmosphere from anthropogenic activities, mostly through increased evaporation from irrigated crops, but also through power plant cooling, and marginally through the combustion of fossil fuel. One may therefore question why there is so much focus on $CO_2$, and not on water vapour, as a forcing to climate change.

Water vapour behaves differently from $CO_2$ in one fundamental way: it can condense and precipitate. When air with high humidity cools, some of the vapour condenses into water droplets or ice particles and precipitates. The typical residence time of water vapour in the atmosphere is ten days. The flux of water vapour into the atmosphere from anthropogenic sources is considerably less than from 'natural' evaporation. Therefore, it has a negligible impact on overall concentrations, and does not contribute significantly to the long-term greenhouse effect. This is the main reason why tropospheric water vapour (typically below 10 km altitude) is not considered to be an anthropogenic gas contributing to radiative forcing.

Anthropogenic emissions do have a significant impact on water vapour in the stratosphere, which is the part of the atmosphere above about 10 km. Increased concentrations of methane ($CH_4$) due to human activities lead to an additional source of water, through oxidation, which partly explains the observed changes in that atmospheric layer. That stratospheric water change has a radiative impact, is considered a forcing, and can be evaluated. Stratospheric concentrations of water have varied significantly in past decades. The full extent of these variations is not well understood and is probably less a forcing than a feedback process added to natural variability. The contribution of stratospheric water vapour to warming, both forcing and feedback, is much smaller than from $CH_4$ or $CO_2$.

The maximum amount of water vapour in the air is controlled by temperature. A typical column of air extending from the surface to the stratosphere in polar regions may contain only a few kilograms of water vapour per square metre, while a similar column of air in the tropics may contain up to 70 kg. With every extra degree of air temperature, the atmosphere can retain around 7% more water vapour (see upper-left insert in the FAQ 8.1, Figure 1). This increase in concentration amplifies the greenhouse effect, and therefore leads to more warming. This process, referred to as the water vapour feedback, is well understood and quantified. It occurs in all models used to estimate climate change, where its strength is consistent with observations. Although an increase in atmospheric water vapour has been observed, this change is recognized as a climate feedback (from increased atmospheric temperature) and should not be interpreted as a radiative forcing from anthropogenic emissions. *(continued on next page)*




**FAQ 8.1, Figure 1 |** Illustration of the water cycle and its interaction with the greenhouse effect. The upper-left insert indicates the relative increase of potential water vapour content in the air with an increase of temperature (roughly 7% per degree). The white curls illustrate evaporation, which is compensated by precipitation to close the water budget. The red arrows illustrate the outgoing infrared radiation that is partly absorbed by water vapour and other gases, a process that is one component of the greenhouse effect. The stratospheric processes are not included in this figure.

*FAQ 8.1 (continued)*

Currently, water vapour has the largest greenhouse effect in the Earth's atmosphere. However, other greenhouse gases, primarily $CO_2$, are necessary to sustain the presence of water vapour in the atmosphere. Indeed, if these other gases were removed from the atmosphere, its temperature would drop sufficiently to induce a decrease of water vapour, leading to a runaway drop of the greenhouse effect that would plunge the Earth into a frozen state. So greenhouse gases other than water vapour provide the temperature structure that sustains current levels of atmospheric water vapour. Therefore, although $CO_2$ is the main anthropogenic control knob on climate, water vapour is a strong and fast feedback that amplifies any initial forcing by a typical factor between two and three. Water vapour is not a significant initial forcing, but is nevertheless a fundamental agent of climate change.

Frequently Asked Questions

## FAQ 8.2 | Do Improvements in Air Quality Have an Effect on Climate Change?

*Yes they do, but depending on which pollutant(s) they limit, they can either cool or warm the climate. For example, whereas a reduction in sulphur dioxide ($SO_2$) emissions leads to more warming, nitrogen oxide ($NO_x$) emission control has both a cooling (through reducing of tropospheric ozone) and a warming effect (due to its impact on methane lifetime and aerosol production). Air pollution can also affect precipitation patterns.*

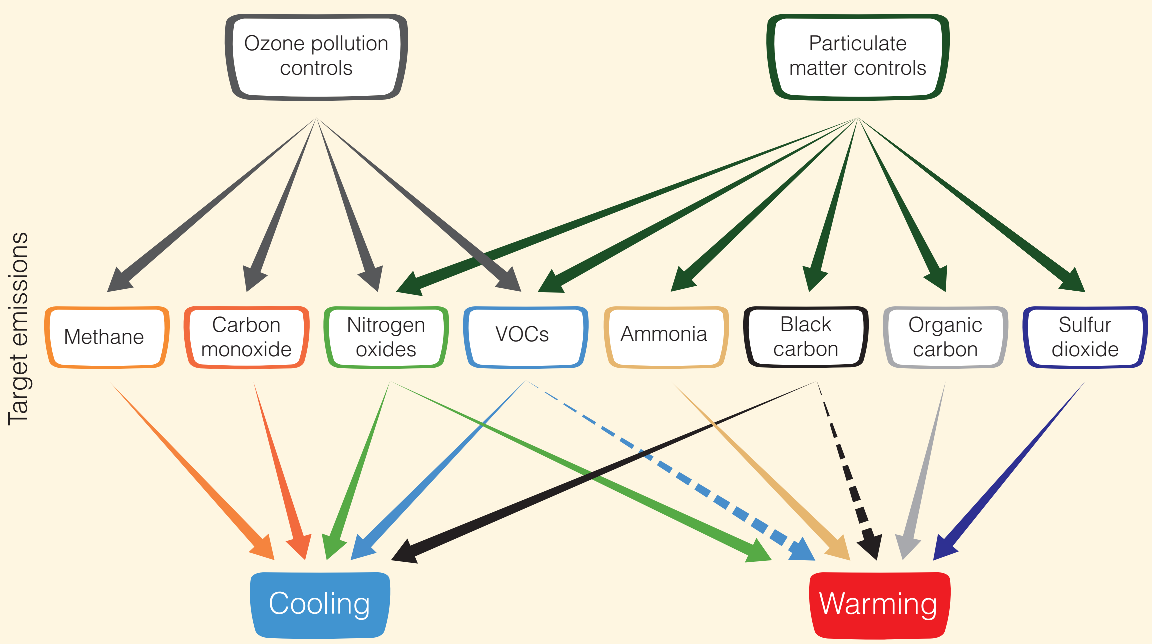
Air quality is nominally a measure of airborne surface pollutants, such as ozone, carbon monoxide, $NO_x$ and aerosols (solid or liquid particulate matter). Exposure to such pollutants exacerbates respiratory and cardiovascular diseases, harms plants and damages buildings. For these reasons, most major urban centres try to control discharges of airborne pollutants.

Unlike carbon dioxide ($CO_2$) and other well-mixed greenhouse gases, tropospheric ozone and aerosols may last in the atmosphere only for a few days to a few weeks, though indirect couplings within the Earth system can prolong their impact. These pollutants are usually most potent near their area of emission or formation, where they can force local or regional perturbations to climate, even if their globally averaged effect is small.

Air pollutants affect climate differently according to their physical and chemical characteristics. Pollution-generated greenhouse gases will impact climate primarily through shortwave and longwave radiation, while aerosols can in addition affect climate through cloud–aerosol interactions.

Controls on anthropogenic emissions of methane (FAQ 8.2, Figure 1) to lower surface ozone have been identified as 'win–win' situations. Consequences of controlling other ozone precursors are not always as clear. $NO_x$ emission controls, for instance, might be expected to have a cooling effect as they reduce tropospheric ozone, but their impact on $CH_4$ lifetime and aerosol formation is more likely instead to cause overall warming.

Satellite observations have identified increasing atmospheric concentrations of $SO_2$ (the primary precursor to scattering sulphate aerosols) from coal-burning power plants over eastern Asia during the last few decades. The most recent power plants use scrubbers to reduce such emissions (albeit not the concurrent $CO_2$ emissions and associated long-term climate warming). This improves air quality, but also reduces the cooling effect of sulphate aerosols and therefore exacerbates warming. Aerosol cooling occurs through aerosol–radiation and aerosol–cloud interactions and is estimated at –0.9 W $m^{-2}$ (all aerosols combined, Section 8.3.4.3) since pre-industrial, having grown especially during the second half of the 20th century when anthropogenic emissions rose sharply. *(continued on next page)*




**FAQ 8.2, Figure 1 |** Schematic diagram of the impact of pollution controls on specific emissions and climate impact. Solid black line indicates known impact; dashed line indicates uncertain impact.

*FAQ 8.2 (continued)*

Black carbon or soot, on the other hand, absorbs heat in the atmosphere (leading to a 0.4 W $m^{-2}$ radiative forcing from anthropogenic fossil and biofuel emissions) and, when deposited on snow, reduces its albedo, or ability to reflect sunlight. Reductions of black carbon emissions can therefore have a cooling effect, but the additional interaction of black carbon with clouds is uncertain and could lead to some counteracting warming.

Air quality controls might also target a specific anthropogenic activity sector, such as transportation or energy production. In that case, co-emitted species within the targeted sector lead to a complex mix of chemistry and climate perturbations. For example, smoke from biofuel combustion contains a mixture of both absorbing and scattering particles as well as ozone precursors, for which the combined climate impact can be difficult to ascertain.

Thus, surface air quality controls will have some consequences on climate. Some couplings between the targeted emissions and climate are still poorly understood or identified, including the effects of air pollutants on precipitation patterns, making it difficult to fully quantify these consequences. There is an important twist, too, in the potential effect of climate change on air quality. In particular, an observed correlation between surface ozone and temperature in polluted regions indicates that higher temperatures from climate change alone could worsen summertime pollution, suggesting a 'climate penalty'. This penalty implies stricter surface ozone controls will be required to achieve a specific target. In addition, projected changes in the frequency and duration of stagnation events could impact air quality conditions. These features will be regionally variable and difficult to assess, but better understanding, quantification and modelling of these processes will clarify the overall interaction between air pollutants and climate.

Frequently Asked Questions

## FAQ 9.1 | Are Climate Models Getting Better, and How Would We Know?

*Climate models are extremely sophisticated computer programs that encapsulate our understanding of the climate system and simulate, with as much fidelity as currently feasible, the complex interactions between the atmosphere, ocean, land surface, snow and ice, the global ecosystem and a variety of chemical and biological processes.*

*The complexity of climate models—the representation of physical processes like clouds, land surface interactions and the representation of the global carbon and sulphur cycles in many models—has increased substantially since the IPCC First Assessment Report in 1990, so in that sense, current Earth System Models are vastly 'better' than the models of that era. This development has continued since the Fourth Assessment, while other factors have also contributed to model improvement. More powerful supercomputers allow current models to resolve finer spatial detail. Today's models also reflect improved understanding of how climate processes work—understanding that has come from ongoing research and analysis, along with new and improved observations.*

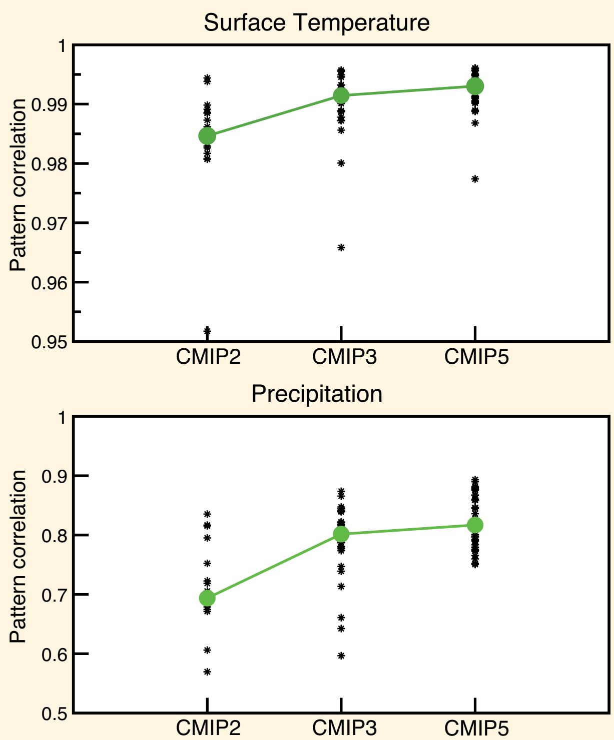
Climate models of today are, in principle, better than their predecessors. However, every bit of added complexity, while intended to improve some aspect of simulated climate, also introduces new sources of possible error (e.g., via uncertain parameters) and new interactions between model components that may, if only temporarily, degrade a model's simulation of other aspects of the climate system. Furthermore, despite the progress that has been made, scientific uncertainty regarding the details of many processes remains.

An important consideration is that model performance can be evaluated only relative to past observations, taking into account natural internal variability. To have confidence in the future projections of such models, historical climate—and its variability and change—must be well simulated. The scope of model evaluation, in terms of the kind and quantity of observations available, the availability of better coordinated model experiments, and the expanded use of various performance metrics, has provided much more quantitative information about model performance. But this alone may not be sufficient. Whereas weather and seasonal climate predictions can be regularly verified, climate projections spanning a century or more cannot. This is particularly the case as anthropogenic forcing is driving the climate system toward conditions not previously observed in the instrumental record, and it will always be a limitation.

Quantifying model performance is a topic that has featured in all previous IPCC Working Group I Reports. Reading back over these earlier assessments provides a general sense of the improvements that have been made. Past reports have typically provided a rather broad survey of model performance, showing differences between model-calculated versions of various climate quantities and corresponding observational estimates.

Inevitably, some models perform better than others for certain climate variables, but no individual model clearly emerges as 'the best' overall. Recently, there has been progress in computing various performance metrics, which synthesize model performance relative to a range of different observations according to a simple numerical score. Of course, the definition of such a score, how it is computed, the observations used (which have their

*(continued on next page)*




**FAQ 9.1, Figure 1 |** Model capability in simulating annual mean temperature and precipitation patterns as illustrated by results of three recent phases of the Coupled Model Intercomparison Project (CMIP2, models from about year 2000; CMIP3, models from about 2005; and CMIP5, the current generation of models). The figure shows the correlation (a measure of pattern similarity) between observed and modelled temperature (upper panel) and precipitation (lower panel). Larger values indicate better correspondence between modelled and observed spatial patterns. The black symbols indicate correlation coefficient for individual models, and the large green symbols indicate the median value (i.e., half of the model results lie above and the other half below this value). Improvement in model performance is evident by the increase in correlation for successive model generations.

*FAQ 9.1 (continued)*

own uncertainties), and the manner in which various scores are combined are all important, and will affect the end result.

Nevertheless, if the metric is computed consistently, one can compare different generations of models. Results of such comparisons generally show that, although each generation exhibits a range in performance, the average model performance index has improved steadily between each generation. An example of changes in model performance over time is shown in FAQ 9.1, Figure 1, and illustrates the ongoing, albeit modest, improvement. It is interesting to note that both the poorest and best performing models demonstrate improvement, and that this improvement comes in parallel with increasing model complexity and an elimination of artificial adjustments to atmosphere and ocean coupling (so-called 'flux adjustment'). Some of the reasons for this improvement include increased understanding of various climate processes and better representation of these processes in climate models. More comprehensive Earth observations are also driving improvements.

So, yes, climate models are getting better, and we can demonstrate this with quantitative performance metrics based on historical observations. Although future climate projections cannot be directly evaluated, climate models are based, to a large extent, on verifiable physical principles and are able to reproduce many important aspects of past response to external forcing. In this way, they provide a scientifically sound preview of the climate response to different scenarios of anthropogenic forcing.

Frequently Asked Questions

## FAQ 10.1 | Climate Is Always Changing. How Do We Determine the Causes of Observed Changes?

*The causes of observed long-term changes in climate (on time scales longer than a decade) are assessed by determining whether the expected 'fingerprints' of different causes of climate change are present in the historical record. These fingerprints are derived from computer model simulations of the different patterns of climate change caused by individual climate forcings. On multi-decade time scales, these forcings include processes such as greenhouse gas increases or changes in solar brightness. By comparing the simulated fingerprint patterns with observed climate changes, we can determine whether observed changes are best explained by those fingerprint patterns, or by natural variability, which occurs without any forcing.*

*The fingerprint of human-caused greenhouse gas increases is clearly apparent in the pattern of observed 20th century climate change. The observed change cannot be otherwise explained by the fingerprints of natural forcings or natural variability simulated by climate models. Attribution studies therefore support the conclusion that 'it is* extremely likely *that human activities have caused more than half of the observed increase in global mean surface temperatures from 1951 to 2010.'*

The Earth's climate is always changing, and that can occur for many reasons. To determine the principal causes of observed changes, we must first ascertain whether an observed change in climate is different from other fluctuations that occur without any forcing at all. Climate variability without forcing—called internal variability—is the consequence of processes within the climate system. Large-scale oceanic variability, such as El Niño-Southern Oscillation (ENSO) fluctuations in the Pacific Ocean, is the dominant source of internal climate variability on decadal to centennial time scales.

Climate change can also result from natural forcings external to the climate system, such as volcanic eruptions, or changes in the brightness of the sun. Forcings such as these are responsible for the huge changes in climate that are clearly documented in the geological record. Human-caused forcings include greenhouse gas emissions or atmospheric particulate pollution. Any of these forcings, natural or human caused, could affect internal variability as well as causing a change in average climate. Attribution studies attempt to determine the causes of a detected change in observed climate. Over the past century we know that global average temperature has increased, so if the observed change is forced then the principal forcing must be one that causes warming, not cooling.

Formal climate change attribution studies are carried out using controlled experiments with climate models. The model-simulated responses to specific climate forcings are often called the fingerprints of those forcings. A climate model must reliably simulate the fingerprint patterns associated with individual forcings, as well as the patterns of unforced internal variability, in order to yield a meaningful climate change attribution assessment. No model can perfectly reproduce all features of climate, but many detailed studies indicate that simulations using current models are indeed sufficiently reliable to carry out attribution assessments.

FAQ 10.1, Figure 1 illustrates part of a fingerprint assessment of global temperature change at the surface during the late 20th century. The observed change in the latter half of the 20th century, shown by the black time series in the left panels, is larger than expected from just internal variability. Simulations driven only by natural forcings (yellow and blue lines in the upper left panel) fail to reproduce late 20th century global warming at the surface with a spatial pattern of change (upper right) completely different from the observed pattern of change (middle right). Simulations including both natural and human-caused forcings provide a much better representation of the time rate of change (lower left) and spatial pattern (lower right) of observed surface temperature change.

Both panels on the left show that computer models reproduce the naturally forced surface cooling observed for a year or two after major volcanic eruptions, such as occurred in 1982 and 1991. Natural forcing simulations capture the short-lived temperature changes following eruptions, but only the natural + human caused forcing simulations simulate the longer-lived warming trend.

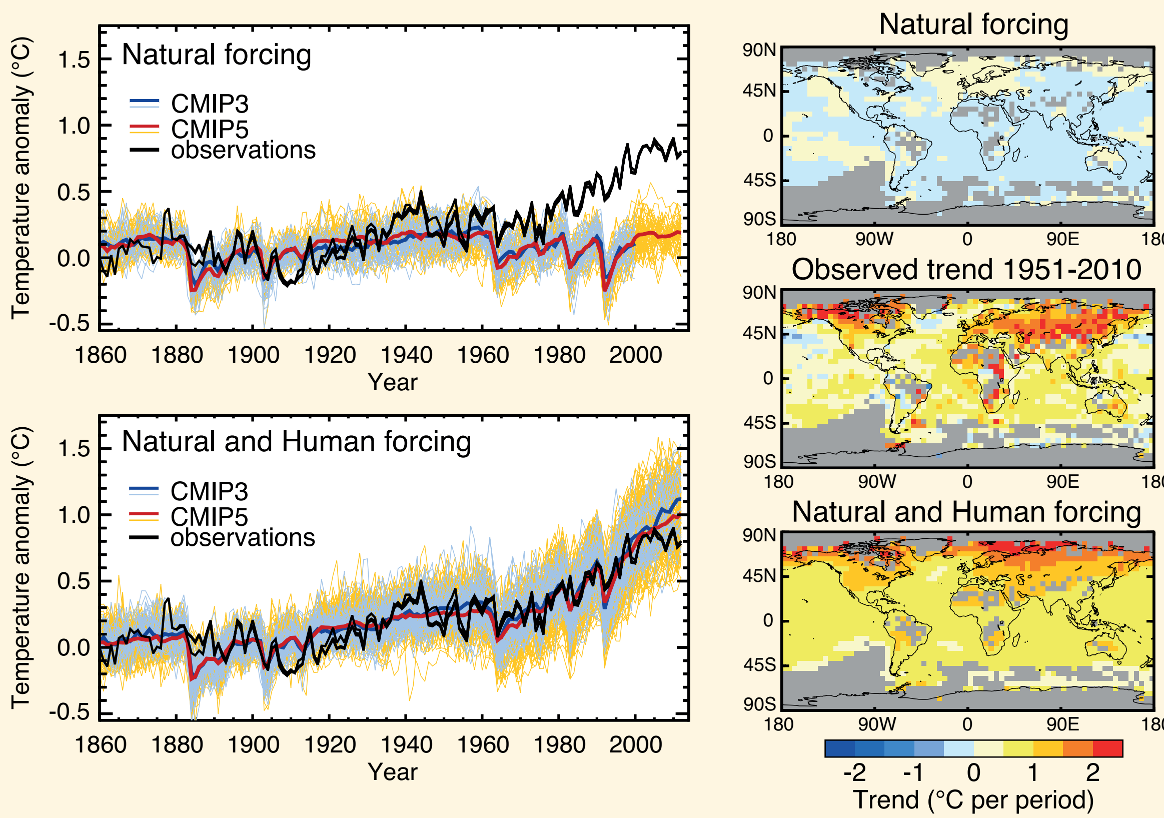
A more complete attribution assessment would examine temperature above the surface, and possibly other climate variables, in addition to the surface temperature results shown in FAQ 10.1, Figure 1. The fingerprint patterns associated with individual forcings become easier to distinguish when more variables are considered in the assessment.

*(continued on next page)*

FAQ

*FAQ 10.1 (continued)*

Overall, FAQ 10.1, Figure 1 shows that the pattern of observed temperature change is significantly different than the pattern of response to natural forcings alone. The simulated response to all forcings, including human-caused forcings, provides a good match to the observed changes at the surface. We cannot correctly simulate recent observed climate change without including the response to human-caused forcings, including greenhouse gases, stratospheric ozone, and aerosols. Natural causes of change are still at work in the climate system, but recent trends in temperature are largely attributable to human-caused forcing.




**FAQ 10.1, Figure 1 |** (Left) Time series of global and annual-averaged surface temperature change from 1860 to 2010. The top left panel shows results from two ensemble of climate models driven with just natural forcings, shown as thin blue and yellow lines; ensemble average temperature changes are thick blue and red lines. Three different observed estimates are shown as black lines. The lower left panel shows simulations by the same models, but driven with both natural forcing and human-induced changes in greenhouse gases and aerosols. (Right) Spatial patterns of local surface temperature trends from 1951 to 2010. The upper panel shows the pattern of trends from a large ensemble of Coupled Model Intercomparison Project Phase 5 (CMIP5) simulations driven with just natural forcings. The bottom panel shows trends from a corresponding ensemble of simulations driven with natural + human forcings. The middle panel shows the pattern of observed trends from the Hadley Centre/Climatic Research Unit gridded surface temperature data set 4 (HadCRUT4) during this period.

FAQ

Frequently Asked Questions

## FAQ 10.2 | When Will Human Influences on Climate Become Obvious on Local Scales?

*Human-caused warming is already becoming locally obvious on land in some tropical regions, especially during the warm part of the year. Warming should become obvious in middle latitudes—during summer at first—within the next several decades. The trend is expected to emerge more slowly there, especially during winter, because natural climate variability increases with distance from the equator and during the cold season. Temperature trends already detected in many regions have been attributed to human influence. Temperature-sensitive climate variables, such as Arctic sea ice, also show detected trends attributable to human influence.*

Warming trends associated with global change are generally more evident in averages of global temperature than in time series of local temperature ('local' here refers generally to individual locations, or small regional averages). This is because most of the local variability of local climate is averaged away in the global mean. Multi-decadal warming trends detected in many regions are considered to be outside the range of trends one might expect from natural internal variability of the climate system, but such trends will only become obvious when the local mean climate emerges from the 'noise' of year-to-year variability. How quickly this happens depends on both the rate of the warming trend and the amount of local variability. Future warming trends cannot be predicted precisely, especially at local scales, so estimates of the future time of emergence of a warming trend cannot be made with precision.

In some tropical regions, the warming trend has already emerged from local variability (FAQ 10.2, Figure 1). This happens more quickly in the tropics because there is less temperature variability there than in other parts of the globe. Projected warming may not emerge in middle latitudes until the mid-21st century—even though warming trends there are larger—because local temperature variability is substantially greater there than in the tropics. On a seasonal basis, local temperature variability tends to be smaller in summer than in winter. Warming therefore tends to emerge first in the warm part of the year, even in regions where the warming trend is larger in winter, such as in central Eurasia in FAQ 10.2, Figure 1.

Variables other than land surface temperature, including some oceanic regions, also show rates of long-term change different from natural variability. For example, Arctic sea ice extent is declining very rapidly, and already shows a human influence. On the other hand, local precipitation trends are very hard to detect because at most locations the variability in precipitation is quite large. The probability of record-setting warm summer temperatures has increased throughout much of the Northern Hemisphere . High temperatures presently considered extreme are projected to become closer to the norm over the coming decades. The probabilities of other extreme events, including some cold spells, have lessened.

In the present climate, individual extreme weather events cannot be unambiguously ascribed to climate change, since such events could have happened in an unchanged climate. However the probability of occurrence of such events could have changed significantly at a particular location. Human-induced increases in greenhouse gases are estimated to have contributed substantially to the probability of some heatwaves. Similarly, climate model studies suggest that increased greenhouse gases have contributed to the observed intensification of heavy precipitation events found over parts of the Northern Hemisphere. However, the probability of many other extreme weather events may not have changed substantially. Therefore, it is incorrect to ascribe every new weather record to climate change.

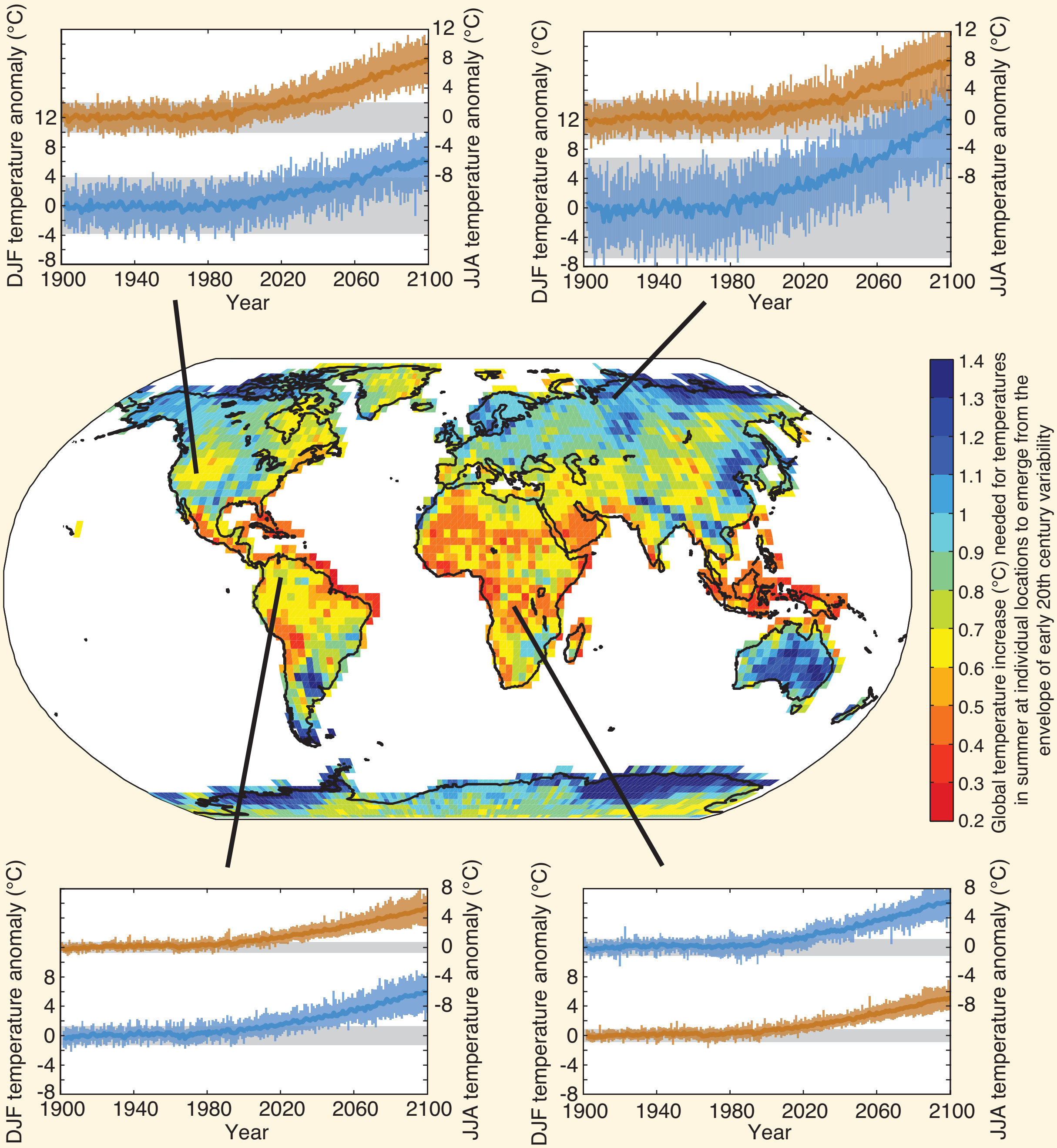
The date of future emergence of projected warming trends also depends on local climate variability, which can temporarily increase or decrease temperatures. Furthermore, the projected local temperature curves shown in FAQ 10.2, Figure 1 are based on multiple climate model simulations forced by the same assumed future emissions scenario. A different rate of atmospheric greenhouse gas accumulation would cause a different warming trend, so the spread of model warming projections (the coloured shading in FAQ 10.2, Figure 1) would be wider if the figure included a spread of greenhouse gas emissions scenarios. The increase required for summer temperature change to emerge from 20th century local variability (regardless of the rate of change) is depicted on the central map in FAQ 10.2, Figure 1.

A full answer to the question of when human influence on local climate will become obvious depends on the strength of evidence one considers sufficient to render something 'obvious'. The most convincing scientific evidence for the effect of climate change on local scales comes from analysing the global picture, and from the wealth of evidence from across the climate system linking many observed changes to human influence. *(continued on next page)*

FAQ

FAQ

*FAQ 10.2 (continued)*




**FAQ 10.2, Figure 1 |** Time series of projected temperature change shown at four representative locations for summer (red curves, representing June, July and August at sites in the tropics and Northern Hemisphere or December, January and February in the Southern Hemisphere) and winter (blue curves). Each time series is surrounded by an envelope of projected changes (pink for the local warm season, blue for the local cold season) yielded by 24 different model simulations, emerging from a grey envelope of natural local variability simulated by the models using early 20th century conditions. The warming signal emerges first in the tropics during summer. The central map shows the global temperature increase (°C) needed for temperatures in summer at individual locations to emerge from the envelope of early 20th century variability. Note that warm colours denote the smallest needed temperature increase, hence earliest time of emergence. All calculations are based on Coupled Model Intercomparison Project Phase 5 (CMIP5) global climate model simulations forced by the Representative Concentration Pathway 8.5 (RCP8.5) emissions scenario. Envelopes of projected change and natural variability are defined as ±2 standard deviations. (Adapted and updated from Mahlstein et al., 2011.)

Frequently Asked Questions

## FAQ 11.1 | If You Cannot Predict the Weather Next Month, How Can You Predict Climate for the Coming Decade?

*Although weather and climate are intertwined, they are in fact different things. Weather is defined as the state of the atmosphere at a given time and place, and can change from hour to hour and day to day. Climate, on the other hand, generally refers to the statistics of weather conditions over a decade or more.*

*An ability to predict future climate without the need to accurately predict weather is more commonplace that it might first seem. For example, at the end of spring, it can be accurately predicted that the average air temperature over the coming summer in Melbourne (for example) will very likely be higher than the average temperature during the most recent spring—even though the day-to-day weather during the coming summer cannot be predicted with accuracy beyond a week or so. This simple example illustrates that factors exist—in this case the seasonal cycle in solar radiation reaching the Southern Hemisphere—that can underpin skill in predicting changes in climate over a coming period that does not depend on accuracy in predicting weather over the same period.*

The statistics of weather conditions used to define climate include long-term averages of air temperature and rainfall, as well as statistics of their variability, such as the standard deviation of year-to-year rainfall variability from the long-term average, or the frequency of days below 5°C. Averages of climate variables over long periods of time are called climatological averages. They can apply to individual months, seasons or the year as a whole. A climate prediction will address questions like: 'How likely will it be that the average temperature during the coming summer will be higher than the long-term average of past summers?' or: 'How likely will it be that the next decade will be warmer than past decades?' More specifically, a climate prediction might provide an answer to the question: 'What is the probability that temperature (in China, for instance) averaged over the next ten years will exceed the temperature in China averaged over the past 30 years?' Climate predictions do not provide forecasts of the detailed day-to-day evolution of future weather. Instead, they provide probabilities of long-term changes to the statistics of future climatic variables.

Weather forecasts, on the other hand, provide predictions of day-to-day weather for specific times in the future. They help to address questions like: 'Will it rain tomorrow?' Sometimes, weather forecasts are given in terms of probabilities. For example, the weather forecast might state that: 'the likelihood of rainfall in Apia tomorrow is 75%'.

To make accurate weather predictions, forecasters need highly detailed information about the current state of the atmosphere. The chaotic nature of the atmosphere means that even the tiniest error in the depiction of 'initial conditions' typically leads to inaccurate forecasts beyond a week or so. This is the so-called 'butterfly effect'.

Climate scientists do not attempt or claim to predict the detailed future evolution of the weather over coming seasons, years or decades. There is, on the other hand, a sound scientific basis for supposing that aspects of climate can be predicted, albeit imprecisely, despite the butterfly effect. For example, increases in long-lived atmospheric greenhouse gas concentrations tend to increase surface temperature in future decades. Thus, information from the past can and does help predict future climate.

Some types of naturally occurring so-called 'internal' variability can—in theory at least—extend the capacity to predict future climate. Internal climatic variability arises from natural instabilities in the climate system. If such variability includes or causes extensive, long-lived, upper ocean temperature anomalies, this will drive changes in the overlying atmosphere, both locally and remotely. The El Niño-Southern Oscillation phenomenon is probably the most famous example of this kind of internal variability. Variability linked to the El Niño-Southern Oscillation unfolds in a partially predictable fashion. The butterfly effect is present, but it takes longer to strongly influence some of the variability linked to the El Nino-Southern Oscillation.

Meteorological services and other agencies have exploited this. They have developed seasonal-to-interannual prediction systems that enable them to routinely predict seasonal climate anomalies with demonstrable predictive skill. The skill varies markedly from place to place and variable to variable. Skill tends to diminish the further the prediction delves into the future and in some locations there is no skill at all. 'Skill' is used here in its technical sense: it is a measure of how much greater the accuracy of a prediction is, compared with the accuracy of some typically simple prediction method like assuming that recent anomalies will persist during the period being predicted.

Weather, seasonal-to-interannual and decadal prediction systems are similar in many ways (e.g., they all incorporate the same mathematical equations for the atmosphere, they all need to specify initial conditions to kick-start

*(continued on next page)*

FAQ

*FAQ 11.1 (continued)*

predictions, and they are all subject to limits on forecast accuracy imposed by the butterfly effect). However, decadal prediction, unlike weather and seasonal-to-interannual prediction, is still in its infancy. Decadal prediction systems nevertheless exhibit a degree of skill in *hindcasting* near-surface temperature over much of the globe out to at least nine years. A 'hindcast' is a prediction of a past event in which only observations prior to the event are fed into the prediction system used to make the prediction. The bulk of this skill is thought to arise from *external forcing*. 'External forcing' is a term used by climate scientists to refer to a forcing agent outside the climate system causing a change in the climate system. This includes increases in the concentration of long-lived greenhouse gases.

Theory indicates that skill in predicting decadal precipitation should be less than the skill in predicting decadal surface temperature, and hindcast performance is consistent with this expectation.

Current research is aimed at improving decadal prediction systems, and increasing the understanding of the reasons for any apparent skill. Ascertaining the degree to which the extra information from internal variability actually translates to increased skill is a key issue. While prediction systems are expected to improve over coming decades, the chaotic nature of the climate system and the resulting butterfly effect will always impose unavoidable limits on predictive skill. Other sources of uncertainty exist. For example, as volcanic eruptions can influence climate but their timing and magnitude cannot be predicted, future eruptions provide one of a number of other sources of uncertainty. Additionally, the shortness of the period with enough oceanic data to initialize and assess decadal predictions presents a major challenge.

Finally, note that decadal prediction systems are designed to exploit both externally forced and internally generated sources of predictability. Climate scientists distinguish between decadal predictions and decadal projections. Projections exploit only the predictive capacity arising from external forcing. While previous IPCC Assessment Reports focussed exclusively on projections, this report also assesses decadal prediction research and its scientific basis.

Frequently Asked Questions

## FAQ 11.2 | How Do Volcanic Eruptions Affect Climate and Our Ability to Predict Climate?

*Large volcanic eruptions affect the climate by injecting sulphur dioxide gas into the upper atmosphere (also called stratosphere), which reacts with water to form clouds of sulphuric acid droplets. These clouds reflect sunlight back to space, preventing its energy from reaching the Earth's surface, thus cooling it, along with the lower atmosphere. These upper atmospheric sulphuric acid clouds also locally absorb energy from the Sun, the Earth and the lower atmosphere, which heats the upper atmosphere (see FAQ 11.2, Figure 1). In terms of surface cooling, the 1991 Mt Pinatubo eruption in the Philippines, for example, injected about 20 million tons of sulphur dioxide ($SO_2$) into the stratosphere, cooling the Earth by about 0.5°C for up to a year. Globally, eruptions also reduce precipitation, because the reduced incoming shortwave at the surface is compensated by a reduction in latent heating (i.e., in evaporation and hence rainfall).*

*For the purposes of predicting climate, an eruption causing significant global surface cooling and upper atmospheric heating for the next year or so can be expected. The problem is that, while a volcano that has become more active can be detected, the precise timing of an eruption, or the amount of $SO_2$ injected into the upper atmosphere and how it might disperse cannot be predicted. This is a source of uncertainty in climate predictions.*

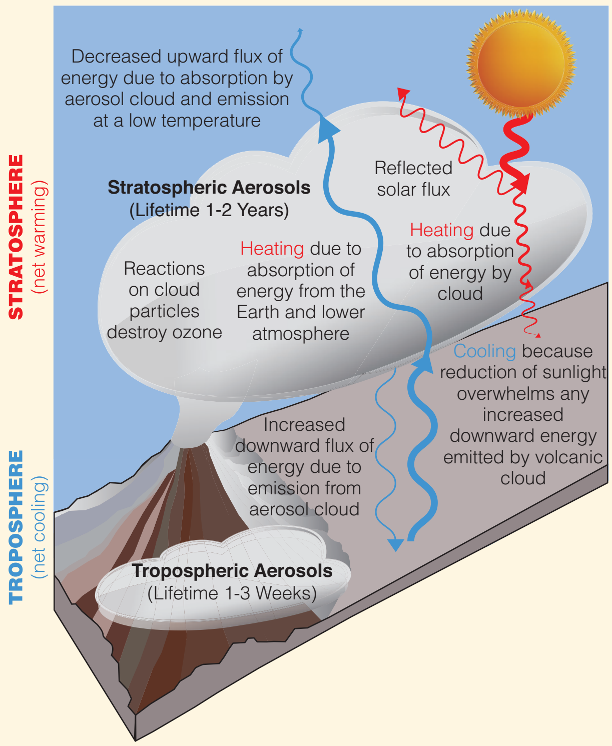
Large volcanic eruptions produce lots of particles, called ash or tephra. However, these particles fall out of the atmosphere quickly, within days or weeks, so they do not affect the global climate. For example, the 1980 Mount St. Helens eruption affected surface temperatures in the northwest USA for several days but, because it emitted little $SO_2$ into the stratosphere, it had no detectable global climate impacts. If large, high-latitude eruptions inject sulphur into the stratosphere, they will have an effect only in the hemisphere where they erupted, and the effects will only last a year at most, as the stratospheric cloud they produce only has a lifetime of a few months.

Tropical or subtropical volcanoes produce more global surface or tropospheric cooling. This is because the resulting sulphuric acid cloud in the upper atmosphere lasts between one and two years, and can cover much of the globe. However, their regional climatic impacts are difficult to predict, because dispersion of stratospheric sulphate aerosols depends heavily on atmospheric wind conditions at the time of eruption. Furthermore, the surface cooling effect is typically not uniform: because continents cool more than the ocean, the summer monsoon can weaken, reducing rain over Asia and Africa. The climatic response is complicated further by the fact that upper atmospheric clouds from tropical eruptions also absorb sunlight and heat from the Earth, which produces more upper atmosphere warming in the tropics than at high latitudes.

The largest volcanic eruptions of the past 250 years stimulated scientific study. After the 1783 Laki eruption in Iceland, there were record warm summer temperatures in Europe, followed by a very cold winter. Two large eruptions, an unidentified one in 1809, and the 1815 Tambora eruption caused the 'Year Without a Summer' in 1816. Agricultural failures in Europe and the USA that year led to food shortages, famine and riots.

The largest eruption in more than 50 years, that of Agung in 1963, led to many modern studies, including observations and climate model calculations. Two subsequent large eruptions, El Chichón in 1982 and Pinatubo in 1991, inspired the work that led to our current understanding of the effects of volcanic eruptions on climate.

*(continued on next page)*




**FAQ 11.2, Figure 1 |** Schematic of how large tropical or sub-tropical volcanoes impact upper atmospheric (stratospheric) and lower atmospheric (tropospheric) temperatures.

FAQ

*FAQ 11.2 (continued)*

Volcanic clouds remain in the stratosphere only for a couple of years, so their impact on climate is correspondingly short. But the impacts of consecutive large eruptions can last longer: for example, at the end of the 13th century there were four large eruptions—one every ten years. The first, in 1258 CE, was the largest in 1000 years. That sequence of eruptions cooled the North Atlantic Ocean and Arctic sea ice. Another period of interest is the three large, and several lesser, volcanic events during 1963–1991 (see Chapter 8 for how these eruptions affected atmospheric composition and reduced shortwave radiation at the ground.

Volcanologists can detect when a volcano becomes more active, but they cannot predict whether it will erupt, or if it does, how much sulphur it might inject into the stratosphere. Nevertheless, volcanoes affect the ability to predict climate in three distinct ways. First, if a violent eruption injects significant volumes of sulphur dioxide into the stratosphere, this effect can be included in climate predictions. There are substantial challenges and sources of uncertainty involved, such as collecting good observations of the volcanic cloud, and calculating how it will move and change during its lifetime. But, based on observations, and successful modelling of recent eruptions, some of the effects of large eruptions can be included in predictions.

The second effect is that volcanic eruptions are a potential source of uncertainty in our predictions. Eruptions cannot be predicted in advance, but they will occur, causing short-term climatic impacts on both local and global scales. In principle, this potential uncertainty can be accounted for by including random eruptions, or eruptions based on some scenario in our near-term ensemble climate predictions. This area of research needs further exploration. The future projections in this report do not include future volcanic eruptions.



Third, the historical climate record can be used, along with estimates of observed sulphate aerosols, to test the fidelity of our climate simulations. While the climatic response to explosive volcanic eruptions is a useful analogue for some other climatic forcings, there are limitations. For example, successfully simulating the impact of one eruption can help validate models used for seasonal and interannual predictions. But in this way not all the mechanisms involved in global warming over the next century can be validated, because these involve long term oceanic feedbacks, which have a longer time scale than the response to individual volcanic eruptions.

Frequently Asked Questions

## FAQ 12.1 | Why Are So Many Models and Scenarios Used to Project Climate Change?

*Future climate is partly determined by the magnitude of future emissions of greenhouse gases, aerosols and other natural and man-made forcings. These forcings are external to the climate system, but modify how it behaves. Future climate is shaped by the Earth's response to those forcings, along with internal variability inherent in the climate system. A range of assumptions about the magnitude and pace of future emissions helps scientists develop different emission scenarios, upon which climate model projections are based. Different climate models, meanwhile, provide alternative representations of the Earth's response to those forcings, and of natural climate variability. Together, ensembles of models, simulating the response to a range of different scenarios, map out a range of possible futures, and help us understand their uncertainties.*

Predicting socioeconomic development is arguably even more difficult than predicting the evolution of a physical system. It entails predicting human behaviour, policy choices, technological advances, international competition and cooperation. The common approach is to use scenarios of plausible future socioeconomic development, from which future emissions of greenhouse gases and other forcing agents are derived. It has not, in general, been possible to assign likelihoods to individual forcing scenarios. Rather, a set of alternatives is used to span a range of possibilities. The outcomes from different forcing scenarios provide policymakers with alternatives and a range of possible futures to consider.

Internal fluctuations in climate are spontaneously generated by interactions between components such as the atmosphere and the ocean. In the case of near-term climate change, they may eclipse the effect of external perturbations, like greenhouse gas increases (see Chapter 11). Over the longer term, however, the effect of external forcings is expected to dominate instead. Climate model simulations project that, after a few decades, different scenarios of future anthropogenic greenhouse gases and other forcing agents—and the climate system's response to them—will differently affect the change in mean global temperature (FAQ 12.1, Figure 1, left panel). Therefore, evaluating the consequences of those various scenarios and responses is of paramount importance, especially when policy decisions are considered.

Climate models are built on the basis of the physical principles governing our climate system, and empirical understanding, and represent the complex, interacting processes needed to simulate climate and climate change, both past and future. Analogues from past observations, or extrapolations from recent trends, are inadequate strategies for producing projections, because the future will not necessarily be a simple continuation of what we have seen thus far.

Although it is possible to write down the equations of fluid motion that determine the behaviour of the atmosphere and ocean, it is impossible to solve them without using numerical algorithms through computer model simulation, similarly to how aircraft engineering relies on numerical simulations of similar types of equations. Also, many small-scale physical, biological and chemical processes, such as cloud processes, cannot be described by those equations, either because we lack the computational ability to describe the system at a fine enough resolution to directly simulate these processes or because we still have a partial scientific understanding of the mechanisms driving these processes. Those need instead to be approximated by so-called parameterizations within the climate models, through which a mathematical relation between directly simulated and approximated quantities is established, often on the basis of observed behaviour.

There are various alternative and equally plausible numerical representations, solutions and approximations for modelling the climate system, given the limitations in computing and observations. This diversity is considered a healthy aspect of the climate modelling community, and results in a range of plausible climate change projections at global and regional scales. This range provides a basis for quantifying uncertainty in the projections, but because the number of models is relatively small, and the contribution of model output to public archives is voluntary, the sampling of possible futures is neither systematic nor comprehensive. Also, some inadequacies persist that are common to all models; different models have different strength and weaknesses; it is not yet clear which aspects of the quality of the simulations that can be evaluated through observations should guide our evaluation of future model simulations. *(continued on next page)*

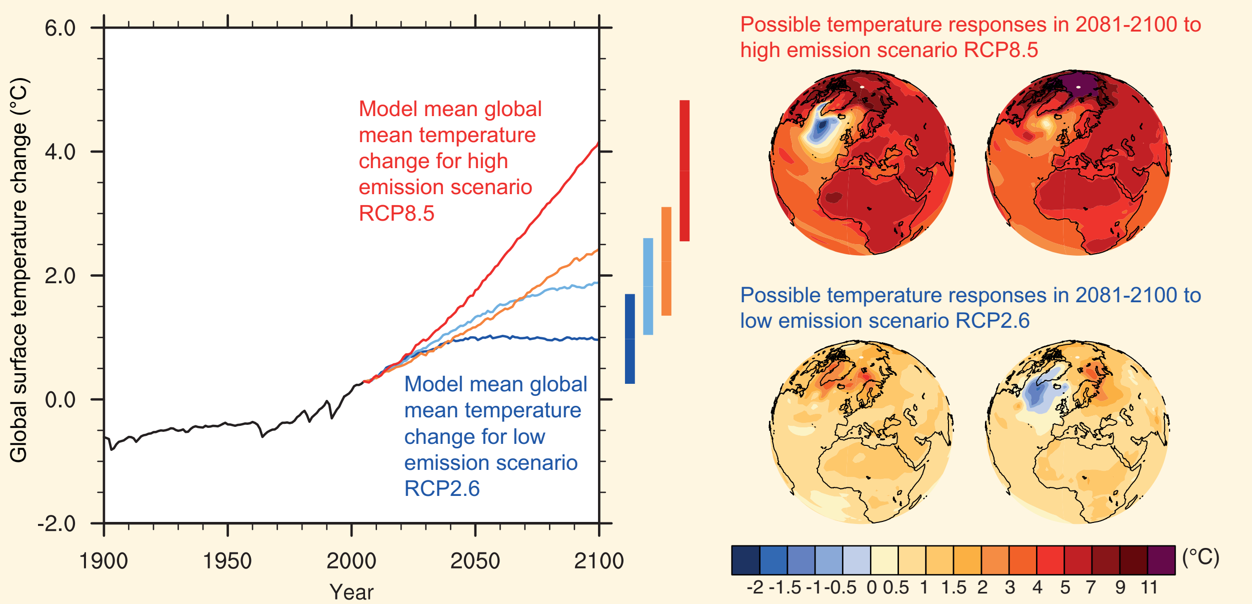
*FAQ 12.1 (continued)*

Models of varying complexity are commonly used for different projection problems. A faster model with lower resolution, or a simplified description of some climate processes, may be used in cases where long multi-century simulations are required, or where multiple realizations are needed. Simplified models can adequately represent large-scale average quantities, like global average temperature, but finer details, like regional precipitation, can be simulated only by complex models.

The coordination of model experiments and output by groups such as the Coupled Model Intercomparison Project (CMIP), the World Climate Research Program and its Working Group on Climate Models has seen the science community step up efforts to evaluate the ability of models to simulate past and current climate and to compare future climate change projections. The 'multi-model' approach is now a standard technique used by the climate science community to assess projections of a specific climate variable.

FAQ 12.1, Figure 1, right panels, shows the temperature response by the end of the 21st century for two illustrative models and the highest and lowest RCP scenarios. Models agree on large-scale patterns of warming at the surface, for example, that the land is going to warm faster than ocean, and the Arctic will warm faster than the tropics. But they differ both in the magnitude of their global response for the same scenario, and in small scale, regional aspects of their response. The magnitude of Arctic amplification, for instance, varies among different models, and a subset of models show a weaker warming or slight cooling in the North Atlantic as a result of the reduction in deepwater formation and shifts in ocean currents.

There are inevitable uncertainties in future external forcings, and the climate system's response to them, which are further complicated by internally generated variability. The use of multiple scenarios and models have become a standard choice in order to assess and characterize them, thus allowing us to describe a wide range of possible future evolutions of the Earth's climate.




**FAQ 12.1, Figure 1 |** Global mean temperature change averaged across all Coupled Model Intercomparison Project Phase 5 (CMIP5) models (relative to 1986–2005) for the four Representative Concentration Pathway (RCP) scenarios: RCP2.6 (dark blue), RCP4.5 (light blue), RCP6.0 (orange) and RCP8.5 (red); 32, 42, 25 and 39 models were used respectively for these 4 scenarios. *Likely* ranges for global temperature change by the end of the 21st century are indicated by vertical bars. Note that these ranges apply to the difference between two 20-year means, 2081–2100 relative to 1986–2005, which accounts for the bars being centred at a smaller value than the end point of the annual trajectories. For the highest (RCP8.5) and lowest (RCP2.6) scenario, illustrative maps of surface temperature change at the end of the 21st century (2081–2100 relative to 1986–2005) are shown for two CMIP5 models. These models are chosen to show a rather broad range of response, but this particular set is not representative of any measure of model response uncertainty.

FAQ

Frequently Asked Questions

## FAQ 12.2 | How Will the Earth's Water Cycle Change?

*The flow and storage of water in the Earth's climate system are highly variable, but changes beyond those due to natural variability are expected by the end of the current century. In a warmer world, there will be net increases in rainfall, surface evaporation and plant transpiration. However, there will be substantial differences in the changes between locations. Some places will experience more precipitation and an accumulation of water on land. In others, the amount of water will decrease, due to regional drying and loss of snow and ice cover.*

The water cycle consists of water stored on the Earth in all its phases, along with the movement of water through the Earth's climate system. In the atmosphere, water occurs primarily as a gas—water vapour—but it also occurs as ice and liquid water in clouds. The ocean, of course, is primarily liquid water, but the ocean is also partly covered by ice in polar regions. Terrestrial water in liquid form appears as surface water—such as lakes and rivers—soil moisture and groundwater. Solid terrestrial water occurs in ice sheets, glaciers, snow and ice on the surface and in permafrost and seasonally frozen soil.

Statements about future climate sometimes say that the water cycle will accelerate, but this can be misleading, for strictly speaking, it implies that the cycling of water will occur more and more quickly with time and at all locations. Parts of the world will indeed experience intensification of the water cycle, with larger transports of water and more rapid movement of water into and out of storage reservoirs. However, other parts of the climate system will experience substantial depletion of water, and thus less movement of water. Some stores of water may even vanish.

As the Earth warms, some general features of change will occur simply in response to a warmer climate. Those changes are governed by the amount of energy that global warming adds to the climate system. Ice in all forms will melt more rapidly, and be less pervasive. For example, for some simulations assessed in this report, summer Arctic sea ice disappears before the middle of this century. The atmosphere will have more water vapour, and observations and model results indicate that it already does. By the end of the 21st century, the average amount of water vapour in the atmosphere could increase by 5 to 25%, depending on the amount of human emissions of greenhouse gases and radiatively active particles, such as smoke. Water will evaporate more quickly from the surface. Sea level will rise due to expansion of warming ocean waters and melting land ice flowing into the ocean (see FAQ 13.2).

These general changes are modified by the complexity of the climate system, so that they should not be expected to occur equally in all locations or at the same pace. For example, circulation of water in the atmosphere, on land and in the ocean can change as climate changes, concentrating water in some locations and depleting it in others. The changes also may vary throughout the year: some seasons tend to be wetter than others. Thus, model simulations assessed in this report show that winter precipitation in northern Asia may increase by more than 50%, whereas summer precipitation there is projected to hardly change. Humans also intervene directly in the water cycle, through water management and changes in land use. Changing population distributions and water practices would produce further changes in the water cycle.

Water cycle processes can occur over minutes, hours, days and longer, and over distances from metres to kilometres and greater. Variability on these scales is typically greater than for temperature, so climate changes in precipitation are harder to discern. Despite this complexity, projections of future climate show changes that are common across many models and climate forcing scenarios. Similar changes were reported in the AR4. These results collectively suggest well understood mechanisms of change, even if magnitudes vary with model and forcing. We focus here on changes over land, where changes in the water cycle have their largest impact on human and natural systems.

Projected climate changes from simulations assessed in this report (shown schematically in FAQ 12.2, Figure 1) generally show an increase in precipitation in parts of the deep tropics and polar latitudes that could exceed 50% by the end of the 21st century under the most extreme emissions scenario. In contrast, large areas of the subtropics could have decreases of 30% or more. In the tropics, these changes appear to be governed by increases in atmospheric water vapour and changes in atmospheric circulation that further concentrate water vapour in the tropics and thus promote more tropical rainfall. In the subtropics, these circulation changes simultaneously promote less rainfall despite warming in these regions. Because the subtropics are home to most of the world's deserts, these changes imply increasing aridity in already dry areas, and possible expansion of deserts. *(continued on next page)*

FAQ

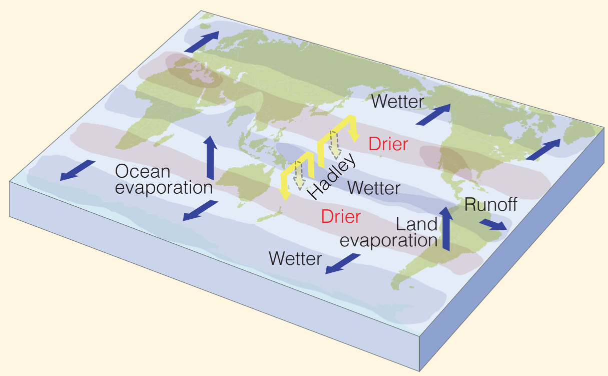
*FAQ 12.2 (continued)*

Increases at higher latitudes are governed by warmer temperatures, which allow more water in the atmosphere and thus, more water that can precipitate. The warmer climate also allows storm systems in the extratropics to transport more water vapour into the higher latitudes, without requiring substantial changes in typical wind strength. As indicated above, high latitude changes are more pronounced during the colder seasons.

Whether land becomes drier or wetter depends partly on precipitation changes, but also on changes in surface evaporation and transpiration from plants (together called evapotranspiration). Because a warmer atmosphere can have more water vapour, it can induce greater evapotranspiration, given sufficient terrestrial water. However, increased carbon dioxide in the atmosphere reduces a plant's tendency to transpire into the atmosphere, partly counteracting the effect of warming.

In the tropics, increased evapotranspiration tends to counteract the effects of increased precipitation on soil moisture, whereas in the subtropics, already low amounts of soil moisture allow for little change in evapotranspiration. At higher latitudes, the increased precipitation generally outweighs increased evapotranspiration in projected climates, yielding increased annual mean runoff, but mixed changes in soil moisture. As implied by circulation changes in FAQ 12.2, Figure 1, boundaries of high or low moisture regions may also shift.

A further complicating factor is the character of rainfall. Model projections show rainfall becoming more intense, in part because more moisture will be present in the atmosphere. Thus, for simulations assessed in this report, over much of the land, 1-day precipitation events that currently occur on average every 20 years could occur every 10 years or even more frequently by the end of the 21st century. At the same time, projections also show that precipitation events overall will tend to occur less frequently. These changes produce two seemingly contradictory effects: more intense downpours, leading to more floods, yet longer dry periods between rain events, leading to more drought.




**FAQ 12.2, Figure 1 |** Schematic diagram of projected changes in major components of the water cycle. The blue arrows indicate major types of water movement changes through the Earth's climate system: poleward water transport by extratropical winds, evaporation from the surface and runoff from the land to the oceans. The shaded regions denote areas more likely to become drier or wetter. Yellow arrows indicate an important atmospheric circulation change by the Hadley Circulation, whose upward motion promotes tropical rainfall, while suppressing subtropical rainfall. Model projections indicate that the Hadley Circulation will shift its downward branch poleward in both the Northern and Southern Hemispheres, with associated drying. Wetter conditions are projected at high latitudes, because a warmer atmosphere will allow greater precipitation, with greater movement of water into these regions.

At high latitudes and at high elevation, further changes occur due to the loss of frozen water. Some of these are resolved by the present generation of global climate models (GCMs), and some changes can only be inferred because they involve features such as glaciers, which typically are not resolved or included in the models. The warmer climate means that snow tends to start accumulating later in the fall, and melt earlier in the spring. Simulations assessed in this report show March to April snow cover in the Northern Hemisphere is projected to decrease by approximately 10 to 30% on average by the end of this century, depending on the greenhouse gas scenario. The earlier spring melt alters the timing of peak springtime flow in rivers receiving snowmelt. As a result, later flow rates will decrease, potentially affecting water resource management. These features appear in GCM simulations.

Loss of permafrost will allow moisture to seep more deeply into the ground, but it will also allow the ground to warm, which could enhance evapotranspiration. However, most current GCMs do not include all the processes needed to simulate well permafrost changes. Studies analysing soils freezing or using GCM output to drive more detailed land models suggest substantial permafrost loss by the end of this century. In addition, even though current GCMs do not explicitly include glacier evolution, we can expect that glaciers will continue to recede, and the volume of water they provide to rivers in the summer may dwindle in some locations as they disappear. Loss of glaciers will also contribute to a reduction in springtime river flow. However, if annual mean precipitation increases—either as snow or rain—then these results do not necessarily mean that annual mean river flow will decrease.

Frequently Asked Questions
## FAQ 12.3 | What Would Happen to Future Climate if We Stopped Emissions Today?

*Stopping emissions today is a scenario that is not plausible, but it is one of several idealized cases that provide insight into the response of the climate system and carbon cycle. As a result of the multiple time scales in the climate system, the relation between change in emissions and climate response is quite complex, with some changes still occurring long after emissions ceased. Models and process understanding show that as a result of the large ocean inertia and the long lifetime of many greenhouse gases, primarily carbon dioxide, much of the warming would persist for centuries after greenhouse gas emissions have stopped.*

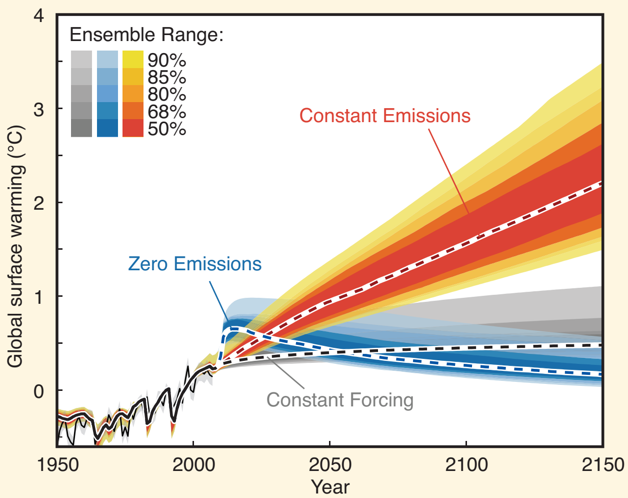
When emitted in the atmosphere, greenhouse gases get removed through chemical reactions with other reactive components or, in the case of carbon dioxide ($CO_2$), get exchanged with the ocean and the land. These processes characterize the lifetime of the gas in the atmosphere, defined as the time it takes for a concentration pulse to decrease by a factor of e (2.71). How long greenhouse gases and aerosols persist in the atmosphere varies over a wide range, from days to thousands of years. For example, aerosols have a lifetime of weeks, methane ($CH_4$) of about 10 years, nitrous oxide ($N_2O$) of about 100 years and hexafluoroethane ($C_2F_6$) of about 10,000 years. $CO_2$ is more complicated as it is removed from the atmosphere through multiple physical and biogeochemical processes in the ocean and the land; all operating at different time scales. For an emission pulse of about 1000 PgC, about half is removed within a few decades, but the remaining fraction stays in the atmosphere for much longer. About 15 to 40% of the $CO_2$ pulse is still in the atmosphere after 1000 years.

As a result of the significant lifetimes of major anthropogenic greenhouse gases, the increased atmospheric concentration due to past emissions will persist long after emissions are ceased. Concentration of greenhouse gases would not return immediately to their pre-industrial levels if emissions were halted. Methane concentration would return to values close to pre-industrial level in about 50 years, $N_2O$ concentrations would need several centuries, while $CO_2$ would essentially never come back to its pre-industrial level on time scales relevant for our society. Changes in emissions of short-lived species like aerosols on the other hand would result in nearly instantaneous changes in their concentrations.

The climate system response to the greenhouse gases and aerosols forcing is characterized by an inertia, driven mainly by the ocean. The ocean has a very large capacity of absorbing heat and a slow mixing between the surface and the deep ocean. This means that it will take several centuries for the whole ocean to warm up and to reach equilibrium with the altered radiative forcing. The surface ocean (and hence the continents) will continue to warm until it reaches a surface temperature in equilibrium with this new radiative forcing. The AR4 showed that if concentration of greenhouse gases were held constant at present day level, the Earth surface would still continue to warm by about 0.6°C over the 21st century relative to the year 2000. This is the climate commitment to current concentrations (or constant composition commitment), shown in grey in FAQ 12.3, Figure 1. Constant emissions at current levels would further increase the atmospheric concentration and result in much more warming than observed so far (FAQ 12.3, Figure 1, red lines).

Even if anthropogenic greenhouses gas emissions were halted now, the radiative forcing due to these long-lived greenhouse gases concentrations would only slowly decrease in the future, at a rate determined by the lifetime of the gas (see above). Moreover, the

*(continued on next page)*




**FAQ 12.3, Figure 1 |** Projections based on the energy balance carbon cycle model Model for the Assessment of Greenhouse Gas-Induced Climate Change (MAGICC) for constant atmospheric composition (constant forcing, grey), constant emissions (red) and zero future emissions (blue) starting in 2010, with estimates of uncertainty. Figure adapted from Hare and Meinshausen (2006) based on the calibration of a simple carbon cycle climate model to all Coupled Model Intercomparison Project Phase 3 (CMIP3) and Coupled Climate Carbon Cycle Model Intercomparison Project (C4MIP) models (Meinshausen et al., 2011a; Meinshausen et al., 2011b). Results are based on a full transient simulation starting from pre-industrial and using all radiative forcing components. The thin black line and shading denote the observed warming and uncertainty.

FAQ

*FAQ 12.3 (continued)*

climate response of the Earth System to that radiative forcing would be even slower. Global temperature would not respond quickly to the greenhouse gas concentration changes. Eliminating $CO_2$ emissions only would lead to near constant temperature for many centuries. Eliminating short-lived negative forcings from sulphate aerosols at the same time (e.g., by air pollution reduction measures) would cause a temporary warming of a few tenths of a degree, as shown in blue in FAQ 12.3, Figure 1. Setting all emissions to zero would therefore, after a short warming, lead to a near stabilization of the climate for multiple centuries. This is called the commitment from past emissions (or zero future emission commitment). The concentration of GHG would decrease and hence the radiative forcing as well, but the inertia of the climate system would delay the temperature response.

As a consequence of the large inertia in the climate and carbon cycle, the long-term global temperature is largely controlled by total $CO_2$ emissions that have accumulated over time, irrespective of the time when they were emitted. Limiting global warming below a given level (e.g., 2°C above pre-industrial) therefore implies a given budget of $CO_2$, that is, higher emissions earlier implies stronger reductions later. A higher climate target allows for a higher $CO_2$ concentration peak, and hence larger cumulative $CO_2$ emissions (e.g., permitting a delay in the necessary emission reduction).

Global temperature is a useful aggregate number to describe the magnitude of climate change, but not all changes will scale linearly global temperature. Changes in the water cycle for example also depend on the type of forcing (e.g., greenhouse gases, aerosols, land use change), slower components of the Earth system such as sea level rise and ice sheet would take much longer to respond, and there may be critical thresholds or abrupt or irreversible changes in the climate system.

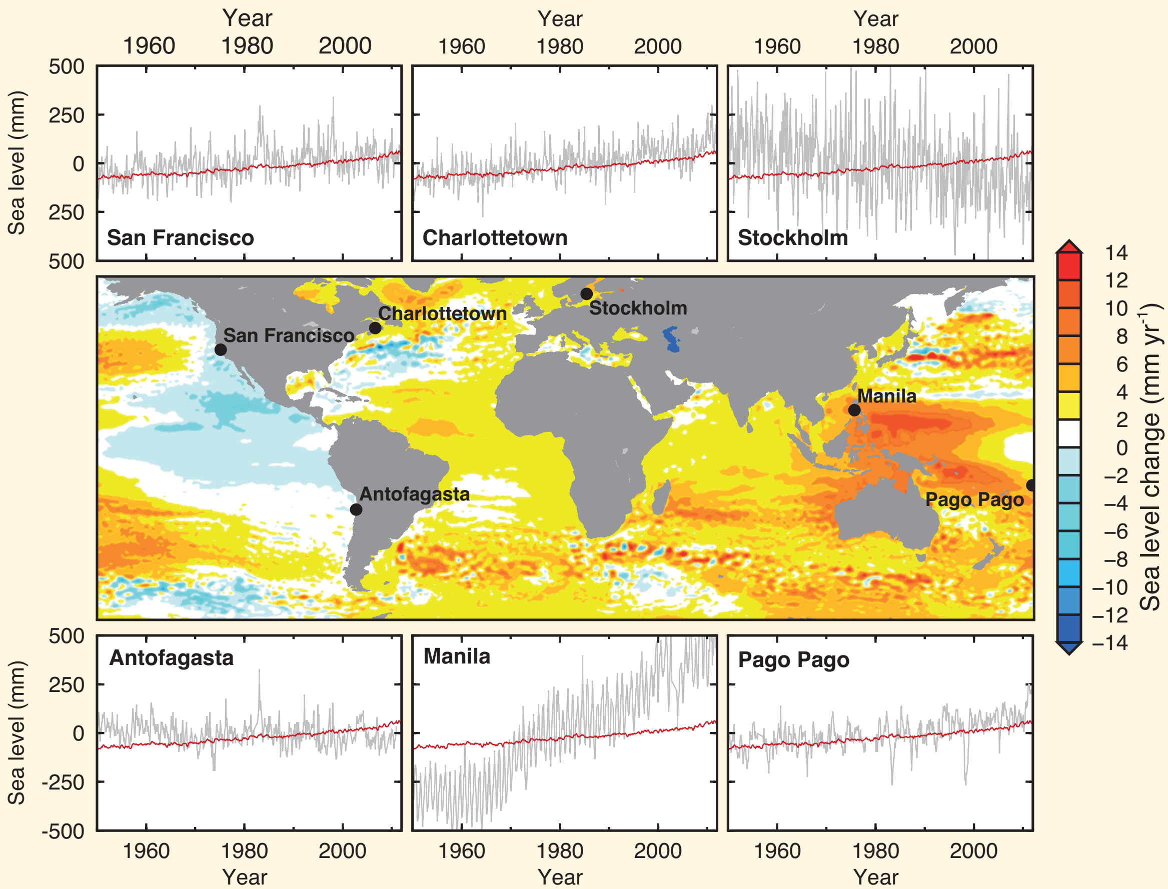
Frequently Asked Questions

## FAQ 13.1 | Why Does Local Sea Level Change Differ from the Global Average?

*Shifting surface winds, the expansion of warming ocean water, and the addition of melting ice can alter ocean currents which, in turn, lead to changes in sea level that vary from place to place. Past and present variations in the distribution of land ice affect the shape and gravitational field of the Earth, which also cause regional fluctuations in sea level. Additional variations in sea level are caused by the influence of more localized processes such as sediment compaction and tectonics.*

Along any coast, vertical motion of either the sea or land surface can cause changes in sea level relative to the land (known as relative sea level). For example, a local change can be caused by an increase in sea surface height, or by a decrease in land height. Over relatively short time spans (hours to years), the influence of tides, storms and climatic variability—such as El Niño—dominates sea level variations. Earthquakes and landslides can also have an effect by causing changes in land height and, sometimes, tsunamis. Over longer time spans (decades to centuries), the influence of climate change—with consequent changes in volume of ocean water and land ice—is the main contributor to sea level change in most regions. Over these longer time scales, various processes may also cause vertical motion of the land surface, which can also result in substantial changes in relative sea level.

Since the late 20th century, satellite measurements of the height of the ocean surface relative to the center of the Earth (known as geocentric sea level) show differing rates of geocentric sea level change around the world (see FAQ 13.1, Figure 1). For example, in the western Pacific Ocean, rates were about three times greater than the global mean value of about 3 mm per year from 1993 to 2012. In contrast, those in the eastern Pacific Ocean are lower than the global mean value, with much of the west coast of the Americas experiencing a fall in sea surface height over the same period. *(continued on next page)*




**FAQ13.1, Figure 1 |** Map of rates of change in sea surface height (geocentric sea level) for the period 1993–2012 from satellite altimetry. Also shown are relative sea level changes (grey lines) from selected tide gauge stations for the period 1950–2012. For comparison, an estimate of global mean sea level change is also shown (red lines) with each tide gauge time series. The relatively large, short-term oscillations in local sea level (grey lines) are due to the natural climate variability described in the main text. For example, the large, regular deviations at Pago Pago are associated with the El Niño-Southern Oscillation.

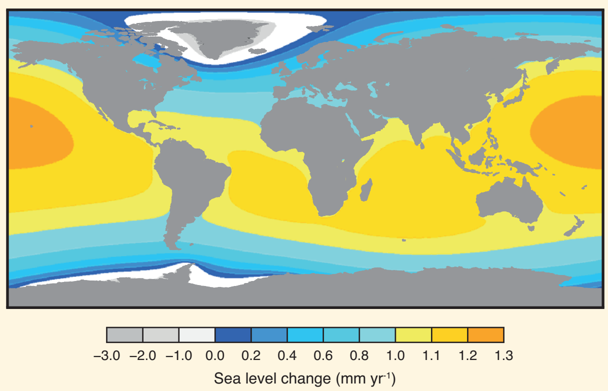
*FAQ 13.1 (continued)*

Much of the spatial variation shown in FAQ 13.1, Figure 1 is a result of natural climate variability—such as El Niño and the Pacific Decadal Oscillation—over time scales from about a year to several decades. These climate variations alter surface winds, ocean currents, temperature and salinity, and hence affect sea level. The influence of these processes will continue during the 21st century, and will be superimposed on the spatial pattern of sea level change associated with longer term climate change, which also arises through changes in surface winds, ocean currents, temperature and salinity, as well as ocean volume. However, in contrast to the natural variability, the longer term trends accumulate over time and so are expected to dominate over the 21st century. The resulting rates of geocentric sea level change over this longer period may therefore exhibit a very different pattern from that shown in FAQ 13.1, Figure 1.

Tide gauges measure relative sea level, and so they include changes resulting from vertical motion of both the land and the sea surface. Over many coastal regions, vertical land motion is small, and so the long-term rate of sea level change recorded by coastal and island tide gauges is similar to the global mean value (see records at San Francisco and Pago Pago in FAQ 13.1, Figure 1). In some regions, vertical land motion has had an important influence. For example, the steady fall in sea level recorded at Stockholm (FAQ 13.1, Figure 1) is caused by uplift of this region after the melting of a large (>1 km thick) continental ice sheet at the end of the last Ice Age, between ~20,000 and ~9000 years ago. Such ongoing land deformation as a response to the melting of ancient ice sheets is a significant contributor to regional sea level changes in North America and northwest Eurasia, which were covered by large continental ice sheets during the peak of the last Ice Age.

In other regions, this process can also lead to land subsidence, which elevates relative sea levels, as it has at Charlottetown, where a relatively large increase has been observed, compared to the global mean rate (FAQ 13.1, Figure 1). Vertical land motion due to movement of the Earth's tectonic plates can also cause departures from the global mean sea level trend in some areas—most significantly, those located near active subduction zones, where one tectonic plate slips beneath another. For the case of Antofagasta (FAQ 13.1, Figure 1) this appears to result in steady land uplift and therefore relative sea level fall.

In addition to regional influences of vertical land motion on relative sea level change, some processes lead to land motion that is rapid but highly localized. For example, the greater rate of rise relative to the global mean at Manila (FAQ 13.1, Figure 1) is dominated by land subsidence caused by intensive groundwater pumping. Land subsidence due to natural and anthropogenic processes, such as the extraction of groundwater or hydrocarbons, is common in many coastal regions, particularly in large river deltas.




**FAQ13.1, Figure 2 |** Model output showing relative sea level change due to melting of the Greenland ice sheet and the West Antarctic ice sheet at rates of 0.5 mm yr$^{-1}$ each (giving a global mean value for sea level rise of 1 mm yr$^{-1}$). The modelled sea level changes are less than the global mean value in areas near the melting ice but enhanced further afield. (Adapted from Milne et al., 2009)

It is commonly assumed that melting ice from glaciers or the Greenland and Antarctic ice sheets would cause globally uniform sea level rise, much like filling a bath tub with water. In fact, such melting results in regional variations in sea level due to a variety of processes, including changes in ocean currents, winds, the Earth's gravity field and land height. For example, computer models that simulate these latter two processes predict a regional fall in relative sea level around the melting ice sheets, because the gravitational attraction between ice and ocean water is reduced, and the land tends to rise as the ice melts (FAQ 13.1, Figure 2). However, further away from the ice sheet melting, sea level rise is enhanced, compared to the global average value.

In summary, a variety of processes drive height changes of the ocean surface and ocean floor, resulting in distinct spatial patterns of sea level change at local to regional scales. The combination of these processes produces a complex pattern of total sea level change, which varies through time as the relative contribution of each process changes. The global average change is a useful single value that reflects the contribution of climatic processes (e.g., land-ice melting and ocean warming), and represents a good estimate of sea level change at many coastal locations. At the same time, however, where the various regional processes result in a strong signal, there can be large departures from the global average value.

Frequently Asked Questions

## FAQ 13.2 | Will the Greenland and Antarctic Ice Sheets Contribute to Sea Level Change over the Rest of the Century?

*The Greenland, West and East Antarctic ice sheets are the largest reservoirs of freshwater on the planet. As such, they have contributed to sea level change over geological and recent times. They gain mass through accumulation (snowfall) and lose it by surface ablation (mostly ice melt) and outflow at their marine boundaries, either to a floating ice shelf, or directly to the ocean through iceberg calving. Increases in accumulation cause global mean sea level to fall, while increases in surface ablation and outflow cause it to rise. Fluctuations in these mass fluxes depend on a range of processes, both within the ice sheet and without, in the atmosphere and oceans. Over the course of this century, however, sources of mass loss appear set to exceed sources of mass gain, so that a continuing positive contribution to global sea level can be expected. This FAQ summarizes current research on the topic and provides indicative magnitudes for the various end-of-century (2081-2100 with respect to 1986-2005) sea level contributions from the full assessment, which are reported as the two-in-three probability level across all emission scenarios.*

Over millennia, the slow horizontal flow of an ice sheet carries mass from areas of net accumulation (generally, in the high-elevation interior) to areas of net loss (generally, the low-elevation periphery and the coastal perimeter). At present, Greenland loses roughly half of its accumulated ice by surface ablation, and half by calving. Antarctica, on the other hand, loses virtually all its accumulation by calving and submarine melt from its fringing ice shelves. Ice shelves are floating, so their loss has only a negligible direct effect on sea level, although they can affect sea level indirectly by altering the mass budget of their parent ice sheet (see below).

In East Antarctica, some studies using satellite radar altimetry suggest that snowfall has increased, but recent atmospheric modelling and satellite measurements of changes in gravity find no significant increase. This apparent disagreement may be because relatively small long-term trends are masked by the strong interannual variability of snowfall. Projections suggest a substantial increase in 21st century Antarctic snowfall, mainly because a warmer atmosphere would be able to carry more moisture into polar regions. Regional changes in atmospheric circulation probably play a secondary role. For the whole of the Antarctic ice sheet, this process is projected to contribute between 0 and 70 mm to sea level fall.

Currently, air temperatures around Antarctica are too cold for substantial surface ablation. Field and satellite-based observations, however, indicate enhanced outflow—manifested as ice-surface lowering—in a few localized coastal regions. These areas (Pine Island and Thwaites Glaciers in West Antarctica, and Totten and Cook Glaciers in East Antarctica) all lie within kilometre-deep bedrock troughs towards the edge of Antarctica's continental shelf. The increase in outflow is thought to have been triggered by regional changes in ocean circulation, bringing warmer water in contact with floating ice shelves.

On the more northerly Antarctic Peninsula, there is a well-documented record of ice-shelf collapse, which appears to be related to the increased surface melting caused by atmospheric warming over recent decades. The subsequent thinning of glaciers draining into these ice shelves has had a positive—but minor—effect on sea level, as will any further such events on the Peninsula. Regional projections of 21st century atmospheric temperature change suggest that this process will probably not affect the stability of the large ice shelves of both the West and East Antarctica, although these ice shelves may be threatened by future oceanic change (see below).

Estimates of the contribution of the Antarctic ice sheets to sea level over the last few decades vary widely, but great strides have recently been made in reconciling the observations. There are strong indications that enhanced outflow (primarily in West Antarctica) currently outweighs any increase in snow accumulation (mainly in East Antarctica), implying a tendency towards sea level rise. Before reliable projections of outflow over the 21st century can be made with greater confidence, models that simulate ice flow need to be improved, especially of any changes in the grounding line that separates floating ice from that resting on bedrock and of interactions between ice shelves and the ocean. The concept of 'marine ice-sheet instability' is based on the idea that the outflow from an ice sheet resting on bedrock below sea level increases if ice at the grounding line is thicker and, therefore, faster flowing. On bedrock that slopes downward towards the ice-sheet interior, this creates a vicious cycle of increased outflow, causing ice at the grounding line to thin and go afloat. The grounding line then retreats down slope into thicker ice that, in turn, drives further increases in outflow. This feedback could potentially result in the rapid loss of parts of the ice sheet, as grounding lines retreat along troughs and basins that deepen towards the ice sheet's interior.

*(continued on next page)*

*FAQ 13.2 (continued)*

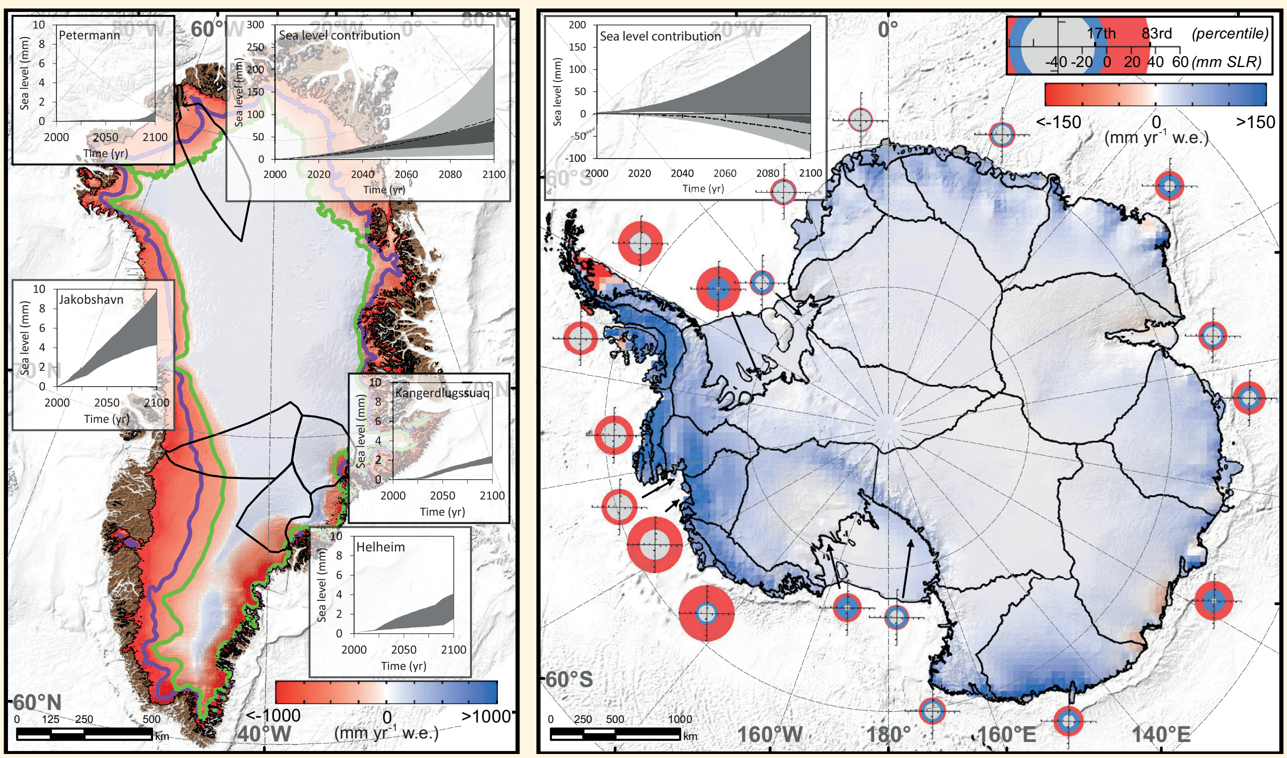
Future climate forcing could trigger such an unstable collapse, which may then continue independently of climate. This potential collapse might unfold over centuries for individual bedrock troughs in West Antarctica and sectors of East Antarctica. Much research is focussed on understanding how important this theoretical concept is for those ice sheets. Sea level could rise if the effects of marine instability become important, but there is not enough evidence at present to unambiguously identify the precursor of such an unstable retreat. Change in outflow is projected to contribute between –20 (i.e., fall) and 185 mm to sea level rise by year 2100, although the uncertain impact of marine ice-sheet instability could increase this figure by several tenths of a metre. Overall, increased snowfall seems set to only partially offset sea level rise caused by increased outflow.

In Greenland, mass loss through more surface ablation and outflow dominates a possible recent trend towards increased accumulation in the interior. Estimated mass loss due to surface ablation has doubled since the early 1990s. This trend is expected to continue over the next century as more of the ice sheet experiences surface ablation for longer periods. Indeed, projections for the 21st century suggest that increasing mass loss will dominate over weakly increasing accumulation. The refreezing of melt water within the snow pack high up on the ice sheet offers an important (though perhaps temporary) dampening effect on the relation between atmospheric warming and mass loss.

Although the observed response of outlet glaciers is both complex and highly variable, iceberg calving from many of Greenland's major outlet glaciers has increased substantially over the last decade, and constitutes an appreciable additional mass loss. This seems to be related to the intrusion of warm water into the coastal seas around Greenland, but it is not clear whether this phenomenon is related to inter-decadal variability, such as the North Atlantic

*(continued on next page)*






**FAQ 13.2, Figure 1 |** Illustrative synthesis of projected changes in SMB and outflow by 2100 for (a) Greenland and (b) Antarctic ice sheets. Colours shown on the maps refer to projected SMB change between the start and end of the 21st century using the RACMO2 regional atmospheric climate model under future warming scenarios A1B (Antarctic) and RCP4.5 (Greenland). For Greenland, average equilibrium line locations during both these time periods are shown in purple and green, respectively. Ice-sheet margins and grounding lines are shown as black lines, as are ice-sheet sectors. For Greenland, results of flowline modelling for four major outlet glaciers are shown as inserts, while for Antarctica the coloured rings reflect projected change in outflow based on a probabilistic extrapolation of observed trends. The outer and inner radius of each ring indicate the upper and lower bounds of the two-thirds probability range of the contribution, respectively (scale in upper right); red refers to mass loss (sea level rise) while blue refers to mass gain (sea level fall). Finally, the sea level contribution is shown for each ice sheet (insert located above maps) with light grey referring to SMB (model experiment used to generate the SMB map is shown as a dashed line) and dark grey to outflow. All projections refer to the two-in-three probability range across all scenarios.

*FAQ 13.2 (continued)*

Oscillation, or a longer term trend associated with greenhouse gas–induced warming. Projecting its effect on 21st century outflow is therefore difficult, but it does highlight the apparent sensitivity of outflow to ocean warming. The effects of more surface melt water on the lubrication of the ice sheet's bed, and the ability of warmer ice to deform more easily, may lead to greater rates of flow, but the link to recent increases in outflow is unclear. Change in the net difference between surface ablation and accumulation is projected to contribute between 10 and 160 mm to sea level rise in 2081-2100 (relative to 1986-2005), while increased outflow is projected to contribute a further 10 to 70 mm (Table 13.5).

The Greenland ice sheet has contributed to a rise in global mean sea level over the last few decades, and this trend is expected to increase during this century. Unlike Antarctica, Greenland has no known large-scale instabilities that might generate an abrupt increase in sea level rise over the 21st century. A threshold may exist, however, so that continued shrinkage might become irreversible over multi-centennial time scales, even if the climate were to return to a pre-industrial state over centennial time scales. Although mass loss through the calving of icebergs may increase in future decades, this process will eventually end when the ice margin retreats onto bedrock above sea level where the bulk of the ice sheet resides.

Frequently Asked Questions

## FAQ 14.1 | How is Climate Change Affecting Monsoons?

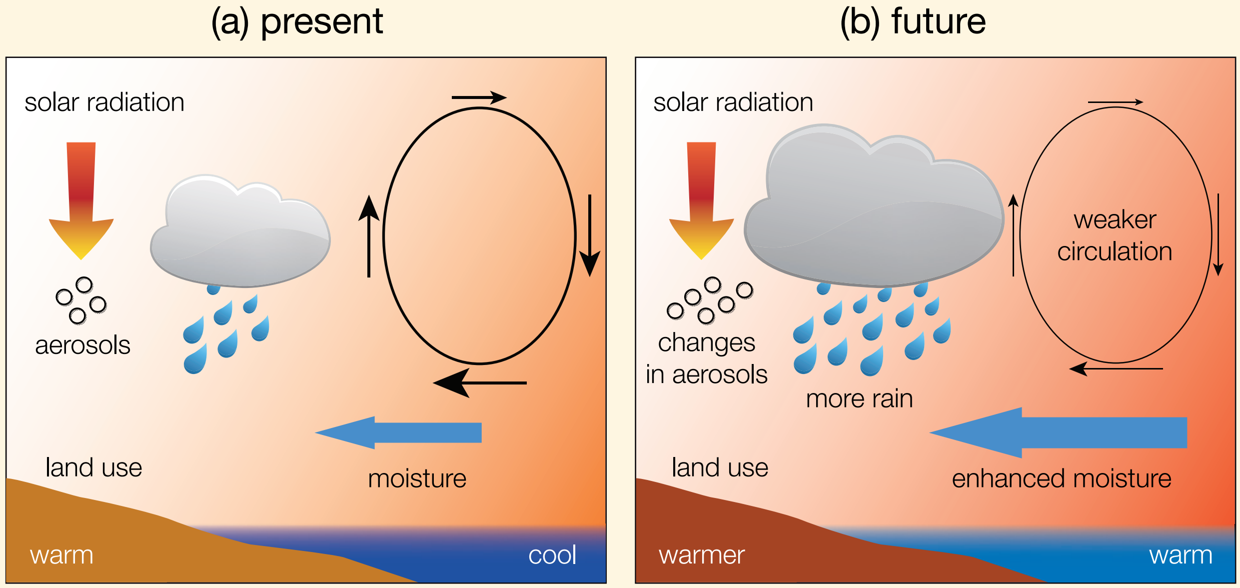
*Monsoons are the most important mode of seasonal climate variation in the tropics, and are responsible for a large fraction of the annual rainfall in many regions. Their strength and timing is related to atmospheric moisture content, land–sea temperature contrast, land cover and use, atmospheric aerosol loadings and other factors. Overall, monsoonal rainfall is projected to become more intense in future, and to affect larger areas, because atmospheric moisture content increases with temperature. However, the localized effects of climate change on* regional *monsoon strength and variability are complex and more uncertain.*

Monsoon rains fall over all tropical continents: Asia, Australia, the Americas and Africa. The monsoon circulation is driven by the difference in temperature between land and sea, which varies seasonally with the distribution of solar heating. The duration and amount of rainfall depends on the moisture content of the air, and on the configuration and strength of the atmospheric circulation. The regional distribution of land and ocean also plays a role, as does topography. For example, the Tibetan Plateau—through variations in its snow cover and surface heating—modulates the strength of the complex Asian monsoon systems. Where moist on-shore winds rise over mountains, as they do in southwest India, monsoon rainfall is intensified. On the lee side of such mountains, it lessens.

Since the late 1970s, the East Asian summer monsoon has been weakening and not extending as far north as it used to in earlier times , as a result of changes in the atmospheric circulation. That in turn has led to increasing drought in northern China, but floods in the Yangtze River Valley farther south. In contrast, the Indo-Australian and Western Pacific monsoon systems show no coherent trends since the mid-20th century, but are strongly modulated by the El Niño-Southern Oscillation (ENSO). Similarly, changes observed in the South American monsoon system over the last few decades are strongly related to ENSO variability. Evidence of trends in the North American monsoon system is limited, but a tendency towards heavier rainfalls on the northern side of the main monsoon region has been observed. No systematic long-term trends have been observed in the behaviour of the Indian or the African monsoons.

The land surface warms more rapidly than the ocean surface, so that surface temperature contrast is increasing in most regions. The tropical atmospheric overturning circulation, however, slows down on average as the climate warms due to energy balance constraints in the tropical atmosphere. These changes in the atmospheric circulation lead to regional changes in monsoon intensity, area and timing. There are a number of other effects as to how

*(continued on next page)*




**FAQ 14.1, Figure 1 |** Schematic diagram illustrating the main ways that human activity influences monsoon rainfall. As the climate warms, increasing water vapour transport from the ocean into land increases because warmer air contains more water vapour. This also increases the potential for heavy rainfalls. Warming-related changes in large-scale circulation influence the strength and extent of the overall monsoon circulation. Land use change and atmospheric aerosol loading can also affect the amount of solar radiation that is absorbed in the atmosphere and land, potentially moderating the land–sea temperature difference.

*FAQ 14.1 (continued)*

climate change can influence monsoons. Surface heating varies with the intensity of solar radiation absorption, which is itself affected by any land use changes that alter the reflectivity (albedo) of the land surface. Also, changing atmospheric aerosol loadings, such as air pollution, affect how much solar radiation reaches the ground, which can change the monsoon circulation by altering summer solar heating of the land surface. Absorption of solar radiation by aerosols, on the other hand, warms the atmosphere, changing the atmospheric heating distribution.

The strongest effect of climate change on the monsoons is the increase in atmospheric moisture associated with warming of the atmosphere, resulting in an increase in total monsoon rainfall even if the strength of the monsoon circulation weakens or does not change.

Climate model projections through the 21st century show an increase in total monsoon rainfall, largely due to increasing atmospheric moisture content. The total surface area affected by the monsoons is projected to increase, along with the general poleward expansion of the tropical regions. Climate models project from 5% to an approximately 15% increase of global monsoon rainfall depending on scenarios. Though total tropical monsoon rainfall increases, some areas will receive less monsoon rainfall, due to weakening tropical wind circulations. Monsoon onset dates are *likely* to be early or not to change much and the monsoon retreat dates are *likely* to delay, resulting in lengthening of the monsoon season.

Future regional trends in monsoon intensity and timing remain uncertain in many parts of the world. Year-to-year variations in the monsoons in many tropical regions are affected by ENSO. How ENSO will change in future—and how its effects on monsoon will change—also remain uncertain. However, the projected overall increase in monsoon rainfall indicates a corresponding increase in the risk of extreme rain events in most regions.

Frequently Asked Questions

## FAQ 14.2 | How Are Future Projections in Regional Climate Related to Projections of Global Means?

*The relationship between regional climate change and global mean change is complex. Regional climates vary strongly with location and so respond differently to changes in global-scale influences. The global mean change is, in effect, a convenient summary of many diverse regional climate responses.*

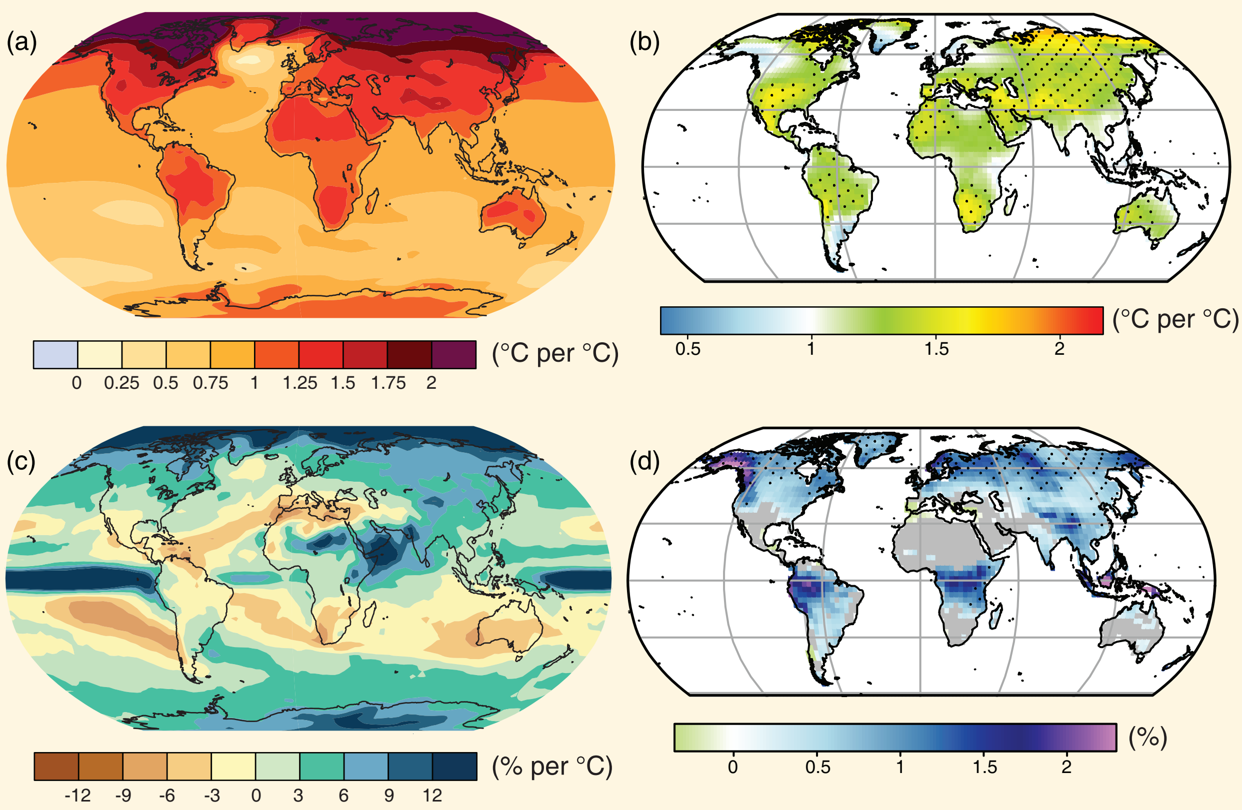
Heat and moisture, and changes in them, are not evenly distributed across the globe for several reasons:

- External forcings vary spatially (e.g., solar radiation depends on latitude, aerosol emissions have local sources, land use changes regionally, etc.).
- Surface conditions vary spatially, for example, land/sea contrast, topography, sea surface temperatures, soil moisture content.
- Weather systems and ocean currents redistribute heat and moisture from one region to another.

Weather systems are associated with regionally important climate phenomena such as monsoons, tropical convergence zones, storm tracks and important modes of climate variability (e.g., El Niño-Southern Oscillation (ENSO), North Atlantic Oscillation (NAO), Southern Annular Mode (SAM), etc.). In addition to modulating regional warming, some climate phenomena are also projected to change in the future, which can lead to further impacts on regional climates (see Table 14.3).

Projections of change in surface temperature and precipitation show large regional variations (FAQ 14.2, Figure 1). Enhanced surface warming is projected to occur over the high-latitude continental regions and the Arctic ocean,

*(continued on next page)*




**FAQ 14.2, Figure 1 |** Projected 21st century changes in annual mean and annual extremes (over land) of surface air temperature and precipitation: (a) mean surface temperature per °C of global mean change, (b) 90th percentile of daily maximum temperature per °C of global average maximum temperature, (c) mean precipitation (in % per °C of global mean temperature change), and (d) fraction of days with precipitation exceeding the 95th percentile. Sources: Panels (a) and (c) projected changes in means between 1986–2005 and 2081–2100 from CMIP5 simulations under RCP4.5 scenario (see Chapter 12, Figure 12.41); Panels (b) and (d) projected changes in extremes over land between 1980–1999 and 2081–2100 (adapted from Figures 7 and 12 of Orlowsky and Seneviratne, 2012).

*FAQ 14.2 (continued)*

while over other oceans and lower latitudes changes are closer to the global mean (FAQ 14.2, Figure 1a). For example, warming near the Great Lakes area of North America is projected to be about 50% greater than that of the global mean warming. Similar large regional variations are also seen in the projected changes of more extreme temperatures (FAQ 14.2, Figure 1b). Projected changes in precipitation are even more regionally variable than changes in temperature (FAQ 14.2, Figure 1c, d), caused by modulation from climate phenomena such as the monsoons and tropical convergence zones. Near-equatorial latitudes are projected to have increased mean precipitation, while regions on the poleward edges of the subtropics are projected to have reduced mean precipitation. Higher latitude regions are projected to have increased mean precipitation and in particular more extreme precipitation from extratropical cyclones.

Polar regions illustrate the complexity of processes involved in regional climate change. Arctic warming is projected to increase more than the global mean, mostly because the melting of ice and snow produces a regional feedback by allowing more heat from the Sun to be absorbed. This gives rise to further warming, which encourages more melting of ice and snow. However, the projected warming over the Antarctic continent and surrounding oceans is less marked in part due to a stronger positive trend in the Southern Annular Mode. Westerly winds over the mid-latitude southern oceans have increased over recent decades, driven by the combined effect of loss of stratospheric ozone over Antarctica, and changes in the atmosphere's temperature structure related to increased greenhouse gas concentrations. This change in the Southern Annular Mode is well captured by climate models and has the effect of reducing atmospheric heat transport to the Antarctic continent. Nevertheless, the Antarctic Peninsula is still warming rapidly, because it extends far enough northwards to be influenced by the warm air masses of the westerly wind belt.