International Cryosphere Climate Initiative

Overshoot and the Cryosphere: Averting Global Meltdown

The Implications of Temperature Overshoot, Peak and Decline for the Cryosphere



UN environment programme

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Suggested citation

International Cryosphere Climate Initiative (2026). Overshoot and the Cryosphere: Averting Global Meltdown. The Implications of Temperature Overshoot, Peak and Decline for the Cryosphere [Kirkham, J. D., Pearson, P., Garbe, J., Hancock, S., and Imdieke, A. (eds.)]. Stockholm, Sweden. DOI: 10.5281/zenodo.23477467

URL: https://iccinet.org/overshoot/

Acknowledgements

The International Cryosphere Climate Initiative (ICCI) extends its gratitude to the authors, reviewers, and chapter coordinators for their contributions to the preparation of this report. Authors and reviewers have contributed to the report in their individual capacities. Their affiliations are only mentioned for identification purposes.

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Abbreviations and Acronyms

AABW	Antarctic Bottom Water
AMOC	Atlantic Meridional Overturning Circulation
AR6	Sixth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC)
CDR	Carbon dioxide removal
CH_4	Methane
CMIP	Coupled Model Intercomparison Project
CO_2	Carbon dioxide
EAIS	East Antarctic Ice Sheet
GHG	Greenhouse gas
GMSL	Global mean sea level
GST	Global Stocktake
ICJ	International Court of Justice
IPCC	Intergovernmental Panel on Climate Change
LDC	Least developed country
MICI	Marine ice-cliff instability
MISI	Marine ice-sheet instability
OECD	Organisation for Economic Co-operation and Development
PPM	Parts per million
RCP	Representative Concentration Pathway
SIDS	Small Island Developing States
SLE	Sea-level equivalent
SSP	Shared Socioeconomic Pathway
UN	United Nations
UNEA	United Nations Environment Assembly
UNEP	United Nations Environment Programme
UNFCCC	United Nations Framework Convention on Climate Change
WAIS	West Antarctic Ice Sheet



Photo credit: L. Hösli

Glossary

In this glossary, selected key terms used in this report are defined. The terms in this glossary are taken from the online glossary of the Intergovernmental Panel on Climate Change (IPCC) https://apps.ipcc.ch/glossary/ unless otherwise indicated.

Active layer
The upper layer of ground above permafrost that thaws each summer and refreezes in winter (adapted from IPCC).

Adaptation
In human systems, the process of adjustment to actual or expected climate and its effects, in order to moderate harm or exploit beneficial opportunities. In natural systems, the process of adjustment to actual climate and its effects; human intervention may facilitate adjustment to expected climate and its effects.

Adaptation limits
The point at which an actor's objectives (or system needs) cannot be secured from intolerable risks through adaptive actions.

Albedo
The proportion of sunlight (solar radiation) reflected by a surface or object, often expressed as a percentage. Clouds, snow and ice usually have high albedo; soil surfaces cover the albedo range from high to low; vegetation in the dry season and/or in arid zones can have high albedo, whereas photosynthetically active vegetation and the ocean have low albedo. The Earth's planetary albedo changes mainly through changes in cloudiness and of snow, ice, leaf area and land cover.

Atlantic Meridional Overturning Circulation (AMOC)
The main current system in the South and North Atlantic Oceans. AMOC transports warm upper-ocean water northwards and cold, deep water southwards, as part of the global ocean circulation system. Changes in the strength of AMOC can affect other components of the climate system.

Atmospheric rivers
Long, narrow (up to a few hundred km wide), shallow (up to a few km deep) and transient corridors of strong horizontal water vapour transport that are typically associated with a low-level jet stream ahead of the cold front of an extratropical cyclone.

Antarctic Bottom Water (AABW)
The coldest, densest water mass in the world's oceans, occupying depths below 4,000 metres in major ocean basins.

Antarctic Slope Current
A narrow, westward-flowing ocean circulation that surrounds the margins of Antarctica along the upper continental slope.

Arctic amplification
Describes the phenomenon where surface temperature change at high latitudes exceeds the global average surface temperature change. The terms Arctic amplification or Antarctic amplification are used when describing the phenomenon occurring at one of the poles.

Black carbon
A fine black powder made of nearly pure elemental carbon, produced by the incomplete combustion or thermal decomposition of gaseous or liquid hydrocarbons.

Carbon budget
The maximum amount of cumulative net global anthropogenic CO_2 emissions that would result in limiting global warming to a given level with a given probability, considering the effect of other anthropogenic climate forcers. This is referred to as the Total Carbon Budget when expressed starting from the pre-industrial period, and as the Remaining Carbon Budget when expressed from a recent specified date.

Carbon dioxide removal (CDR)
Anthropogenic activities removing CO_2 from the atmosphere and durably storing it in geological, terrestrial or ocean reservoirs, or in products. It includes existing and potential anthropogenic enhancement of biological or geochemical CO_2 sinks and direct air carbon dioxide capture and storage (DACCS) but excludes natural CO_2 uptake not directly caused by human activities.

Cascading impacts
Cascading impacts from extreme weather/climate events occur when an extreme hazard generates a sequence of secondary events in natural and human systems that result in physical, natural, social or economic disruption, whereby the resulting impact is significantly larger than the initial impact. Cascading impacts are complex and multi-dimensional, and are associated more with the magnitude of vulnerability than with that of the hazard.

Circumpolar Deep Water (CDW)
The largest volume water mass in the Southern Ocean, formed by the mixing of North Atlantic Deep Water, Antarctic Bottom Water, Antarctic Intermediate Water, and recirculated deep waters from the Indian and Pacific Oceans.

Cryosphere
The components of the Earth system at and below the land and ocean surface that are frozen, including snow cover, glaciers, ice sheets, ice shelves, icebergs, sea ice, lake ice, river ice, permafrost and seasonally frozen ground.

Deep convection
The process by which cold, salty, dense surface waters sink into the deeper ocean, carrying heat, oxygen, carbon and nutrients with them. Deep convection acts as an important driver of the deep and abyssal ocean overturning circulation, regulating the long-term storage and redistribution of heat and carbon (definition by the authors of this report).

Deoxygenation
The decline in the amount of dissolved oxygen in seawater. This process reduces the amount of oxygen available to marine organisms and is driven by factors including ocean warming and changes in ocean circulation (definition by the authors of this report).

Earth System Model (ESM)
A coupled atmosphere–ocean general circulation model (AOGCM) in which a representation of the carbon cycle is included, allowing for interactive calculation of atmospheric carbon dioxide (CO_2) or compatible emissions. Additional components (e.g., atmospheric chemistry, ice sheets, dynamic vegetation, nitrogen cycle, but also urban or crop models) may be included.

Loss and Damage, and losses and damages
Research has taken Loss and Damage (capitalised letters) to refer to political debate under the United Nations Framework Convention on Climate Change (UNFCCC) following the establishment of the Warsaw Mechanism on Loss and Damage in 2013, which is to 'address loss and damage associated with impacts of climate change, including extreme events and slow onset events, in developing countries that are particularly vulnerable to the adverse effects of climate change.' Lowercase letters (losses and damages) have been taken to refer broadly to harm from (observed) impacts and (projected) risks and can be economic or non-economic.

Maladaptation
Actions that may lead to increased risk of adverse climate-related outcomes, including via increased greenhouse gas (GHG) emissions, increased or shifted vulnerability to climate change, more inequitable outcomes, or diminished welfare, now or in the future. Most often, maladaptation is an unintended consequence.

Mitigation (of climate change)
A human intervention to reduce emissions or enhance the sinks of greenhouse gases.

Marine heatwave
A prolonged period of abnormally high ocean temperatures and spans large regions.

Mass balance
Difference between the mass input (accumulation) and the mass loss (ablation) of an ice body (e.g., a glacier or ice sheet) over a stated time period, which is often a year or a season. Surface mass balance refers to the difference between surface accumulation and surface ablation.

Net-zero CO_2 emissions
Condition in which anthropogenic carbon dioxide (CO_2) emissions are balanced by anthropogenic CO_2 removals over a specified period.

Net-zero GHG emissions
Condition in which metric weighted anthropogenic GHG emissions are balanced by metric-weighted anthropogenic GHG removals over a specified period. The quantification of net-zero GHG emissions depends on the GHG emission metric chosen to compare emissions and removals of different gases, as well as the time horizon chosen for that metric.

Net-negative CO_2 emissions
Condition in which anthropogenic CO_2 removals exceed anthropogenic CO_2 emissions over a specified period.

Net-negative GHG emissions
A situation of net negative GHG emissions is achieved when metric-weighted anthropogenic GHG removals exceed metric-weighted anthropogenic GHG emissions. Where multiple GHG are involved, the quantification of net emissions depends on the metric chosen to compare emissions of different gases (such as the global warming potential), as well as the chosen time horizon (definition from the UNEP Limiting Overshoot Report).

Ocean acidification/corrosive conditions
Ocean acidification causes a decline in pH through absorption of CO_2 into the ocean from the atmosphere, producing carbonic acid. Declining pH can be harmful to marine organisms as stable pH is necessary for many biological functions and processes.

Ocean acidification also causes a decline in carbonate ions as a result of a shift in marine carbonate chemistry. Carbonate ions are needed to form calcium carbonate minerals, which form the basis of many organisms' shells, skeletons and other hard parts. As ocean acidification progresses, the carbonate ion concentration decreases, lowering the calcium carbonate saturation state. Once the saturation state is below 1 (the corrosive threshold), the mineral is more likely to dissolve, and the seawater is considered corrosive (definition by the authors of this report).

Ocean freshening
Ocean water is saline (saltwater), but the entry of freshwater from rivers and melting ice can decrease the salinity to lower levels, including levels where ocean-based organisms cannot survive (definition by the authors of this report).

Ocean overturning circulation
The global system of ocean currents where surface water cools and sinks to the deep ocean, eventually recirculating back to the surface via deep ocean circulation pathways. The overturning circulation is also controlled by salinity effects: salty water is heavier, so sinks while freshwater/less saline water floats above and stays at the surface (definition by the authors of this report).

Ocean stratification
Refers to the density layering of the water column and the degree to which adjacent layers can mix. Higher stratification means less mixing and can slow or prevent ocean circulation. This can also prevent nutrients and oxygen from reaching organisms at different levels. This is especially important for the health and survival of marine organisms (definition by the authors of this report).

Overshoot, peak and decline pathways
Sometimes abbreviated as Overshoot pathways. Pathways that first exceed a global warming level (e.g., 1.5°C) and then return to or below that level again before the end of a specified period (e.g., before 2100). Sometimes the magnitude and likelihood of the overshoot are also characterized. The overshoot duration can vary but typically lasts for at least multiple decades. (adapted from IPCC AR6 by the UNEP Limiting Overshoot Report).

A 1.5°C overshoot, peak and decline pathway comprises a multidecadal sequence involving four distinct stages:

- **Exceedance of 1.5°C:** the period in which global warming surpasses 1.5°C but before it reaches its peak.
- **Plateauing stage:** the period when global warming reaches and plateaus at its maximum level above 1.5°C for a period before starting to decline. The maximum level is called peak warming.
- **Decline stage:** the period in which global temperatures begin to decline from their peak level.
- **Stabilization stage:** reached once global temperatures return to below 1.5°C after peak and reversal and stabilize at this level.

For the purpose of this report, Overshoot, when mentioned alone, refers to the exceedance of a specified global warming level. The world is considered in overshoot when the multidecadal average global surface air temperature increase exceeds a specified warming level (e.g., 1.5°C), not when the global temperature in a single year exceeds that warming level temperature[1].

Permafrost
Ground (a mixture of soil or rock and any ice and organic material) that remains at or below 0°C for two consecutive years or longer.

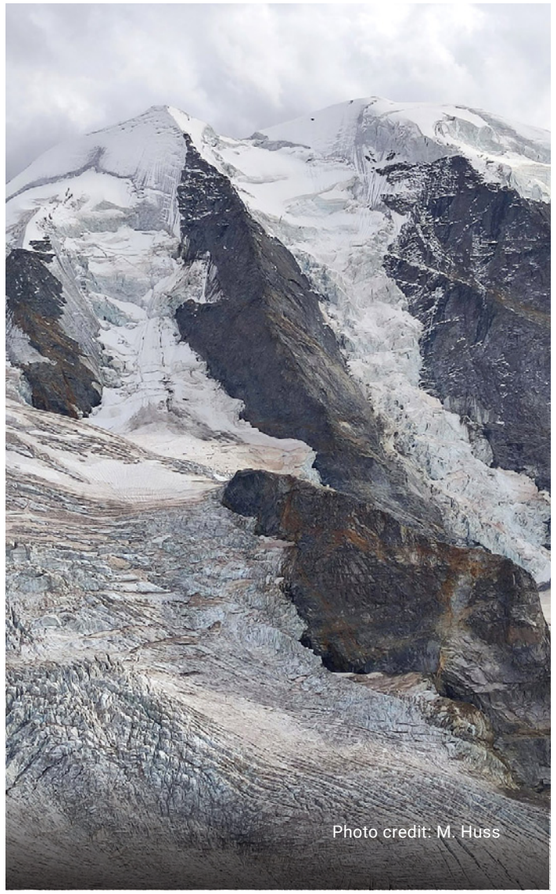
Primary production
The production of organic matter by organisms such as phytoplankton using sunlight and nutrients. It forms the foundation of marine food webs and plays an important role in removing and storing carbon from the atmosphere (adapted from IPCC).

Talik
An unfrozen layer or area that forms below the active layer but above permafrost. This middle layer forms because winter freezing does not reach deep enough to reconnect with permafrost (adapted from IPCC).

Thermal erosion
A process in which thawing of permafrost and physical removal of thawed ground by moving surface water are happening simultaneously, quickly developing landscapes that can contain deep channels and gullies (adapted from IPCC).

Thermokarst
A process that involves subsidence of the ground surface caused by melting of excess ice in thawing permafrost and resulting in the formation of depressions that may create lakes, wetlands or thaw slumps (adapted from IPCC).

Tipping point
A critical threshold beyond which a system reorganizes, often abruptly and/or irreversibly.


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Summary for Policymakers

Executive Summary

Exceeding the 1.5°C limit is termed "overshoot†," and brings large — and often irreversible — melting, thawing and loss in the cryosphere and polar oceans: the areas of the planet with snow, ice and frozen soils.

The impacts of cryosphere loss with overshoot will be global, and in many cases irreversible on timescales relevant to human society. The reality of these irreversible impacts, including threats which may exceed the limits of adaptation and cause extensive loss and damage, needs to be made clear to policy makers and the general public.

There remains a persistent myth that overshoot to 2°C, 2.5°C or even 3°C and above is manageable, or even preferable to the "costs" of urgent mitigation. Cryosphere-mediated global impacts make clear this is not the case, even in the near-term.

† The Paris Agreement established a framework to limit the rise in global temperature to well below 2°C and to pursue efforts to limit warming to 1.5°C. These limits are relative to global mean surface temperatures during the pre-industrial period (1850–1900). The Intergovernmental Panel on Climate Change (IPCC) uses a 20-year average to define global warming levels, with the crossing time of particular warming levels as the mid-point of the 20-year period. The first calendar year above 1.5°C occurred in 2024. Although a single year above 1.5°C does not imply that the Paris Agreement limit has been formally breached, recent research suggests that the first year above 1.5°C is likely to occur within the 20-year period where average warming reaches that same level [2].

For policy, the following implications are clear:

1. **Exceeding 1.5°C is not safe for the cryosphere. Global impacts and risks from cryosphere loss increase with every fraction of a degree of warming.**
2. **Overshoot must be kept as short and low as possible. Efforts to lower long-term temperatures back to 1.5°C, and preferably below 1°C, are of critical importance to reduce cryosphere-driven loss and damage.**
3. **Returning to 1.5°C after a brief period of overshoot remains feasible if efforts to reduce greenhouse gas emissions are intensified immediately, supported by deep reductions in short-lived climate forcers and significantly increased carbon removals from the atmosphere once net zero emissions are reached.**
4. **Many impacts from cryosphere loss will be irreversible on human timescales even if global temperatures are returned to 1.5°C, especially if thresholds and tipping points are crossed. These include sea-level rise from the melting of ice sheets and glaciers, water security challenges from glacier and snow loss, greenhouse gas emissions from permafrost thaw, polar ocean warming, acidification, ocean circulation disruption, and ecosystem and biodiversity loss.**
5. **Increased efforts and resources are urgently needed to meet growing adaptation needs, build resilience, and respond to loss and damage as a result of exceeding 1.5°C because of cryosphere loss.**
6. **The ultimate scale of impacts and economic costs from cryosphere loss is up to today's political leaders. Stringent near-term emissions reductions, focused on rapidly transitioning away from fossil fuels, to limit overshoot as close to 1.5°C as possible and for as short a time as possible can still slow many losses, enable effective adaptation, and avoid the most destructive impacts on the global community.**

Overshoot of 1.5°C is now unavoidable due to our collective failure to act. Recent modelling, however, points to measures that could limit potential overshoot to 1.7°C for around just two decades as net zero greenhouse gas emissions are reached, with sustainable carbon drawdown thereafter leading to temperature decline to 1.5°C or below (about 1.2°C) by 2100‡.

This places temperatures back on the lowest emissions pathways assessed in the IPCC's Sixth Assessment Report (AR6); with the potential to fall below 1°C by 2150 if carbon drawdown continues at scale. **These are the measures, needed and readily available, to avoid the worst destructive impacts, triggered by overshoot, from global cryosphere loss.**

‡ For example, see:
Grant et al., (2025), Rescuing 1.5°C: New evidence on highest possible ambition to deliver the Paris Agreement.
https://climateanalytics.org/publications/rescuing-1-5c

UNEP Emissions Gap Report 2026: The point of return – Urgent emissions cuts needed to come back from overshoot of 1.5°C.
https://doi.org/10.59117/20.500.11822/50038

Main Findings

Scientific understanding of the sensitivity of the cryosphere and polar oceans to rising atmospheric greenhouse gas concentrations and global temperatures has grown enormously in recent decades.

As we enter a period of temperature overshoot above the Paris 1.5°C warming limit, humans and the environment will face greater risks from a melting cryosphere and degradation of the polar oceans. Melting land ice accelerates sea-level rise that already threatens hundreds of millions of people, as well as infrastructure, agricultural land, ecosystems and cultural heritage situated along coastlines globally. The cryosphere and polar oceans also directly provide essential and economically valuable resources to billions of people through provision of drinking water, agricultural irrigation, power generation, transport routes over sea ice and by supporting diverse ecosystems. Snow, glaciers, sea ice, ice sheets, permafrost and the polar oceans help maintain a safe and liveable climate on Earth by regulating ocean and atmospheric temperatures, including ocean circulation and regional weather patterns.

Overshooting 1.5°C therefore poses a growing range of risks from the Earth's cryosphere.

The polar ice sheets covering Greenland and Antarctica are melting about six times faster today than in the 1990s. The response of these ice sheets to overshoot will depend on the peak temperature and the duration of the overshoot period: the higher the warming and the longer it is sustained, the greater the level of committed long-term sea-level rise. Higher temperatures also increase the risk of triggering ice sheet instabilities that could cause sea levels to rise by several metres over the next few centuries, potentially at rates of up to several (1-4) centimetres per year at times, well beyond adaptation limits for many coastal and low-lying communities.

Even in scenarios where the duration of overshoot is very brief and global warming returns to 1.5°C, ice loss will continue. Current scientific evidence suggests that a truly 'safe' long-term temperature limit for the polar ice sheets lies much closer to 1.0°C, and possibly even lower.

Loss of mountain glaciers and snowpack will drive increased environmental hazards for populations situated in and downstream of mountain regions, ranging from water insecurity to sudden disaster events like glacial lake outburst floods or ice/rock avalanches, even with appropriate adaptation and disaster prevention measures. Even temporary overshoot of 1.5°C will irreversibly decrease snow and ice reserves on human timescales, with substantially greater losses at higher levels of overshoot that will result in extreme and rapid variations in water availability. Scenarios in which temperatures return to 1.5°C after several centuries of overshoot show notably higher losses of glacier ice than projections with no overshoot. A return to well below 1.5°C could initiate some glacier regrowth on time scales of decades to centuries with sufficient snowfall and regional climate recovery. Less downstream water availability is still expected however compared to minimal or non-overshoot scenarios, affecting populations, industries, and mountain and downstream ecosystems reliant on meltwater from glaciers and snow for centuries. New land areas and ecosystems are also emerging as glaciers retreat, presenting new governance challenges to manage risks in deglaciating landscapes.

Overshoot of both temperature and the equivalent atmospheric CO_2 levels (>430 ppm and above) will also have profound impacts on the Arctic and Southern oceans. These polar oceans are the foundation of the global ocean circulation system that redistributes heat, carbon, oxygen and nutrients around the world. They play a disproportionately large role in absorbing and storing heat and CO_2, helping to regulate the global climate, while also supporting diverse ecosystems and economically important fisheries. Both polar oceans are already experiencing rapid, interconnected physical, chemical and ecological changes due to rising CO_2 levels and temperature, including acidification from absorption of CO_2, warming and freshening of ocean waters, and large-scale changes in ocean circulation patterns and currents such as the Atlantic Meridional Overturning Circulation (AMOC). Overshoot pathways consistently project long-lasting changes in the polar oceans, on scales of decades to multiple centuries, even after temperatures and atmospheric CO_2 levels decline. These include persistently warmer polar oceans, more corrosive waters harmful to shelled life, slowdown and other changes in ocean currents, sea-level rise and widescale polar ecosystem disruption. Reversal of these impacts, especially after significant overshoot, will not occur on timescales relevant to societies, ecosystems, fisheries, biodiversity, and climate stability.

Declining sea ice coverage and thickness occurred in the Arctic well before global temperatures exceeded 1.0°C of warming; and Antarctic sea ice recently shifted from a highly variable pattern to a period of steep decline that today equals the Arctic in scale. Even if temperatures stabilize at 1.5°C, occasional ice-free conditions may occur in the Arctic in September (the end of the polar summer), while the thermal inertia of the Southern Ocean is expected to drive continued Antarctic sea ice decline for centuries even after global temperatures stabilize. Overshoot to 2°C will result in several months of very low Arctic sea ice conditions in the summer and frequent ice-free Septembers; temperatures towards 3°C will result in months of ice-free conditions each year and severely thinner and less extensive sea ice even in winter.

Decreased polar sea ice will carry high economic costs globally due to accelerated rise in global temperatures in the absence of reflective sea ice, greater ice loss from Greenland and especially Antarctica due to related collapse of protective ice shelves, changes in ocean currents, and impacts on sea ice-based ecosystems and human communities. While the impacts of overshoot on sea ice are probably more reversible than for other components of the cryosphere, recovery under global cooling will likely take decades to centuries due to the excess heat stored in the ocean, especially during longer and higher periods of overshoot.

The thawing of permafrost — frozen ground in high-latitude and high-elevation regions, including near-coastal seabed — also began well below 1°C. Such thaw today releases greenhouse gases annually on the scale of large industrialized nations, destabilizes the foundations of infrastructure and settlements built on frozen land, changes local and regional water cycles, and impacts vegetation including the intensity and frequency of wildfires. Permafrost thaw also increases the risk of mountain hazards such as the slope collapse which occurred in Nepal in August 2026, resulting in thousands of lives lost and economic damage equivalent to 10% of Nepal's economy in a single event. Every fraction of a degree of overshoot will cause more widespread permafrost thaw and related impacts. Although some permafrost may eventually refreeze if global temperatures decline, the carbon released from permafrost thaw is effectively irreversible and will further accelerate global warming. Overshoot also will fundamentally reshape hydrological networks in these regions for decades even after temperatures have begun to decline. During the overshoot peak, wildfire weather conditions are expected to intensify through longer fire seasons, more frequent lightning ignitions, and deeper burning of organic soils, which may take decades to centuries to recover. Any subsequent cooling will slow vegetation growth, reducing the amount of carbon that can be reabsorbed from the atmosphere into former permafrost soils.

A Call for Urgent Action

Temporarily exceeding 1.5°C does not mean that the goals of the Paris Agreement are irrelevant but instead makes them more urgent, specifically to avert or slow catastrophic and often irreversible global damage from cryosphere loss. Current scientific understanding shows that:

- Returning to 1.5°C or below by 2100 remains possible with sufficient political ambition and urgency, specifically by speeding the transition away from fossil fuel use to reach net zero reach net zero by mid-century by mid-century; together with sustainable carbon drawdown at scale.
- If such sustainable carbon drawdown continues, temperatures could decline further to 1.0°C or below by the middle of next century, with CO_2 levels dropping well below 400 ppm.
- Such urgent action can prevent and even reverse some cryosphere loss. Where changes are irreversible, due to Earth system feedbacks or the crossing of tipping points, urgent action can still significantly slow cryosphere loss and its global impacts.



Photo credit: J. Kirkham

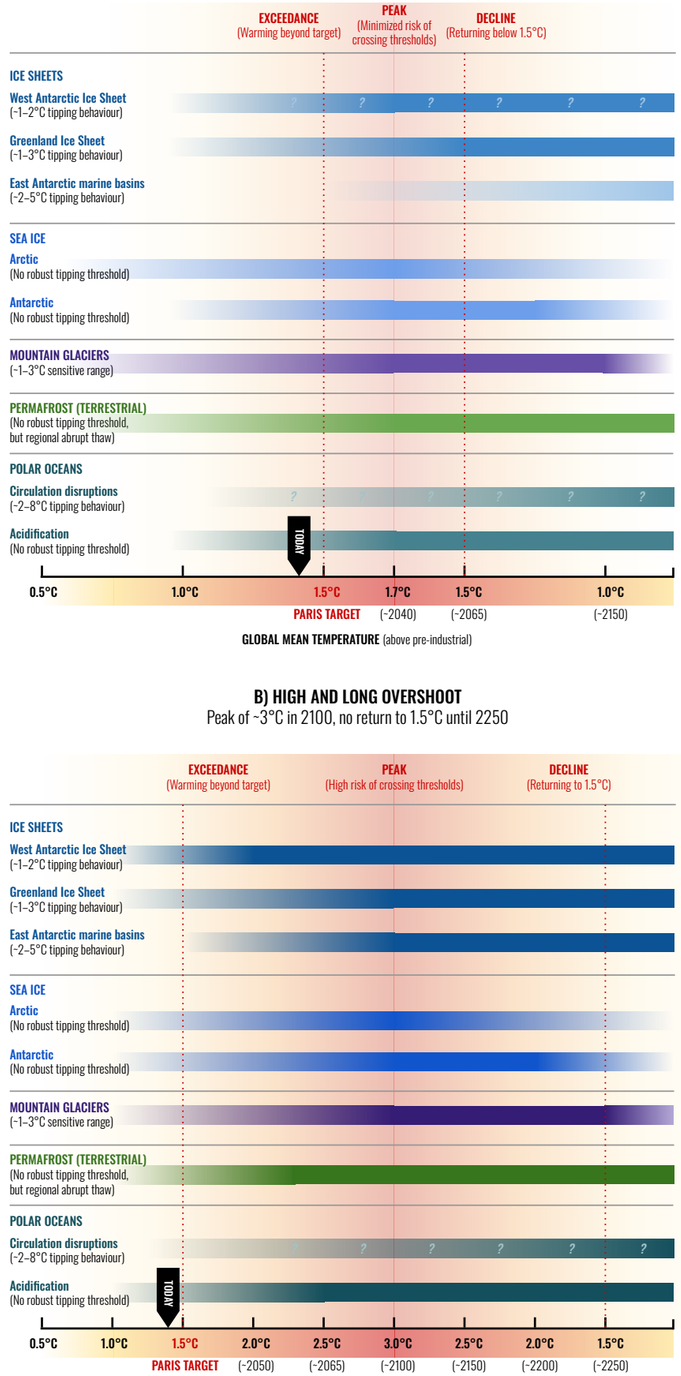


Figure SPM.1: ***Illustrative summary of the response of the cryosphere to (A) limited overshoot versus (B) a high and long overshoot.*** *Darker colours mean greater impacts, including higher risks of crossing thresholds and tipping points. Nearly all of these cryosphere elements began responding to warming at temperature levels between 0.5–1.0ºC above pre-industrial. Impacts from ice sheet loss, permafrost thaw, and polar ocean acidification are essentially irreversible on human timescales, especially with higher and longer overshoot. Illustrative overshoot, peak and decline pathways are shown in Figure 1.1. Note the different time and temperature scales between the two panels.*



01 Introduction

Overshoot, Peak, and Decline

Overshooting 1.5°C will result in large, and often irreversible, losses and changes to the cryosphere: Earth's frozen regions on land and at sea. This is because the cryosphere is one of the most sensitive and rapidly changing environments on Earth. Today, around 10% of Earth's land area is covered by glaciers or ice sheets. Permafrost underlies large areas of high latitude and high elevation regions. Sea ice covers vast areas of the Arctic and Southern oceans, and seasonal snow blankets over 30% of the Earth's land surface area each year [3].

The cryosphere keeps the planet in balance. The high albedo of its reflective white surfaces reflects solar radiation back out to space, helping to regulate the Earth's energy balance (the exchange of heat between Earth's surface and the atmosphere), with impacts on ocean and atmospheric circulation, regional weather patterns, and the transport and storage of heat, water and carbon around the globe.

Global warming is rapidly approaching 1.5°C, the lower Paris Agreement limit§. Over the last decade (2016–2025), global warming averaged 1.26°C above pre-industrial (1850–1900) levels[5], and individual calendar years are approaching and exceeding 1.5°C [4, 6].

Due to our collective failure to cut greenhouse gas emissions fast enough, it is no longer possible to prevent long-term global warming from exceeding 1.5°C above pre-industrial levels in the next few years without a period of temperature overshoot [2, 7, 8].

Rapid changes, extreme events, and abrupt shifts in environmental conditions are already occurring to the cryosphere even at today's levels of warming. These include the increasing loss of ice from the Greenland and Antarctic ice sheets [9, 10]; the retreat and disappearance of mountain glaciers around the globe[11]; declining sea ice extent at both poles [12-14]; widespread thaw and degradation of permafrost in high-latitude and high-elevation regions [15]; the acidification, warming and freshening of polar ocean waters; growing wide-scale changes to global ocean circulation and ocean currents in the polar regions [3]; and increasing heatwaves and temperature extremes [16, 17].

Cryosphere loss has global impacts. Loss of ice stored in glaciers and the vast ice sheets covering Greenland and Antarctica causes sea levels to rise, threatening low-lying coastal settlements and environments worldwide, especially island nations.

§ The Paris Agreement established a framework to limit the rise in global temperature to well below 2°C and to pursue efforts to limit warming to 1.5°C. These limits are relative to global mean surface temperatures during the pre-industrial period (1850–1900). The Intergovernmental Panel on Climate Change (IPCC) uses a 20-year average to define global warming levels, with the crossing time of particular warming levels as the mid-point of the 20-year period. The first calendar year above 1.5°C occurred in 2024, reaching 1.55°C above pre-industrial levels [4]. Although a single year above 1.5°C does not imply that the Paris Agreement limit has been formally breached, recent research suggests that the first year above 1.5°C is likely to occur within a 20-year period where average warming reaches that same level [2].

Changing meltwater supply from glaciers and snow affects the water resources that billions of people and downstream ecosystems rely on. Sea ice in the Arctic and around Antarctica helps to maintain the equator-to-pole atmospheric temperature gradient that drives global weather patterns, as well as the heat and salinity gradients of ocean waters that drive the global ocean circulation system. The thawing of permafrost releases additional greenhouse gas emissions that affects the global carbon cycle, destabilizes the foundations of infrastructure and settlements built on permafrost, alters water cycles, and increases the risk of hazards from ground and slope collapse. Finally, the polar oceans are vital in regulating Earth's climate by absorbing heat and carbon and acting as engines for global ocean circulation. This circulation redistributes heat and nutrients around the world, providing a critical foundation for marine food webs [3, 18]. The positive global benefits provided by the Arctic and Southern oceans have been valued at hundreds of billions of dollars each year [19, 20]. All of these essential planetary functions become increasingly at risk in a world that exceeds 1.5°C of global warming.

Exceeding 1.5°C does not render the goals of the Paris Agreement irrelevant; it makes them even more urgent. The recent Advisory Opinion of the International Court of Justice (ICJ) on the Obligations of States in Respect of Climate Change ruled that it "considers the 1.5°C threshold to be the parties' agreed primary temperature goal […] under the Paris Agreement" [21]. The pursuit of 1.5°C therefore endures as a legal and ethical imperative of the Paris Agreement, even in a period of overshoot [22]. Countries now need to pursue the Paris goals from above 1.5°C rather than below, with the aim of eventually bringing temperatures back to, or below, this level of warming.

Reversing global warming after exceeding 1.5°C is feasible. However, it will require significantly increased sustainable carbon drawdown from the atmosphere combined with an immediate intensification of efforts to reduce greenhouse gas emissions, supported by deep reductions in short-lived forcers including methane and (especially relevant to cryosphere) black carbon [8]. Urgent near-term emissions reductions are of the utmost importance to limit peak warming and keep the duration of overshoot as short as possible. These actions are vital to ensure that 1.5°C remains in reach, because current global carbon dioxide removal (CDR) capacity remains extremely limited [23]. Even with optimistic scaling, CDR may only lower temperatures by a few tenths of a degree this century [8], so limiting peak overshoot temperatures to no more than ~1.7–1.8°C is essential to keeping the 1.5°C goal within reach [24].

Returning global warming to or below 1.5°C requires action across three response phases. First, an immediate response where emissions are rapidly reduced to slow warming, while urgent adaptation measures are put in place to protect the most vulnerable communities. Second, a coping and containment phase where alongside measures to strengthen global resilience to climate change impacts, deep decarbonization results in net zero greenhouse gas emissions to stabilize temperatures at peak levels. Third, a long-term recovery phase where carbon drawdown brings temperatures back below 1.5°C, during which it remains essential that any persistent adaptation gaps are closed [8]. These actions will allow the world to implement an "overshoot peak-and-decline pathway," where temperatures: (1) first exceed 1.5°C, before (2) global warming peaks and plateaus; (3) then declines; and (4) eventually stabilizes below 1.5°C (Figure 1.1) [1, 25].

Limiting global losses and damages from the cryosphere however requires returning to well below 1.5°C, preferably to near or below 1.0°C as quickly as possible. The central challenge to global climate policy is therefore to limit the magnitude of peak warming as much as possible, keep the duration of overshoot as short as possible, and reduce global temperatures back to safer levels as quickly as possible [8].

However, current international climate policies and actions remain far off track. Current climate policies are set to raise global temperatures 2.8°C by 2100 (range 1.6–4.5°C) [7], and that only if fully implemented. Such sustained high temperatures will release devastating impacts from the cryosphere that will impact the lives and livelihoods of billions worldwide. Minimising overshoot is therefore essential to slow negative global impacts from cryosphere loss triggered by global warming, especially above 1.5°C.

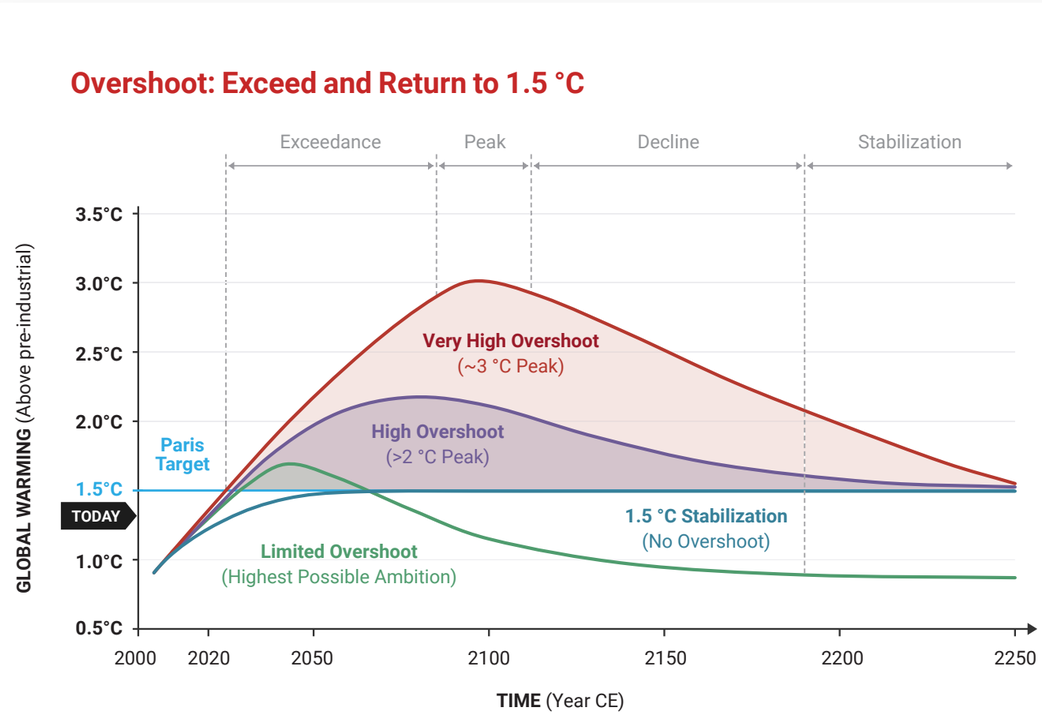


Figure 1.1: ***Illustrative depiction of "overshoot, peak and decline" pathways.*** *Three illustrative temperature overshoot scenarios are displayed that temporarily exceed 1.5°C before returning and stabilising near or below this level after several decades to centuries. The very high (red) and high (purple) overshoot peak and decline pathways are loosely based on the High-to-Low and Medium-to-Low ScenarioMIP-CMIP7 pathways [26]. The 'Highest Possible Ambition' pathway (green), where peak warming is limited to 1.7°C and declines below 1.5°C before 2100 and down to 1.0°C by 2150 is based on ref.[24]. A (now unachievable) reference pathway where global temperatures stabilize at 1.5°C global warming without overshoot is shown in dark blue.*

This Report

The consequences of overshooting 1.5°C requires urgent global attention. As part of a series of technical reports on Limiting Overshoot [8], in collaboration with the UN Environment Programme, the International Cryosphere Climate Initiative (ICCI) has coordinated this report on the risks and consequences of exceeding 1.5°C for the cryosphere. The following chapters summarize the global impacts of overshoot for five key components of the cryosphere and polar oceans, as well as what can still be saved by urgent and elevated global climate action to peak and then reverse global warming.

In Chapter 2, the report outlines the impact of overshoot on **long-term sea-level rise commitment from the Greenland and Antarctic ice sheets.** Chapter 3 examines the implications of **glacier and snowpack loss** (and potential regrowth) under overshoot for downstream water security, hazards in mountain regions, and ecosystems. In Chapter 4, the response and potential reversibility of **Arctic and Antarctic sea ice** is discussed. Chapter 5 describes the impacts of warming and overshoot on **permafrost**, including on the global carbon cycle, infrastructure, hazards, communities, the regional hydrological cycle and on ecosystems and wildfires. In Chapter 6, the response of the **polar oceans** to overshoot is described through its effects on ocean warming, freshening of ocean waters, acidification, and large-scale modifications to global ocean circulation. Chapter 7 discusses the **interactions and feedbacks in cryosphere systems** that will occur under overshoot pathways. The report concludes with a discussion of **key policy actions and research priorities** that are urgently needed to minimize and prepare for the impacts of cryosphere loss in this new era of overshoot (Chapter 8).

Photo credit: L. Hösli

02 Polar Ice Sheets

Executive Summary

The Greenland and Antarctic ice sheets are highly sensitive to every fraction of a degree of warming. The higher the warming and the longer it continues, the greater the long-term sea-level commitment. Limiting global warming to 1.5°C, even after a brief overshoot, will not prevent future ice-sheet retreat and long-term sea-level rise in the order of 0.5–1 metre by 2100 and 2-9 metres once Earth's ice sheets have fully adjusted to this new global temperature. It will, however, likely slow that full committed loss to occur over several centuries to millennia by reducing the atmospheric and oceanic warming that drives ice loss and sea-level rise, especially from vulnerable sectors of Greenland and Antarctica. Every year above 1.5°C increases the risk of triggering ice sheet instabilities that may lead to earlier and more rapid sea-level rise. If such instabilities are triggered, especially at higher peak overshoot temperatures, rates of sea-level rise could reach several centimetres per year, well beyond adaptation limits of most coastal communities and ecosystems. Regardless of its ultimate level and rate, sea-level rise from ice sheet loss will be irreversible on millennial timescales, even if temperatures later stabilize or even cool. Indeed, current scientific evidence suggests that a 'safe' temperature limit for today's ice sheets lies much closer to 1.0°C, and possibly even lower.



Photo credit: J. Kirkham

Suggested citation (this chapter): Colleoni, F., Garbe, J., Ghomsi, F. E. K., Li, D., Stokes, C. R., and Zhao, C. (2026). Polar Ice Sheets. In: Overshoot and the Cryosphere: Averting Global Meltdown. The Implications of Temperature Overshoot, Peak and Decline for the Cryosphere [Kirkham, J.D., Pearson, P., Garbe, J., Hancock, S., Imdieke, A. (eds.)]. International Cryosphere Climate Initiative. DOI: 10.5281/zenodo.23477467

Background

Ice loss from the Greenland and Antarctic ice sheets is a major contributor to accelerating sea-level rise and will increasingly determine the long-term risks faced by coastal communities and low-lying nations. While historically the thermal expansion of the oceans as well as the melting of mountain glaciers were the dominant drivers of sea-level rise, mass loss from both ice sheets, which together hold enough ice to raise global sea-level by 65 metres, has increased in recent decades [9, 10, 27]. Based on detailed satellite observations, these ice sheets are now losing mass about six times faster than in the 1990s, with hundreds of billions of tonnes of ice lost each year and further losses already committed [10, 28, 29]. This acceleration of ice loss, combined with the potential for feedbacks and instabilities, leads to highly uncertain sea level outcomes for coastal communities towards the end of this century and beyond, making adaptation planning difficult [30].

In Antarctica, interaction with warm ocean waters has driven the thinning, acceleration, and mass loss of glaciers in West Antarctica, the Antarctic Peninsula, and parts of East Antarctica (Figure 2.1;) [9, 10, 31, 32]. Greenland, by contrast, loses a substantial portion of its mass due to atmospheric warming. Approximately half of the mass lost from the Greenland Ice Sheet is related to surface melting, with the remainder lost via marine-terminating outlet glaciers through processes including iceberg calving, ice margin retreat, and increased ice discharge [33-35] (Figure 2.1).

Both the Greenland and Antarctic ice sheets are sensitive to self-reinforcing feedbacks that amplify their response to climate change [27, 29, 36]. In Greenland, ice loss is enhanced by the melt–albedo feedback, in which increased surface melting reduces the reflectivity of snow and ice, allowing the surface to absorb more solar radiation and promoting further melt.

This surface darkening can be driven by melt itself, as well as by the storage of meltwater in supraglacial lakes, the deposition of black carbon and dust, the growth of surface algae, and/or rainfall [37, 38]. Another important mechanism is the melt–elevation feedback. In this process, as the ice sheet loses mass, its surface lowers into warmer elevations, further enhancing surface melting and accelerating ice loss [39-42]. This process also takes time to unfold and 'catch up' with present climate and, as a result, Greenland is already committed to contribute at least a further 27 cm to future sea level rise relative to 2019 levels, irrespective of any further warming [28]. That sea-level rise commitment grows with warming, with some suggesting that Western Greenland is approaching a critical threshold or 'tipping point' [43].

For Antarctica, concerns arise primarily in sectors where the ice sheet rests on ground that is much deeper than sea level; the case for much of the West Antarctic Ice Sheet (WAIS; 4.3 m total sea-level equivalent (SLE), of which 3.4 m is marine based), large sectors of the East Antarctic Ice Sheet (EAIS, of which 19.2 m SLE is marine based) and parts of the Antarctic Peninsula (0.1 m SLE is marine based) [44]. In those marine-based sectors, outlet glaciers extend out into the ocean to form ice shelves that are sensitive to both atmospheric and ocean warming. Ice shelves play a critical stabilizing role by exerting a buttressing force that restrains the flow of inland ice towards the ocean [45, 46]. The intrusion of warm ocean water below the ice shelves enhances basal melting, thinning the ice shelves and weakening their buttressing effect [47, 48]. Ice-shelf thinning can trigger grounding-line retreat on beds that slope towards the ice sheet interior, leading to a process known as the Marine Ice Sheet Instability (MISI) – an amplifying feedback that drives accelerating ice discharge, retreat, and long-term sea-level rise [49, 50]. Areas where ice is only lightly grounded on a relatively flat bed may also be prone to very rapid retreat [51].

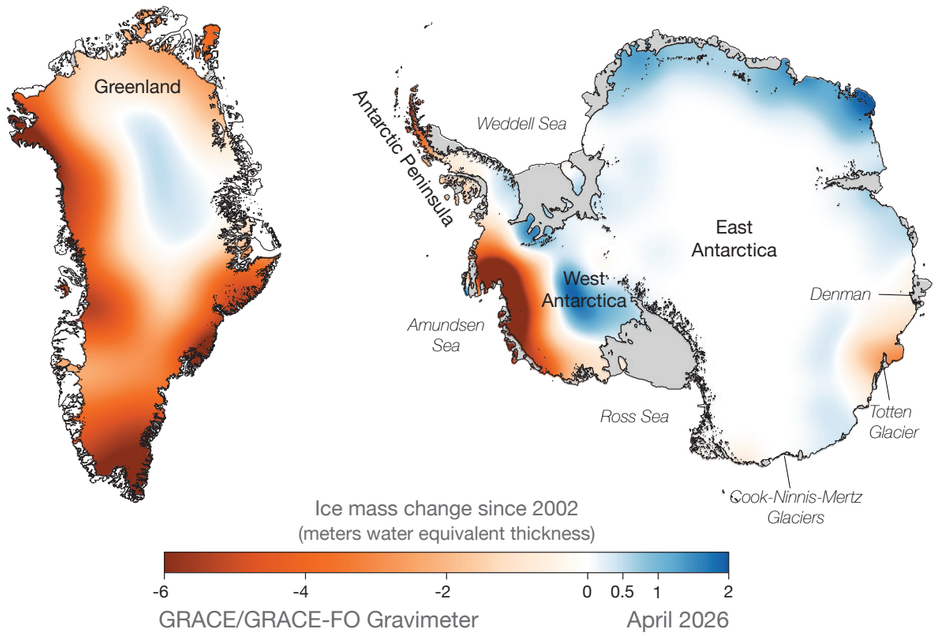


Figure 2.1: ***Spatial distribution of Greenland and Antarctica land ice mass changes since 2002 as of April 2026 from satellite gravimeter GRACE/GRACE-FO [110].*** *Blue colours indicate a mass gain, while orange to brown colours indicate a mass loss. Ice shelves are indicated in grey, and regions mentioned in the text are labelled.*

A further mechanism that could dramatically accelerate ice loss, known as the Marine Ice Cliff Instability (MICI), has also been proposed [52, 53]. Under MICI, if the buttressing ice shelves surrounding the Antarctic Ice Sheet are lost, tall and potentially unstable ice cliffs may be exposed around the ice-sheet margins that could fail rapidly.

Ice-sheet sectors grounded in basins more than 1 km below sea level would be particularly susceptible to this process, but whether retreat becomes self-sustaining depends on local stabilising factors such as bed topography, ice-shelf buttressing, mélange, and lateral drag [54-56]. Consequently, it is unlikely that MICI will play a role in WAIS dynamics in the 21st century [57], but the role of MICI in driving future WAIS retreat cannot be ruled out.

Consequences of Overshooting 1.5°C

The response of the Greenland and Antarctic ice sheets to different levels of sustained global warming is informed by both palaeo-records from past warm periods and numerical modelling of future change [41, 58]. Geological records indicate that global temperatures around 1.5°C above pre-industrial levels or even lower, when sustained over centennial to millennial timescales, have been sufficient to reduce the volume of the Greenland and Antarctic ice sheets substantially, raising sea level by several metres [27, 59, 60].

The IPCC [18] has assessed the likely increase in global mean sea level (GMSL) that would be committed at various levels and durations of warming (Figures 2.2 and 2.3). Put simply, the longer any given warming level is sustained, the higher the long-term sea level commitment. At +1.5°C, for example, ice sheets are likely to commit around 0.5 to 1 m to GMSL over a 100-yr timescale (Figure 2.3). This increases to around 2-3 m over a 2000-yr timescale, and around 6-7 m over a 10,000-yr timescale, consistent with palaeo-ranges of 5-10 m during the Last Interglacial (around 129,000 to 116,000 years ago).

At +2°C, these values increase to around 2-6 m over a 2000-yr timescale and 8-13 m over 10,000 years. Warming of +3°C would commit around 4-10 m over a 2,000-yr timescale and around 10-24 m over a 10,000-yr timescale, which is consistent with palaeo-records from the Mid-Pliocene Warm Period (2.9 to 3.3 million years ago). At +4°C, a commitment of 12-16 m over 2,000 years and 19-33 m over 10,000 years is expected, with +5°C of warming committing 19-22 m over a 2,000-year period and 27-37 m over 10,000 years[18]. For comparison, a sea-level rise of just 2 m is estimated to displace around 650 million people based on the present-day population distribution [61].

The Last Interglacial provides an important analogue for ice-sheet sensitivity to relatively low levels of global warming. During this period, atmospheric CO_2 concentrations remained below 300 ppm (~280 ppm) and global mean surface temperatures were about 1°C warmer than during pre-industrial times (see Table 1 in ref. [27]). Nevertheless, geological and glaciological evidence point to a global mean sea level about 2 to 9 m higher than today [27, 41, 59]. Identifying which ice sheet, or regions of an ice sheet, contributed to the rise in sea level is more challenging, especially considering that the ice sheets of Greenland and Antarctica may have responded asynchronously [62, 63]. The central portion of Greenland remained largely intact, but the southern region deglaciated, contributing at least 1 m to sea level and possibly more [27]. In West Antarctica, seaways may have opened between the Ronne, Amundsen and Ross embayments during the Last Interglacial, following a major retreat or partial collapse of parts of the ice sheet [64, 65]. A recent review of the evidence suggests a maximum Antarctic contribution of 4-6 m, but the precise configuration of the collapsed ice remains unresolved [66].

Greenland, and parts of Antarctica, also responded strongly to past intervals of sustained warmth that were driven by slow changes in Earth's orbit [67-69]. Sediments recovered from beneath the Greenland Ice Sheet indicate that substantial retreat occurred during Marine Isotope Stage 11, around 400,000 years ago, when atmospheric CO_2 concentrations were close to 280 ppm (current levels as of 2026 are around 427 ppm) but warm conditions persisted for tens of thousands of years. These records suggest that Greenland contributed at least 1.4 m to global sea level during that period [69]. Together, these palaeoclimate constraints demonstrate that the long-term stability of both ice sheets cannot be inferred from temperature alone, but depends strongly on how long a given level of warming is sustained and whether non-linear processes are triggered during a period of warmth.

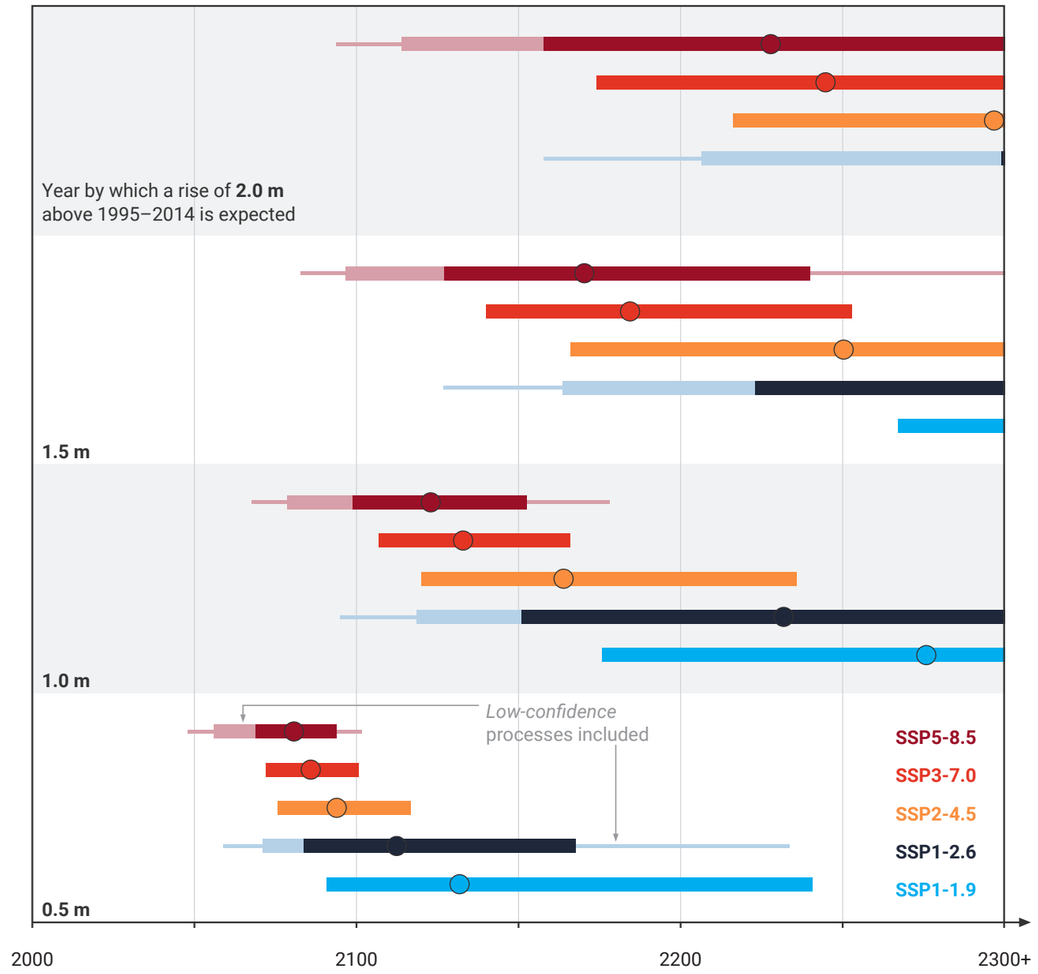


Figure 2.2: ***Timing of exceedance of global mean sea level thresholds of 0.5, 1.0, 1.5 and 2.0 m, under different SSPs.*** *Lightly shaded thick/thin bars show 17th–83rd/5th–95th percentile low-confidence ranges for SSP1-2.6 and SSP5-8.5. Figure is adapted from the IPCC AR6 WGI Technical Summary for Policymakers [18].*

Although the geological record provides important constraints on the magnitude of past ice loss, the rate at which this loss occurred remains poorly understood. This distinction is crucial because the impacts of sea-level rise depend not only on the eventual amount of committed ice loss, but also on how quickly that loss unfolds. Very rapid rates of sea-level rise, for example above 10 mm yr^{-1} (i.e., 1 m per century) would be very challenging to accommodate, especially if they occurred more abruptly than envisioned (e.g., through marine ice sheet instability) because many coastal planning and protection measures require decades to implement effectively [70]. Meanwhile, it has been noted that the rate of sea level rise has doubled over the last three decades [71] and, if the recent acceleration were to continue, sea-level rise could approach 10 mm yr^{-1} by the end of this century, although long-term extrapolations are more uncertain. Furthermore, the palaeo-record indicates that sea-level rise can reach rates up to 4 times higher. Rapid ice sheet collapse during the last deglaciation is thought to have generated rates as high as 35-40 mm yr^{-1} around 14.6 thousand years ago (i.e., 3-4 centimetres per year; up to 4 m per century) but this also involved the deglaciation of now extinct ice sheets that covered large parts of North America and North-West Europe [72].

Often informed by these palaeo-records, numerical modelling projections increasingly suggest that substantial ice-sheet loss could occur at temperatures that may be reached this century. For Greenland, previous studies estimate that a threshold for irreversible surface melting may be crossed at approximately 1°C to 3°C of warming above pre-industrial levels, potentially committing Greenland to long-term ice loss [39, 73, 74]. Several studies place the onset of large and irreversible Greenland ice loss below 2°C of warming, and possibly closer to 1.5°C, although estimates vary across models and depend on assumptions about atmospheric forcing, surface mass balance, and ice-sheet geometry [36, 39, 75].

Under very high emissions (RCP8.5), some models predict that most of Greenland could be mostly ice-free in less than 1000 years [76].

The vulnerability of the Antarctic Ice Sheet varies strongly among drainage basins [77-80]. A warming threshold as low as 1–2°C above pre-industrial levels could lead to the long-term irreversible loss of a large fraction of marine-based ice in West Antarctica [80]. Marine-based sectors in East Antarctica, representing ~5 m of potential sea-level rise, are at risk of losing stability at 2–5 °C, where tipping thresholds could be crossed in relation to MISI [77, 80, 81].

The WAIS is considered especially vulnerable because much of it is grounded below sea level, making it sensitive to irreversible ocean-driven retreat at warming levels close to, or even below, 1.5°C of warming above pre-industrial levels [36, 60, 77]. Recent modelling suggests that large parts of West Antarctica may already be committed to future long-term loss even under today's warming levels of approximately 1.3°C above pre-industrial [77, 80, 82-84]. In addition, recent coupled ocean and ice-sheet modelling indicates that committed ocean warming from past and ongoing fossil-fuel emissions will continue to drive extensive melting beneath West Antarctic ice shelves even under very low-emission trajectories [85, 86]. Overall, the Antarctic Ice Sheet is expected to continue to retreat throughout the twenty-first century, losing mass even if aggressive emissions reductions are implemented [87]. Reversing this retreat would require global cooling substantially below the temperature level at which the initial retreat was triggered, and in some cases potentially below pre-industrial conditions [77, 82].

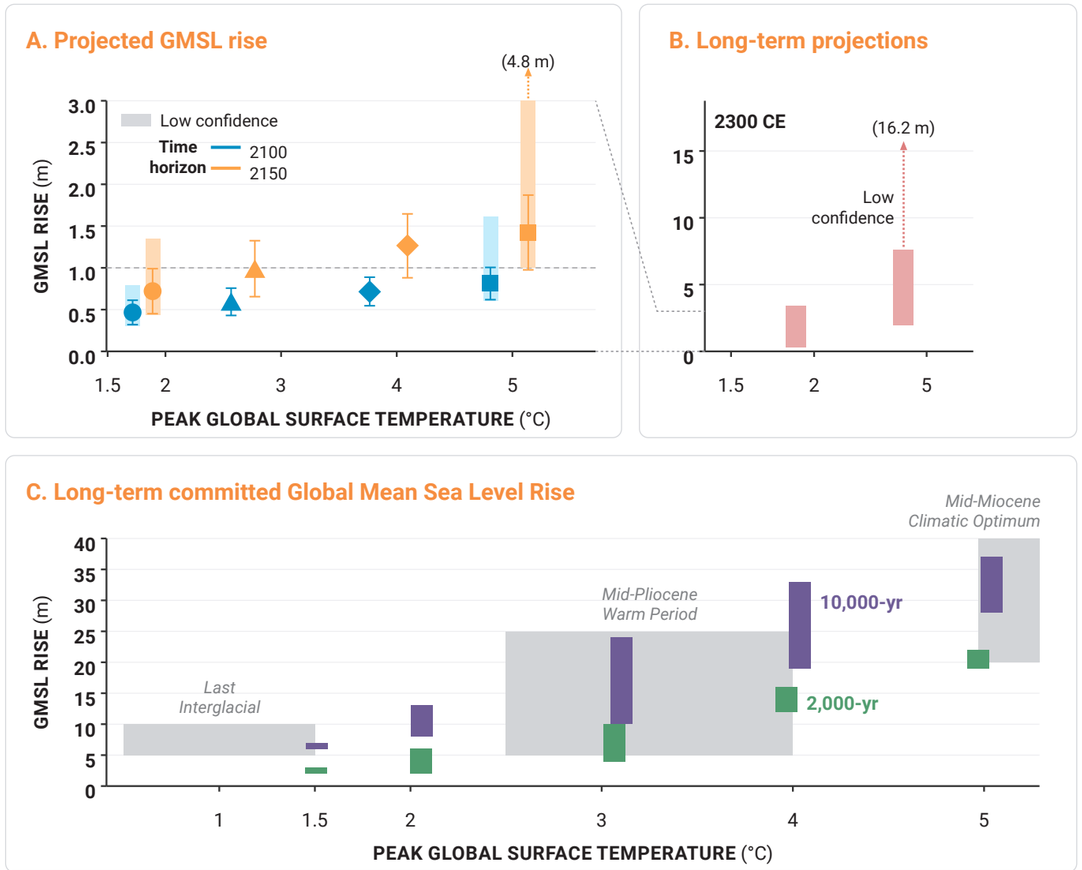


Figure 2.3: ***A.*** *Projected global mean sea level (GMSL) rise in 2100 and 2150 for four emission scenarios, modified from Arias et al. [18]: SSP1-2.6 (~1.5-2°C), SSP2-4.5 (~2.5-3°C), SSP3-7.0 (~2.8-5.5°C), and SSP5-8.5 (~4.4-5°C). Medium confidence projections (symbols) and low confidence projections (bars) are shown in the 17th-83rd percentile range. Low-confidence projections accounts for structured expert judgement [111] and one model study including MICI processes for Antarctica [53]. Data are IPCC AR6 Chapter 9 [41].* ***B.*** *Projected long-term GMSL rise in 2300 CE from low-confidence scenarios for SSP1-2.6 (~1.5-2°C) and SSP5-8.5 (~4.4-5°C). Bars correspond to the likely range of projections while the dashed arrow corresponds to projection from ref.[53] with Antarctic contribution alone up to 14 metres. Data are from IPCC AR6 Chapter 9 [41].* ***C.*** *Committed GMSL rise on multi-centennial to multi-millennial scale adapted after IPCC AR6 Technical Summary [18]: on 2,000-year (green) and 10,000-year (purple) time scales as a function of global surface mean temperature, relative to 1850–1900. For longer-term commitments, warming is indexed by peak warming above 1850–1900 reached after cessation of emissions. Grey shades show palaeo'-constraints on global surface temperature and GMSL for the Last Interglacial (~130 - 116 ka), Mid-Pliocene Warm Period (~3.3 - 3Ma) and for the Mid-Miocene Climatic Optimum (~14 Ma). For this latter period, the reconstructed global warming exceeds the temperature range of the graph (+5-9°C), but reconstructed sea level range is +20-40 m [112]. Note the different GMSL scales in all three panels.*

Reversibility After Overshoot

The consequences of temporarily exceeding and then returning below a given temperature level depend not only on the magnitude of warming, but also on the duration of the overshoot, the final temperature upon decline, and the response timescales of different sectors of the ice sheets [88]. While surface mass loss and outlet glacier acceleration respond on much shorter timescales and are contributing to the recent acceleration in sea-level rise, the full committed response of the polar ice sheets is likely to unfold over centuries to millennia [77, 89-91]. This means that a temperature threshold may be temporarily exceeded without resulting in a full change of system state, if the period of overshoot is short compared to the effective response timescale of the ice sheets [88, 92]. However, if warming is sustained for long enough above a critical level, tipping dynamics within the ice sheet system may be triggered, locking in irreversible ice loss for centuries to millennia, even if temperatures are stabilized or even reduced [39, 53, 77, 90, 91]. However, particularly for Antarctica, it is important to note that even a decadal-scale period of warming might be sufficient to nudge the ice sheet into MISI that could then unfold for several centuries, even if temperatures are subsequently cooled [86, 89]. This makes both the peak temperature and duration of any temporary temperature overshoot particularly important for ice-sheet stability [53, 73, 75, 91].

Modelling suggests that exceeding a critical threshold for ice sheet stability does not necessarily imply complete or irreversible loss of the Greenland Ice Sheet [73]. In some simulations, reducing the climatic forcing after an overshoot allows at least partial regrowth of the Greenland Ice Sheet over millennia, indicating that some fraction of the ice-sheet response may remain reversible under sufficiently rapid and sustained cooling [73, 91, 93-95]. However, reversibility is not guaranteed.

In some overshoot scenarios, atmosphere–ice-sheet feedbacks, particularly the surface mass balance–elevation feedback, can inhibit regrowth once the ice-sheet surface has lowered into warmer atmospheric layers [36, 73, 94, 95]. Conversely, increased snowfall in a warmer climate [96], or greater incidence of large amounts of precipitation from atmospheric rivers [97], may help to dampen Greenland mass loss feedbacks.

Only a limited number of modelling studies have explicitly simulated the response of the Antarctic Ice Sheet to global warming overshoot scenarios [98, 99] and/or the impacts of delayed global mitigation action [53]. These studies suggest that the long-term consequences of overshoot may be particularly severe for marine-based sectors of Antarctica such as the WAIS (~3.5 m sea-level equivalent) [44] and parts of the EAIS (~19 m marine-based SLE), where grounding-line retreat into deeper basins can become self-sustaining owing to the MISI. The time required to trigger large-scale ice sheet retreat after crossing a critical temperature threshold is currently highly uncertain, with possible onsets ranging from decades to centuries to millennia into the future [53, 83, 100]. Accordingly, it may be possible to limit the extent of committed retreat if temperatures were to decline sufficiently quickly below the critical temperature threshold, but how quickly this would need to be achieved is currently poorly understood and will depend on the relative importance of atmospheric versus ocean warming. Whilst air temperatures are expected to stabilize within decades of reaching net zero greenhouse gas emissions [101, 102], the ocean will experience a delayed response that will unfold over centuries owing to thermal inertia [103, 104]; increasing the likelihood that ice sheet tipping points are crossed. Like Greenland, parts of Antarctica are likely to receive increased snowfall in a warming climate, especially East Antarctica [105, 106], but this is unlikely to compensate for the substantial mass loss that would unfold if the same atmospheric warming caused ice shelves to collapse through enhanced surface melting and instabilities were triggered.

These results imply that returning global mean temperature below a given level after a period of overshoot may reduce the magnitude of centennial to multi-millennial sea-level rise from ice sheet mass loss and thus decrease the associated long-term risks. However, while some of the most severe potential outcomes, such as the complete loss of the Greenland or West Antarctic ice sheets, may be avoided if global mean temperature is rapidly reduced below 1.5°C after a short period of overshoot [73, 99], every additional decade at which temperatures remain above the critical threshold will lead to additional sea-level rise and associated impacts, and will increase the risk of triggering large-scale ice retreat.

Even without additional warming above today's level, Greenland and Antarctica are expected to contribute several metres to sea-level rise over the coming centuries to millennia [18, 27]. Preserving long-term ice-sheet stability may therefore require global mean temperatures to be reduced substantially below present-day levels, potentially towards 1°C above pre-industrial or lower [27, 107, 108]. In overshoot scenarios, lowering temperatures after the peak would reduce the thermal forcing on both ice sheets and thereby limit subsequent ice loss (Figure 2.4). Every additional century at which temperatures exceed 1.5°C is predicted to add an additional 40 cm to the median sea-level rise commitment by 2300 compared with non-overshoot conditions [98, 109]. Ultimately, the consequences of the overshoot are unlikely to be fully reversible. Long-term sea-level rise may not be avoided, but its magnitude and rate can still be reduced.

Photo credit: J. Kirkham

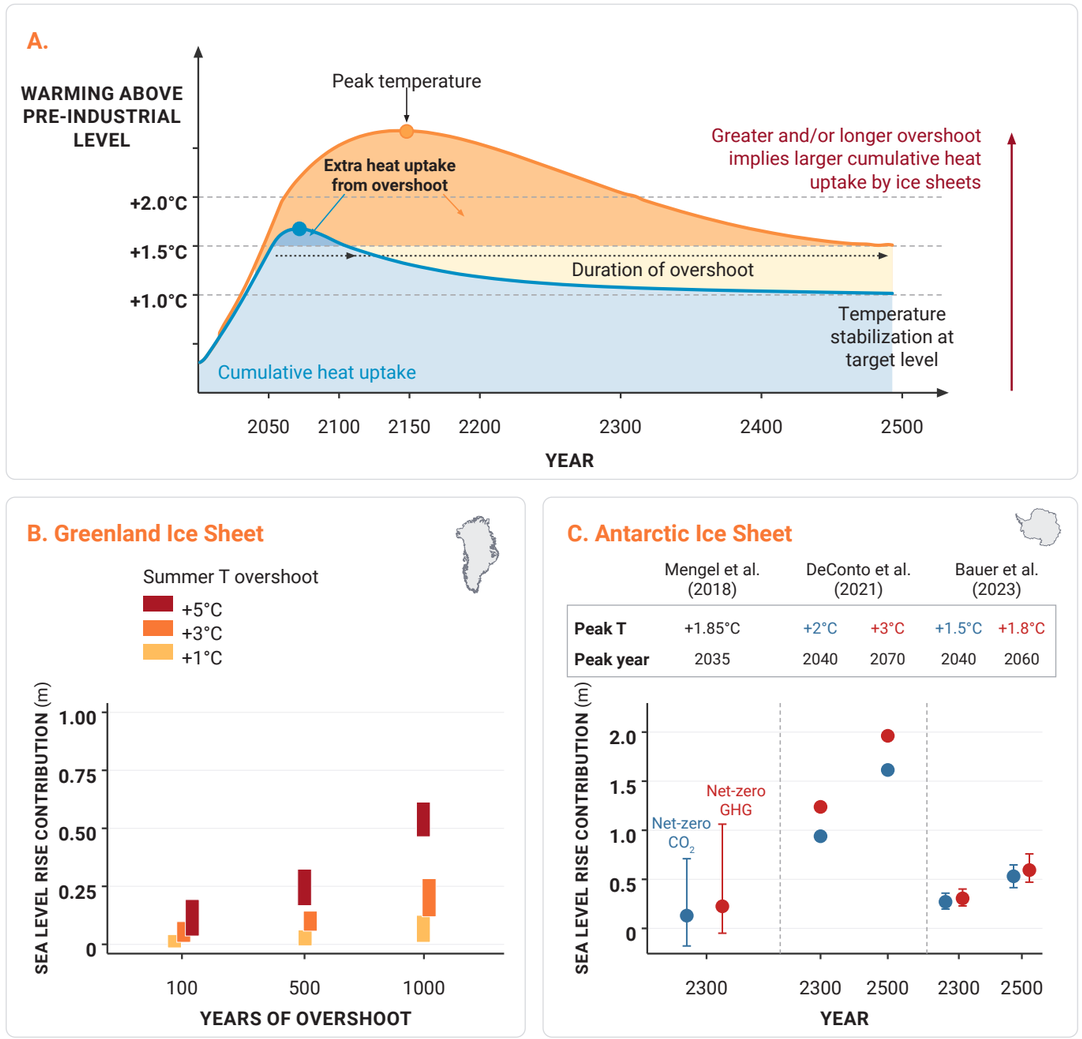


Figure 2.4: ***idealized ice-sheet responses to temporary overshoot of 1.5°C. A.*** *Conceptual representation of an overshoot, peak and decline pathway. Global temperature temporarily exceeds a target level before declining and stabilising at a lower level. The magnitude of the peak warming and the duration of the overshoot determine the additional cumulative heat taken up by the climate system and, consequently, the long-term response of the ice sheets.* ***B.*** *Maximum simulated Greenland Ice Sheet contribution to global mean sea-level rise under regional summer-temperature overshoots of +1°C, +3°C and +5 °C above present over different overshoot durations (based on ref. [73]). Vertical ranges span the simulations run with two different ice sheet models.* ***C.*** *Antarctic Ice Sheet contributions under overshoot pathways assessed by ref.[98], ref.[53] and ref.[99]. Ref. [98] values correspond to the median across nine net zero CO_2 or net zero greenhouse-gas pathways and associated 17th–83rd percentile intervals. Ref. [99] uncertainty ranges reflect sensitivity to upper and lower climate model temperature responses used to force the ice sheet model. T — temperature.*

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03 Mountain Glaciers and Snowpack

Executive Summary

The accelerated loss of glaciers and snowpack, especially under higher and longer overshoot of 1.5°C, will drive growing environmental hazards for populations situated in and downstream of mountain regions. These range from water insecurity and drought to sudden disaster events such as glacial lake outburst floods or ice/rock avalanches that may surpass the limits of adaptation and disaster risk reduction measures. Even temporary overshoot of 1.5°C will irreversibly decrease snow and ice reserves on human timescales. Higher and longer peak warming will result in extreme loss, with glaciers disappearing almost entirely in some regions as temperatures approach 3°C and rapid changes in water availability that could push some communities beyond the limits of adaptation. A return to well below 1.5°C could initiate some glacier regrowth on time scales of decades to centuries if accompanied by sufficient snowfall recovery. Although declines in snowfall under overshoot are theoretically quicker to reverse than glacier loss with cooler temperatures, changes in precipitation patterns after return from overshoot are complex and remain a challenging research gap. Even if glaciers and snow can be partly restored with declining temperatures, the loss of available downstream water is expected to persist after higher overshoot compared to minimal or non-overshoot scenarios. This would disrupt populations, industries and economic sectors, and mountain and downstream ecosystems reliant on meltwater from glaciers and snow for centuries.



Photo credit: M. Huss

Suggested citation (this chapter): Kang, S., Kaser, G., Llambi, L. D., Maharjan, A., Marshall, A., Schuster, L., Tielidze, L., and Zhang, Q. (2026). Mountain Glaciers and Snowpack. In: Overshoot and the Cryosphere: Averting Global Meltdown. The Implications of Temperature Overshoot, Peak and Decline for the Cryosphere. [Kirkham, J.D., Pearson, P., Garbe, J., Hancock, S., Imdieke, A. (eds.)]. International Cryosphere Climate Initiative. DOI: 10.5281/zenodo.23477467

Background

Mountain cryosphere provides critical services to nearly two billion people who live both locally and downstream of these regions [113]. Often referred to as the “water towers” of the world, mountain glaciers and snowpack provide water for drinking, agricultural irrigation and hydroelectricity as well as transportation [114]. They also foster local habitats and ecosystems [115], support tourism, and serve as a source of cultural heritage, placed-based connection, and spirituality for Indigenous Peoples, local communities, groups and individuals [116, 117]. Loss of mountain cryosphere therefore has cascading impacts on the food, water, economic, social and political security of people both within and far outside of mountain regions [118].

This chapter encompasses seasonal snow and alpine glaciers, which share a close relationship. Snowfall in the accumulation zone of glaciers is essential for nourishing and sustaining their mass balance, as the accumulated snow is buried and compressed over many years to form glacial ice. The balance between snow accumulation and the amount of ice melt determines the overall health of a glacier. Even small snowfall events can heavily reduce the melting of glaciers during the melt season due to snow’s high albedo. Snow that does not fall on glaciers contributes to seasonal snowpack that accumulates and melts annually. Such seasonal snow is the largest single component of the cryosphere by area and responds much faster to climatic changes than glaciers. Permafrost is also a crucial component of the high mountain cryosphere, especially through its role in maintaining the stability of mountain slopes, and is discussed in more detail in Chapter 5.

Many of the world’s mountainous regions are warming faster than the global average [119-121]. This means that even if planetary warming were to be limited globally to 1.5°C by 2100, the temperature increase in mountain regions is likely to exceed 1.5°C of warming. For example, global warming of 1.5°C translates to between 1.8-2.2°C of average warming in the Hindu Kush Himalaya [122, 123].

Global glacier mass has declined on average by 5% in the first two decades of this century [11, 124] (2000-2023), and losses have been much greater in regions such as the European Alps (39% ice mass lost), the Caucasus (35%), the Southern Alps of New Zealand (29%), North Asia (23%), Rocky Mountains of North America (23%), and the Tropics (20%) [11]. Critically, more than 40% of mass loss recorded since the mid-1970s has occurred in the last decade, and six of the highest mass-loss years ever were recorded between 2018-2025 [125]. Global declines in seasonal snow have also been observed, particularly at lower elevations [126]. These long-term seasonal snow declines [127, 128] as well as episodic “snow droughts” [129-131] have been attributed to anthropogenic climate change [132].



Photo credit: M. Huss



Photo credit: M. Huss

Glacier and snow loss create an increased risk of environmental hazards, and impact dependent species and ecosystems, such as alpine wetlands that depend on glacial runoff [133]. Glacier retreat and permafrost thaw can destabilize hillslopes and lead to the development of lakes which can result in glacial lake outburst floods, ice and rock avalanches, and landslides from over steepened mountain slopes. These events pose extreme risks to downstream infrastructure, economies and communities exposed to such hazards, as demonstrated by recent events in the Nepal Himalaya in 2026 [134], Switzerland in 2025 [135, 136], the Caucasus in 2023 [137] and the Indian Himalaya in 2021 [138]. Similarly, a lack of snow cover can make land more vulnerable to drought [139, 140], wildfires [141-144], agricultural risks [145, 146], and disrupted ecosystem function [147]. Snow also helps to regulate Earth's energy balance (the exchange of heat between Earth's surface and the atmosphere) and impacts regional weather patterns. Additionally, glacier losses can pose governance-related challenges associated with fragmented regulatory frameworks, institutional incoherence, and spatial planning instruments that may not yet be in place to manage the governance of new spaces, landscapes and postglacial ecosystems that are emerging as glaciers retreat [148-150].

Some glaciers are sentient beings for many Indigenous Peoples and local communities across the Americas, East Africa and the Hindu Kush Himalaya, and the loss of glacial ice affects the cultural identity, spirituality and heritage of these communities [108, 151-153]. The economic and social well-being of people and many animal and plant species depend on seasonal snow cover. As snow recedes, for example, yak herders in India and Nepal face challenges maintaining traditional migration timings, routes and campsites [154] due to increased water shortages at high elevation camps or rain-on-snow events, with higher animal death from starvation and terrain instability [155].

Consequences of Overshooting 1.5°C

Glaciers

Every fraction of a degree temperature rise matters for glacier preservation [156]. Even with strong mitigation of greenhouse gas emissions, all mountain regions of the world will experience considerable glacier loss in the coming decades as these ice stores continue to adjust to a warmer climate. This is because most glaciers globally are currently out of balance even with today's climate and will continue to shrink even if warming were to cease immediately [157, 158]. Under current policies, global temperatures are set to reach 2.8°C by 2100 (range 1.6–4.5°C) [7]; if this temperature is maintained over centuries, globally only ~25% of today's glacier ice is predicted to remain [158]. However, the amount and timeline for this loss varies greatly by region. While massive polar glaciers (excluding the Greenland and Antarctic ice sheets) are projected to take centuries to reach a new balance with the climate, mountain glaciers in mid-latitude and tropical regions would decline far more rapidly, experiencing 80% of their forced loss within decades [158] (Table 3.1).

Limiting global temperature rise to 1.5°C above pre-industrial levels will not prevent further committed ice loss, but will allow around 54% of current global glacier ice to be retained – around twice as much as would be left under current policies [158]. A world with minimal overshoot and return to 1.5°C this century is also crucial to ensure that glaciers will still exist in some regions of the world (Table 3.1).

For example, in Scandinavia and western North America, even long-term warming levels of 2.0°C (well below current policies) would result in near-total ice loss in these regions, whereas an emissions trajectory in line with 1.5°C would preserve at least ~20% of today's ice in these locations. Should temperatures approach 3.0°C of warming, even the higher altitude Central and Eastern Himalayas are projected to retain just 10% of their ice mass once glaciers fully equilibrate, compared to around 35% under a 1.5°C emissions scenario [158]. In the Western Himalaya, Hindu Kush and Karakoram regions, which in recent decades were near stable [11], only 60% of the glacier mass would remain in a 2.0°C future, but 85% under a 1.5°C pathway [158].

Under a 1.5°C emissions pathway, the global rate of glacier extinction would peak in the 2040s with around 2000 glaciers disappearing annually before the rate of extinction slows. In contrast, under very high emissions (4.0°C in 2100) up to 4000 glaciers would vanish annually at the peak rate of extinction in the mid-2050s [159]. In addition to implications for water security, the disappearance of individual glaciers has impacts on culture, tourism, spiritual practices and communal identity [116, 160, 161].

A very low emissions pathway is therefore essential to safeguard glacier-dependent resources (both economic and cultural), and ecosystems and biodiversity – especially given that specialized cold-adapted species already face severe habitat loss and local extinctions [148, 162]. Low emissions are also essential to minimize the risk of severe hazards such as glacial lake outburst floods and ice/rock avalanches related to ice loss [163-165].

Snowpack

Warming trajectories that exceed 1.5°C will decrease snowfall and seasonal snowpack. Annual snow accumulations are determined in large part by air temperature and precipitation conditions [166, 167], although other factors such as humidity [168] and snowfall intensity and frequency [169] also play a role. Because seasonal snow melts annually, it responds much more quickly to different warming levels than mountain glaciers. Warming affects seasonal snow by reducing the fraction of precipitation falling as snow [170, 171], shortening snow cover duration [172, 173], and increasing mid-winter melt [174]. The greatest snow sensitivity to warming occurs in regions with average winter temperatures warmer than -8°C [175].

Higher emission scenarios and temperatures are widely understood to increase snow loss. One simplified model suggests that in a 1.5°C warming scenario, the area of mountain regions with wintertime snowfall will decrease by about 9.5% (1 million km^2; an area approximately four times the size of the United Kingdom) relative to 1982-2020 levels [176]. This loss would double in a 2°C warming scenario and more than quintuple in a 4°C scenario [176]. On longer timescales, global climate models project that snow-cover duration across mountain regions will decrease by 30% by the end of the century in 1.5°C scenarios, by 50% in moderate emissions scenarios, and by 80% with very high emissions [3]. These changes correspond to an 8% decline in Northern Hemisphere spring snow cover extent per degree of warming [177]. In the European Alps, each degree of global warming translates into a reduction of one month in average snow cover duration at elevations of around 1500 m [178].

The ability for projected increases in precipitation (snowfall) to offset reductions in snow depth [179] is only expected to apply in the coldest regions (at very high elevations or high latitude) [180, 181].

Increasing rainfall in high elevation regions may also negatively affect the stability of traditional houses designed for snowfall, resulting in loss of traditional architecture [182]. Continued declines in annual snowpack will result in negative economic impacts for water resources [183], agriculture [146], hydropower [184], and tourism [185].

Reversibility After Overshoot

Even a temporary overshoot of 1.5°C will have effectively irreversible consequences over human timescales because glaciers will not recover for centuries. The long-term recovery outlook depends on both the peak temperature reached and the total duration of the overshoot period. One modelling study in which global warming temporarily reached 3°C by the year 2150 and returned to 1.5°C around 2300 found that global glacier mass loss was up to 16% larger than if no overshoot of 1.5°C occurred [157]. Some degree of recovery for glaciers is predicted to be possible with sufficient temperature decline, as long as glaciers still exist in those regions when cooling begins. Regrowth would be especially visible for steep, fast-responding glaciers in areas such as in the European Alps, High Mountain Asia, and the Tropics, which could show early signs of slow regrowth only a few decades after global cooling back towards 1.5°C levels has begun [157] (Figure 3.1).



Photo credit: M. Huss

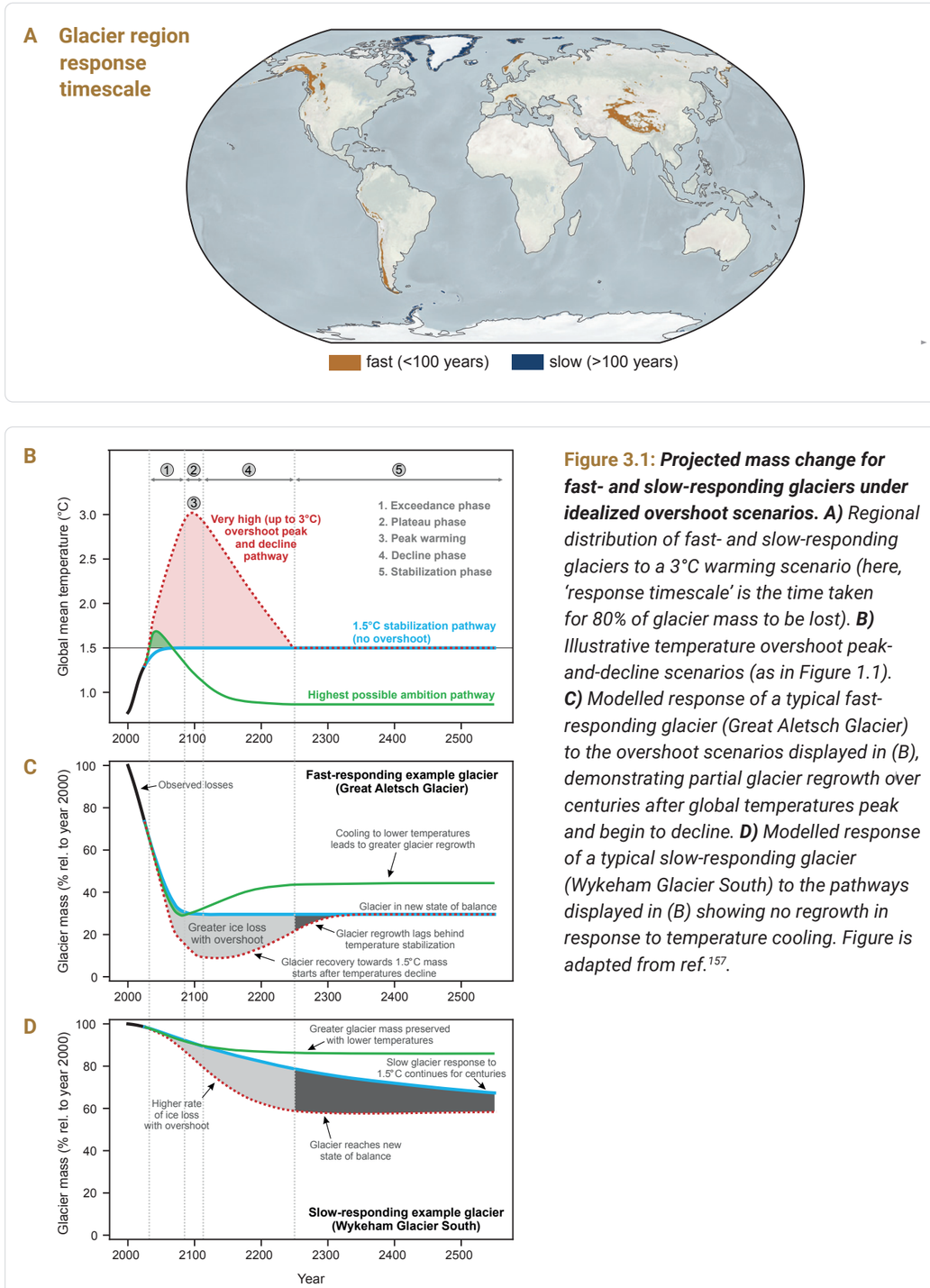


Figure 3.1: ***Projected mass change for fast- and slow-responding glaciers under idealized overshoot scenarios. A)*** *Regional distribution of fast- and slow-responding glaciers to a 3°C warming scenario (here, 'response timescale' is the time taken for 80% of glacier mass to be lost).* ***B)*** *Illustrative temperature overshoot peak-and-decline scenarios (as in Figure 1.1).* ***C)*** *Modelled response of a typical fast-responding glacier (Great Aletsch Glacier) to the overshoot scenarios displayed in (B), demonstrating partial glacier regrowth over centuries after global temperatures peak and begin to decline.* ***D)*** *Modelled response of a typical slow-responding glacier (Wykeham Glacier South) to the pathways displayed in (B) showing no regrowth in response to temperature cooling. Figure is adapted from ref.*[157].

However, because of the great sensitivity of glaciers to global warming even at today's temperatures, long-term global temperatures well below 1.5°C would be required to maximise glacier preservation and regrowth. For example, preserving 70% of current glacier ice would require bringing temperatures down to 0.8°C over centennial timescales [158]. Rates of glacier regrowth may also be hindered by several positive feedback mechanisms which remain poorly understood (See Chapter 8: Recommendations). These include the thinning and fragmentation of heavily crevassed glaciers as observed in North America and the Himalaya [186-188], accelerating ice volume loss on plateau icefields due to thinning and warming air temperatures that drive greater melt at lower elevations, and also the declining albedo of low-slope icefields due to rising snowlines [189]. Reversing global temperature rise also does not guarantee that local climates will recover to their previous state in terms of both local temperature and snowfall conditions [109].

Glacier recovery in the return phase of an overshoot peak-and-decline trajectory strongly depends on the recovery of snowfall. While declining temperatures may reduce melt, the regrowth of glaciers would require increased precipitation in the form of snow falling on the glacier (or former glacier) area. No scientific literature currently exists specifically assessing seasonal snow recovery and its impact on reversing deglaciation and loss of global and local albedo. However, the underlying physical processes suggest that snowfall declines are likely reversible to the same extent as local precipitation and temperature because snowfall is closely related to these factors. However, local temperature and precipitation changes may not be fully reversible during temperature decline, with pathway-dependent effects that vary regionally [190-195].



Photo credit: M. Huss

While snow trajectories remain understudied in the overshoot context, changes in seasonal snow may be relatively reversible. While snowfall reversibility - on glaciers or on the land surface - depends primarily on the reversibility of regional temperature and precipitation, seasonal snow by definition accumulates and melts annually, meaning that it does not have the multi-decadal inertia that glacial systems do. In an overshoot decline period, snow melt dynamics could be affected by additional factors altered by peak warming including forest land cover [196], incidence of wildfires [197], and dust [198]. The possibility of seasonal snowfall recovery is therefore plausible but remains uncertain in its magnitude, spatial coverage, and rate.

The impacts of overshoot on ecosystems and downstream hydrology during the post-overshoot recovery will be profound and qualitatively different from those during the warming phase [199]. Overshoot pathways that peak at high temperatures are expected to force extreme, rapid variations in glacier runoff. An initial meltwater surge would be followed by a steep drop in meltwater during the cooling phase as glaciers regrow, reducing water availability from glacier melt for decades to centuries after peak warming compared to scenarios in which 1.5°C was not exceeded [157]. Even a temporary water shortage may lead to long-lasting consequences in watersheds that are largely fed by meltwater, potentially disrupting industries and populations dependent on meltwater from glaciers and snow, downstream ecosystems, and nutrient cycles including offshore and coastal ecosystems near glacier-fed river deltas. Transitions from glacier- and snowmelt-fed to rainfall-derived runoff represents a fundamental regime shift, with uncertain long-term consequences for key species like salmon in the Pacific Northwest [200] and the creation of new ecosystems in deglaciated areas [201-203]. The potential for enduring impacts from adaptations to these conditions during the peak warming or decline phase of overshoot remains almost entirely unknown. Limiting the peak and duration of overshoot is therefore essential to prevent larger hydrological fluctuations and extreme events, which may challenge the adaptation capacity of downstream communities and ecosystems.



Photo credit: J. Kirkham

Table 3.1: *Committed regional glacier mass losses at different warming levels if global temperatures were to stabilize at these levels for long enough for glaciers to fully equilibrate with the climate. All numbers are in reference to 2020 glacier mass levels. Data is collated from ref. [158]. * Here, 'response timescale' is the time taken for 80% of glacier mass to be lost.*

REGION	PERCENTAGE OF GLACIER MASS REMAINING AT 1.5°C	PERCENTAGE OF GLACIER MASS REMAINING AT 2°C	PERCENTAGE OF GLACIER MASS REMAINING AT 2.7°C (CLOSE TO CURRENT POLICIES)	RESPONSE TIMESCALE* AT 1.5°C WARMING (YEARS)	RESPONSE TIMESCALE* AT 3°C WARMING (YEARS)
Alaska	59	42	31	100	75
Arctic Canada (North)	65	53	31	640	395
Arctic Canada (South)	13	5	1	240	120
Caucasus & Middle East	55	41	18	30	20
Central Asia	65	43	23	100	65
Central Europe	35	14	7	40	20
Greenland Periphery	37	32	14	370	220
Iceland	35	23	4	320	130
Low Latitudes	54	32	13	20	10
North Asia	30	14	7	50	25
New Zealand	76	66	39	35	25
Russian Arctic	25	12	1	450	160
Scandinavia	18	0	0	170	55
Southern Andes	55	50	36	90	80
South Asia East (Central and Eastern Himalaya)	37	26	16	50	30
South Asia West (Western Himalaya, Karakorum & Hindu Kush)	83	62	41	140	70
Sub- & Antarctic Islands	56	44	33	830	550
Svalbard & Jan Mayen	41	18	6	320	180
Western Canada & US	19	7	2	60	30



Photo credit: M. Huss

04 Sea Ice

Executive Summary

Sea ice can eventually recover when global temperatures decrease after overshoot of 1.5°C, but this recovery will not be instantaneous. Excess heat stored in the ocean will delay recovery by decades to centuries, especially with higher peak temperatures and longer overshoot durations. During overshoot, the Arctic Ocean will experience several months of very low sea ice concentrations each year in the summer by 2°C, including frequent ice-free conditions in September (the peak of summer sea ice loss), and possibly well below that temperature; even 1.5°C may result in occasional ice-free summer conditions. Ice-free periods would last for several months with a higher and longer period of overshoot in the Arctic. Sea ice conditions in the Southern Ocean are harder to predict. Antarctic sea ice recently shifted from a highly variable pattern to one of steep decline, signalling a potential regime shift. While studies remain scarce, current evidence suggests that the large thermal inertia of the Southern Ocean will continue to drive Antarctic sea ice loss for decades to centuries even once temperatures stabilize under net zero greenhouse gas emissions. Crucially, even a temporary loss of sea ice during overshoot risks triggering abrupt and irreversible increases in coastal erosion, loss of sea-ice dependent ecosystems and human settlements, marine ecosystem collapse, and the disintegration of ice shelves, especially around Antarctica.



Photo credit: J. Kirkham

Suggested citation (this chapter): Doddridge, E., Jahn, A., Raphael, M., and Stroeve, J. (2026). Sea Ice. In: Overshoot and the Cryosphere: Averting Global Meltdown. The Implications of Temperature Overshoot, Peak and Decline for the Cryosphere. [Kirkham, J.D., Pearson, P., Garbe, J., Hancock, S., Imdieke, A. (eds.)]. International Cryosphere Climate Initiative.
DOI: 10.5281/zenodo.23477467

Background

Sea ice forms at the intersection of ocean and atmosphere and is strongly affected by both. The future of sea ice at both poles is closely tied to our emissions trajectory; as our planet warms, sea ice declines, triggering a cascading series of ecological and economic impacts. In the Arctic, late summer (September) sea ice extent has decreased by over 40% since 1979 [204], with extent and thickness declining across all seasons despite some year-to-year variability [204-206]. Along with this decline, the region has transitioned away from a stable multi-year ice regime towards one dominated by thinner, faster-moving seasonal ice floes that are less likely to survive the summer melt season [204, 207].

Historically characterized by substantial decadal variability, including a slight expansion that peaked in 2014, recent observations suggest that Antarctic sea ice has also entered a period of decline [12, 208]. The summer minima in 2017, 2022, and 2023 all set new records for the lowest observed extent [12, 209], while the wintertime sea ice coverage in 2023 was well below any previous observations since at least 1979, signalling a potential regime shift in the Southern Ocean [12, 210].

In both the Arctic and the Antarctic, the loss of sea ice impacts the physical environment and ecosystems. Sea ice forms a habitat for many species, and its loss threatens polar food webs, ranging from phytoplankton to penguins and polar bears. The International Union for Conservation of Nature (IUCN) classified polar bears as vulnerable in 1982, and in 2008 the U.S. Fish and Wildlife Service listed polar bears as threatened.

Several seal species, including hooded, harp, ring and bearded seals are on the IUCN red list due to rapid Arctic sea ice loss. The recent listing of emperor penguins as endangered was largely motivated by the loss of suitable Antarctic sea ice habitat [211, 212].

Beyond its biological impacts, sea ice protects coastal margins from damaging waves. In the Arctic, the loss of sea ice is leading to enhanced coastal erosion [213]. In the Antarctic, sea ice buffers Antarctic ice shelves, protecting them from wave damage [214, 215]. The loss of sea ice leads to enhanced iceberg calving [216], ice shelf disintegration [217], and can affect the flow of the Antarctic Ice Sheet with implications for rates of sea-level rise [32, 214]. On a planetary scale, sea ice creates a barrier between the ocean and atmosphere, reducing evaporation and heat loss from the ocean to the atmosphere [218]. It also increases the surface albedo and contributes to maintaining the equator-to-pole temperature gradient that drives global weather patterns. Shifts in current weather patterns around the world, such as wetter winters in the Mediterranean and drier winters in the southwestern United States are tied to disruption of this temperature gradient [219]. Sea ice formation and melt additionally modify the density of seawater and play a critical role in driving global-scale ocean circulations that redistribute heat, salt, and carbon around the globe [220]; see Chapter 6, Polar Oceans). Sea ice therefore plays a critical role in the earth system, and even a temporary loss of sea ice over new areas during an overshoot peak-and-decline pathway may promote abrupt and potentially irreversible changes in ice-dependent ecosystems as well as in other parts of the climate system.

Consequences of Overshooting 1.5°C

Climate model simulations of how sea ice extent will respond to global warming are characterized by a large spread of possible outcomes in both hemispheres. For the Arctic, despite this spread, models consistently agree that sea ice extent will continue to decline as global temperatures rise [221-223]. In contrast, there is much lower confidence in projections for the Southern Ocean due to persistent biases in how models represent Antarctic sea ice [224, 225]. While the rapid loss of Antarctic sea ice coverage since its 2014 peak has brought observations more in line with model projections [226], a comprehensive analysis of CMIP6 models has shown that current climate models are still not able to simulate the seasonal expansion and retreat of Antarctic sea ice realistically [227]. Since climate models are our best-available tools for exploring future scenarios, including the impacts of overshoot pathways on sea ice and associated systems, the lack of accuracy in climate models in simulating the observed Antarctic sea ice evolution remains a critical limitation.

Arctic

Even if global temperatures can be limited to 1.5°C, Arctic air temperatures are expected to exceed levels considered 'extremely warm' under pre-industrial conditions on nearly 300 days of the year [14]. Faced with these elevated temperatures, projections indicate that the summer extent of Arctic sea ice will generally stabilize around the record low conditions first observed in 2012 [14], when minimum Arctic sea-ice extent fell below 4 million km^2 for the first time in modern history [228]. However, even at this warming level, occasional ice-free September conditions (defined as having sea ice extent below 1 million km^2) may still occur [229, 230]. By 2°C, projections indicate that the Arctic Ocean will experience several months of very low sea ice concentrations each year in the summer [230], including frequent ice-free conditions in September [231].

If global warming approaches 3°C, the Arctic Ocean is projected to experience ice-free conditions in September every year, with ice-free conditions possibly lasting several months. Changes will not be limited to the summer; the winter sea ice will be much thinner than today and often no longer extend to the coasts because of the shorter sea-ice growth season and large amounts of winter warming [14]. Furthermore, it is important to note that once net-zero greenhouse gas emissions are reached, Arctic sea ice is expected to continue to decline until it reaches equilibrium with the new climate [232-234]. This continued decline is due to the long equilibration time of the climate system to external forcing [235, 236], which has been found to be longer for the surface polar oceans compared to the surface mid-latitude oceans [237].

Antarctic

The Southern Ocean exhibits significant inertia in the climate system; it is slow to commence warming [238] and continues to warm long after the global-mean surface temperature is stabilized [239]. Consistent with this long-term warming, modelling studies have shown that Antarctic sea ice continues to decline after net-zero CO_2 emissions is reached [13, 240]. Current evidence suggests that the Southern Ocean will take centuries to stabilize under a global-mean temperature stabilization scenario, much longer than the Arctic Ocean. Simulations that stabilize at higher global-mean temperatures exhibit higher rates of regional warming and more rapid sea ice loss in the Southern Ocean (e.g., Figure 3 in ref. [240]). As such, even under a temperature stabilization scenario, Antarctic sea ice is expected to continue declining with the associated impacts becoming more severe over time.

Reversibility After Overshoot

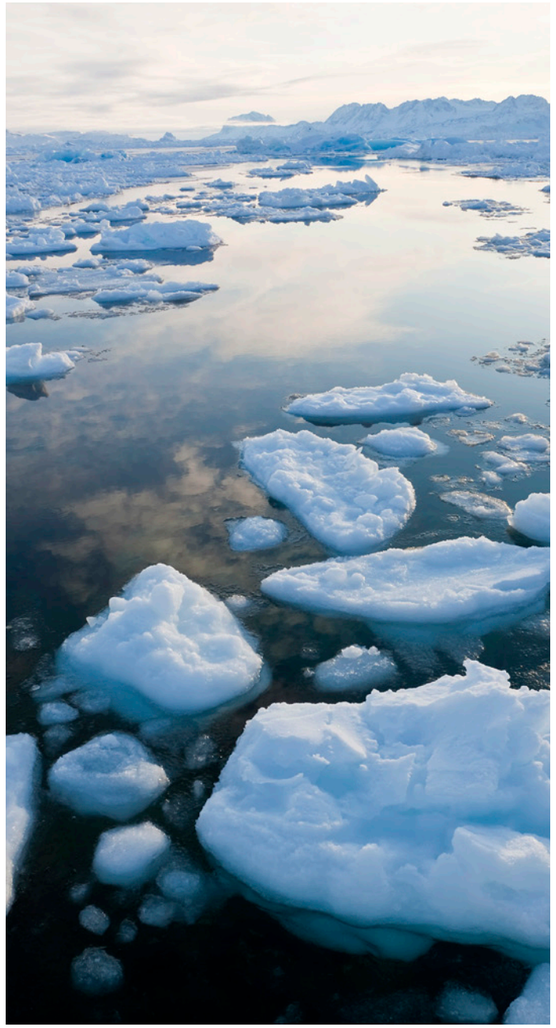
Arctic

Studies have generally shown that Arctic sea ice expands again after overshoot when global temperatures decrease [232-234, 241-244]. However, this recovery is not instantaneous, with studies showing delays of decades to several centuries for sea ice recovery [232-234, 244]. One of the primary drivers for this delay in sea ice recovery is the ocean heat inertia, i.e., the excess heat stored in the upper layers of the Arctic Ocean after periods of reduced or no sea ice cover [232, 241, 244, 245]. In addition, enhanced downward longwave radiation over the central Arctic Ocean due to greater moisture influx from more frequent atmospheric rivers in a warmer climate reaching the Arctic Ocean may delay sea ice recovery [244]. Furthermore, the delay in sea ice recovery during the winter season is expected to last longer than summer sea ice recovery due to the larger influence of ocean heat transport and storage on winter sea ice formation [232, 241, 244].

Notably, in model simulations the Arctic sea ice cover after an overshoot is generally lower than the pre-overshoot sea ice cover at the same CO_2 level [232, 234, 244]. This is due to slow processes in the climate system that take time to fully equilibrate [235, 236] and hence lead to a difference in the truly stable versus transient Arctic sea ice cover [232]. Furthermore, the length of the equilibration time in the Arctic differs between warming and cooling, with a much longer stabilization time for cooling [246]. The difference in transient versus equilibrated sea ice state, however, does not indicate general irreversibility in the Arctic sea ice system on centennial to millennial timescales [232, 233, 241-244]. However, on human timescales (years to decades), Arctic sea ice cover may look fundamentally different in a world with an overshoot compared to a stable 1.5°C world without overshoot [232, 234].

Furthermore, the cascading impacts of increased sea ice loss during overshoot are in some cases irreversible.

For example, more open water has been shown to lead to more coastal erosion [213] and permafrost thaw [247] along the Arctic coast. These impacts are irreversible on human timescales (see Chapter 5: Permafrost and Chapter 8: Interactions and Feedbacks in a Cryosphere under Overshoot) and can have societal impacts when these force relocation of human communities from locations where they existed for centuries. Changes in Arctic ecosystems are also already ongoing in response to Arctic sea ice loss [248, 249] and may not be reversible, especially with sea-ice dependent species lost during the overshoot period.



Antarctic

Upwelling of abyssal ocean waters limits the rate at which the Southern Ocean responds to global temperature change during both the warming and subsequent cooling phases of an overshoot pathway. Global simulations have shown that Antarctic sea ice is slower to recover than Arctic sea ice [241]. The ocean's ability to store and later upwell heat means that Southern Ocean sea ice can also exhibit dramatic changes centuries after the peak in global surface temperature [250]. A comparison of overshoot scenarios with peak warmings of 2.0°C, 2.5°C, and 3.0°C and return to 1.5°C, shows that the Southern Ocean surface temperature warms slightly more during overshoot when the peak temperature is higher, and that post-overshoot, the surface temperatures eventually converge with the stable 1.5°C simulation without overshoot [251]. While these simulations showed little impact during the transient overshoot peak on Antarctic sea ice extent, all scenarios show a continued decline in Antarctic sea ice extent post-overshoot, with a persistent and widespread thinning of Antarctic sea ice [104, 251].

While studies remain scarce, current evidence suggests that Antarctic sea ice extent will likely take many centuries to stabilize after an overshoot [13]. During these centuries, Antarctic sea ice extent and volume will continue to decline at a rate that is more closely tied to the final stabilization temperature than the peak temperature [13, 240, 251]. However, the loss of Antarctic sea ice, even temporarily, has the potential to trigger abrupt changes and tipping points in associated systems [13]. While some of these changes may be reversible following peak warming, others are expected to be either completely irreversible or irreversible on human timescales. For example, species extinctions due to loss of habitat are not reversible [252, 253], and Antarctic Ice Sheet destabilisation due to ice shelf loss [254], and ocean heat uptake in the absence of sea ice [251] will take centuries to millennia to reverse.

Photo credit: J. Kirkham

05 Permafrost

Executive Summary

Overshoot pathways are expected to increase greenhouse gas emissions from permafrost thaw compared to pathways that stabilize at 1.5°C with no or minimal overshoot, substantially reducing the remaining carbon budget to meet the goals of the Paris Agreement. During overshoot, permafrost regions will be exposed to higher temperatures for longer, accelerating thaw as well as destabilizing previously frozen ground in the Arctic and high elevation regions and mountain slopes. This can destabilise the foundations of infrastructure and settlements built on permafrost, alter water cycles, and increase risk of hazards from rapid ground and slope collapse. During overshoot, wildfires are expected to intensify through longer fire seasons, more frequent lightning ignitions, and deeper burning of organic soils, which may take decades to centuries to recover. Although permafrost will refreeze if global temperatures are eventually returned to 1.5°C or below, the carbon released during overshoot is effective irreversible, especially because the subsequent cooling phase will create less favourable conditions for vegetation growth and carbon drawdown into former permafrost soils. Overshoot will fundamentally reshape hydrological networks in permafrost regions for decades, even after global air temperatures have stabilized or begun to decline.



Photo credit: B. Radosavljevic

Suggested citation (this chapter): Duan, Z., Kuklina, V., Romanovsky, V., Ruiz Pereira, S., Schuur, E. A. G., Steinert, N.J., and Stuenzi, S. M. (2026). Permafrost. In: Overshoot and the Cryosphere: Averting Global Meltdown. The Implications of Temperature Overshoot, Peak and Decline for the Cryosphere. [Kirkham, J.D., Pearson, P., Garbe, J., Hancock, S., Imdieke, A. (eds.)]. International Cryosphere Climate Initiative. DOI: 10.5281/zenodo.23477467

Background

Permafrost is frozen ground found across high-latitude and high-mountain regions, technically defined as a mixture of soil or rock and any ice and organic material that remains at or below 0°C for two consecutive years or longer [3]. Today, permafrost is widespread in the Arctic, sub-Arctic, high mountain regions and in ice-free areas of the Antarctic and sub-Antarctic. Terrestrial permafrost underlies approximately 14 million km^2 (15%) of the Northern Hemisphere [255, 256]. Submarine permafrost is also found on continental shelves offshore of Siberia and North America [257]. Permafrost stores at least 1,700 billion metric tonnes of frozen and thawing carbon on land [258, 259] and more than 4300 billion tonnes if subsea permafrost is included [260]. This is several times the amount of carbon currently stored in the atmosphere; consequently, even the release of just a small fraction of the carbon stored as permafrost could substantially increase emissions of carbon dioxide and methane, further accelerating the rate of climate change [14].

Permafrost warming and thawing has profound impacts on people, ecosystems and infrastructure that use permafrost as a foundation. Surface permafrost (located at depths of 0-3 metres beneath the ground) is the quickest to respond to warming and has the most influence on ecosystems and people [14]. Recent permafrost warming started in the Northern Hemisphere in the late 1970s. It was not uniform in space and time, with the warming rate averaging between 0.1 and 1.0°C each decade for the last 40 years; higher rates are observed at higher latitudes and higher elevations [15, 261, 262]. Permafrost surface thaw has increased the depth at which soils thaw during summer and then re-freeze during winter, called the “active layer” [263, 264], as well as the formation and deepening of taliks [265]. Permafrost thaw emits greenhouse gases, triggers natural hazards, increases coastal erosion in populated regions, and can destabilize infrastructure [266-268].

In addition to the impact of rising global temperatures, this thaw can be accelerated by natural and human-made surface disturbances such as wildfires, agricultural activities, urban, extractive, transportation, and engineering infrastructure development as well as by landfills and wastelands in permafrost regions [269-272].

Once permafrost begins to thaw, microbes begin decomposing previously frozen organic matter, releasing carbon into the atmosphere as both carbon dioxide and methane, as well as smaller amounts of nitrous oxide (N_2O, a highly potent climate forcer) [266, 269, 273]. Not all feedbacks amplify processes can continue emitting for centuries after initial thaw. The formation of taliks can accelerate this decomposition by introducing a subsurface environment with year-round above-freezing conditions. Permafrost also controls how water and carbon are stored and how they move through the landscape. Thawing permafrost can change the timing and distribution of water in soils, wetlands, lakes, rivers, and the movement of groundwater into aquifers beneath permafrost layers, with consequences for ecosystems, freshwater resources, and communities that depend on them [274-276].

Consequences of Overshooting 1.5°C

Greenhouse Gas Emissions

Already today, the northern permafrost region, comprising Arctic and boreal ecosystems, is a CO_2 source to the atmosphere (on the order of +0.106 Gt C yr^{-1}) [277, 278]. This includes emissions from fires, carbon released and transported by rivers, coastal erosion and the exchange of greenhouse gases between frozen soils, tundra vegetation, and Earth’s atmosphere. Future CO_2 and CH_4 emissions may total 230 Gt C CO_2-eq (CO_2-equivalent, which combines CH_4 and CO_2 emissions) by 2100 under high warming approaching 3-4°C [259]. In contrast, the total amount by 2100 with warming around 2°C could total around 55 Gt C CO_2-eq.

These greenhouse gas emissions can be considered in comparison to those from today's largest emitting nations, ranging from India (which emitted 1.23 Gt C CO_2-eq in 2025) to the United States or China (1.64 and 4.35 Gt C CO_2-eq in 2025, respectively) [279]. Accounting for the impacts of abrupt thaw and wildfire further increases emissions beyond those from gradual permafrost thaw alone, reducing the remaining carbon budget to meet the goals of the Paris Agreement by 25% for 1.5°C, and 17% for 2°C [280]. Furthermore, as organic matter contained within permafrost continues to decompose once thawed, these emissions will continue after 2100 for multiple centuries [281].

Carbon loss from gradual top-down permafrost thaw is expected to increase approximately linearly with global warming [109, 282-284]. Since overshoot pathways increase temperature exposure relative to low-temperature stabilization scenarios, on the one hand, overshoot increases permafrost carbon loss [285]. On the other hand, if global temperatures can be peaked at 1.5°C and then lowered rapidly, overall warming exposure may be lower than scenarios in which global temperatures stabilize at 1.5°C. The additional emissions released from permafrost carbon as a result of overshoot may be more complex than linear warming projections, reducing, for example, remaining carbon budgets by 45-85% in order to stabilize at 1.5°C, or 15-25% in order to stabilize at 2.0°C [286]. The effect of permafrost and peatland emissions on 2300 temperatures has been estimated to increase by 0.02°C with every additional century of overshoot [109]. Additionally, nonlinear or sudden processes, including abrupt permafrost thaw, drive continued greenhouse gas emissions that persist beyond the temperature peak of the overshoot pathway [269, 287-289].

The additional warming produced by this cycle of rising temperatures, permafrost thaw, decomposition, and carbon release – known as the "permafrost carbon–climate feedback" – increases the risk that tipping points elsewhere in the Earth system (such as the polar ice sheets) may be crossed during overshoot that would not otherwise be crossed in a climate without this feedback [92, 284].

While some recovery of permafrost extent may be possible if global temperatures are brought back to 1.5°C or below, a large portion of the carbon lost from thawed permafrost is effectively irreversible on human timescales [288, 290-293] (Figure 5.1). The resulting carbon–climate feedback may, however, be partly compensated by increased or sustained ocean carbon uptake [294], and the expansion of fast-growing, northward-spreading vegetation which acts as a carbon sink [295]. However, because rates of vegetation growth drop as temperatures cool in the temperature decline phase of an overshoot peak-and-decline pathway, vegetation expansion is only temporary and may be less effective in dampening warming feedbacks compared to trajectories that stabilize at a higher constant temperature [296, 297]. Thus, while an overshoot trajectory may reduce the magnitude of multi-century carbon emissions from permafrost thaw in northern high latitude regions, it also brings unique challenges depending on the magnitude, duration and rate of temperature change.

Infrastructure, Physical Hazards and Socio-Economic Impacts

At a local scale, the transition at around 0°C of permafrost and frozen ground to a thawed state represents an identifiable threshold that can cause tipping behaviour in ecosystems, infrastructure, and permafrost-dependent livelihood systems, especially those of Arctic and mountain Indigenous Peoples. This is seen most dramatically and devastatingly where degradation and failure of frozen ground occurs together with large glacier cover, as seen in recent ice and rock avalanches in the Indian Himalaya [138], the European Alps [135, 136], and in Nepal in August 2026 [134], which all led to extensive infrastructure damage and loss of life. Permafrost responds to every increment of warming, which means that there is no "safety margin" for acceptable permafrost thaw [282, 298]. Such loss and damage from permafrost thaw in mountain regions can be expected to increase under overshoot conditions (Figure 5.1).

In the Arctic, around 5 million people live in settlements built on permafrost, with 1 million of those along coastlines potentially vulnerable to erosion [299, 300]. At least 120,000 buildings, 40,000 km of roads and 9,500 km of pipelines are located regions of the northern hemisphere that contain permafrost [272, 301]. The economic impact of building and road damage from Arctic permafrost thaw is projected to be in the order of ~$70-270 billion USD by mid-century depending on the level of overshoot and its duration [302-304].

The most immediate impact of permafrost thaw is a decrease in bearing capacity of building foundations built on piles inserted into permafrost. Rapid thawing of ice-rich permafrost can produce substantial differences in subsidence of the ground surface (thermokarst, see Glossary). The rate of subsidence in the most vulnerable locations can exceed 5 cm per year [305].

Such large and uneven subsidence affects numerous civil and industrial infrastructure with a significant high cost to repair and maintain the function of community buildings, transportation routes, and utility services [272, 301, 306-308]. Such impacts are already observed around buildings, roads, airports, and pipelines in the Arctic and mountain regions [272, 309, 310].

For many Indigenous and nomadic communities, loss of permafrost landscapes leads to irreversible loss of cultures that evolved in close relationship with the environment [311-313]. In addition, thermokarst and thermal erosion may cause lake drainage, imposing an additional hardship on populations that rely on lakes for their drinking water supply [314, 315], and increase the rate of coastal and riverbank erosion, posing a threat to those communities [300, 316, 317]. Permafrost thaw also affects communities using ice cellars to store food in summertime [318].

For high-elevation mountain regions such as the Qinghai–Tibet Plateau, limiting warming to 1.5°C instead of 2°C has been estimated to reduce the cost of infrastructure damage from permafrost thaw by $1.32 billion USD by 2100 [319]. Despite this, even under moderate emissions, nearly two-thirds of the permafrost area in the Qinghai-Tibet Plateau is expected to become a "high-hazard" zone this century [319].

Because a large portion of the massive ground ice in permafrost was accumulated during the last Ice Age, its thaw is irreversible on time scales of thousands of years. As a result, infrastructure damage from additional permafrost thawing during the peak of an overshoot cannot be reversed by the cooling that follows the overshoot, making this damage irreversible as well. Moreover, refreezing of taliks formed during the peak of an overshoot period may lead to substantial frost heaves that cause newly frozen soils to expand and flex upward, damaging existing infrastructure even further as temperatures cool.

The Impacts and Recovery Time of Permafrost in Response to Overshoot

Reversible with sufficient cooling over decades

Reversible with sufficient cooling over centuries

Largely irreversible on human timescales

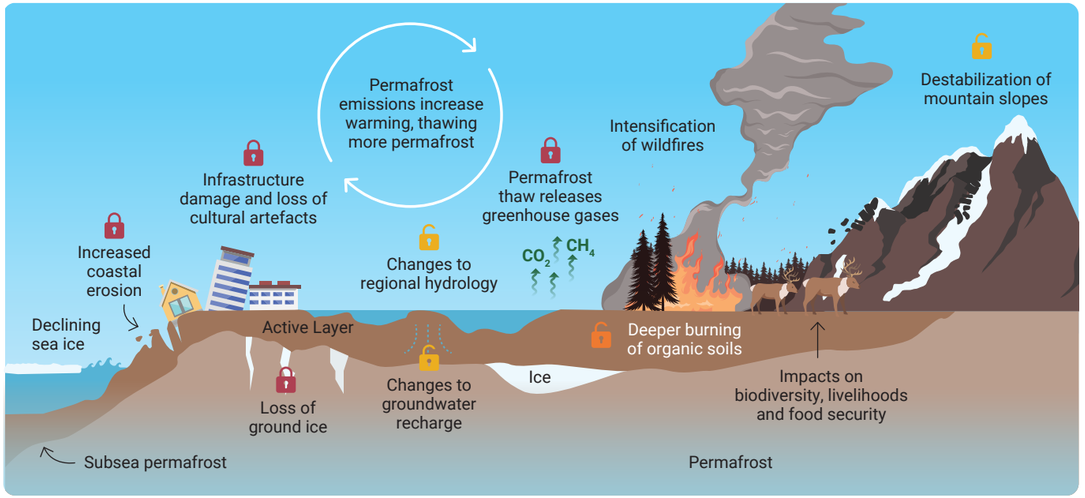


Figure 5.1. ***Illustration of the impacts and reversibility of permafrost thaw in response to overshoot.*** *Every fraction of a degree of overshoot will cause more widespread permafrost thaw and related impacts. Although some permafrost may eventually refreeze if global temperatures decline, many impacts are largely irreversible on human timescales, and some could even further accelerate global warming.*

Hydrology

Permafrost is an important regulator of water movement and storage in cold regions. Frozen ground limits the movement of water into deeper soils and helps to control how water is stored and released, affecting river flows, wetlands, lakes, soil moisture, groundwater recharge, and the seasonal availability of water. Fundamental shifts in water movement from surface runoff to deeper subsurface pathways are already being observed in the Arctic, Sub-Arctic, and high mountain regions due to permafrost thaw [275, 276, 320, 321]. The magnitude and persistence of these hydrological changes depend on both the level and the duration of warming, making overshoot pathways particularly important [291, 294].

Permafrost degradation can substantially alter soil moisture, with responses varying across landscapes: ground-ice melt can make soil conditions wetter locally, but continued thaw can also enhance drainage and lead to drier near-surface conditions in some areas [263, 274]. As the active layer deepens and taliks form and expand, water can infiltrate to greater depths, increasing ground infiltration, winter baseflow and, in some regions, summer peak runoff [276, 322]. Because deep soils respond slowly to changes in climate, hydrological responses may continue to evolve for decades after global air temperatures have stabilized or begun to decline. Although post-overshoot cooling may allow some permafrost to refreeze, recovery of pre-thaw hydrological conditions is likely to lag behind temperature recovery.

The loss of excess ground ice can permanently alter surface topography through subsidence, while changes in subsurface flow pathways may persist for extended periods depending on local ground-ice conditions, landscape evolution, and the extent of permafrost degradation [291, 294]. Groundwater responses can also be delayed. As permafrost thaws, increased groundwater recharge and storage can delay changes in groundwater contributions to streams and rivers by decades or longer [323]. Deeper groundwater flow also modifies the temperature of groundwater discharge, potentially leading to cooler summer headwater streams than under shallow, permafrost-controlled conditions [324].

Under climate overshoot scenarios, continued permafrost degradation may expand and prolong these hydrological and ecological changes, even after temperatures stabilize or decline. Freshwater ecosystems may be affected through changes in stream temperature, water availability, and the transport of carbon, nutrients, and other dissolved constituents [323, 324]. In addition, greater infiltration and evapotranspiration may reduce streamflow from headwater catchments unless offset by increased precipitation in a warmer climate, with implications for aquatic habitats, biogeochemical cycling, and ecosystem productivity [325]. The dryland belt of Northern Eurasia would be exposed to increased risk of desertification under these changes [326].

The consequences of these changes differ between landscapes. In Arctic lowlands, thermokarst lakes are particularly important due to the extensive area occupied — around half of the landscape in some instances [327]. Thaw of ice-rich permafrost can cause ground subsidence and promote the formation or expansion of lakes. However, continued thaw can also open drainage pathways that can drain these lakes, resulting in loss of water storage [321, 327-329]. These dynamic changes in lake extent can in turn alter the regional energy balance through shifts in land surface albedo, carbon cycling, habitats for wildlife, and drinking water supply for remote Arctic communities [287, 329-331]. Loss of soil moisture can also contribute to vegetation browning in northern permafrost ecosystems [332]. On the Qinghai–Tibet Plateau, permafrost thaw can result in increased infiltration and deeper subsurface drainage, contributing to drier conditions with declining near-surface soil moisture and river flows that could decrease seasonal water availability [275, 322].

Permafrost thaw can also affect water quality by affecting the transport of dissolved organic carbon, nutrients, sediments, and other dissolved materials into rivers and downstream aquatic ecosystems [328, 333, 334]. Recent observations of “rusting rivers” in the Arctic further demonstrate that permafrost thaw can change river chemistry through the release of iron and other metals and potentially affect water quality and freshwater ecosystems [335].

Ecosystems, Vegetation and Wildfires

Extreme events are an increasingly important driver of permafrost thaw. Heat extremes intensify with warming [336], and extreme rainfall accelerates active-layer thaw [337]. Because the thaw triggered by a single extreme can persist for years [338], these impacts accumulate as extremes grow more frequent. The growing incidence of wildfire, extreme heatwave and precipitation events can increase abrupt thaw. Once initiated, large and persistent changes in ecosystem state can result, encompassing processes from thermokarst and thermal erosion to wildfire-driven thaw [339], raising permafrost emissions further.

An overshoot peak-and-decline pathway does not simply mean a return to conditions before the peak. Rather, overshoot will leave lasting damage, including irreversible permafrost loss. While changes to fire regimes and vegetation may recover as temperatures decline, the carbon combusted and the ground ice melted during the peak cannot be recovered on human timescales. During the overshoot peak, fire regimes will intensify due to longer fire seasons, increased lightning ignition [340], and deeper burning into organic soils, including overwintering fires [341]. Combustion of the insulating organic layer removes the surface thermal buffer, exposing legacy carbon (i.e., organic soil carbon that escaped burning in previous fires [342]) and potentially initiating abrupt thaw [269, 343, 344]. Both the forest canopy and the organic surface layer thermally protect permafrost ground, and dieback or fire removes these buffers, which could take decades to centuries to recover after peak overshoot [345, 346]. The deeper seasonal thaw and talik formation that follows exposes deeper soil organic carbon, some of it millennia old, to decomposition. By lowering surface albedo and drying near-surface soils, this deeper thaw intensifies summer warming and atmospheric aridity, leading to conditions that are more conducive to further fires [347].

The organic carbon held in permafrost soils far exceeds the biomass carbon lost with burned vegetation, so soil carbon dominates fire-related emissions through two routes: direct combustion of soil organic matter during the wildfire [342], and the added thaw and decomposition that fire and canopy loss subsequently unlocks [348]. In tundra, the controlling mechanism differs: expanding tall shrubs trap snow, and the thicker, more insulating snowpack suppresses cold-season heat loss from the ground, warming shallow soils in winter so that thaw and related emissions increase [349, 350].

Not all feedbacks amplify permafrost thaw and its wider impacts. Repeated fires are expected to drive vegetation shifts from coniferous to deciduous plants in boreal North America, and deciduous forests burn with lower carbon losses, dampening the fire-climate feedback [351, 352]. Post-thaw peat accumulation and vegetation regrowth can also re-sequester carbon through processes that can take centuries [353]. Sustained but spatially uneven Arctic greening offers a further partial offset through increased carbon uptake in vegetation and summer shading [354, 355], though it is locally offset by browning, as described above [356].

These offsets are real but partial: they operate on timescales far longer than the overshoot peak, and they do not restore ground ice. Against these partial offsets, fire is now a dominant control on the region's net carbon balance rather than a local disturbance. Once fire emissions are included, any increasing Arctic–boreal CO_2 sink due to higher vegetation growth is no longer statistically significant and the permafrost region is approximately CO_2-neutral [278], consistent with studies that include fire and abrupt-thaw emissions alongside methane and N_2O emissions [277].

Both overshoot magnitude and duration matter: cumulative severe fire years, not peak temperature alone, determine committed carbon loss. Incorporating permafrost thaw and wildfire roughly doubles projected permafrost emissions this century, cutting the remaining 1.5°C carbon budget by about a quarter [280].



The sheared-away glacier and mountainside below Tsangbu Peak, Nepal, that triggered the deadly Rasuwa floods in August 2026.

Photo credit: G. Zheng

06 Polar Oceans

Executive Summary

The Arctic and Southern Oceans already show rapid, interconnected physical, chemical and ecological changes due to continuing greenhouse gas emissions, including the warming and freshening of polar waters, acidification, and large-scale changes to ocean circulation. Overshoot pathways will leave greater and longer-lasting changes in the polar oceans even after temperature and atmospheric CO_2 levels decline, compared to pathways with no or lower overshoot levels; and are associated with a higher risk of crossing critical thresholds or tipping points. Impacts irreversible on human timescales, ranging from decades to centuries and millennia, include persistent ocean warming and freshening, acidification, circulation changes and ecosystem disruption. Each of these phenomena is more pronounced and more persistent the longer lasting and higher the overshoot. Acidification, causing harm to shelled and other marine life, is especially persistent in polar waters under overshoot scenarios. Slowdown of ocean currents, after passing still-uncertain tipping thresholds, may similarly persist for centuries to millennia, and resultant ecosystem changes may drive extinction of some cold-dependent polar species. For policy, the implication is clear: avoiding or minimising overshoot is essential. Delayed atmospheric CO_2 and temperature decline will not reverse polar ocean warming, acidification, circulation changes or ecosystem loss on timescales relevant to societies, fisheries, conservation or climate stability.

Suggested citation (this chapter): England, M. H., Findlay, H. S., Henley, S. F., Monteiro, T., Nicholson, S., and Zhang, Z. (2026). Polar Oceans. In: Overshoot and the Cryosphere: Averting Global Meltdown. The Implications of Temperature Overshoot, Peak and Decline for the Cryosphere [Kirkham, J.D., Pearson, P., Garbe, J., Hancock, S., Imdieke, A. (eds.)]. International Cryosphere Climate Initiative. DOI: 10.5281/zenodo.23477467

Background

The polar and subpolar oceans, encompassing the Arctic, North Atlantic and Southern oceans, are central regulators of the Earth system. These polar oceans are the foundation of the global ocean circulation system that redistributes heat, carbon, oxygen and nutrients around the world. Although distant from many global population centres, they play an outsized and largely under-appreciated role in absorbing and storing heat and CO_2, driving the behaviour of global ocean currents and helping to regulate the global climate, while also supporting diverse ecosystems and economically important fisheries. Polar oceans are already undergoing rapid, interconnected physical, chemical and ecological change as the world approaches the Paris Agreement temperature limit of 1.5°C warming, with global mean surface temperatures now averaging about 1.26°C above pre-industrial levels over 2016–2025 [357], while atmospheric CO_2 averages 427 ppm (parts per million) [358, 359]. At these levels, rapid and abrupt environmental shifts are already being seen across both polar regions [13, 207].

Cold and dense polar waters can more easily take up and store carbon, though this increases their vulnerability to acidification [360]. Polar oceans also buffer global atmospheric temperature rise by absorbing excess heat from the atmosphere. This heat, along with carbon, oxygen, and nutrients, is redistributed globally by polar ocean circulation [3, 232, 361]. Polar marine food webs support biodiversity, fisheries and carbon sequestration by species such as Antarctic krill [362] and primary producers [363]. However, by acting as efficient and important climate buffers, these regions are at greater risk of rapid change, resulting in faster warming, freshening, deoxygenation, and acidification compared to other ocean regions [364-367].

The Arctic has warmed up to four times faster than the global average [368], and Arctic upper-ocean warming exceeds the global mean rate [369]. Sea-ice loss and warm water inflows are increasing Arctic marine heatwave frequency and duration [370, 371], while the presence of low oxygen regions is growing in parts of the Arctic Ocean [367]. Marine heatwaves and ocean warming are also increasingly affecting several sectors of the Southern Ocean including the Ross, Amundsen-Bellingshausen and Davis Seas [372], where sea ice has decreased extensively since 2015/16 (see Chapter 4: Sea Ice). These heat extremes can alter ecosystem functions and biodiversity [373-376]. Such ecological changes matter for the global climate because marine organisms help sequester carbon over hundreds to thousands of years as they sink to the deep ocean [362, 363, 377].

Alongside warming, levels of acidity are increasing in both surface polar waters and deeper in the water column due to atmospheric CO_2 uptake. The current rate of acidification is essentially unprecedented in the geological record [378] and far greater in polar regions than in temperate and tropical oceans [379]. In both the Arctic and coastal Antarctica, acidification is further increased by growing freshwater input from melting glaciers and ice sheets, which adds even more CO_2 to polar waters through various chemical processes [380-383]. Many polar organisms already live close to the limits of their adaptability to acidification [384, 385], and corrosive conditions have been observed across parts of the Arctic Ocean [364, 365] and the Southern Ocean [386]. A recent assessment concluded that ocean acidification has crossed a planetary boundary in several polar and subpolar regions [387].

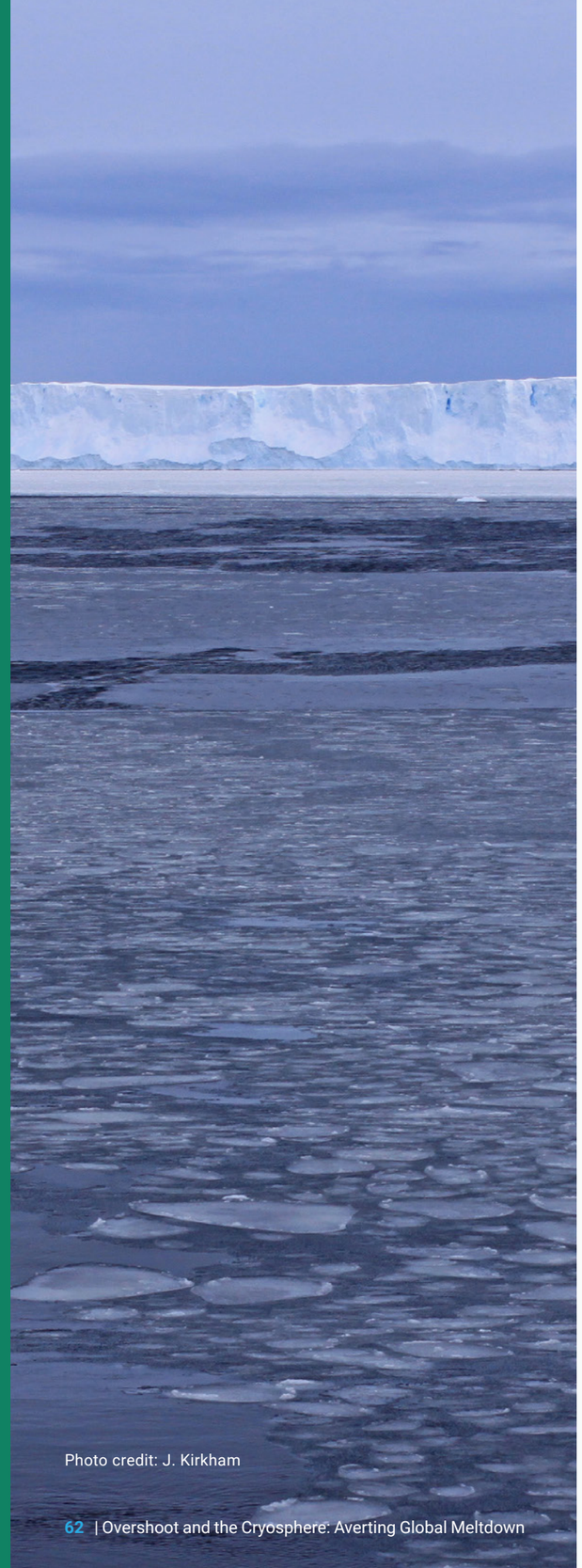


Photo credit: J. Kirkham

Polar-driven global ocean circulation systems are threatened by both warming and increased meltwater input. In the Arctic, higher levels of Greenland ice melt and river runoff have reduced the redistribution of heat, nutrients, oxygen and carbon by the oceans [232, 388-391]. Warming and freshwater forcing also contribute to concern over a weakening of the Atlantic Meridional Overturning Circulation (AMOC) [392-396], which plays a pivotal role in global climate regulation through the redistribution of heat between the tropical and polar regions. The AMOC is susceptible to large-scale instabilities — or 'tipping points' — that would have severe global climate impacts if triggered [75]. Evidence of this occurring is seen in both models [397-401] and palaeoclimate archives [402], and concerns have been raised about an irreversible threshold being crossed potentially this century [43, 401, 403, 404]. The likelihood and timing of a full AMOC collapse remain uncertain [395, 405, 406], but substantial weakening would affect regional climate, sea level, ecosystems and carbon uptake [401, 407].

Around Antarctica, meltwater-driven changes can substantially weaken Antarctic Bottom Water (AABW) formation and export [232] and alter the Antarctic Circumpolar and Antarctic Slope currents [408, 409], with potential impacts on global ocean circulation. These changes may also affect the ability of the ocean to take up CO_2 from the atmosphere, accelerating warming globally (see also Chapter 7: Interactions and Feedbacks in a Cryosphere under Overshoot) [361, 410]. Recent evidence further highlights that AMOC changes and variations in the Southern Ocean overturning are strongly linked, suggesting that disruptions in one region can amplify or alter circulation changes in the other [411].

Climate change is already impacting both polar oceans. The future of these important polar ocean processes will depend on the trajectory of greenhouse gas emissions and global temperatures, and their eventual stabilization levels.

Consequences of Overshooting 1.5°C

Ocean Warming

The polar oceans will continue to warm even after net-zero emissions are reached and global atmospheric temperatures stabilize [3, 239]. Warming is not just limited to the surface ocean; deep water warming will accelerate over the next 30 years especially under high emissions scenarios that overshoot to 3–4°C by 2100 [232]. The extent of ocean warming depends on the length and duration of overshoot and the stabilization temperature, but it is clear that the ocean will take longer to stabilize than the atmosphere due to delayed ocean heat uptake. In addition, changes in ocean circulation may sustain warming even after temperatures stabilize [239]. Continued ocean warming will add to long-term sea-level rise through both thermal expansion and by accelerating the loss of the polar ice sheets (also see Chapter 2: Polar Ice Sheets)[3].

In overshoot scenarios that return global average temperatures to 1.5°C by the year 2500, models suggest that substantial residual ocean surface warming persists in the polar and sub-polar waters, with highest residual temperatures in the North Atlantic (up to +3.1°C in the 3°C overshoot scenario compared to the 1.5°C stabilization scenario) and the Southern Ocean (+1.2°C) [251]. Persistent high-latitude ocean warming is also predicted due to long-lived impacts on Atlantic and Southern ocean overturning [251]. Scenarios that overshoot and then return to about 2°C by 2300 also show persistent warming, especially in the Southern Ocean, compared to scenarios with minimal overshoot and returning to 1.5°C by 2100 [412], as residual warming penetrates into the ocean interior via North Atlantic Deep Water, Antarctic Intermediate Water and AABW pathways [99].

Climate models also predict more frequent and intense marine heatwaves in both polar regions. Arctic marginal seas are proving to be surface marine heatwave hotspots. Subsurface heatwave events are likely just as intense and longer-lasting, but remain under-studied [413]. Processes unique to these regions shape polar marine heatwaves, including sea ice impacts, stratification of the water column, and upward fluxes of warm subsurface water [413].

The frequency and duration of marine heatwaves under scenarios with both minimal and near-3°C overshoot by the end of the century is greater in the Arctic than the global average due to Arctic amplification processes, i.e., the much greater warming there [414]. For the Southern Ocean, projections with minimal or near-3°C overshoot also indicate an intensification of warming and marine heatwaves, particularly between 40°S and 55°S in the Antarctic Circumpolar Current region. These changes are projected to be stronger with overshoot nearing 3°C, while minimal overshoot with temperatures returning below 1.5°C by 2100 would substantially reduce long-term heat stress on Southern Ocean ecosystems [415]. However, polar-specific studies have not yet quantified how marine heatwave frequency, duration, and intensity differ between overshoot and minimal overshoot pathways reaching the same endpoint, making this a priority knowledge gap.

Freshening

Overshoot scenarios affect polar ocean salinity through freshwater entering the ocean from sea ice, ice shelf and ice sheet melt. Even with warming held below 2°C, both the Greenland and Antarctic ice sheets are still expected to lose mass during this century, and longer-term critical thresholds for ice-sheet stability may lie near 1.5°C (see also Chapter 2: Polar Ice Sheets)[36]. In the Arctic, the dominant impact of overshoot is increased freshening by meltwater of the Greenland coastal ocean, Arctic gateways, and subpolar North Atlantic. Greenland runoff accelerates with warming because of Arctic amplification, expansion of ice sheet surface melt, and other feedbacks, with sustained ice loss occurring once global warming exceeds about 2.7°C [416]. Greenland meltwater input can freshen the subpolar North Atlantic, increase stratification, and weaken deep convection [417]. One study [418] suggests the North Atlantic can freshen even without meltwater impacts from ice sheets, implying that ocean-circulation changes alone can drive substantial decreases in salinity. In their overshoot scenario, which temporarily reaches 2°C of warming around mid-century before returning to 1.5°C by 2100, surface salinity in this region declines by ~0.57 g kg^{-1}. The Arctic also receives freshwater from the many rivers that drain into the Arctic Ocean basin, and discharge from these rivers seems to be increasing as the climate warms [419].

In Antarctica, in addition to surface melt, warming increases freshwater input from loss of ice shelves — floating platforms of ice that extend from the main ice sheet into the ocean around much of the continent. Even with 1.5–2°C warming, about 14–18% of Antarctic ice shelf area may become vulnerable to collapse from surface melt [420]. Melting of the undersides of Antarctic ice shelves is also expected to increase in response to warming oceans, especially in West Antarctica, even under 1.5–2°C scenarios [85]. Freshwater release from the entire Antarctic Ice Sheet is projected to increase only slightly (~12%) with low or minimal overshoot, but double by 2100 with global warming approaching 4°C [421].

Meltwater input both freshens the upper surface waters and strengthens stratification, meaning that the cold fresh water remains at the surface and prevents warmer subsurface ocean water from rising [322, 422]. This will reduce the formation of AABW, slowing global meridional overturning circulation and influencing global ocean currents [390, 423, 424].

At the same time, increases in salinity may occur below this fresh surface layer from relatively warm, saline Circumpolar Deep Water [423, 425]. Thus, responses of polar ocean salinity to overshoot scenarios can differ both vertically in the water column and regionally: surface freshening dominates freshwater-input regions, whereas greater salinity can emerge in other regions with greater intrusion of saline Circumpolar Deep Water [239]. Comparable vertical restructuring occurs in the Arctic, where fresh polar water may comingle with increasingly saline Atlantic-origin waters [426, 427].

This cold freshwater leads to surface cooling but subsurface water warming; the latter can increase melting of ice shelves from below and promote an amplifying feedback effect on ice-shelf retreat, while the surface cooling may act as a competing dampening feedback [322, 428]. By cooling the surface waters, this feedback may also preserve greater amounts of sea ice. One study showed 83% more Antarctic sea ice compared to simulations without freshwater forcing from warming to about 4°C by 2100 [429]. Another study found that with such high overshoot, models that include even modest Antarctic meltwater input dampens surface climate change in much of the Southern Hemisphere, reducing surface warming outside the tropics in 2100 by ~15–30% and Antarctic sea ice concentration loss by ~20–60% in some regions, compared to models that do not include meltwater feedbacks [430]. These initial studies emphasize the complexity of highly interconnected polar ocean systems and the urgent need for more research to understand responses in a changing climate.

Acidification

There is currently no viable or scalable practical way to reverse ocean acidification without removing CO_2 from the atmosphere. Natural buffering processes such as rock weathering operate over tens of thousands of years, and as long as CO_2 levels remain higher in the atmosphere than the ocean, acidification of the oceans will persist [431]. Even with atmospheric CO_2 removal, ocean acidification, especially in the polar regions, will not immediately return to previous historic levels, instead taking decades to centuries to recover [412, 432-434].

A 1.5°C stabilization scenario requires atmospheric CO_2 concentrations between 425–450 ppm by 2100, which means an increase in average global surface ocean acidity of about 50% compared to pre-industrial levels, with an average pH of 8.01 [418]. Both polar regions will have crossed critical thresholds for ocean acidification at these levels [387]. In practical terms, large parts of the Arctic and Southern oceans will spend increasing periods each year in conditions corrosive to shell-forming organisms, with direct consequences for food webs and fisheries.

Stabilization at 2°C will result in a global ocean acidity increase of about 78% (year 2100 compared to pre-industrial levels), with an average pH of 7.94 [418]. This would result in widespread expansion of corrosive waters, with over 50% of the Arctic surface waters exceeding the corrosive threshold level [435]. The Southern Ocean is also expected to see widespread expansion of corrosive waters by the end of the century under warming scenarios approaching 3°C by 2100 [436, 437]. These corrosive waters are already harming marine life by impairing shell-building foundational species such as pteropods that already show substantial shell damage across both polar oceans, and also at the global scale [438, 439]. Pteropods serve as important ocean acidification indicators and a source of food for economically important fish species including young salmon, Arctic cod and char [384, 440, 441]. In addition to long-term overall acidification, polar organisms will increasingly be exposed to seasonal and episodic extremes in pH and carbonate saturation, which may determine biological risk more directly than average annual conditions [442].



Photo credit: J. Kirkham

Relatively few analyses of the response and impacts of ocean acidification under overshoot scenarios have been done to-date [418, 432, 434]. One of the few studies looking at this suggests that a 2°C overshoot at mid-century, returning to 1.5°C by 2100 (i.e. CO_2 concentrations peaking at about 465 ppm and then declining back to around 430 ppm), would cause more ocean acidification compared to the 1.5°C stabilization scenario, with about an 88% increase in acidity compared to pre-industrial levels in mid-century, before returning to about a 54% increase in acidity by the end of the century [418]. However, these are global ocean acidification figures; acidification in the polar oceans will be greater and persist for longer. Recent model assessments show that acidification in the Arctic Ocean does not recover as CO_2 decreases [432, 443]. Instead, surface waters remain comparatively more acidic in scenarios where CO_2 emissions are ramped up and then decline, with corrosive conditions persisting until atmospheric CO_2 drops about 120 ppm below the corrosive threshold, requiring a drop from about 500 ppm at the peak of overshoot to about 380 ppm to reverse (380 ppm represents global temperatures close to 1°C of warming) [434]. Other work demonstrates that elevated acidification levels persist in the Arctic in 2300 in higher overshoot scenarios compared to scenarios where temperatures peak at about 1.8°C but decline to 1.5°C by 2100 [412]. This delayed and amplified acidification means that an overshoot pathway can leave polar waters more acidic, and corrosive conditions more persistent, than a pathway reaching the same end-of-century CO_2 or temperature level without a period of overshoot. Instead, coastal communities and ecosystems will face a prolonged period of heightened risk, even in scenarios where ambitious mitigation action is taken after overshoot.

Ocean Circulation

Even minimal overshoot scenarios may have a significant influence on ocean circulation changes compared to scenarios with no overshoot, including when temperatures return to 1.5°C by 2100. Ocean circulation impacts depend on the magnitude, duration, and timing of projected warming, ice melt, freshening, and stratification, because these take a significant time to stabilize in climate model simulations. Of particular concern are irreversible changes to the AMOC, the North Atlantic subpolar gyre, and the Southern Ocean overturning circulation [444]. For the Atlantic Ocean, some studies find that even an overshoot pathway that returns to 1.5°C by 2100 projects a weaker AMOC compared to a pathway with no overshoot of 1.5°C at all, and with little evidence of recovery with return to this lower temperature level [99, 418]. A key factor controlling these changes is not just how rapidly emissions are reduced [404], but also whether models simulate large-scale changes to polar ocean density structure and deep-water formation, which both are likely to be irreversible on near-term timescales.

Under continued warming, major ocean currents in the North Atlantic, most notably the AMOC and the North Atlantic subpolar gyre, are expected to continue to weaken [445], raising concerns about a threshold being crossed that is difficult to reverse [446, 447]. Weakened ocean currents in the North Atlantic would have significant impacts on regional climates, particularly in the northern hemisphere, but also across the planet. A substantially weaker AMOC would result in a profound cooling effect on Northwestern Europe due to severe (>80%) reductions in heat release from the North Atlantic to the atmosphere [407, 448], reduced precipitation [407], and more intense cold extremes, particularly in winter. Large day-to-day temperature fluctuations and an increase in the frequency of winter storms are also expected under a substantially weaker AMOC [407].

The rate, timing and extent of future AMOC weakening remains highly uncertain [405, 449], with some climate models predicting substantial weakening by the end of this century [450] while others imply only limited weakening [451]. Recent work has highlighted that AMOC stability is highly sensitive to the rate of emissions increase, with faster rates of CO_2 growth associated with AMOC collapse at lower temperature thresholds [404]. This makes rapid limiting of emissions crucial for reducing the risk of AMOC collapse. However, over longer timescales, ongoing weakening of the AMOC may be difficult to prevent, with some analyses suggesting that the AMOC could enter an extremely weak state beyond year 2100 even under intermediate or low emissions scenarios [396, 398].

Changes in the AMOC and the North Atlantic subpolar gyre occur on different timescales: the North Atlantic subpolar gyre can change in less than a decade, whereas a full AMOC collapse involves triggering larger-scale feedbacks that play out over centuries [88, 396, 452]. Long-term committed changes in the North Atlantic could include sustained freshening, reduced deep convection, weaker and shallower overturning, and greater sea-level rise in eastern North America in addition to a far colder and more extreme climate in northern Europe [453-456]. Upper ocean circulation changes might recover most rapidly with return to lower temperatures, provided freshwater forcing does not persist or tipping thresholds are not crossed [418]. Nonetheless, some models suggest that a collapse of deep convection in the North Atlantic subpolar gyre could occur even with very minimal overshoot (~1.6°C) and return below 1.5°C by 2100, with potential impacts on western Europe [448].

For the AMOC, with its likely slow onset to tipping behaviour and collapse, recovery over centuries may be possible if the overshoot — i.e., the duration of time spent above critical tipping thresholds — is short enough [88, 457]. Current Earth system models of AMOC strength following overshoot however exhibit a wide range of possible AMOC futures [458].

Some modelling studies [418] show a partial recovery, but not on the same timescale as the decline in atmospheric temperature because recovery depends on the evolving North Atlantic salt budget, not simply on the return of temperatures below a given threshold [457, 459]. The response of the AMOC to overshoot is highly sensitive to both the rate at which freshwater input increases and then decreases, and the initial strength of the AMOC before freshwater begins to decline [457, 460]. Faster declines in freshwater input after the peak are more likely to allow the AMOC to eventually recover. Conversely, prolonged exposure to high freshwater forcing significantly increases the risk of a transition to a collapsed state, which does not recover [457]. Palaeo-climate evidence suggests that rapid recovery from a weak to a strong AMOC state within a few centuries is also possible [461]. This may however be problematic for future societies adapted to climate conditions associated with a weaker AMOC.

Around Antarctica, the dominant ocean circulation risk is a weakening of Dense Shelf Water formation and AABW export, which ventilates the deep ocean and influences global heat, carbon, oxygen and nutrient storage [462]. Antarctic sea ice, a key driver of AABW formation, is expected to continue to decline for some period, even following the reversal of overshoot under net-zero emissions (see Chapter 4: Sea Ice) [104, 251]. Model simulations suggest that projected Antarctic meltwater input will reduce AABW formation [390, 423], open pathways for warm Circumpolar Deep Water to access the continental shelf, and warm and age the deep ocean, with impacts that could last for centuries [422]. Observations and modelling evidence support this finding, with several studies [463-465] reporting a substantial reduction in regional bottom water volumes linked to changes over the continental shelf. In overshoot scenarios, the abyssal ocean response is likely to be among the least reversible elements, because Antarctic Ice Sheet and ice shelf melt, and associated ocean freshening, can persist long after atmospheric temperatures stabilize, and deep ocean renewal occurs on centennial timescales.

Other Southern Ocean circulation systems can also be profoundly impacted by the melting of Antarctic ice. For example, the Antarctic Circumpolar Current is projected to weaken due to polar freshening, with one model exhibiting a slowdown of about 20% by 2050 under strong meltwater forcing [408, 423]. In that same model, the meltwater input can substantially strengthen the Antarctic Slope Current [409]. A resulting increase in warm eddy heat transport toward Antarctic ice shelves could further erode them, accelerating ice sheet loss.

In summary, polar ocean circulation changes under overshoot and return scenarios will vary greatly at different ocean depths. Surface and wind-driven features may respond relatively quickly to temperature return and stabilization, and some Atlantic overturning recovery is possible if the overshoot is short and freshwater forcing weakens.

However, the deep circulation components — convective overturning of North Atlantic Deep Water, the AMOC, Dense Shelf Water formation, AABW export, and deep ocean overturning — are much more likely to retain a long-lived overshoot legacy. The duration of overshoot, the cumulative freshwater input from the polar ice sheets, and the time spent near critical thresholds will all determine which circulation changes can be avoided, which changes may recover, and which changes will remain effectively locked in for centuries (Table 6.1; Figure 6.1).

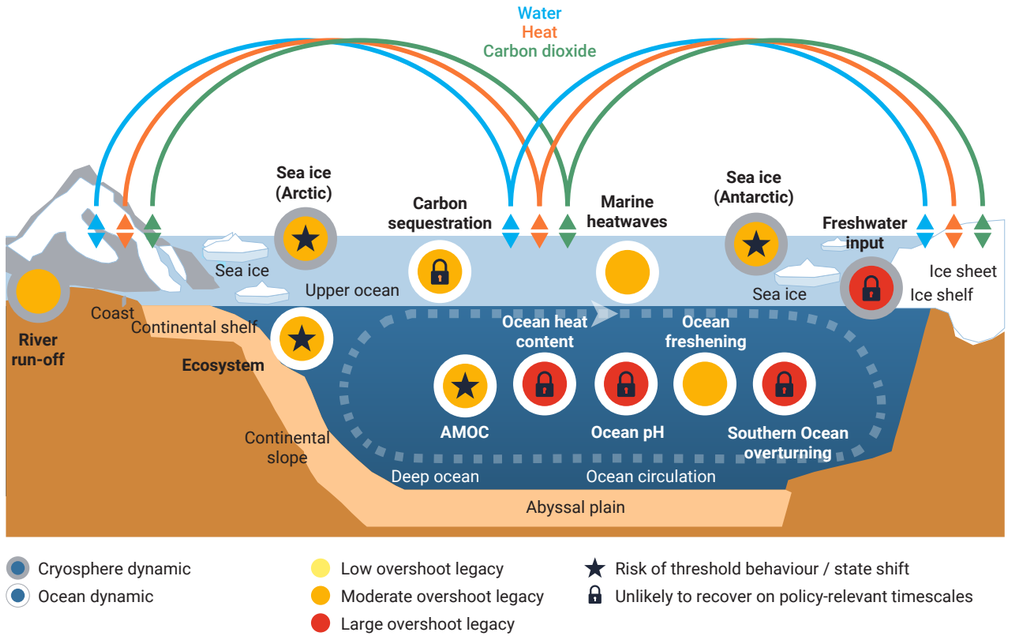


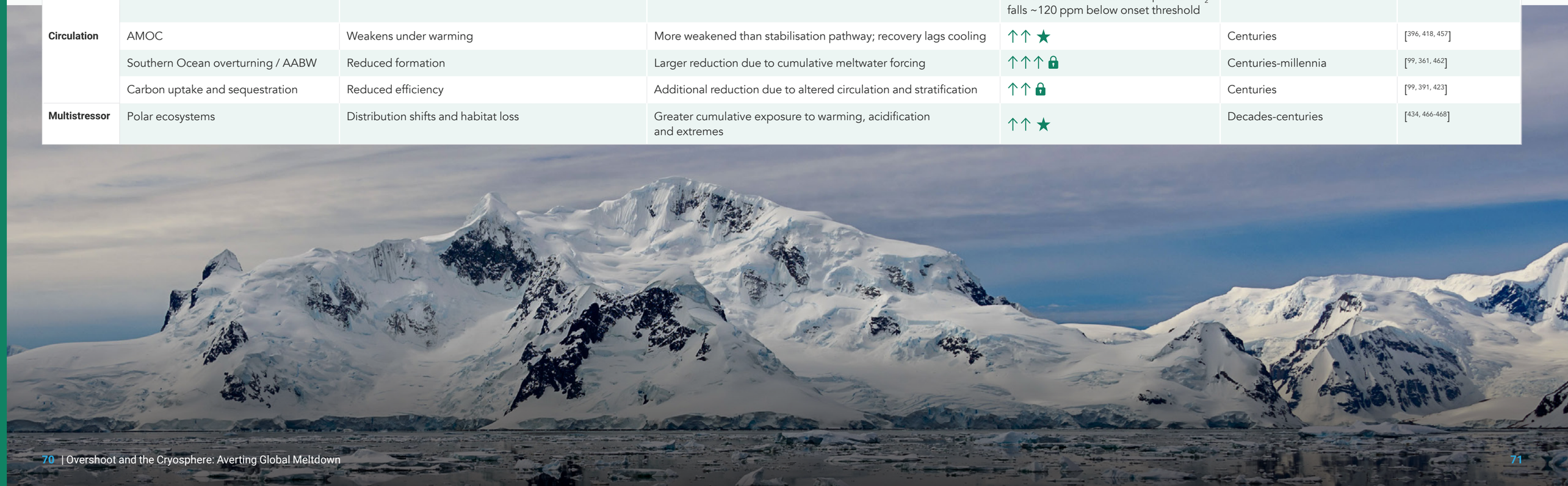
Figure 6.1: ***Overview of legacy impacts resulting from overshoot pathways that temporarily exceed 1.5°C above pre-industrial levels on various components of the polar oceans compared to a temperature stabilization at 1.5°C or below.*** *Redrawn and adapted from IPCC[3] Box 1.1 Figure 1.*



Table 6.1: *Legacy impacts on the different components and dynamics of the polar oceans resulting from temporary climate overshoot exceeding 1.5°C above pre-industrial levels compared to a scenario in which global warming stabilizes around 1.5°C.*

Visual scale: ↑ = small overshoot legacy; ↑↑ = moderate overshoot legacy; ↑↑↑ = large overshoot legacy; ★ = risk of threshold behaviour / state shift; 🔒 = unlikely to reverse on policy-relevant timescales.

SYSTEM COMPONENT		STABILISATION PATHWAY (1.5°C ENDPOINT)	OVERSHOOT PATHWAY	DIFFERENCE ATTRIBUTABLE TO OVERSHOOT	RECOVERY TIMESCALE	KEY EVIDENCE [REF. NUMBER]
Warming	High-latitude ocean temperature	Warming gradually stabilizes	Residual warming remains after temperatures return to 1.5°C	↑↑↑🔒 North Atlantic up to +3.1°C; Southern Ocean up to +1.2°C warmer than stabilization pathway	Centuries	[99, 251, 412]
	Ocean heat content	Elevated heat storage	Greater cumulative heat uptake and deeper penetration	↑↑↑🔒	Centuries-millennia	[99, 251, 412]
	Marine heatwaves	Increased frequency and duration	Additional extreme-event exposure during overshoot period	↑↑	Decades-centuries	[371, 414, 466]
Freshening	Surface salinity / freshening	Freshening under warming	Additional meltwater input and circulation adjustment	↑↑ ~0.57 g kg^{-1} additional freshening in one overshoot simulation	Century+	[417, 418]
	Antarctic freshwater input	Elevated freshwater fluxes	Greater cumulative ice-sheet and ice-shelf melt	↑↑↑🔒	Centuries	[421, 423]
Acidification	Ocean pH	Persistent acidification under 1.5°C	More severe peak acidification and delayed recovery	↑↑↑🔒 pH minimum ~7.91 versus less severe decline in stabilization pathway	Decades-centuries	[418, 432, 434]
	Corrosive waters	Expansion into polar surface waters	Larger extent and longer persistence	↑↑↑🔒 aragonite-corrosive conditions persist in the Arctic Ocean until atmospheric CO_2 falls ~120 ppm below onset threshold	Century+	[434, 435, 443]
Circulation	AMOC	Weakens under warming	More weakened than stabilisation pathway; recovery lags cooling	↑↑★	Centuries	[396, 418, 457]
	Southern Ocean overturning / AABW	Reduced formation	Larger reduction due to cumulative meltwater forcing	↑↑↑🔒	Centuries-millennia	[99, 361, 462]
	Carbon uptake and sequestration	Reduced efficiency	Additional reduction due to altered circulation and stratification	↑↑🔒	Centuries	[99, 391, 423]
Multistressor	Polar ecosystems	Distribution shifts and habitat loss	Greater cumulative exposure to warming, acidification and extremes	↑↑★	Decades-centuries	[434, 466-468]



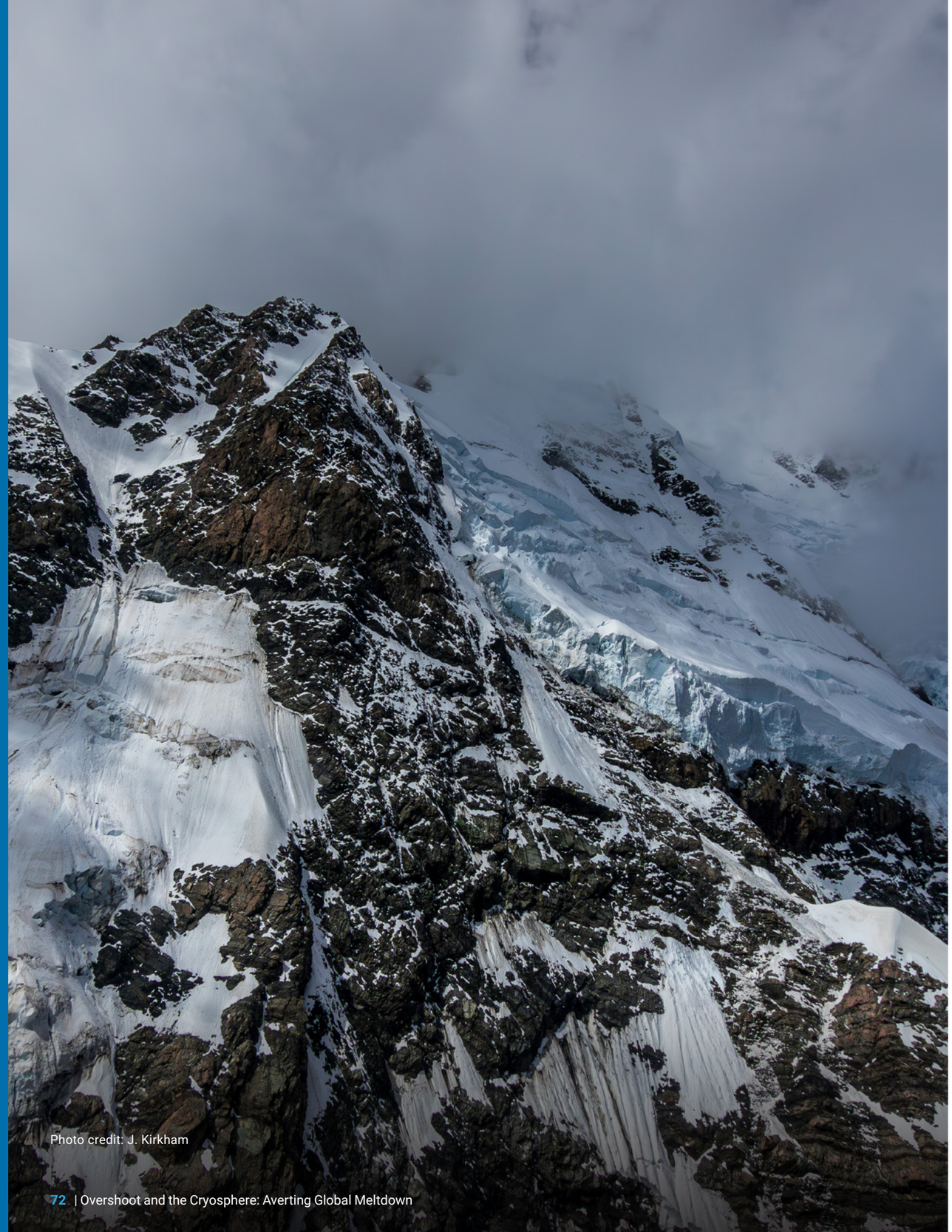


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07 Interactions and Feedbacks in a Cryosphere Under Overshoot

The response of the different elements of the cryosphere and the polar oceans described in Chapters 2–6 to overshooting 1.5°C will not occur in isolation. Rather, they will be subject to amplifying feedbacks and interactions that may lock in further irreversible losses and trigger cascading impacts. Effective policy measures must take into account these interactions, feedbacks, and increasing non-linear responses between various cryosphere elements under overshoot.

Interactions Between Cryosphere Elements

The cryosphere and polar oceans are closely connected. Ocean waters surround the Greenland and Antarctic ice sheets and directly influence the rate at which ice is lost, by several processes. Here, outlet glaciers often extend out into the ocean to form floating ice shelves that buttress the flow of ice towards the coast and into the ocean, supporting the stability of the larger inland ice sheets [46, 469, 470]. These ice shelves are highly vulnerable to ocean warming. Although ice loss in Greenland is dominated by surface melting from atmospheric warmth, intrusions of warm Atlantic water still play a key role in driving the retreat of remaining outlet glaciers extending into the ocean [29, 471]. In West Antarctica, intrusion of warm circumpolar deep water is widely understood to be the main driver of current ice loss in West Antarctica, and may also play a role in marine-based portions of East Antarctica [47, 472-474].

In Antarctica, committed ocean warming from past as well as current fossil fuel emissions is expected to continue for an extended period even after net zero greenhouse gas emissions are reached, resulting in unavoidable West Antarctic ice-shelf melting over the twenty-first century [85, 86].

During peak overshoot conditions, Antarctic ice shelves will also be exposed to greater surface melting [420] which risks triggering surface melt-albedo feedbacks that would maintain higher mass losses and reduce ice shelf stability long after global temperatures begin to decline [79, 475]. As ice sheet meltwater around Antarctica increases and sea-ice formation reduces, the production of Antarctic Bottom Water is set to slow [476], opening a pathway that allows warmer Circumpolar Deep Water greater access to the continental shelf and the margins of the Antarctic Ice Sheet [422, 462, 477]. Greater inflow of warm water will further enhance the basal melting of ice shelves, freshening the ocean in a self-reinforcing feedback [478, 479]. The combination of ocean surface freshening and reduced brine rejection from diminished Antarctic sea-ice formation during the winter freeze up will increase vertical stratification of the upper ocean layer, trapping warm water beneath the surface [480].

Antarctic sea ice also plays an important role in maintaining the stability of the margins of the ice sheet by protecting ice shelves from potentially damaging ocean waves and swells, and declining sea ice extent can also lead to warmer surface waters, increasing the rate of ocean-driven ice-shelf melt [481, 482]. Even a temporary loss of Antarctic sea ice during overshoot is therefore expected to reduce buffering of surface waves, leading to higher risk of ice shelf disintegration and increased Antarctic ice loss [218]. In addition, the decades-to-centuries greater exposure of the Southern Ocean surface waters during reduced sea-ice conditions to peak atmospheric CO_2 levels will drive severe, persistent ocean acidification long after atmospheric temperatures have stabilized, with long-lasting impacts on polar marine ecosystems [483].

In the Arctic, sea ice protects coastlines from damaging waves and its loss is leading to greater coastal erosion [213]. Arctic coastal erosion is also increasing due to permafrost thaw along coastal margins, as well as subsidence of previously frozen soils [484]. The decline of Arctic sea ice will create longer periods of open ocean water, increasing the heat uptake of the upper ocean and leading to higher rates of near-coastal sub-sea permafrost thaw, marine heatwaves and further sea-ice decline [3, 371, 485]. Decreasing sea ice cover and greater coastal permafrost erosion has also been shown to reduce the Arctic Ocean's capacity to take up atmospheric CO_2, thereby further increasing warming [267]. On land, permafrost vegetation feedbacks may delay the recovery of seasonal snowpack due to the influence of changing forest and tundra land cover, wildfire frequency, and dust and black carbon deposition [196-198]. These factors may potentially maintain the long-term effects of an overshoot period on Arctic snow- and glacier-related processes even after temperatures decline, delaying recovery.

Perhaps most seriously for efforts to return to lower temperatures, the additional release of greenhouse gases from permafrost thaw during overshoot will also drive further climate warming. In addition to increasing peak temperatures, these higher temperatures potentially may trigger other Earth system tipping points that would not have been crossed without this permafrost emissions feedback [92, 284]. Permafrost emissions will continue for centuries after initial thaw, which means these will need to be offset by even greater carbon drawdown [266].

In addition, freshening and warming of the polar oceans threaten the stability of large-scale ocean circulation systems. Ice lost from the polar ice sheets, combined with Arctic river runoff increase upper-ocean stratification and reduce the exchange of heat, nutrients, oxygen and carbon throughout the water column [388-391, 477]. Sea ice formation and melt also changes the salinity and thereby the density of seawater, and thus plays a critical role in driving global-scale ocean circulation [486, 487]. In the North Atlantic, increased freshwater input, alongside warming, has contributed to concern over weakening of the Atlantic Meridional Overturning Circulation (AMOC) [392-395], which would severely cool Northern Europe and affect weather patterns worldwide.

Given these feedbacks and connections, overshooting 1.5°C increases the risk of triggering a complex cascade of impacts between different cryosphere elements and other Earth system tipping elements [92, 284, 488]. These feedbacks essentially make the impacts more irreversible, when viewed as an interconnected system, compared to when individual elements of the cryosphere are considered in isolation.

Impacts on Cryosphere-Dependent Ecosystems

Overshoot will profoundly impact cryosphere-dependent ecosystems in both marine and terrestrial environments. On land, loss of mountain glaciers and declining snow cover will result in long-lasting disruption to ecosystems dependent on water from the mountain cryosphere. Changes in water availability will affect landscape and ecosystem structure such as wetland composition, food webs, and nutrient cycles including offshore and coastal ecosystems [199, 201]. Specialized species in glacier-dependent ecosystems are already facing severe habitat loss and local extinctions under current warming [148, 162, 489, 490]. Overshoot is likely to pose additional challenges to ecosystem adaptation, as species will need to adapt to both the warming phase and declining temperatures during the recovery period. Their response is likely to lag behind rates of glacier retreat by decades to centuries, further hampering the ability of alpine plant communities to successfully adapt [202, 491].

For ecosystems particularly reliant on glacier- and snow-fed runoff, water supply may need to shift entirely to terrestrial-derived runoff, at least on the timescale of many decades, dependent on the rate of glacier and snowpack recovery. This represents a fundamental regime shift for mountain and coastal ecosystems that may have long-term consequences for key species like salmon [108]. Continued glacier retreat is also expected to expose hundreds of square kilometres of new land that may produce new terrestrial, marine and freshwater ecosystems [148, 162, 492]. In deglaciated areas, these emerging ecosystems may offer refuge for other cold-adapted species unable to survive warming in other regions, but rates of species establishment can be extremely slow [202, 491]. Currently, less than half of glacial areas expected to retreat are located in protected areas [148]; this will pose emerging governance challenges in newly deglaciated areas exposed as a result of overshoot.

Overshoot will also fundamentally reshape hydrological networks in permafrost regions. Greater amounts of newly thawed ground during the overshoot temperature peak will trigger the expansion or drainage of lakes and wetlands [321, 327, 328], which may further impact specially adapted ecosystems [324].

In marine settings, loss of sea ice will have severe and potentially irreversible impacts on local ecosystems [248, 493]. Sea-ice loss and warmer inflows are increasing the frequency and duration of marine heatwaves [370, 371], which can alter primary production, plankton, and krill dynamics [32, 374, 494]. A warmer and fresher surface ocean strengthens stratification, limiting the vertical resupply of nutrients from colder, denser waters and threatening the base of Arctic food webs [371, 388, 389]. Greater light availability following sea-ice decline can initially increase primary production in some regions [495], but stronger stratification and nutrient limitation may limit productivity and alter phytoplankton composition, trophic transfer and biological carbon export [375, 496], leading for example to the algae and plankton blooms that have been seen off southeastern Greenland in recent years.



Warming also reduces oxygen solubility, driving rapid deoxygenation in Arctic gateway regions at rates up to six times the global mean, with circulation spreading this deoxygenated water into deeper layers [367]. Similarly, warming reduces the amount of CO_2 that seawater can hold, weakening the ocean's ability to absorb CO_2 from the atmosphere and store carbon in the ocean interior. Together, warming, freshening, acidification, and deoxygenation increase physiological stress on organisms and can weaken food web stability [497] and carbon sequestration. This matters not just for marine ecosystems, but for climate mitigation because absorption of carbon by marine organisms that descend into the deep ocean when they die stores carbon there on millennial timescales [362, 363], helping decrease atmospheric CO_2.

In the Southern Ocean, warming, marine heatwaves, acidification, and ice melt also generate coupled feedbacks. Marine heatwaves can alter primary production, plankton dynamics and carbon uptake by organisms [374, 494] and reduce the suitability of Antarctic krill habitat under future climate scenarios [290].

Regional sensitivity is not uniform: around the West Antarctic Peninsula, limiting warming below 2°C would reduce further sea-ice redistribution and land-ice retreat, whereas higher warming levels increase the risk of sustained ice loss and irreversible ecosystem damage, including local species extinctions such as penguin colonies on policy relevant timescales [32]. Projections from a coupled ocean–sea ice–biogeochemistry model indicate that, under strong warming of about 4°C by 2100, the rate of deep-ocean carbon sequestration in the southern Weddell Sea could decline abruptly by 2100 to only 40% of its 1990s level [498].

Overshoot therefore adds a legacy dimension that is not captured by the final temperature or atmospheric CO_2 endpoint alone. A polar ocean that has been exposed for decades to additional heat, freshwater input, acidification and ecosystem stress may not resemble a polar ocean that has reached the same endpoint without overshoot, particularly if irreversible critical thresholds or tipping points are crossed. The magnitude and duration of overshoot — and not just the eventual return to 1.5°C — are therefore critical for polar ocean ecosystem risks.

Irreversibility of Losses

The impact of overshooting 1.5°C on the cryosphere will be felt for centuries to millennia and the irreversible nature of cryosphere loss and its ensuing impacts is a central theme throughout this Report. Sea-level rise from ice loss and ocean thermal expansion will be irreversible on millennial timescales, even if temperatures later stabilize or even cool. Even 0.5–1 metre of future sea-level rise will result in catastrophic impacts to many vulnerable low-lying communities; especially Small Island Developing States (SIDS) and Least Developed Countries (LDCs), but also in many middle-income and OECD nations exposed to coastal risks. Around 250 million people already occupy land less than 1 metre above current high tide lines [61]. In mountain regions and communities downstream, even temporary overshoot of 1.5°C will irreversibly decrease glacier ice reserves on human timescales, with recovery taking decades to centuries, providing that snowfall and precipitation can even recover following temperature decline. Once thawed, the carbon released from permafrost during overshoot is effectively lost irreversibly even if temperatures later reduce, and vegetation, fire and hydrological regimes in former permafrost lands will take decades to centuries to recover even after the ground refreezes permanently with temperature decline. Similarly, the physical, chemical and ecological changes that overshoot will drive in the polar oceans will take centuries or longer to recover even with declining atmospheric CO_2 levels and temperatures. While some components of the cryosphere like seasonal snow cover and Arctic sea ice may recover faster as global temperatures decrease, the secondary impacts of snow and sea ice loss during the overshoot period can also trigger abrupt and irreversible harm to species, ecosystems, coastlines, Indigenous cultures, human settlements and other components of the cryosphere. Many of those in harm's way, in coastal and downstream communities, already exist near the limits of adaptation.

Overshoot will therefore leave a persistent and irreversible legacy on the cryosphere and the human and natural systems dependent on it. It is therefore crucial to minimize the extent and duration of overshoot to slow or avert these essentially permanent global impacts from cryosphere loss on society and the planet.



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08 Recommendations

Policy Recommendations
A rapid, ambitious, and coordinated policy response focused on known solutions

- **The reality of irreversible impacts, including those which may exceed the limits of adaptation leading to growing loss and damage, needs to be made clear to policy makers and the general public.** There remains a persistent myth that overshoot to 2°C, 2.5°C and 3°C and above is manageable, or even preferable to the "costs" of urgent mitigation. The global impacts of cryosphere loss — including those that are already occurring such as the 2026 Nepal glacier/mountain collapse disaster — make clear this is not the case, even in the near-term.

- **Efforts must be taken to minimize overshoot peak and duration through urgent reductions in carbon emissions, which primarily arise from use of CO_2-rich fossil fuel sources for energy, transport and manufacturing.** This includes over-performance on current Nationally Determined Contributions to reach net zero greenhouse gas emissions by 2050. Greater detail on sectoral priorities, and the Highest Possible Ambition (HPA) Pathways[24] are available in publications such as the 2026 UNEP Emissions Gap Report[7].

- **Greater investment and research is urgently needed to scale up existing and incoming clean and sustainable carbon removal mechanisms.** An urgent need exists for not just reaching net zero greenhouse gas emissions, but net negative greenhouse gas emissions through declining atmospheric CO_2 levels to bring temperatures back to 1°C or below in the next century to slow or prevent cryosphere-mediated loss and damage. Carbon sequestration and drawdown should be reserved for this purpose rather than to offset continued excessive use of fossil fuels.

- **In addition to high-CO_2 carbon sources like fossil fuel emissions, reductions in methane and, especially for the cryosphere, black carbon emissions should be urgently addressed.** Black carbon emissions reductions that impact snow and ice have demonstrably increased melting and cryosphere loss, and their reduction has strong health and food security co-benefits for local populations.

- **Great emphasis should be placed on improving measures to address the actual cause of warming carbon emissions such as the greater roll out of renewables and the expansion of sustainable carbon drawdown options.** Greater resources additionally are needed to support feasible adaptation and, where cryosphere impacts exceed the limits of adaptation such as extreme land and glacier collapse events, desertification or irreversible sea-level rise, supporting populations harmed by now-inevitable loss and damage from overshoot. Other examples include water resource changes and loss affecting agriculture and energy generation, loss of sea ice transport pathways and ecosystems. Such financial support is especially essential for Indigenous Peoples facing loss of cultural heritage and resources.

- **Investment into concepts that only attempt to mask carbon emissions or purportedly hold back cryosphere loss should be re-directed into the improvement and scale-up of known feasible solutions, in particular carbon drawdown, given known negative side effects[499], lack of feasibility and sustainability[500], and especially damage to global oceans burdened with additional acidification from the continued rise in CO_2 concentrations.** Approaches that set region against region, or cryosphere loss against ocean damage, are especially dangerous to effective and sufficiently rapid action and to global security.

- **Urgent investment is also needed into the development and implementation of Early Warning Systems, both in terms of protection of local downstream, coastal and permafrost-based communities and in more remote regions with potential for unusually rapid global impact.** The latter include the close monitoring of marine-based portions of the Antarctic Ice Sheet, ocean current conditions near Greenland, Antarctica and in the North Atlantic, and remaining tracts of land and near-coastal permafrost for signs of abrupt thaw. (More detail on research priorities are outlined below.) These efforts are essential to provide as much preservation of life, property and cultural heritage as possible, even in the face of events that may ultimately prove so extreme as to overwhelm such systems. Mitigation efforts must be integrated and combined with improved disaster awareness and adaptation to improve resilience.

- **Given the many inevitable impacts of cryosphere loss even with brief and limited overshoot, planning should now begin to adequately meet projected adaptation needs, as well as support for loss and damage financing, especially where relocation appears necessary.** This means both private and public investment, including the Green Climate Fund, Adaptation Fund, the Loss and Damage Fund, the Global Environmental Facility and programs of the multilateral development banks.

- **Planning for protection of cryosphere-impacted ecosystems, such as those vulnerable to desertification from snowpack and glacier loss, riverine and coastline flooding, high mountain ecosystems, sea ice-based ecosystems and others is also essential, especially to protect threatened biodiversity, species threatened with extinction, and new ecosystems that emerge as glaciers retreat.** Possible steps may include the creation of protected areas and the recognition and integration of glacial and postglacial ecosystems into national and international legal systems (laws, treaties, and public policies, for example).

- **Cryosphere and the implications of its loss must be included explicitly in UNFCCC, UNEA, and other UN processes.** Despite the 2019 IPCC Special Report on Oceans and Cryosphere in a Changing Climate (SROCC) and attempts by individual governments and the scientific community to raise these concerns, cryosphere has remained essentially absent from UNFCCC processes and negotiation tracks, except for occasional mention as one among many components of the global climate system. Given the global and irreversible impacts of overshoot from cryosphere degradation, these physical realities need to be explicitly incorporated across negotiation tracks (especially the Second Global Stocktake and Third Periodic Review) and also within relevant UN bodies, and in regional and national planning and governance.

Research Recommendations

The impact of global warming on cryosphere loss is already well understood. Risks escalate with every fraction of a degree of warming, and a continued failure of political action will lock in further losses and negative impacts on populations and nature for decades to centuries. However, the concept of overshooting and then returning below 1.5°C requires urgent further investigation to ensure that policies relating to mitigation, adaptation, resilience building and loss and damage are informed by the best available science in the face of overshoot.

Large multi-national, multi-stakeholder, and multidisciplinary efforts such as the International Polar Year, the Decade of Action for Cryospheric Sciences, Antarctica InSync, concerted numerical modelling efforts such as the Ice Sheet Modelling Intercomparison Project (ISMIP) and the Tipping Point Modelling Intercomparison Project (TIPMIP) and others provide prime areas where critical measurements and research questions relating to overshoot can be addressed.

Key Recommendations

- Successfully advancing knowledge on this topic will depend on sustained, improved, and well-coordinated observational networks: without this, research risks being fragmented, misdirected and/or inaccurate.

- More simulations of how the cryosphere will respond to overshoot and subsequent temperature decline are needed. In particular, understanding of the potential for the cryosphere to surpass irreversible thresholds or tipping points under overshoot, lock in amplifying feedbacks, and trigger cascading impacts on other elements of the climate system needs to be improved.

- Due to the extremely long timescales that cryosphere loss can unfold over in response to even low levels of warming, simulations will need to extend over multiple centuries (i.e., to the year 2300 and beyond) in order to truly visualize the impact of overshoot on cryosphere loss. This will require a coordinated modelling effort utilising multiple model ensembles that examine the response of the cryosphere to a range of plausible peak temperatures and overshoot durations. The new generation of greenhouse gas emission and concentration scenarios used for global projections under CMIP7 provide an excellent foundation for this work [26].

There are no benefits to exceeding and remaining above 1.5°C [8], but with some amount of overshoot now unavoidable, research must now examine what 'safer' overshoot pathways look like, in terms of:

- Losses that can be reduced or avoided by a swift peak-and-decline pathway, trade-offs between the peak warming reached and the duration of overshoot, and the ability to reduce the risk or prevent the crossing of tipping points if global temperatures can be lowered within a certain timeframe.

- Better quantification of the risk and damage caused by higher levels of overshoot and delayed climate action will also be highly relevant information for shaping policy, including the sources and magnitude of deep uncertainties associated with these analyses and estimates.

- Impacts of undertaking the highest possible ambition in terms of climate action, where temperatures decline below 1.5°C and — as this reports notes as necessary — to around 1.0°C or below in terms of maintaining cryosphere stability and lowering the scale of global impacts [27, 107, 108, 501].

Even in a world where temperatures can be returned to 1.5°C, populations must be prepared for extensive cryosphere losses and escalating impacts that will continue through the coming decades and centuries:

- Adaptation and disaster prevention measures must not only be fit for purpose in reducing overshoot risks, but must include impacted communities from the beginning and be embedded within the socio-cultural contexts in which such measures are designed for and implemented [150]. This is especially important given the impacts of cryosphere loss disproportionately affect communities with fewer financial, institutional and technical resources to adapt.

- Research should explore the human dimensions of adaptation and resilience in overshoot – including their limits, given the already lived experiences with cryosphere loss and the expected future continuation of those changes. Additionally, research on the governance and institutional regimes to manage emerging landscapes as glaciers retreat, are particularly relevant in view of adaptation trade-offs that are necessary under accelerating climate change [148].

- Research that purposefully draws on anticipatory approaches for future-oriented planning, thereby moving beyond reactive and crisis-driven responses, is fundamental if scientifically rigorous research insights are to be relevant, inclusive, and actionable [150].

Specific Research Priorities

Ice Sheets

To better constrain estimates of how the polar ice sheets will respond to overshoot, more studies are required that systematically vary both the peak forcing (e.g., in ocean or air temperatures) and the duration of overshoot, and examine how these factors affect sea-level rise outcomes over many centuries.

Improvements in the precision and certainty surrounding critical ice sheet thresholds (both in terms of absolute temperature and time required to trigger threshold responses) are needed. These factors may vary between different ice-sheet sectors [80]. Efforts should focus on improving understanding and representation of critical processes including subglacial hydrology, instabilities related to ice sheet geometry, ice shelf hydrology and calving, ice-ocean interactions (including within ice shelf cavities), feedbacks and teleconnections between ice sheets and other parts of the climate system, and more detailed bed geometry measurements at scales useful to ice sheet and ocean modelling, particularly around and upstream of current grounding zones [17, 84, 479, 502-504]. The role of short-term (decadal or shorter) variations in atmospheric or ocean conditions as amplifiers of longer-term ice sheet change should also be refined [29].

Mountain Cryosphere

Determining with confidence how mountain glaciers, snowpack and permafrost will respond to overshoot requires improving representation of ice dynamics and local feedback mechanisms within numerical models, which may otherwise struggle to capture the full complexity of tipping dynamics, such as topographic controls, elevation or albedo mass balance feedbacks, and physical glacier fragmentation [189, 505]. Many feedbacks that could amplify ice loss and dampen glacier regrowth following overshoot – such as surface darkening from black carbon and snow loss [506], dust and algae, as well as accelerated calving [507] – are not yet fully integrated into large-scale glacier models, suggesting that current projections of glacier loss are likely conservative [157]. The use of higher-order 3D ice flow models should be prioritized to accurately resolve threshold-crossing behaviours and potential irreversibility across all glaciated regions.

Emphasis should be placed on improving understanding of the hydrological risks and natural hazards posed by glacier decline, snow loss and mountain permafrost destabilization in response to overshoot, especially at the regional level. Although many cryospheric processes exhibit decadal-scale hysteresis, meaning cooling the planet will not immediately return water systems to their pre-warming states, the long-term impacts on mountain water resources remain poorly understood [199]. One consequence is a temporary period of depleted glacier runoff that occurs as glaciers begin to regrow [157]. Failing to evaluate possible impacts of overshoot trajectories on water supply may lead to maladaptive governance and infrastructure decisions, as systems designed for steady warming may prove ineffective under the unique conditions of a post-overshoot cooling phase. Research is therefore needed on actionable adaptation options for mountain contexts to manage risks, specifically exploring the human dimensions of adaptation and resilience in overshoot – including their limits.

These research needs include the types of governance and anticipatory approaches needed to ensure fit-for-purpose planning and the effective management of future risks [149, 150].

Further assessment of seasonal snow reversibility in overshoot peak-and-decline pathways should include both temperature and precipitation reversibility, as well as land cover processes that affect snow accumulation and melt. Cascading impacts of snow loss on processes such as fire activity and vegetation cover also require better quantification [508]. Further research is needed into the implications of snow loss and potential recovery for adaptation strategies, both in terms of annual water management decisions [509, 510] and long-lived water infrastructure [511-513]. The benefits and trade-offs of these approaches to managing snow-dependent water resources in overshoot peak-and-decline scenarios are presently completely unquantified.



Photo credit: J. Kirkham

Sea Ice

Improving understanding of the response of sea ice to overshoot requires a coordinated modelling effort utilising multiple model ensembles across long timescales and sustained observations (which should also include research collaboration with local and Indigenous communities [514]). To date, all overshoot simulations of the response of sea ice to overshoot have relied on single model simulations. The inclusion of overshoot scenarios in CMIP7 [26], will support more robust multi-model studies in the future for Arctic and Antarctic sea ice and the wider climate system response. Importantly, future Antarctic sea ice simulations need to be assessed carefully to determine if they accurately represent the observed sea ice evolution in the Southern Ocean over the observational period, in order to improve confidence in future Antarctic sea ice projections.

Future simulations of the impact of overshoot should also take into account the long equilibration timescale of the climate system, including sea ice, which continues to hinder our ability to quantify the equilibrium response of Antarctic and Arctic sea ice to changes in global-mean temperature. For example, ocean dynamics and memory mean that it is possible for interior ocean warmth to resurface and cause rapid loss of Antarctic sea ice centuries after peak warming [515].

The potential timescales and magnitudes of these impacts are not well understood and require further investigation. In addition, a number of important impacts and feedbacks associated with sea ice loss are not currently captured in global climate models. These processes will need to be evaluated and accounted for outside of the CMIP7 projections. For example, the impact of sea ice on iceberg calving and ice shelf disintegration is not captured, but recent observational evidence [482, 516] suggests that this is a significant effect and will affect the future evolution of the Antarctic Ice Sheet. Landfast sea ice, i.e., ice that is "stuck fast" to coastal topography or ice shelves, is also only included in a small number of sea ice models and is currently not represented in coupled climate models.

However, for accurate projections of future sea ice cover, especially at high resolution, including landfast sea ice in coupled climate models is important. Overall, future research should prioritize identifying and mitigating persistent biases within current climate models, and assessing the effect of different magnitudes and durations of overshoots on Arctic and Antarctic sea ice through coordinated multi-model overshoot simulations across long timescales.



Permafrost

For permafrost, attention should focus on fully accounting for permafrost emissions when estimating the remaining global carbon budget, understanding how vegetation mediates permafrost responses to temperature overshoot, and how hydrological changes will affect water resources and the transport of sediments and nutrients. Representation of permafrost carbon in Earth System Models (ESMs) will need to be improved [517]. For example, only 2 of 11 CMIP6 ESMs used in IPCC AR6 include permafrost carbon. These assessments also focus only on CO_2 and methane emissions, but nitrous oxide emissions from thawing permafrost carry higher global warming potential and remain poorly constrained. Most large-scale permafrost carbon models still represent only gradual thaw [517], while missing abrupt thaw processes (e.g., thermokarst, thermal erosion, thaw slumps, lake drainage), fire processes, and the impact of saline coastal permafrost thaw that can increase carbon release substantially [280, 518].

Furthermore, vegetation is currently only crudely represented in most models despite it being an important control on the response of permafrost to overshoot. To improve understanding, the magnitude of permafrost hysteresis in response to overshoot should be better quantified, along with the asymmetry between rapid vegetation establishment during the warming peak and slow die-back during the cooling phase of overshoot, and the role that disturbance–thaw feedbacks such as fire may have on permafrost recovery. Further research should therefore prioritize better resolving canopy structure in coupled models, observationally constraining recovery timescales against flux-tower and remote-sensing records, and quantifying how much committed, irreversible carbon loss these vegetation-mediated processes add beyond current estimates.

In addition, estimates of the extent of permafrost-dependent infrastructure that is at risk of damage due to overshoot need to be improved [519].

Similarly, linkages between permafrost thaw, water resources and the transport of sediments, nutrients and dissolved organic carbon into rivers, lakes, and downstream aquatic ecosystems, are currently poorly represented in ESMs — especially changes in groundwater to surface water connectivity, abrupt thaw processes, and coupled water–carbon interactions. Future research should prioritize long-term hydrological observations across Arctic, Sub-Arctic, and high mountain permafrost regions, together with improved representation of hydrological processes and their interactions with carbon cycling in ESMs, to better assess future risks under both temperature stabilization and overshoot scenarios.



Photo credit: G. Kappenberger

Polar Oceans

Addressing the implications of climate overshoot for the polar oceans requires improved understanding of how physical, chemical and biological changes interact, how rapidly they may unfold, and which changes can be reversed following a return below 1.5°C. A priority is to better constrain the drivers, characteristics and consequences of polar marine heatwaves, including their coupling between the surface and subsurface ocean. Research should assess how marine heatwaves affect ecosystems and ocean carbon uptake, and whether such impacts persist or amplify under overshoot scenarios. Improved understanding of mixed-layer depth variability and stratification is also needed, particularly because changes in these processes strongly influence upper-ocean heat storage and exchange. For ocean acidification, research should address how overshoot affects the magnitude, spatial distribution and persistence of acidification in polar waters, and how these changes will affect vulnerable organisms, ecosystems and biogeochemical processes. Changes in large-scale ocean circulation, particularly in the polar regions, also require greater attention. Priorities include better constraining how freshwater input, sea-ice loss and ocean warming alter stratification, water-mass formation and transport, including Antarctic Bottom Water and Southern Ocean overturning, as well as Arctic circulation and exchanges with lower latitudes. Improved understanding is needed of the potential for these changes to influence global ocean circulation and climate, including possible nonlinear responses and thresholds. In addition, the long-term response of ocean circulation systems like the AMOC to scenarios in which temperatures fall below 1°C in the next century requires greater attention, congruent with research that indicates 1°C as a safer stabilization temperature for other cryosphere elements such as the polar ice sheets [27, 107]. Greater emphasis should be placed on interactions and feedbacks across the polar ocean–cryosphere–atmosphere system.

This includes the effects of Arctic river discharge on freshening, nutrient and carbon export, and primary production; feedbacks between sea-ice change and marine carbon uptake and storage; how acidification interacts with ocean warming, freshening and sea-ice loss; and interactions between ocean warming, circulation, ice-shelf melt and ice-sheet stability. Earth system models should therefore better represent coupled ice–ocean processes and incorporate improved biogeochemistry and ecosystem dynamics. Identifying where, when and under what conditions nonlinear changes or tipping behaviour may occur is particularly important for assessing overshoot risks.

The Essential Role of Observations and Improved Modelling

These advances all depend on sustained observations and continued numerical modelling efforts. Cryosphere observing systems should be expanded to enable earlier detection of change, with greater coverage of the subsurface ocean and year-round measurements, particularly during winter. Autonomous observing technologies can help overcome logistical constraints but require sustained, long-term investment. Additional coordinating capacity, capability building and data sharing are needed to strengthen observation networks that can also involve local and Indigenous communities. Finally, research should strengthen the evidence base for local and regional adaptation and ecosystem protection, including assessment of feasible mitigation measures and nature-based approaches. Such assessments should explicitly account for the extreme vulnerability, remoteness and limited resilience of polar and mountain cryosphere ecosystems.



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