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Elevation dependent climate change in mountain environments

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Abstract

Mountain regions exhibit rapid environmental changes under anthropogenic warming. The rates of these changes are often stratified by elevation, leading to elevation-dependent climate change (EDCC). In this review, we examine evidence of systematic change in the elevation profiles of air temperature and precipitation (including snow). On a global scale, mountain–lowland trend differences for temperature, precipitation and snowfall are 0.21 °C century^{-1} (enhanced mountain warming), –11.5 mm century^{-1} (enhanced mountain drying), and –25.6 mm century^{-1} (enhanced mountain snow loss), respectively for 1980–2020 based on averaging available datasets. This EDCC is primarily driven by changes in surface albedo, specific humidity and atmospheric aerosol concentrations. Throughout the 21st century, most models predict that enhanced warming in mountain regions will continue (at 0.13°C century^{-1}), but precipitation changes are less certain. Superimposed upon these global trends, EDCC patterns can vary substantially between mountain regions. Patterns in the Rockies and Tibetan plateau are more consistent with the global mean than other regions. In situ mountain observations are skewed towards low elevations and understanding of EDCC is biased towards mid-latitudes. Efforts to address this uneven data distribution and increase the spatial and temporal resolution of models of mountain processes are urgently needed to understand the impacts of EDCC on ecological and hydrological systems.

[H1] Introduction

Although global mean air temperature has increased ~1.1°C since pre-industrial levels[1], the rates of anthropogenic warming are highly spatially heterogeneous. Arctic regions, in particular, have experienced warming 2-4 times the global average[2] owing to ice–albedo[3,4], snow–albedo[5] and vegetation–atmosphere feedbacks[6,7]—a process termed Arctic Amplification. As elevated and mountain regions share many thermal characteristics with polar regions[8,9], they too tend to exhibit warming rates higher than the global average[10], albeit with complexity related to topographically-driven microclimates[11]. Such positive elevational dependent warming (EDW) is evident at historical[12] and longer[13] timescales, and despite much variation according to time of day, season and location[14], mean warming at high elevations is ~50% faster than the global mean since 1950(ref.[15]).

However, elevation dependence is not restricted to temperature. Broader elevation dependent climate change (EDCC) is becoming apparent (**Fig. 1**), encompassing atmospheric moisture content[16], precipitation[17,18], snow cover and surface albedo[19,20], radiation and energy balances[21], and wind regimes[22]. Indeed, gridded datasets demonstrate systematic changes in precipitation and snowfall with elevation.

EDCC will have marked local, regional and global impacts. For instance, over 1 billion people depend, at least in part, on mountain snow and glaciers as a storage for water resources[23–25], strongly affected by changes in temperature and precipitation regimes. Relatedly, diminishing snow[26–28] and glaciers[29,30] make snowmelt-driven regimes flashier, pluvial and, thus, less predictable[31], increasing the likelihood of flooding[32] and drought[33]. EDCC will also change the potential of mountain hazards such as landslides[34,35], rockfalls, avalanches[36] and glacial lake outburst floods (GLOFs)[37–39]. EDCC could also exacerbate extreme events in mountain regions[40–42], including heat waves[43], rain on snow events[44], forest fires[45], flooding[46,47] and pest outbreaks[48], each with their own impacts. Yet, despite the substantial impacts, EDCC remains little examined.

In this Review, we synthesise knowledge of EDCC in a range of climate variables. We begin by outlining physically based perspectives on EDCC for temperature and precipitation (including snow-related variables). We next document observed patterns of historical EDCC across mountain ranges, before turning to model predictions of future changes. In the final section, we outline future research priorities.

[H1] Physical-based perspectives on EDW and EDCC

EDCC refers to any systematic change in climate-related trends across an elevation gradient and can, thus, be broadly applied. The elevation gradient refers to the slope surface and not height in the free atmosphere. EDCC can be used to assess so-called response variables (for instance, temperature or precipitation) and drivers (for instance, snow albedo or downwelling longwave radiation); the distinction between response and driver is blurred by absence of a universal definition (for instance, surface albedo can be either driver or response, dependent on perspective). Here, expected changes in temperature, precipitation and snow (which includes albedo)—the most commonly-assessed variables—are examined from expert review using first principles (**Fig. 2**). In each case, profiles represent a general expectation, with understanding that local conditions could diverge given mountain complexities. For each reference transect, variables are evaluated at the mountain surface (along the elevation gradient). Although largely restricted to temperature and precipitation to date, other variables can be expected to exhibit EDCC (**Box 1**).

[H2] Temperature

Surface air temperature warming is anticipated to be faster at higher elevations, representing positive EDW (**Fig 2a**). Two prevailing types of EDW profile are expected: an overall increase in warming rate with elevation; a peak warming rate concentrated in a specific elevation band; or some combination thereof. Enhanced warming rates are observed at mean annual temperatures ~0°C(ref.[11]). As the 0°C isotherm broadly corresponds with the mean annual snowline[20], this enhanced warming supports snow cover loss (or the snow albedo feedback; SAF) as the dominant EDW driver[49], particularly near the freezing point. SAF has an important role as snow ages and melts[50]. Conversely, increases in downward longwave radiation (DLR), which is non-linearly dependent on specific humidity (**Box 1**)[51,52] are expected to promote increased warming at higher elevations. Reductions in aerosol loads can increase the solar flux reaching the surface. Although sometimes transported to high altitudes[53], aerosols usually impact low elevations more strongly[54,55] ; thus, their reduction causes increased warming at lower elevations. Owing to the difference balance of surface albedo, specific humidity or changes in other variables[10], the shape of temperature trend–elevation curves will vary spatially and temporally.

[H2] Precipitation

Mean precipitation is generally theorised to increase in a warmer world (**Fig. 2b**). Mean precipitation (climatology) peaks around the lifting condensation level (LCL) in an idealised convective system. Orographic or convective lifting above this level results in precipitation, but the precipitation recorded at the surface will decrease with elevation above the LCL. Below the LCL, not all air parcels have the required uplift to reach the LCL and form precipitation, so precipitation decreases. Although the LCL varies from day to day, there is a climatological mean applicable to most locations, broadly coincident with the elevation of maximum precipitation on the mountain slope[56]. Based on increasing temperatures alone, water vapour will condense at higher altitudes, raising the mean LCL. This change would lead to an increase in precipitation above (and just below) the new LCL, and a decrease in precipitation below (and just above) the old. However, rising temperatures will also cause an increase in total precipitable water in the atmospheric column[57], shifting the change curve further towards increasing precipitation (**Fig. 2b**). The exact elevation at which precipitation increases rather than decreases depends on the balance between the new LCL being reached less often by moist near surface air parcels, and the larger total precipitable water available, with potential for more intense maximum precipitation[58]. This argument is most applicable in the tropics, where convective processes with strong vertical lifting produce most precipitation[59,60]. At mid- to-high latitudes, precipitation changes on slopes with strong horizontal advection are more complex, with strong contrasts between windward and leeward slopes being influenced by shifting synoptic patterns[61]. Changes in upslope winds (which contribute to moist air lifting well above the LCL) or downslope winds (which subdue precipitation through enforced subsidence) add to the complexity of change.

[H2] Snow Cover and Surface Albedo

Snow cover is expected to retreat to higher elevations in a warming climate, decreasing surface albedo at the current snowline **(Fig. 2c)**. Indeed, as snow shifts to higher elevations and a future (higher) snowline develops, high-albedo snow and ice will be replaced with lower-albedo bare ground or vegetation, establishing SAF. Snow metamorphism, dust or aerosol deposition, and freeze–thaw processes also mean weathered old snow (commonly found at lower elevations nearer the snowline) has a lower albedo compared to fresh snow[62,63,64]. These changes mean the maximum albedo decrease will occur near the current snowline—the region of greatest snow cover change integrated over the year. Any adjacent snow-covered area above the area of snow loss will also exhibit an albedo decrease owing to increased snow metamorphosis.

Surface albedo changes can also be expected from vegetation changes. Specifically, negative changes in albedo are expected at the current treeline[65] as once grassy surfaces are replaced with lower albedo forest vegetation. Even above the future treeline, alpine vegetation could become denser, decreasing albedo by a smaller amount relative to sub-alpine vegetation[66]. However, albedo is also dependent on vegetation species[67], succession and phenology[68]. In vegetated areas between the future treeline and the current snowline, the albedo decrease is theorised to be slightly larger at higher elevation where shrubs replace less reflective rock rather than different vegetation. In certain mountain ranges (especially in higher and mid latitudes), the treeline and snowline overlap for part of the year, amplifying any albedo decrease at those elevations. Below the treeline, land use and human impact confound systematic predictions of changes in surface albedo.

[H1] Observed EDCC

Although physically based considerations suggest substantial EDCC across temperature, precipitation (including snow related variables) and surface albedo, additional factors can modify observed patterns. Historical changes are now discussed to assess these potential discrepancies. Previously published estimates of EDCC are supplemented with quantifications using CRU[69], GISTEMP[70], BERKELEY[71], GPCC[72], ERA5[73] and CMIP5[74] from 1980-2020 (**Table 1; Supplementary Tables 1-4).** Using the same gridded datasets in principle ensures comparable treatment of mountain regions worldwide. However, these results must be treated with caution owing to the unequal distribution of measurements, which are especially lacking at high elevations. The regional assessment of EDCC focuses on regions with the best data availability: the Andes, Tibetan Plateau and High Mountain Asia (TP and HMA), Greater Alpine Region (GAR, encompassing The Alps and parts of the Apennines and Dinaric Alps) and the Rocky Mountains.

[H2] Global Analyses

EDW is broadly prevalent at the global scale [75]. Indeed, in-situ weather station data highlight enhanced warming at higher elevations (mountains) compared to lower elevations (lowlands) based on paired comparisons[14]. Gridded analyses confirm this tendency for positive EDW at the global scale, with remarkable consistency in sign across all datasets (**Table 1; Fig. 3a; Supplementary Table 1**). The specific magnitude of the EDW, however, varies depending on the dataset: from 1980-2020, mountain-lowland trend differences range from 0.12°C century^{-1} in CRU to 0.36°C century^{-1} in ERA5 (**Table 1**). There are suggestions this warming difference might be increasing over time (**Supplementary Table 1**), but limitations in data quality and natural variability prevent accurate determination of this evolution. Despite this tendency for EDW, the station-based paired comparison indicates no systematic difference between high- and low-elevation warming when the data were aggregated. Thus, factors beyond elevation are just as influential in controlling absolute warming rate at the global scale, and EDW is often a locally expressed phenomenon[14].

Precipitation also shows suggestions of elevation dependent changes. At the global scale, there is a tendency towards a reduction in orographic precipitation gradients in the gridded datasets since 1980 (**Table 1**)[14] —that is, there is negative elevation-dependent precipitation change (EDPC), reflecting larger precipitation changes in lowlands compared to mountains (**Fig.3b)**. Like for temperature, this elevational weakening in precipitation is dataset dependent, with the mountain–lowland difference ranging from 0 mm yr^{-1} century^{-1} for ERA5 to –42 mm yr^{-1} century^{-1} for CRU (**Table 1; Supplementary Table 2**); CMIP5, a model hindcast, has contrasting patterns. Since 1960 there is broad consistency in sign across all gridded datasets, lending support to this global-scale EDPC. In contrast, to date, no systematic global difference between mountain and lowland precipitation trends has been reported for in situ stations [14].

Snowfall changes—strongly linked to precipitation—similarly exhibit elevation sensitivity. In general, global snow cover extent and snow cover duration have decreased by -3.6 ± 2.7% and 15.1 ± 11.6 days over 1982-2020 in mountain regions[19]. Gridded datasets also provide evidence of elevation dependent snow change, or a weakening of the orographic snowfall gradient (**Table 1; Supplementary Table 3**). Although snow is decreasing in both mountain and lowland regions, the absolute decrease is stronger in mountain regions, likely because there is greater snowfall in the first place.

[H2] Andes

EDW trends in the Andes show much variability owing to the large contrasts between western and eastern slopes, and the elongated nature of the mountain range, which encompasses tropical, subtropical and mid-latitude zones. Temperatures in the coastal zone are often decoupled from higher elevations and the eastern slopes owing to the effect of the cold Pacific Ocean to the west of the mountain range. Consequently, the coastal zone has cooled since 1981, while higher elevations have continued to warm, enhancing positive EDW on the western slope[76]. Conversely, on the eastern slopes much weaker EDW is reported because trends at low elevation are more similar to those at the higher elevations[77]. In the tropical Andes (7–20°S) between 1,000–

5,000 m EDW is amplified during the day in the winter[78] and is less pronounced at night. This daytime amplification might have important consequences for the melting of snow and ice[27]. Such spatial and temporal contrasts in EDW create highly uncertain results when averaged at the mountain range scale (**Table 1; Fig. 3c**) with mostly negative EDW in CRU (–0.16 °C century^{-1}) and BERKELEY (–0.23 °C century^{-1}) during 1960–2020 and more positive or neutral EDW in GISTEMP (0.2 °C century^{-1}) and ERA5 (0.68 °C century^{-1}). However, GISTEMP and ERA5 results evolve toward negative or less positive EDW of –0.35 °C century^{-1} and 0.1 °C century^{-1}, respectively during 1980–2020.

The tropical and sub-tropical regions show different precipitation trends. ERA5 data show a general lack of EDPC in the tropical region during 1951–2020(ref.[79]). In contrast, precipitation is increasing more rapidly at the highest elevations (>4000 m) in the subtropical region leading to a broadly positive EDPC[79]. However, the mountain–lowland difference at the mountain range scale using the gridded datasets (tropical and subtropical regions combined) is inconsistent, with wildly contrasting estimates in ERA5 (-77mm century^{-1}) and in GPCC (113mm century^{-1}) over 1980-2020 (**Table 1; Fig. 3d**).

Changing snow cover in the Andes and its elevation dependence has not been widely investigated. In the tropical Andes the snowline is extremely high (>5000 m), and consistent observations are few. Regions with reduced precipitation in the subtropical Andes on mid-elevation slopes (2000-3000 m) correlate with decreased snow cover persistence. For example, snow cover duration decreased by 2-5 days yr^{-1}, snowline elevation increased 10–30 m yr^{-1}, and snow losses were most pronounced on the eastern leeward slopes[80] during 2000–2016 between 29 and 36°S.

[H2] Tibetan Plateau and High Mountain Asia

Enhanced mountain warming appears to be clear in the TP and HMA. Most global datasets show strengthening positive EDW, although there is some disagreement as to the magnitude, with mountain–lowland differences of 0.47 °C century^{-1} and 0.36 °C century^{-1} for GISTEMP and BERKELEY, respectively and 0.16 °C century^{-1} for ERA5 during 1980–2020 (**Table 1; Fig. 3e**). Additionally, station data suggest that EDW is especially pronounced in the Third Pole region[81]. Warming peaks at the snowline (4,500–5,000 m) and declines above this elevation[82–84]. There is large spatial and temporal variability in EDW in HMA[85]; however, an EDW trend emerges across HMA as a whole. Increased warming at high elevations is especially pronounced for minimum temperatures during winter and maximum temperatures during spring, owing to increases in DLR and SAF processes, caused by increased specific humidity and reduction in snow cover, respectively [86,87]. Elevation dependent temperature changes are also observed in individual ranges such as the Tien Shan[88], Himalaya[89,90], and NyenchenTanglha[83] and in mountain chains along the margins of the TP[91].

Patterns of precipitation across the TP and HMA are complex, with precipitation regimes differing among mountain ranges (which mostly form the edges of the plateau). Analyses of EDPC and how it has evolved across the whole region are lacking from the wider literature. The mountain–lowland precipitation trend difference at the plateau-wide scale varies between gridded datasets, ranging from –137 mm yr^{-1} century^{-1} in ERA5 to 57 mm yr^{-1} century^{-1} in GPCC during 1980–2020 (**Table 1; Fig. 3f**). Precipitation on the southern side of the Himalaya increases with elevation up to a maximum (~3000 m), then decreases between 3000-5600 m but the profile stays relatively flat between 3500 and 5000 m(ref.[92,93]). On the TP itself, much precipitation is sourced from uptake in adjacent or upwind terrain; therefore, change in evapotranspiration-related variables across altitude and change in precipitation are connected, leading to much local recycling of moisture[94].

Snowfall is being replaced by rainfall which is extending to higher elevations across the TP and HMA. Analyses of station data since around 1960 show that the fraction of snowfall (as a proportion of total precipitation) decreased in the Tien Shan between 1,500 and 2,500 m(ref.[95]) or 3,500 m(ref.[96]). This decrease is mainly attributed to snowfall overall increasing less rapidly than total precipitation[97]. Snow loss on the TP is migrating to higher elevations. Station observations[87] during 1973–2002 revealed a loss of snow cover up to around 3,600 m but stable or even increased cover above this elevation. By 1989–2018 snow loss had occurred at all elevations up to 5,000 m, with loss concentrated at higher elevations. This elevational gradient in snow cover change has intensified positive EDW. Since around 1850 onwards snow loss has been exacerbated by aerosol deposition of black carbon on snow[98], which reduces snow albedo[99,100]. Between 2001 and 2016 the mean snowline elevation derived from satellite data increased the most in areas with very low and very high snowlines (5.15 m y^{-1} in the Eastern Tien Shan, 8.52 m yr^{-1} in the Eastern Himalaya), with much smaller changes or even elevation decrease in the Pamir, Hindu Kush and Western Himalaya [101–103]. Additionally, the TP exhibits elevation dependent vegetation changes with increased greening at low elevations and increased browning at high elevations. The impact of these changes on surface albedo is unclear[104].

[H2] Greater Alpine Region and Caucasus

EDW trends in the European Alps are diverse and vary by season and time of day[16] . The European Alps act as a natural barrier to weather systems, and thus different sub-regions exhibit strong microclimates, particularly the surrounding valleys[105,106]. Since the end of the 19th century the Alpine range has warmed twice as much as the global average[107–109] but any EDW signal is weak in most global datasets with only ERA5 showing significant positive EDW of 0.66 °C $century^{-1}$ during 1980–2020 (**Table 1; Fig. 3g**). Locally, Switzerland[12] exhibits faster warming at low elevations (1981-2017) in autumn and winter with less distinct patterns in spring and summer, although strong spring daytime warming has occurred at higher elevations coincident with snow loss. The negative EDW in autumn and winter, which is also present in the south-eastern Italian Alps[110], is attributed to a reduction in aerosol loads (most relevant in valleys such as the Po and Adige), which preferentially increases sunshine at lower elevations[111,112]. Additional temperature assessments at a global scale confirm that regions with faster warming rates at lower elevations tend to show a reduction in aerosols and cloud since the 1980s(ref.[65, 66]). The effect of SAF on EDW is likely to be small, most influential in spring, and possibly hidden by the effects of aerosols and clouds[12]. In the Caucasus, temperatures have increased in all seasons at low elevations, but only in summer at high elevations[113]. The strongest warming in the Lesser Caucasus occurred in the valleys (for example, the Ararat valley)[113]. However, the few climate stations in this region are biased towards low elevations, which are influenced by urban effects, leading to unreliable EDW assessments[114,115].

Unlike many other mountain regions, the orographic precipitation gradient in the Alps appears to be increasing. The CRU and GPCC identify mountain–lowland precipitation differences of 102 mm yr^{-1} $century^{-1}$ and 85 mm yr^{-1} $century^{-1}$, respectively during 1980–2020 (**Table 1; Fig. 3h**). Additionally, in situ stations measured around a 10% increase in the mountain–lowland annual precipitation ratio between 1961–1990(ref.[18]), possibly owing to a reduction in anthropogenic aerosol load at lower elevations. Between 1961-1990, the elevation gradient in precipitation increased during winter with a 25% reduction in lowland precipitation and a 25% increase in high elevation precipitation. Precipitation trends vary by season; for example, short-term convective precipitation has increased substantially in summer over higher elevations[116].

Average monthly snow depth in the Alpine region is decreasing[28] with a greater loss of snow at lower elevations. Data from over 100 snowfall time series from 1980–2020 in the north-eastern Italian Alps, indicate a negative trend in winter snowfall at low elevations and positive trends at high elevations; although there is no clear trend at intermediate elevations (1,000–2,000 m)[117]. Additionally, these data reveal a significant decrease in April snowfall ($P<0.05$) at all elevations. This decrease in snowfall at low elevations is attributed to increasing mean temperatures, while the slight increase at high elevations in winteris associated with increased precipitation. Similarly, in Switzerland the ratio of snowfall days to precipitation days decreased substantially at lower elevations but the trends were inconsistent at higher elevations[118].

[H2] Rocky Mountains

The Rocky Mountains exhibit a positive EDW trend (**Fig. 3i**). All global gridded temperature datasets and early analyses of temperature changes in the western USA[119] identify this positive signal and the global datasets suggest that this trend is increasing. Indeed, the positive EDW signal in the CRU dataset increased from 0.2 °C $century^{-1}$ in 1960-2020 to 0.73 °C $century^{-1}$ 1980-2020; similarly GISTEMP increased from 0.27 °C $century^{-1}$ to 0.93 °C $century^{-1}$ (**Table 1**). However, inhomogeneity of station data[120] and the changing distribution of stations[121] have made it difficult to quantify EDW from in situ stations in the western USA. Free air moistening at 500 mb is thought to be a strong control of enhanced warming around 5,500–6,000 m in the Alaskan St. Elias mountains, particularly at night[122]. Nighttime warming in boreal winter at high elevations is attributed to increased specific humidity, especially in traditionally drier inland ranges[52].

Negative EDPC has been observed in the Rocky Mountains since the 1940s. The Rocky Mountains and the Cascades are north–south mountain ranges that act as barriers to the prevailing westerly atmospheric flow. The synoptic climatology of the mountains is important for precipitation, causing contrasting lee and windward-specific climates and slope gradients that are dependent on jet steam interactions[123] . Reductions in the strength of the jet stream and westerlies are thought to have reduced the orographic precipitation gradient [123] and contributed to a decrease in precipitation near the crest of the western slope of the Cascade range and a slight increase to the east. Global precipitation datasets consistently show amplified drying at high elevations although the magnitude of the mountain–lowland precipitation difference ranges from -192 mm yr^{-1} century^{-1} in CRU to -106 mm yr^{-1} century^{-1} in GPCC during 1980-2020 (**Table 1; Fig. 3j**). This trend is confirmed by ERA5, which suggests that precipitation has decreased at the highest elevations[79], with a pronounced reduction in the orographic precipitation gradient during 1951-2020.

There is a complex relationship between snow water equivalent trends and elevation in the western North American mountains. In this region, snow water equivalent in maritime areas is primarily controlled by temperature and loss of snowpack has been greatest at relatively low elevations[124]. In some continental high-elevation sites in the California Sierra and Colorado, snowpack is more sensitive to precipitation than temperature and has therefore declined less rapidly. Loss of winter snowpack and earlier snowmelt in spring throughout the western USA have been given as contributing factors to an upslope migration in wildfire occurrence[45]. Wildfire smoke, atmospheric dust and snow algae reduce the surface albedo, contributing to glacial melting and EDCC[125,126]. The radiative effect of black carbon on glacier mass balance is three times larger than that of dust[127], both impacting snowmelt and hydrology[128].

[H1] Modelling Future EDCC

Most simulations of 20th and 21st century temperature change include a relationship between positive EDW and increased CO_2 concentrations. CMIP and CORDEX models agree that the rate of future temperature change positively depends on elevation in the TP–Himalaya[50,129,130], Rocky Mountains[131], European Alps[132] or all of these regions[51]. Models can also explore drivers of simulated changes in elevation gradients because they rarely confine analysis to a single variable (**Supplementary Tables 5,6**). The main drivers of EDW are often identified as SAF, incoming solar radiation and cloud cover, but long-wave radiation and near surface specific humidity also contribute, particularly in winter (**Supplementary Table 6**). There is however, still much work to be done in quantifying the role of certain drivers including land cover change and atmospheric moisture, particularly as explanations of precipitation and snow cover change rather than solely EDW. This section first outlines modelled trends identified in gridded datasets before evaluating changes reported in the literature for specific regions.

[H2] Trends in gridded datasets

Gridded datasets can be used to explore modelled future mountain and lowland temperature and precipitation changes (**Figure 4, Supplementary Figure 1, Supplementary Table 7**). Analysis using an ensemble mean from 35 CMIP5 models[74] (representative concentration pathway 4.5, RCP 4.5) running until 2100 predicts positive EDW and EDPC trends on a global scale by the end of the 21st century (**Fig. 4a,b**). The positive EDW signal is also particularly strong over the TP and HMA (**Fig. 4e**), is weaker for the Andes (**Fig. 4c**) and GAR (**Fig. 4g**), but, perhaps surprisingly, turns negative for the Rocky Mountains (**Fig. 4i**). For precipitation (**Figure 4,** right hand column) enhanced mountain precipitation (positive EDPC) is forecast on a global scale and for all regions except perhaps the GAR (**Fig .4h**) by the end of the 21st century. Although a global comparison of common gridded datasets is informative, future projections for regions use a diverse range of datasets with modelling simulations downscaled as appropriate. The changes identified in the literature do not always agree with the those reported in the gridded datasets.

[H2] Trends in literature

[H3] Andes

Positive EDW is forecast in the tropical and subtropical Andes throughout the 21st century; however, precipitation trends remain unclear. Early projections under the IPCC-SRES A2 and B2 scenarios showed enhanced tropical warming around 4,000–4,500 m on the western and eastern slopes of the tropical Andes by the end of the 21st century, with a rapid decline in warming rate below these elevations, particularly on the western slope[133] . CORDEX-CMIP5 temperature projections continue to indicate positive EDW for T_{max} (daily maximum temperature) on each side of the tropical and subtropical Andes in all seasons, often driven by a reduction in the surface albedo at higher elevations[134]. The projected EDW trends for T_{min} (daily minimum temperature) vary, with positive EDW in the subtropics, but negative EDW in the tropics throughout the year. In the inner tropics, future changes in T_{min} and T_{max} are mostly correlated with changes in DLR and shortwave radiation, respectively. In the subtropics, changes in both T_{min} and T_{max} are driven primarily by loss of snow cover and decreases in surface albedo, especially in winter[135], but changes in longwave radiation and humidity are also drivers, depending on the season and time of day[134]. Patterns of future precipitation change are complex and show less stratification by elevation[133].

[H3] Tibetan Plateau and High Mountain Asia

Most models predict positive EDW over the TP, particularly in autumn[51,129], despite considerable uncertainty about the intensity of this warming. The gridded analysis (**Fig. 4e**) also confirms clear positive EDW by 2100. EDW is often strong for T_{min} in winter and spring and for T_{max} in summer and autumn[50]. CORDEX-EA regional climate simulations project the greatest warming around 5,000–5,500 m in most seasons during 2031–2060, with slightly less warming around 4,500 m in spring [129]. This EDW is primarily driven by SAF; although DLR also contributes, it is sometimes reported as a major driver[51,52] and other times as a secondary driver (with a model-dependent contribution). In regional climate model simulations, the structure and magnitude of projected EDW is sensitive to the physics of the model (in particular the cumulus parameterisation scheme) and the driving general circulation model (GCM), because both alter the projections of snow cover and albedo, which modulate the simulated SAF and its effect[136].

Changes in the elevation dependence of precipitation are varied. An analysis examining the Indian Himalayan region predicts increased precipitation at low elevations and reduced precipitation at high elevations, along with increased warming above 3000 m, by day and night during all seasons except the monsoon[90]. Increased DLR caused by increased humidity is the primary feedback responsible for the enhanced warming at higher elevations[90]. In addition, reductions in cloud cover above 3,000 m increase net solar radiation received at the surface, with associated snow melt and reduced snow depth leading to reduced surface albedo and increased absorption of solar radiation at higher elevations[130]. SAF and atmospheric water content contribute to future EDCC in the Himalaya, with SAF being more important during the daytime and in spring, and atmospheric water content dominating in winter and at night. Additionally, the effect of SAF is concentrated in locations and times at which snow is melting, whereas moisture increase is more widespread[137].

[H3] Greater Alpine Region

There is much variation in modelled EDW and EDPC in the GAR. Modelled positive EDW is often connected with a decrease in snow cover and albedo at higher elevations, especially in spring. Precipitation trends are primarily influenced by local synoptic forcing[132,138]. The WRF convection-permitting regional model projects a drying over all seasons in the lowlands but much smaller variations in the mountains[139]. Additionally, using an ensemble of regional climate models, extreme daily precipitation is projected to increase with elevation, possibly owing to an increase in summer convection[17,18]. Snowfall is predicted to substantially decrease with reductions of up to 80% forecast for the Alpine Forelands by the end of the 21st century[140]. However, winter precipitation is projected to increase at very high elevations; therefore winter snowpack at the highest elevations (>3000 m) might remain relatively stable[132].

[H3] Western North American Mountains

Substantial positive EDW has been predicted throughout the 21st century in the Rockies in literature. For example, CMIP5 GCMs project substantial positive EDW in winter and spring in boreal midlatitudes[86]. Additionally, climate projections of the mid-21st century based on the EC-Earth GCM find seasonally varying EDW trends, with T_{max} and T_{min} exhibiting positive EDW in autumn but not in winter, and in spring and summer the EDW trends depend on the spatial resolution of the model[51]. Simulations using the high-resolution convection-permitting WRF model over the Rockies project that warming rates will monotonically increase with elevation or peak at a certain elevation[131]. The EDW signal is mainly associated with elevation-dependent changes in albedo, surface humidity and/or DLR. The strongest warming is usually associated with a reduction in surface albedo and snow loss, and elevation dependence of free tropospheric warming is of secondary importance.

[H1] A summary of geographic variations in EDCC

There is substantial variation in observed and modelled changes in EDCC. Mountain climates vary geographically owing to latitude, location, topography, prominence, and proximity to other summits. For a detailed list of examples of observed EDCC see **Supplementary Table 8**.

There is a broad agreement between the mean EDW observed across the gridded datasets (boxes in Fig. **5a**) and most individual studies (circles in Fig. **5a**), but the patterns differ between mountain regions. Positive EDW is extensively reported in the Rockies and TP and HMA, but less so in the Andes and GAR. Further research is required to explore the reasons for this difference, including the possible role of hypsometry. The Rockies and the TP and HMA are spatially extensive with large areas of high land; therefore, EDW trends are largely influenced by surface processes such as SAF, which might dominate through the mass elevation effect [141]. Conversely, elevational changes in regions with more isolated peaks such as in the GAR, should be more strongly related to free atmospheric drivers[142]. The EDW pattern in the GAR is complex with disagreement between individual studies and unclear trends in the gridded datasets. This complex pattern might partly relate to changes in aerosol composition, which sometimes favours enhanced warming at lower elevations. In the Andes there appears to be negative EDW, although there is a general lack of data. In the tropical Andes, SAF is expected to be a major driver of EDCC because of the effects of strong sunlight coupled with the deposition of atmospheric black carbon on snow from biomass burning in the Amazon Basin[143]. However, the lack of permanent snow in the tropics limits SAF to extremely high elevations. Increased atmospheric moisture content is also expected to influence EDCC in the tropics, but this contribution is unclear owing to a lack of model studies.

EDPC trends differ between mountain ranges and even between individual studies in the same range. There is a lack of agreement in EDPC trends from individual studies and the gridded datasets over HMA and the TP, with observational studies showing mostly positive EDPC contrasting with mixed results from the gridded datasets (**Fig. 5b**). In the GAR positive EDPC dominates in both gridded data and individual studies. Observational studies using variables such as snow depth, snowfall and snow cover find that snow is almost universally decreasing, with most snow loss occurring at lower mountain elevations (**Fig. 5c**). Often in studies focussed in high mountains, the most rapid snow loss only occurs up to a certain elevation, above which snow is typically more stable. This finding appears to contradict the decreasing orographic snowfall gradient identified in nearly all regions in the gridded analyses which ranges from –5mm yr^{-1} $century^{-1}$ (CRU, Tibetan Plateau) to –91mm yr^{-1} $century^{-1}$ (ERA5, GAR) (**Table 1**). However, the gridded analysis is solely based on snowfall, and takes mountains as a whole and compares the mountain snowfall loss with lowland regions where snowfall is usually a scarcity. Thus, more loss is observed in the higher elevation band, reducing the orographic gradient.

[H1] Summary and future perspectives

Contrasting warming rates at differing elevations are usually associated with changes in elevation gradients of many climate variables, including albedo, precipitation, atmospheric moisture, cloud, radiation balance components and wind speed; therefore, the concept of EDW has been expanded to EDCC. Systematic changes in these gradients are commonly associated with land cover changes, including uphill shifts of snowlines and treelines, the intensification of the hydrological cycle, and changes from snow to rain. Observed EDW can often be attributed to drivers such as surface effects (SAF) and free air effects (atmospheric humidity and DLR), that depend on season and time of day. There are regional differences in observed EDCC over the historical period (1950-present), including a latitudinal contrast, contrasts within mountain ranges (leeward versus windward slopes), and differences due to mountain orientation and topography. Climate models simulate future EDCC throughout the 21st century that is characterised by strong regional variation, due to the variable impact of different drivers in diverse regions and over different timescales. Further work is needed to better understand the trends, drivers and implications of EDCC, as discussed later in this section.

[H2] Limited coverage of mountain in-situ observations

Current in situ observations are inadequate for understanding the complexities of EDCC. Observations are highly skewed towards lower elevations, and particularly biased towards mid-latitudes and the northern hemisphere[144]. Additionally, many observations have inadequate spatio-temporal resolution and can be inaccurate for solid precipitation[145]. Therefore, continued and expanded monitoring is needed to overcome these deficiencies[146], with a focus on developing an agreed protocol for performing measurements along the elevational gradient. Examples of such protocols include the UHOP (Unified High Elevation Observation Platform) protocol, which is under development[147]. There are also several global and regional initiatives improving high elevation observations such as the Long-Term Ecological Research Network (LTER), GLORIA network, Virtual Alpine Observatory, and EvK2CNR, but maintaining funding and resources for long-term measurements is extremely challenging. In addition to elevation, topography is a critical control of mountain climate, especially concavity and convexity, which controls the exposure of the site to the free atmosphere. However, so far topography has not been very prominent in EDCC research.

Satellite datasets have great potential for the analysis of EDCC; however, they are currently underused. Satellite datasets cover the Earth's surface at a sufficiently high resolution to offer valuable information on patterns of EDCC. They also can be used as a proxy for numerous variables including air temperature[148], albedo[149], emissivity and radiation balance, vegetation health (for example, Normalized Difference Vegetation Index), precipitation, cloud and snow properties[150]. Concerns over homogeneity, the relatively short observation period, and other data limitations have hindered the use of these datasets in EDCC research. If in situ measurements can be used for validation, satellite datasets could help to overcome the challenge of the limited coverage of in situ observations.

Work is also needed to understand why different datasets show contrasting EDCC trends. Many gridded datasets, which depend on different algorithms to fill in missing observations, disagree on the sign of EDW for example in mountain regions such as the Andes and GAR (**Table 1**). Datasets show slightly better agreement on the sign of EDPC, especially at the global scale and in the Rockies, but differences remain in the other regions (**Fig. 3**). Model hindcasts such as CMIP5 often obtain different EDW trends to other gridded observations (**Table 1, Fig. 3**); thus, confidence in the future predictions made by such models is limited. Such differences could arise from contrasts in spatial resolution, timescale and/or data sources included. It will be important to quantify and understand uncertainties due to data quality and or model limitations, to reduce the uncertainty of predictions of future EDCC.

[H2] Uneven geographical data coverage

There is a strong mid-latitude bias in the current understanding of EDCC and a bias in observations towards the northern hemisphere. Most tropical mountains (notwithstanding the tropical Andes) have not been examined to the same extent as the regions discussed in this Review. Station density in the tropics is often an order of magnitude lower than in the boreal mid-latitude regions of Europe, Asia and North America[151]. There are also very few high-resolution model simulations of mountain regions in eastern and southern Africa, Indonesia and central America. Owing to the harsh terrain, there is also a lack of studies at high latitudes, particularly in Greenland and Antarctica (much of which are at high elevations).

It is important to understand EDCC in the tropics to gain insight on past and future environmental changes. Differential moisture patterns are expected to be a major control of EDCC in tropical mountains[152] with SAF making a limited contribution owing to the lack of snow and ice [153,154]. Additionally, aspect has minimal effect on slope heating because the sun is often overhead, although east–west slope differences in cloud cover occur owing to diurnal convection cycles. Understanding EDCC in the tropics can provide insight on possible modification and expansion of the Hadley cell in a warmer world[155]. Most models simulate enhanced warming and moistening of the tropical troposphere[156], which would reduce the mean free-air lapse rate [157]. This weakening of the free air lapse rate is expected to lead to positive EDW in the free tropical troposphere; however, it is unclear whether such EDW will occur on mountain slopes in the tropics [158]. Additionally, changes in the free air lapse rate could have important implications for the future of montane cloud forest ecosystems [159,160]. On longer timescales, the tropical lapse rate steepened during the last glacial maximum[161], which corresponded to dry conditions in the tropics, but lower snowlines[162]. Coupling of lapse rate changes with changes in moisture availability is a strong control of past environmental changes at high elevations in the tropics, with wet or dry conditions synonymous with warm or cold conditions in high mountains, respectively. More work is also needed to understand the links between the El Niño–Southern Oscillation (ENSO) and the Indian Ocean Dipole (IOD)[163] and EDCC in the tropics.

It is also important to understand EDCC at high latitude to inform projections of future sea level rise. Peripheral Greenland and Antarctica have their own mountain chains which project above the ice sheets, reaching 3,694 m (Greenland) and 4,892 m (Antarctica) respectively. Work has been done to explore the strong spatial gradients in precipitation and their effect on glacier mass balance, in Greenland[164] and Antarctica[165] . However, there has been little research into long-term EDCC and latitudinal and elevation dependent warming over Antarctica are conflated. Warming is most rapid on the Antarctic peninsula (at relatively low elevations), leading to some findings of negative EDW[166]. The rate of warming and its elevational profile across high latitudes controls the melting of land-based ice in Greenland and Antarctica, which contributes to estimates of global sea levels over the next century[167]. However, much ice-sheet modelling is based on temperatures extrapolated across the ice sheets using hypothetical lapse rates and more in situ data are urgently needed.

[H2] Understudied variables and drivers of EDCC

Although seasonal patterns of EDCC are beginning to become clearer in the Alps[110], this level of detail is not common in other mountain ranges. In mid-latitudes the mean slope lapse rate is often shallower in winter[168], and temperature inversion formation in mountain valleys is an important control of the winter temperature profile[169]. Thus, any reduction in cold air pools that might occur would favour negative EDW[170]. This effect might be expected to be less substantial in summer, although it would depend on synoptic climatology. In summer, precipitation is likely to be dominated by convection, whereas in winter, frontal rainfall (primarily influenced by atmospheric circulation) is likely to dominate, potentially leading to contrasting seasonal patterns in EDPC; however, further examination is needed.

EDCC research mostly reports changes in temperature, precipitation and snow- or albedo-related variables. Knowledge of how other variables, drivers and mechanisms stratify by elevation is largely limited to theoretical reasoning. Important changes will include humidity and cloud stratification, surface winds on mountain slopes, aerosols and their deposition on the surface, and surface energy balance components (**Box 1**). Despite its importance, atmospheric humidity is often overlooked in climate change research. Air humidity controls cloud formation and the free-air lapse rate[171] and low vapour pressure can impact evapotranspiration, vegetation growth, forest fires and snowmelt[172]. Additionally, atmospheric humidity directly influences shortwave and longwave radiation transfer, and therefore surface energy balance, potentially acting as a link between SAF and DLR.

Climate models should be used to investigate the roles and interconnections between drivers, but current simulations are limited by insufficient spatial resolution. Kilometre-scale simulations that can partly resolve relevant mountain processes such as the valley–slope wind system and localised convection have been developed[173,174]. However, robust ensembles of multi-decadal climate simulations at such high resolutions are not yet available owing to limitations in computer resources. Until such simulations are achievable, it is difficult to represent detailed land-surface processes in complex terrain, three-dimensional boundary layer turbulence or localised radiative effects in long-term climate predictions.

[H2] The definition and quantification of EDCC

Although EDW and EDCC are being increasingly discussed, there is no universally agreed method for defining and quantifying them. The mountain climate community needs to define the variables required to measure the phenomenon, and decide on quantitative indicators to represent the strength of EDCC that can be applied in all mountain regions. On the first point, a list of essential mountain climate variables has been proposed[175] based on expert knowledge, with surface albedo, water vapour, land cover and precipitation scoring highly. On the second point, trends can be stratified by elevation band to provide a qualitative profile[90,130,176], or by deriving the linear regression gradient of trend magnitude versus elevation[110,134]. In other cases, mountain and lowland trends are compared[14]; however, there is currently no accepted definition of what elevations correspond to mountains or lowland. The elevation gradient over which EDCC is examined can be subjective in terms of elevation range and spatial extent and it is important to understand how such choices influence the quantification of EDCC.

Mountain climate data are often fragmented between institutions and countries and hard to access. Sharing of available EDCC datasets needs to be improved with the development of agreed protocols for the collection, format and storage of data in which the main focus is change along the elevation gradient. The development of the Mountain Research Initiative (MRI) database on mountain observations[177] is a positive step forward.

[H2] Understanding the broader impacts of EDCC

There are many unanswered questions about the broader implications of EDCC for mountains and wider climate systems. The teleconnections and synergies between EDCC and large thermal systems such as the Asian monsoon and mid-latitude jet stream need further research (**Box 2**). Impacts of EDCC range from more rapid loss of the cryosphere in high mountains to increased mountain hazards and increasing susceptibility to extreme weather events. Changes in the frequency and severity of extreme events, such as short duration high intensity precipitation extremes, heatwaves and intense mountain wind storms will be particularly important. Such changes will influence mountain communities and subsistence economies common in mountain environments, but also millions of people living downstream who rely on water supply from mountain streams[29,178]. However, research into elevation-dependent changes in such extremes remains in the early stages[139,179,180] and needs to be made a priority to support future water resource management. Such research will require increased cooperation between mountain field scientists, climate modelers, and mountain societies and stakeholders.

Understanding EDCC will require the consideration of a range of viewpoints and international and regional collaboration. Mountain ranges often span international borders or combine different economic or political regions, making holistic assessment challenging. Different parts of a mountain range can experience different cultures with contrasting indigenous knowledge systems. This difference provides an opportunity, creating a diversity of ideas and approaches, but combining such varied insights, management systems and viewpoints can be politically difficult due to multiple actors, researchers and communities. Research into understanding EDCC offers a good opportunity to further international and regional collaboration.

Related links

Long-Term Ecological Research Network (LTER) https://lternet.edu/
GLORIA network https://www.gloria.ac.at/network/general

Virtual Alpine Observatory https://www.vao.bayern.de/
EvK2CNR https://www.evk2cnr.org/

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The authors declare no competing interests

Author contributions
N.P. conceived the original idea, wrote large sections of the manuscript, and coordinated the overall submission. M.A., J.K. and S.T. coordinated the writing of individual sections of the manuscript. S.W., L.H., A.N., S.T., E.P. and J.S. produced figures and/or tables. E.A. performed the analysis of trend dependence on elevation and produced corresponding figures/tables. All authors provided input to group discussions and/or comments on the manuscript draft.

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Display items:

Table 1. Trends in mountain–lowland difference in temperature, precipitation and snowfall.

Region	Dataset	1960-2020			1980-2020		
		Temperature (°C century^{-1})	Precipitation (mm yr^{-1} century^{-1})	Snow (mm yr^{-1} century^{-1})	Temperature (°C century^{-1})	Precipitation (mm yr^{-1} century^{-1})	Snow (mm yr^{-1} century^{-1})
Global	CRU [69]	0.06	–19*	–12**	0.12	–42**	–15**
	GISTEMP [70] /GPCC [72] [a]	0.15*	–3	NA	0.17	–11	NA
	BERKELEY [71]	0.06	NA	NA	0.18	NA	NA
	ERA5 [73]	0.22*	–28*	–37**	0.36*	0	–39**
	CMIP5 [74]	0.08	–2	–20**	0.23**	7	–23**
Andes	CRU	-0.16	5	-11*	-0.73**	40	-26*
	GISTEMP/GPCC[a]	0.20*	17	NA	–0.35*	113	NA
	BERKELEY	–0.23*	NA	NA	–0.59*	NA	NA
	ERA5	0.68**	–301**	–29*	0.10	–77	–87**
	CMIP5	0.13**	50**	–16**	0.13**	51*	–16**
TP and HMA	CRU	-0.20*	2	3	-0.15	-32	-5
	GISTEMP/GPCC[a]	0.23*	7	NA	0.47**	57	NA
	BERKELEY	0.11	NA	NA	0.36*	NA	NA
	ERA5	0.35	–48	–22**	0.16	–137**	–43**
	CMIP5	0.06**	–9*	–29**	0.17**	3	–42**
GAR	CRU	-0.04	11	-56**	-0.13	102*	-44
	GISTEMP/GPCC[a]	–0.05	3	NA	–0.13	85	NA
	BERKELEY	0.04	NA	NA	–0.07	NA	NA
	ERA5	0.69**	–15	–106**	0.66**	68	–91*
	CMIP5	–0.07**	–14*	–4	–0.09**	–3	–10
Rockies	CRU	0.20	-81	-18	0.73	-192	-23
	GISTEMP/GPCC[a]	0.27	–52	NA	0.93	–106	NA
	BERKELEY	0.05	NA	NA	0.62	NA	NA
	ERA5	0.68*	–153**	–66*	1.11	–126	–30
	CMIP5	–0.01	–28**	–31**	–0.24	–18	–33**

NA, not applicable; GAR, Greater Alpine Region. The raw mountain and lowland values and data for other periods are available in Supplementary Tables 1-3. [a]GISTEMP for temperature GPCC for precipitation. **significant at $p<0.01$, *significant at $p<0.05$.

Figure captions

Figure 1: Factors and processes associated with elevation-dependent climate change in mountain environments. A schematic figure of the processes associated with, resulting from, or contributing to elevation-dependent climate changes. Where changes are dominantly in one direction these are indicated by + and – for positive and negative changes, respectively. 1) Upper-level prevailing winds interact with the mountains – prevailing winds create windward and leeward effects with differential gradients. 2) Change in cloud base height and moisture, with associated heavy precipitation and changes in cloud forest elevation. 3) Changes in cloud properties and specific humidity, which reduce incoming solar flux but increase downward longwave flux. 4) Strengthening upslope anabatic winds and weakening downslope katabatic winds. 5) Increasing streamflow extremes and variability. 6) Glacier retreat influencing downstream fluvial regimes and habitats. 7) Increase in snowline elevation and changes in snowpack above snowline. 8) Snow albedo decrease. 9) Migration and possible expansion or contraction of vegetation zones upslope and changes in vegetation

composition, including from wildfire. 10) Anthropogenic aerosol declines but wildfire aerosol increases. 11) High elevation plateau creates mass elevation effect and reduces the influence of free air processes. 12) Changing frequencies of temperature inversion formation and cold-air pools in mountain valleys. 13) Permafrost degradation leading to mass wasting. Many of the processes tend to amplify mountain change rather than subdue it.

Figure 2: Expected vertical profiles of important mountain variables in a warmer world. a, The best estimate of the rate of change in air temperature *T* over time *t* with elevation and warming. The profile represents the global mean; regional profiles could differ owing to local processes. Not drawn to scale. **b**, As in panel a but for precipitation, *P*. LCL, lifting condensation level. **c**, As in panel a but for albedo, *α*. Different variables show markedly different profiles.

Figure 3: Observed temperature and precipitation changes in mountain regions. a, Global-mean, 10-year running temperature (T) anomalies[70] (with respect to a 1986-2005 baseline) for mountains (based on[181]; red) and lowlands (blue); shading represents one standard deviation around the mean. Grey circles represent the mountain–lowland difference for individual years, and the grey line the difference with a smoothed spline. **b,** As in panel a, but precipitation (Pr)[69] anomalies for mountains (green) and lowlands (brown). **c-d**, as in a-b, but for the Andes. **e-f**, as in a-b, but for the Tibetan Plateau. **g-h**, as in a-b, but for the Greater Alpine Region. **i-j**, as in a-b, but for the Rockies. Patterns of elevation dependent climate change are mixed, with enhanced mountain warming in the Tibetan Plateau and Rockies, and enhanced precipitation change in the Andes and Greater Alpine Region

Figure 4: Projected temperature and precipitation changes in mountain regions. a, CMIP5 ensemble mean[74] global-mean 10-year running temperature (T) anomalies (with respect to a 1986-2005 baseline) under RCP4.5 for mountains (based on[181]; red) and lowlands (blue); shading represents one standard deviation around the mean. Grey circles represent the mountain–lowland difference for individual years, and the grey line the difference with a smoothed spline, shown on a magnified scale on the right axis. **b,** As in panel a, but simulated precipitation (Pr) anomalies for mountains (green) and lowlands (brown). **c-d**, as in a-b, but for the Andes. **e-f**, as in a-b, but for the Tibetan Plateau. **g-h**, as in a-b, but for the Greater Alpine Region. **i-j**, as in a-b, but for the Rockies. Most mountain regions exhibit a tendency towards enhanced mountain warming and precipitation compared to lowland regions, although the magnitudes differ.

Figure 5: Synthesis of observed elevational dependent climate change. a, Published evidence of elevation dependent warming for individual mountain studies (small circles, representing only the sign; **Supplementary Table 8** and mountain range scale studies (larger circles, representing only the sign; **Supplementary Table 8**, the location of the circle is approximate) and the mean warming signal inferred from gridded datasets for entire mountain regions (large boxes, the size indicating the spatial scale for averaging, colour representing the sign and magnitude of the change; **Table 1 and Supplementary Tables 1-3**). Insert provides a more detailed view of the European Alps. **b**, As in panel a, but for precipitation. **c**, as in panel a, but for snow and without the gridded datasets. Patterns of elevation dependent climate change (EDCC) are mixed and depend on the variable and region considered.

Box 1: Expected elevational profiles of changes in other variables in a warmer world

Although height dependence of temperature, precipitation and albedo changes with warming are well-examined, other important variables are less so. These include specific humidity, surface winds, aerosols and solar radiation.

[bH1] Specific humidity

With warming, specific humidity, *q,* is theorised to increase at all elevations (see figure; d*q*/d*t*, rate of change in *q* with time *t*)**.** However, assuming a constant temperature increase with height, absolute increases in *q* will be greater at low elevations. This difference emerges because of higher temperatures at lower elevations that non-linearly increase moisture holding capacity (saturation vapour pressure) of the atmosphere[182,183]. Yet, changes in *q* will have a larger relative effect at higher elevations owing to the lower initial *q*, with implications for cloud formation and other processes. Patterns in *q* will be modified regionally by global scale climate dynamics leading to spatial variations in moisture.

[bH1] Surface winds

Changes are also expected in surface winds, *U*, including an intensification of daytime upslope winds and a weakening of nighttime downslope winds (see figure). These changes are theorised to be largely driven by shifts in snowlines and treelines[184]. During the day, rising snowlines lead to uncovered bare, rocky ground at higher elevations, driving surface heating that switches downslope katabatic winds over snow to upslope anabatic winds over rock; as maximum surface temperature change between rock and snow occurs under solar heating during the day, the strongest wind changes will occur during this time. Around the treeline, the replacement of bare rock with forest (and corresponding effects on evaporative flux, surface roughness and albedo [185,186]) will have the net effect of decreasing upslope wind speeds. Thus, daytime upslope winds will increase at high elevations (from snowline changes) but decrease at lower elevations (from treeline changes).

Different changes emerge at night. At this time, the switch from snow to rock at higher elevations decreases radiative cooling at the surface, weakening nocturnal downslope winds. Likewise, the combination of increased roughness and reduced longwave radiation efficiency (lower sky view factor) reduces surface cooling, weakening downslope flow below the new treeline. Accordingly, nighttime downslope winds will weaken at all elevations.

[bH1] Anthropogenic aerosols

Anthropogenic aerosols, such as black carbon and sulfates, are theorised to decline in the future at all elevations (see figure; AOD, aerosol optical depth). A reduction in aerosols is expected at all elevations due to a combination of more stringent air pollution control policies and technological advances[187,188]. However, the decline will be relatively larger at low elevations because of the greater initial concentration associated with proximity to population centres and the short lifespan[189]. This reduction ignores any changes in thermal circulations and any strengthening of upslope winds (panel b), which could transfer a larger proportion of the remaining aerosols upslope (even if not a larger absolute amount than at present).

[bH1] Solar radiation

Incoming solar radiation, *S*, at the mountain surface is theorised to decline (see figure). As *q* is the first-order driver of solar radiation reaching the surface via impacts on cloud optical depth[190,191], its increases with rising temperatures lead to a proportional and inverse decrease in *S* at the mountain surface. The largest decreases will occur at low elevations where the atmospheric column is deepest. Any change will depend on relative rates of cloud optical thickness change at different altitudes. Given that the majority of aerosols scatter radiation[192], and absorption and scattering of light by aerosols is influenced by the uptake of water[193], the strength of the effect of *q* will be modified by the concentration of aerosols throughout the atmosphere.

Changes in other radiation components are also important. Reflected solar radiation will be strongly related to any changes in surface albedo. Outgoing longwave will primarily depend on surface temperature which is expected to increase more at high elevations. Downwelling longwave radiation has a non-linear dependence on q and is expected to increase more rapidly at higher elevations[52].

Box 2: Teleconnections with the broader climate system

Changes in mountain precipitation influence the cryosphere and in turn the downstream hydrological regime[194] and water resources[195]. Loss of snowpack and a change from snowfall to rainfall will increase the likelihood of mountain hazards such as landslides, flooding and drought (periods of high or low flow). Although the broad downstream implications are well known, research into teleconnections between mountain climate changes and broader climate changes outside the mountains remains underdeveloped. Of particular concern is how elevation-dependent climate change (EDCC) will influence global circulation systems.

EDCC on the Tibetan Plateau (TP) is expected to influence the Asian monsoons. Changes in surface slope rates and free air contrasts have opposing influences on monsoons; however, it is unclear which effect will dominate. Elevation dependent warming in the Third Pole region will enhance high elevation surface heating, decreasing the surface temperature difference between the TP and surrounding lowlands. This change in temperature gradient would weaken any monsoonal circulation that is dependent on slope-based lapse rates (or surface temperature differences). Sensible heat flux is decreasing at high elevations[21], possibly owing to the wetting of the high plateau and increased cloudiness and precipitation[196,197]. However, if the high plateau warms more than the surrounding free atmosphere (at the same elevation), particularly where snow is lost, the upper-level circulation could change. Summertime heating of the plateau reduces the meridional temperature gradient to the south, weakens the westerly circulation and encourages the development of the East Asian and South Asian monsoons[81]. Thus, a warmer TP and an increased anomalous heat source at 500 mb would strengthen this process.

In parallel with arctic amplification (AA), EDCC could influence the meridional temperature gradient and thus the mid-latitude jet stream, and its control over extreme events. AA influences the climate of mid-latitudes through interactions with the mid-latitude jet[198]. A decreased meridional temperature gradient resulting from AA has been proposed to increase large Rossby Wave undulations and strong polar vortex outbreaks in lower latitudes[199]. Associated increased atmospheric blocking can lead to floods and droughts in unlikely places[200]. Similar mechanisms could result from EDCC if the strength of the meridional temperature gradient were to change owing to enhanced high elevation warming. Such effects are more likely in Eurasia where mountains run west to east and EDCC has a latitudinal component. Although the Andes and Rockies respond to changes in the mid-latitude jet and have differential climate responses on their windward and leeward sides[123,201], it is speculated that EDCC is less likely to influence the jet in this way. However, such ideas concerning the effect of EDCC on monsoons and the mid-latitude jet stream are in the early stages of development and remain under debate[202,203].

TOC blurb

Environmental changes in mountains often depend on elevation. This Review outlines how past and future temperature, precipitation and snowfall trends vary between mountains and lowlands across various mountain regions and discusses the drivers responsible.