Pepin N, Bradley RS, Diaz HF, Baraer M, Caceres EB, Forsythe N, Fowler H, Greenwood G, Hashmi MZ, Liu XD, Miller JR, Ning L, Ohmura A, Palazzi E, Rangwala I, Schöner W, Severskiy I, Shahgedanova M, Wang MB, Williamson SN, Yang DQ. **Elevation-dependent warming in mountain regions of the world**. *Nature Climate Change* 2015, 5(5), 424-430.

DOI link to article:

http://dx.doi.org/10.1038/nclimate2563

Date deposited:

16/06/2015

Embargo release date:

23 October 2015

Elevation-Dependent Warming in Mountain Regions of the World

Mountain Research Initiative EDW Working Group[a]

Abstract

It is generally accepted that the observed warming of our climate system is amplified in the Arctic. In contrast, despite model predictions of future amplified warming in high elevation regions, analyses of limited mountain observations available do not always agree. This is partly because of inherent complexities in mountain climate, but also because current monitoring is inadequate and skewed towards lower elevations. We review important physical mechanisms which could potentially contribute to elevation-dependent warming (EDW) that include: a) snow albedo and surface-based feedbacks, b) water vapour changes and latent heat release, c) surface water vapour and radiative flux changes, d) surface heat loss and temperature change, and e) aerosols. Mechanisms either encourage an enhancement of warming with elevation, or enhance warming in a critical elevation band. The variable combination of mechanisms may cause contrasting patterns of EDW in different regions. We propose future needs to improve knowledge of mountain temperature trends and their controlling mechanisms through improved observations, use of satellite data, and model simulations. Improved understanding of EDW is of critical societal importance because EDW can enhance climate change impacts in mountain ecosystems by accelerating changes in cryospheric systems, hydrological regimes, ecosystem response and biodiversity.

Introduction

It is well-known that the rate of temperature change with increased levels of greenhouse gases in the atmosphere is amplified at high latitudes, but there is also growing evidence that the rate of warming is amplified with elevation, such that high mountain environments are experiencing more rapid changes in temperature than at lower elevations. This "elevation-dependent warming" (EDW) has important implications for the mass balance of the high altitude cryosphere and associated runoff, for ecosystems and farming communities in high mountain environments, as well as for species that reside in restricted altitudinal zones within a mountain range. However, because of sparse high elevation observations, there is a danger that we may not be monitoring some of the regions of the globe that are warming the most. Here we review the evidence for EDW, and examine the mechanisms that may account for this phenomenon. We conclude with a strategy for future research that is needed to reduce current uncertainties and to

[a] N. Pepin, R.S. Bradley, H.F. Diaz, M. Baraer, E.B. Caceres, N. Forsythe, H. Fowler, G. Greenwood, M.Z. Hashmi, X.D. Liu, J. Miller, L. Ning, A. Ohmura, E. Palazzi, I. Rangwala, W. Schoener, I. Severskiy, M. Shahgedanova, M.B. Wang, S.N. Williamson and D.Q. Yang

ensure that the changes taking place in remote high elevation regions of the planet are adequately observed and accounted for.

Evidence for EDW

In theory, it ought to be a simple matter to document the rate and geographical pattern of warming with elevation over recent decades. However, many factors make it extremely difficult to determine the rate of warming in mountainous regions. First and foremost, long-term meteorological stations (with >20 years of record) are extremely sparse at high elevations. For example, in the GHCNv3 database of homogenized stations, out of 7297 stations, only 191 (3%) are above 2000m and 54 (0.7%) above 3000m, and long-term data are simply non-existent above 5000m in any mountain range[1]. Unlike in the Arctic, which is relatively homogenous, mountain temperatures also suffer from extreme local variability due to factors such as topography, slope, aspect and exposure. Furthermore, observations within mountain regions are often taken from valleys with distinct microclimates prone to cold air drainage[2,3], which makes it hard to separate noise from trend.

Evidence for EDW could also come from satellite data, atmospheric reanalysis or model studies but these sources also have limitations. Satellites generate surface ("skin") temperature, which is not generally recorded at surface meteorological stations. Moreover, satellite data are of limited duration[4] and are not well validated in high elevation regions where clouds are common. Reanalysis data sets are heavily dependent on free air (not just surface) data, and they are not homogenised for climate trend analysis. Models generally have poor spatial resolution and require observational data for validation, making it difficult to be sure that simulations are accurate.

Notwithstanding these limitations, there have been many studies that have attempted to quantify EDW (Table 1). More detail concerning individual studies (including metadata) is given in the Supplementary Information. Here we summarize the literature as a whole. A majority of studies suggest that warming is more rapid at higher elevations (first row of Table 1) but there are a number of studies which show no relationship or a more complex situation (other three rows). This is particularly true for observational studies which are in less agreement than model simulations. This may be because most models integrate trends over a long time period (typically up to the end of the 21st century) when EDW may become more widespread than it has been so far. It is also noticeable that minimum temperatures show a stronger tendency towards EDW than maximum temperatures, and so separate mechanisms may be at work during the day and night. Of the studies listed in Table 1, there are relatively few global studies that have examined surface (~2m) temperature data for evidence of EDW[5-8]. Overall there has also been a lack of consistency in the methods and data used to quantify the rate and patterns of warming. Differences in the time periods examined, the stations compared, the elevational range selected, and the temporal resolution of the data (i.e. daily vs monthly or annual temperatures) all vary (see Supplementary Information) and thus contribute to differences in trends. Many studies are relatively short (<50 years) and so strong inter-decadal variability often contributes to observed trends. Although some data homogenization has been achieved for station records in Europe and North America, there is a particular problem with the mountain data in the tropics which is both sparse and inhomogeneous.

The most striking evidence for EDW is from Asia. Yan & Liu[9] investigated warming trends (1961-2012) using 139 stations on and around the Tibetan Plateau, the most extensive high elevation area in the world. Figure 1 shows mean warming rates (ºC/decade) for contrasting periods for stations in 500m wide altitudinal bands starting at 1000m. Systematic increases in warming rate with elevation are uncovered for annual mean temperature, and warming rates have increased in recent decades (Figure 1a). Mean minimum temperatures also show EDW on an annual basis (Figure 1b) as do mean temperatures in autumn (Figure 1c) and winter (Figure 1d). There was no strong elevational effect in other seasons, or for mean maximum temperatures, a result that is consistent with findings in other areas.

Model simulations have also been used to identify EDW, both in historical and future projections[10-12]. In general, most GCMs indicate that EDW in the free atmosphere is a characteristic of low latitude regions, with a maximum warming signal around 400-200 hPa in the AR5 RCP scenarios for example[13] (Fig 12.12 therein). This EDW does not extend into the mid and high latitudes, and in most models warming is concentrated at the surface (>800 hPa) n the high latitudes, especially in the northern hemisphere. Few studies have been specifically designed to investigate the geographical pattern of and processes associated with EDW at the mountain surface which is sometimes dissimilar to the free atmosphere[14]. The climate response to greenhouse gas forcing within complex mountain topography can only be adequately captured using models with a spatial resolution of 5 km or less[15] that generally implement non-hydrostatic equations for the atmosphere. Such simulations can be done for both historical case studies (using reanalyses as boundary forcing) and future scenarios (using GCM projections); and although the computational demands are high at present, novel and selective experimental designs to conduct studies at such high resolutions are feasible[16,17]. Of those modeling studies that have identified EDW, the most common explanation is associated with the snow/albedo feedback mechanism, which is often discussed in terms of the upslope movement of the zero-degree isotherm[10], although other factors such as elevation-dependent changes in cloud cover and soil moisture have also been suggested as possible forcings[18].

Regional variability in EDW could also be influenced by inter-annual to decadal-scale variability in large-scale circulation, e.g. ENSO, NAO, PDO, etc. For example, it has been found that during positive NAO winters in the late 1980s and early 1990s, there was enhanced warming at high elevations in the European Alps, associated with warm wet winters[19]. This pattern was reversed during years with a negative NAO index. In the tropical Andes, increasing freezing level heights between 1958 and 1990 were related to El Niño conditions (warmer sea surface temperatures in the eastern tropical Pacific) and attendant changes in clouds and atmospheric water vapor, which can drive regional variability in EDW[20]. Because of all of these factors, we can only say at present that there is evidence that many, but not all, mountain ranges show enhanced warming with elevation. Understanding the physical mechanisms that drive EDW is therefore essential to explain the regional variations, as discussed in the next section.

Hypotheses and Mechanisms for EDW

Temperature change at the earth's surface is primarily a response to the energy balance, and therefore factors which preferentially increase the net flux of energy to the surface along an elevation gradient would lead to enhanced warming as a function of elevation. Here we discuss various mechanisms and processes that have been linked to EDW[21]. The physical shape of the associated elevational signal is expressed in Figure 2.

a) Albedo

Snow albedo feedback is an important positive feedback in Arctic amplification[22,23], but it is also relevant in snow-dominated high elevation regions[10,24,25] where the seasonal timing of snow cover varies with elevation, and maximum warming rates commonly occur near the annual 0°C isotherm[6,26]. In the Swiss Alps the daily mean 2m temperature of a spring day without snow cover is 0.4°C warmer than one with snow cover[27] (mean value for 1961-2012). The current snowline, which varies in elevation across different mountain ranges, is expected to retreat to higher elevations as the overall climate system warms. Furthermore, the elevation dependency of snow cover duration (and also of the snow/rain ratio) is non-linear so the rate of snowline retreat may increase as temperatures rise[28]. This will result in significant increases in the surface absorption of incoming solar radiation around the retreating snowline (approximated by the 0°C isotherm), initially causing enhanced warming at that elevation[29]. As the snowline migrates upslope, this effect will extend to increasingly higher elevations (Figure 2a). A similar process is expected to result from an upslope migration of tree-lines, owing to the accompanying reduction in surface albedo through greening[30,31]. Changes in the ratio of snow to rain are also likely to occur over a wider elevation band, as already noted in the Tibetan Plateau/Himalaya region, where the snow cover season has shortened, and more precipitation is now falling as rain[32-34]. The snow/albedo mechanism has a stronger influence on maximum than minimum temperatures because of the increase in absorbed solar radiation, as noted by Kothawale et al.[35] who compared maximum and minimum temperature trends in the western Himalayas between 1971 and 2007. A regional climate model study[10] also found EDW in the Alps and suggested that the increasing influence of the snow-albedo feedback mechanism, primarily during spring and summer, was responsible. The specific temperature response (Tmin vs. Tmax) will depend on soil moisture; if the increased surface shortwave absorption is balanced by increases in sensible heat fluxes (latent heat fluxes), the response will be more prominent in Tmax (Tmin). A more amplified response has been found in Tmin relative to Tmax in lower elevation regions (1500-2500m) of the Colorado Rocky Mountains during winter from regional climate models[36.] This was caused in part, by the increases in the absorbed solar radiation at the surface, primarily balanced by increases in the latent heat fluxes caused by the increases in surface soil moisture from snowmelt.

b) Clouds

Observations of long-term changes of clouds and cloud properties are sparse, particularly in high elevation regions, and there are few studies that discuss how changes in clouds might affect EDW[37,38]. Changes in cloud cover and cloud properties affect both shortwave and longwave radiation and thus the surface energy budget. They also affect warming rates in the atmosphere through condensation. A band of enhanced warming caused by latent heat release is expected near the condensation level, which could be further augmented because of higher atmospheric water vapor content[20] resulting from global warming (Figure 2b). If the condensation level rises (which may occur if temperatures also warm and dew point depression increases at sea-level) then a band of reduced warming would occur immediately below the new cloud base level (dotted line on Figure 2b) with enhanced warming above. Thus, the overall implications of a warmer and moister atmosphere support enhanced warming at high elevations[39,40]. For the Tibetan Plateau between 1961 and 2003, decreasing cloud cover during the daytime, but increasing low level clouds at night, has caused minimum temperatures to increase[41]. Using weather stations and high-resolution climate model output, Liu et al.[38] found that cloud-radiation effects were partly responsible for EDW on the Tibetan Plateau. A similar response was observed in the Alps[19] with an altitudinal dependence of temperature anomalies, except that lower elevations were affected by changes in fog and stratus clouds.

c) Water Vapor and Radiative Fluxes

Processes associated with the relationships between longwave radiation, moisture and thermal regimes along an elevation gradient are expected to lead to an elevation-dependent warming. These include i) the sensitivity of downward longwave radiation (DLR) to specific humidity (q), and ii) the relationship between temperature and outgoing longwave radiation (OLR). DLR increases in response to increasing specific humidity, however this relationship is non-linear (Figure 2c) with substantially higher sensitivities at low levels of specific humidity, especially below 2.5g kg^{-1}, which are found in many high elevation regions[40,42,43]. These high sensitivities occur because, below a certain specific humidity threshold, the air becomes optically under-saturated in the longwave water vapor absorption lines. In such conditions, small water vapor increases can have a substantial influence on DLR, resulting in a significantly greater warming response at higher elevations. Both observations[40,44] and climate model simulations[12,25] suggest that this mechanism has contributed to EDW. In the Alps, the DLR-q sensitivity is particularly high for q below 5g kg^{-1}, conditions that are more likely at higher elevations and during the cold season in the extra-tropics[42]. A similar finding was also reported for the Colorado Rocky Mountains[45], which also showed that clouds have limited effect on DLR-q sensitivities.

Another mechanism related to radiative fluxes is a direct consequence of the functional shape of blackbody emissions. For a given heat flux exchange (e.g., an increase in OLR), this relationship will result in a larger temperature change at lower temperatures[46]. OLR is one of the major mechanisms through which the land surface loses heat and is proportional to the fourth power of temperature (the Stephan-Boltzmann law). Therefore, for a given change in radiative heating, higher elevations will experience enhanced warming rates along an elevational gradient. This effect will not be sensitive to seasons or geography (Figure 2d).

d) Aerosols

Most of the atmospheric loading of aerosol pollutants (e.g. atmospheric brown clouds and associated black carbon in Asia) is concentrated at relatively low elevations (<3km)[47], which would be expected to decrease the flux of shortwave radiation to lower mountain slopes (surface dimming effect), but have limited or no effects on higher mountains, above the polluted layer (Figure 2e). However, there has been little systematic investigation of the elevational signal of anthropogenic pollutants in mountain regions. High levels of black carbon have been found at 5000m in the Himalayas during the pre-monsoon season[48], but whether such conditions reflect local up-valley transport or more regional conditions is unclear. Certainly, aerosols are transported to elevations above 10km over Tibet and much of central Asia during convective monsoon activity[49,50].

Black carbon affects the radiation budget in two ways. It absorbs radiation (principally in the mid to lower troposphere) and decreases the surface albedo when deposited on snow[51-53]. It has been suggested[47] that black carbon could account for half the total warming in the Himalayas during the last several decades. Aerosols such as dust depend on other factors such as land-use change. For example, the disturbance of ground cover in the western U.S. causes dust to be readily transferred into the atmosphere and transported by wind into the mountains where it settles on snow, reduces albedo and leads to enhanced warming at higher elevations[54].

Combination of Mechanisms

The resulting response to all these factors and their interactions is complex. Some factors will be more influential than others in certain parts of the globe, and at certain times of year and this may partly account for the differences in EDW reported in Table 1. Albedo feedback will be strongest wherever snowlines are retreating but their elevational focus will be narrowed (and more prominent) in the tropics where there is reduced seasonal variation in snowline elevation. The cloud feedback (latent heat release) will be enhanced at high temperatures and is therefore also dominant in the tropics, particularly in the free atmosphere where it is responsible for amplified warming in most IPCC simulations[55]. In contrast, aerosol loading is concentrated in mid-latitudes (particularly Asia) and the water vapor/radiative feedbacks are enhanced at low temperatures so should be dominant at higher latitudes, at night and in winter. All the physical processes either point to an expected increase in warming with elevation, or an enhanced band of sensitivity that will move upslope with time. Although the physical reasons for the existence of EDW are strong, the observational evidence for EDW is not yet as clear cut. Therefore, we urgently need to develop a more systematic method of climate monitoring at high elevations to quantify the extent of the phenomenon and its spatial variability.

Future needs

The surface in-situ climate observing network needs to be expanded to cover data poor regions, and to include more variables. Air temperature observations at ground stations are essential but many high-altitude areas (e.g., greater than 4000m) are still heavily under-sampled. Therefore, special efforts should be made to extend observations upward to the highest summits with transects of highly accurate instruments (as used, for example, in NOAA's Climate Reference Network) measuring a broad array of variables (e.g., humidity, radiation, clouds, precipitation, soil moisture and snow cover, besides temperature) in order to fill the pronounced high-altitude observational gap. Targeted field campaigns should be devised to detect and better understand EDW in areas where its signal is expected to be strongest (see the previous discussion on mechanisms). For example radiation measurements, together with measurements of humidity, albedo, temperature and soil moisture could be focused around tree-line, and above and below the snowline in areas with high interannual variation in snowline position, to determine the partitioning of energy fluxes; black carbon and other aerosols should be monitored more widely to determine their relationship to other meteorological parameters. More details are given in the Supplementary Information.

Much high elevation data that have been collected are also largely inaccessible due to poorly managed data archives in different countries, and this issue has limited efforts to create homogenous high elevation global datasets. Thus, in addition to new observations, we need a strong effort to locate and evaluate what already exists. Powerful methods for homogenizing climate data are available (see e.g. www.homogenisation.org) but better metadata (e.g. station characteristics such as topography) and information on station location and instrument changes needs to be incorporated more systematically into high elevation datasets. Finally we need agreed metrics to measure and quantify EDW once data are collected.

To further overcome limitations arising from the sparseness of in-situ stations in under-sampled high-altitude regions, spatially continuous remotely-sensed land surface temperature (LST) data from satellites could be used[4]. The availability of new products based on a combination/merging of remotely-sensed land-surface temperature and in situ air temperature data represents an essential ingredient for the study of EDW and should be strongly encouraged for studies of high mountain regions. However, new approaches may be required to resolve temperature trends in complex topographic settings, where cloud cover is often present.

To complement an enhanced observational network, both global and fine-scale regional climate model simulations (historical and future) should be analyzed for sensitive regions in order to (1) investigate how well the models represent the specific climate variables as well as the interactions among them, and (2) identify and quantify the processes which are responsible for EDW. For regional studies, both statistically and dynamically downscaled model simulations may be useful, although multi-decadal simulations will be needed to fully investigate future projections of EDW. Models can also be used to perform sensitivity studies to investigate and quantify the role of specific climate variables. For example, although models have been used to quantify the magnitude of the snow/albedo feedback in the Arctic[56-58], there has been little attempt to do this in high elevation regions[12]. By doing a detailed analysis of regional energy budgets, one can potentially determine which energy budget component(s) are most responsible for temperature change and then in turn which climate variable(s) are most responsible for the change in energy budget. However, many of the variables interact with each other, and it can be difficult to untangle the dependencies. There have been statistical analyses that attempt to quantify specific processes (e.g. Naud et al.[45] who used a neural network method to quantify the sensitivity of downward longwave fluxes to changes in water vapor) but the dependencies among different variables remains problematic and new techniques are still required to improve such analyses.

Conclusion

Elevation-dependent warming is a poorly observed phenomenon that requires urgent attention, to ensure that potentially important changes in high mountain environments are adequately recorded by the global observational network. More rapid changes in high mountain climates would have consequences far beyond the immediate mountain regions, as mountains are "water towers" and the major source of water for large populations in lower elevation regions[59]. The social and economic consequences of enhanced warming in mountain regions could therefore be large, and this alone requires that close attention be paid to the issue. In addition, mountains provide habitat for many of the world's rare and endangered species, and the presence of many different ecosystems in close proximity enhances the ecological sensitivity of mountains to environmental change. Understanding how future climatic changes may impact the zonation of ecosystems in high mountain regions provides an equally compelling argument for fully understanding this issue. A strategy that combines a network of field observations, satellite remote sensing and high resolution climate modeling is required to fully address the problem.

References

1. Lawrimore, J.H. *et al*. An overview of the Global Historical Climatology Network monthly mean temperature data set, version 3. *J. Geophys. Res.* 116, D19121, doi:10.1029/2011JD016187 (2011).

2. Whiteman, C.D. *Mountain Meteorology: Fundamentals and Applications*. Oxford: Oxford University Press, 355pp (2000).

3. Pepin, N.C. Daly, C. & Lundquist, J. The influence of surface versus free-air decoupling on temperature trend patterns in the western United States. *J. Geophys. Res.* 116, D10109, doi:10.1029/2010JD014769 (2011).

4. Qin, J. Yang, K. Liang, S. & Guo, X. The altitudinal dependence of recent rapid warming over the Tibetan Plateau. *Clim. Chang.* 97, 321-327 (2009).

5. Diaz, H.F. & Bradley, R.S. Temperature variations during the last century at high elevation sites. *Clim. Chang.* 36, 253–279 (1997).

6. Pepin, N.C. & Lundquist, J.D. Temperature trends at high elevations: Patterns across the globe. *Geophys. Res. Lett.* 35, L14701, doi:10.1029/ 2008GL034026 (2008).

7. Ohmura, A. Enhanced temperature variability in high-altitude climate change. *Theor. Appl. Climatol.* 110, 499–508 (2012).

8. Wang, Q. Fan, X. & Wang, M. Recent warming amplification over high elevation regions across the globe. *Clim. Dyn.* 43, 87-101 (2014).

9. Yan, L. & Liu, X. Has Climatic Warming over the Tibetan Plateau Paused or Continued in Recent Years? *Journal of Earth, Ocean and Atmospheric Sciences*, 1, 13-28 (2014)

10. Giorgi, F. Hurrell, J. Marinucci, M. & Beniston, M. Elevation dependency of the surface climate change signal: a model study. *J. Clim.* 10, 288–296 (1997).

11. Bradley, R.S. Keimig, F.T. & Diaz, H.F. Projected temperature changes along the American cordillera and the planned GCOS network. *Geophys. Res. Lett.* 31, L16210, doi:10.1029/2004GL020229 (2004).

12. Rangwala, I. Amplified water vapour feedback at high altitudes during winter. *Int. J. Climatol.* 33, 897–903 (2013).

13. Collins, M. *et al.* Long-term Climate Change: Projections, Commitments and Irreversibility. In: *Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change* [Stocker, T.F. *et al.* (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, pp. 1029–1136, doi:10.1017/CBO9781107415324.024 (2013).

14. Pepin, N. & Seidel, D.J. A global comparison of surface and free-air temperatures at high elevations. *J. Geophys. Res.* 110, D03104 (2005).

15. Rasmussen, R. *et al.* High-Resolution Coupled Climate Runoff Simulations of Seasonal Snowfall over Colorado: A Process Study of Current and Warmer Climate. *J. Climate*, 24 (12), 3015–3048 (2011).

16. Rasmussen, R. *et al.* Climate Change Impacts on the Water Balance of the Colorado Headwaters: High-Resolution Regional Climate Model Simulations. *Journal of Hydrometeorology*, 15:3, 1091-1116 (2014).

17. Kendon, E.J. *et al.* Heavier summer downpours with climate change revealed by weather forecast resolution model. *Nature Climate Change*, 4(7), 570-576 (2014).

18. Kotlarski,S. Bosshhard, T. Luthi, D. Paul, P. & Schar, C. Elevation gradients of European climate change in the regional climate model COSMO-CLM. *Clim. Change*, 112: 189-215 (2012)

19. Beniston, M. & Rebetez, M. Regional behavior of minimum temperatures in Switzerland for the period 1979–1993. *Theor. Appl. Climatol.* 53, 231–243 (1996).

20. Diaz, H.F. & Graham, N.E. Recent changes in tropical freezing heights and the role of sea surface temperature. *Nature* 383, 152–155 (1996).

21. Rangwala, I. & Miller, J.R. Climate change in mountains: a review of elevation-dependent warming and its possible causes. *Clim. Chang.* 114, 527–547 (2012).

22. Chapin, F.S. *et al.* Role of land-surface changes in arctic summer warming. *Science* 310, 657–660 (2005).

23. Jeong, J.-H. *et al.* Greening in the circumpolar high-latitude may amplify warming in the growing season. *Clim. Dyn.* 38(7-8), 1421-1431 (2012).

24. Fyfe, J.C. & Flato, G.M. Enhanced climate change and its detection over the Rocky Mountains. *J. Clim.* 12, 230–243 (1999).

25. Rangwala, I. Miller, J.R. Russell, G.L. & Xu, M. Using a global climate model to evaluate the influences of water vapor, snow cover and atmospheric aerosol on warming in the Tibetan Plateau during the twenty-first century. *Clim. Dyn.* 34, 859–872 (2010).

26. Ceppi, P. Scherrer, S.C. Fischer, A.M. & Appenzeller, C. Revisiting Swiss temperature trends 1959-2008. *Int. J. Climatol.* 32, 203–213 (2010).

27. Scherrer, S. Ceppi, P. Croci-Maspoli, M. & Appenzeller, C. Snow-albedo feedback and Swiss spring temperature trends. *Theor Appl Climatology* 110, 509-516 (2012).

28. Hantel, M., Ehrendorfer, M. & Haslinger, A. Climate sensitivity of snow cover duration in Austria. *Int. J of Climatol.* 20, 615-640 (2000).

29. Barnett, T.P. *et al.* Human-Induced Changes in the Hydrology of the Western United States, *Science* 319, 1080-1083 (2008).

30. Holtmeier, F-K. & Broll, G. Sensitivity and response of northern hemisphere altitudinal and polar treelines to environmental change at landscape and local scales. *Glob. Ecol. Biogeogr.* 14(5), 395–410 (2005).

31. Jiang, D. Zhang, Y. & Lang, X. Vegetation feedback under future global warming. *Theor. Appl. Climatol.* 106, 211–227 (2011).

32. Rikiishi, K. & Nakasato, H. Height dependence of the tendency for reduction in seasonal snow cover in the Himalaya and the Tibetan Plateau region, 1966-2001. *Ann. Glaciol.* 43, 369–377 (2006).

33. Archer, D.R. & Fowler, H.J. Spatial and temporal variations in precipitation in the Upper Indus Basin, global teleconnections and hydrological implications. *Hydrol. Earth Syst. Sci.* 8, 47–61 (2004).

34. Bhutiyani, M. Kale, V. & Pawar, N. Climate change and the precipitation variations in the northwestern Himalaya: 1866–2006. *Int. J. Climatol.* 30, 535–548 (2010).

35. Kothawale, D. Munot, A. & Kumar, K.K. Surface air temperature variability over India during 1901-2007, and its association with ENSO. *Clim. Res.* 42, 89–104 (2010).

36. Rangwala, I. Barsugli, J. Cozzetto, K. Neff, J. & Prairie, J. Mid-21st century projections in temperature extremes in the southern Colorado Rocky Mountains from regional climate models. *Clim Dyn*. doi:10.1007/s00382-011-1282-z (2012).

37. Vuille, M. Bradley, R. Werner, M. & Keimig, F. 20th century climate change in the tropical Andes: observations and model results. *Clim. Chang.* 59, 75–99 (2003).

38. Liu, X. Cheng, Z. Yan, L. & Yin, Z.-Y. Elevation dependency of recent and future minimum surface air temperature trends in the Tibetan Plateau and its surroundings. *Glob. Planet. Chang.* 68, 164–174 (2009).

39. Held, I. M. & Soden, B.J. Robust responses of the hydrological cycle to global warming. *J. Clim.* 19, 5686–5699 (2006).

40. Rangwala, I. Miller, J. & Xu, M. Warming in the Tibetan Plateau: Possible influences of the changes in surface water vapor. *Geophys. Res. Lett.* 36, L06703 (2009).

41. Duan, A. & Wu, G. Change of cloud amount and the climate warming on the Tibetan Plateau. *Geophys. Res. Lett.* 33, L22704 (2006).

42. Ruckstuhl, C. Philipona, R. Morland, J. & Ohmura, A. Observed relationship between surface specific humidity, integrated water vapor, and longwave downward radiation at different altitudes. *J. Geophys. Res.* 112, D03302 (2007).

43. Naud, C. M. Miller, J. R. & Landry, C. Using satellites to investigate the sensitivity of longwave downward radiation to water vapor at high elevations. *J. Geophys. Res.-Atmos.* 117, D05101, doi:10.1029/2011JD016917 (2012).

44. Philipona, R. Dürr, B. Ohmura, A. & Ruckstuhl, C. Anthropogenic greenhouse forcing and strong water vapor feedback increase temperature in Europe. *Geophys. Res. Lett.* 32, L19809 (2005).

45. Naud, C. M. Chen, Y. Rangwala, I. & Miller, J. R. Sensitivity of downward longwave surface radiation to moisture and cloud changes in a high elevation region. *J. Geophys. Res* 118,10,072–10,081, doi:10.1002/jgrd.50644. (2013).

46. Ohmura, A. On the cause of "Fram" type seasonal change in diurnal amplitude of air temperature in polar regions. *J. Climatol.* 4, 325-338 (1984).

47. Ramanathan, V. & Carmichael, G. Global and regional climate changes due to black carbon. *Nat. Geosci.* 1(4), 221–227 (2008).

48. Bonasoni, P. *et al.*, Atmospheric Brown Clouds in the Himalayas: first two years of continuous observations at the Nepal-Climate Observatory at Pyramid (5079 m). *Atmos. Chem. Phys. Discuss*., 10, 4823-4885 (2010).

49. Vernier, J-P., Thomason, L.W. & Kar, J. CALIPSO detection of an Asian tropopause aerosol layer, *Geophysical Research Letters*, 38, L07804, doi:10.1029/2010GL046614 (2011)

50. Lawrence, M.G., Asia under a high-level brown cloud. *Nature Geoscience*, 4, 352-353 (2011).

51. Xu, B. *et al.* Black soot and the survival of Tibetan glaciers. *Proc. Natl. Acad. Sci. U.S.A.* 106, 22114–22118 (2009).

52. Lau, W. Kim, M. Kim, K. & Lee, W. Enhanced surface warming and accelerated snow melt in the Himalayas and Tibetan Plateau induced by absorbing aerosols. *Environ. Res. Lett.* 5 (2), 025204 (2010).

53. Gautam, R. Hsu, N.C. Lau, W.K. Yasunari, T. Satellite observations of desert dust-induced Himalayan snow darkening *J. Geophys.,Res*, 40(5),988-993. DOI 10.1002/grl.50226 (2013).

54. Painter, T.H. *et al.* Impact of disturbed desert soils on duration of mountain snow cover. *Geophys. Res. Lett.* 34, L12502 (2007).

55. Santer, B.D. et al., Contributions of anthropogenic and natural forcing to recent tropopause height changes. *Science*, 301, 479–483 (2003).

56. Holland, M.M. & Bitz, C.M. Polar amplification of climate change in coupled models, *Clim. Dyn*. 21, 221–232 (2003).

57. Winton, M. Amplified Arctic climate change: What does surface albedo feedback have to do with it?, *Geophys. Res. Lett*., 33, L03701, doi:10.1029/2005GL025244 (2006).

58. Crook, J. A. Forster, P.M. & Stuber, N. Spatial patterns of modeled climate feedback and contributions to temperature response and polar amplification, *J. Clim.*, 24(14), 3575–3592, doi:10.1175/2011JCLI3863.1 (2011).

59. Viviroli, D. Dürr, H.H. Messerli, B. Meybeck, M. & Weingartner, R. Mountains of the world, water towers for humanity: Typology, mapping, and global significance. Water Resources Research, 43 (7), W07447 http://dx.doi.org/10.1029/ 2006WR005653 (2007).

60. Liu, X. & Chen, B. Climatic Warming in the Tibetan Plateau during recent decades. *Int. J. Climatol.* 20, 1729-1742 (2000).

Full list of authors and affiliations

Nick Pepin, University of Portsmouth, UK

Ray Bradley, University of Massachussetts, USA

Henry Diaz, CIRES, National Oceanic and Atmospheric Administration, Boulder, USA

Michel Baraer, ETS – University of Quebec, Canada

Ernesto Caceres, Instituto Nacional de Meterología e Hidrología, Ecuador

Nathan Forsythe, University of Newcastle, UK

Hayley Fowler, University of Newcastle, UK

Greg Greenwood, University of Berne, Switzerland

Muhammad Zia Hashmi, Water Resources Section, Global Change Impact Studies Centre (GCISC), Islamabad, Pakistan

Xiaodong Liu, SKLLQG, Institute of Earth Environment and CETES, Chinese Academy of Sciences, China

Jim Miller, Rutgers University, USA

Liang Ning, University of Massachussetts, USA

Atsumu Ohmura, ETH, Zurich, Switzerland

Elisa Palazzi, Institute of Atmospheric Sciences and Climate-National Research Council (ISAC-CNR), Italy

Imtiaz Rangwala, Western Water Assessment; Physical Sciences Division, NOAA ESRL, Boulder, USA

Wolfgang Schöner, University of Graz, Austria

Igor Severskiy, Institute of Geography, Kazakhstan

Maria Shahgedanova, University of Reading, UK
Mengben Wang, Shanxi University, China
Scott Williamson, University of Alberta, Canada
Daqing Yang, National Hydrology Research Centre (NHRC), Canada

Acknowledgements

We acknowledge the Mountain Research Initiative for funding an international workshop in Payerbach Austria in April 2014 on Elevation-Dependent Warming at which the idea for this paper was conceived and drafted.

Author Contributions

NP, RB, HF acted as editors for collation and editing of material; NP, HF, EP, IR, JM wrote substantial sections of the main text; NP, SW, EP, NF wrote substantial sections of the supplementary material; SW helped with referencing; XL provided Figure 1; MB provided Figure 2; IR provided Tables. Other group members provided comments and editorial suggestions as appropriate.

Competing Financial Interest Statement

There are no competing financial interests.

Figure Captions

Figure 1: **Elevation-dependent warming (EDW) over and around the Tibetan Plateau**

a) from 1961 to 1990 for annual mean surface air temperature (TA, *Liu and Chen*[60]) compared with more recent trends updated to 2012; and from 1961 to 2012 for b) annual mean minimum temperature (TN), c) autumn mean surface temperature, and d) winter mean surface temperature. Bars represent elevation and trend magnitude is plotted on the y axis (presentation format similar to the original paper).

Figure 2: **Schematic representation of the relative vertical profile in atmospheric warming expected to result from various mechanisms**.

The details of each mechanism are discussed in the text. The y axis represents elevation, and x axis the rate of warming (or expected trend magnitude). The curves are relative in that the absolute value on the x axis is not important, rather the shape of the signal in terms of contrast with elevation is the key illustration.

Table 1. Results from studies that investigated altitudinal gradient in warming rates (updated from Rangwala and Miller 2012).

Altitudinal gradient in the warming rate	Observations			Models		
	T_{min}	T_{max}	T_{avg}	T_{min}	T_{max}	T_{avg}
Increases with elevation	Annual[2,6a,31] Winter[2,5g,14] Spring[21] Autumn[21, 26e]	Annual[2,24] Summer[21]	Annual[13,19e,23b,17c,29,12] All Seasons[21] Winter[26e]	Annual[14] Winter[14,20,9,11] Spring[14,20]	Winter[20] Spring[20] Autumn[9]	Annual[22] Winter[3,4,8,22] Spring[3,8,22] Summer[31]
Decreases with elevation	Winter[2]	Winter[21]	Annual[15,27] Winter[3g] Autumn[3g]	Summer[27]	-	Annual[10f] Spring[10f] Autumn[10f]
No significant gradient	-	Annual[1]	Annual[3,30,7d,25] All Seasons[28, 30]	-	-	Annual[3]
No significant gradient but largest warming rates at an intermediate elevation	-	Annual[16]	Annual[18,20c] Spring[3g]	-	-	Spring[3]

Superscript letters refer to:

[a]No significant gradient but greater warming at higher elevations relative to regions between 0-500m
[b]Radiosonde data; clearest signal in the tropics
[c]65% of the regional groups examined showed fastest trends at highest elevations and 20% showed fastest trends at intermediate elevations
[d]High-elevation trends based on borehole data
[e]Satellite-derived temperature estimations
[f]Reanalyses
[g]Gridded Data

Superscript numbers refer to the following studies:

[1]Bhutiyani et al. (2007), [2]Beniston and Rebetez (1996), [3]Ceppi et al. (2010), [4]Chen et al. (2003), [5]Diaz and Eischeid (2007), [6]Diaz and Bradley (1997), [7]Gilbert and Vincent (2013), [8]Giorgi et al. (1997), [9]Hu et al. (2013), [10]Hu et al. (2014), [11]Im and Ahn (2011), [12]Li et al. (2011), [13]Liu and Chen (2000), [14]Liu et al. (2009), [15]Lu et al. (2010), [16]McGuire et al. (2012), [17]Ohmura (2012), [18]Pepin and Lundquist (2008), [19]Qin et al. (2009), [20]Rangwala et al. (2012), [21]Rangwala et al. (2009), [22]Rangwala et al. (2010), [23]Seidel and Free (2003), [24]Shrestha et al. (1999), [25]Tang and Arnone (2013), [26]Tao et al. (2013), [27]Vuille and Bradley (2000), [28]Vuille et al. (2003), [29]Wang et al. (2014), [30]You et al. (2010), [31]Zubler et al. (2013)

References (Table 1)

1 Beniston, M. & Rebetez, M. Regional behavior of minimum temperatures in Switzerland for the period 1979–1993. *Theoretical and applied climatology* **53**, 231-243 (1996).

2 Bhutiyani, M., Kale, V. & Pawar, N. Long-term trends in maximum, minimum and mean annual air temperatures across the Northwestern Himalaya during the twentieth century. *Climatic Change* **85**, 159-177 (2007).

3 Ceppi, P., Scherrer, S., Fischer, A. & Appenzeller, C. Revisiting Swiss temperature trends 1959–2008. *International Journal of Climatology*, doi:10.1002/joc.2260 (2010).

4 Chen, B., Chao, W. & Liu, X. Enhanced climatic warming in the Tibetan Plateau due to doubling CO2: a model study. *Climate Dynamics* **20**, 401-413 (2003).

5 Diaz, H. & Eischeid, J. Disappearing 'alpine tundra,'Köppen climatic type in the western United States. *Geophysical research letters* **34**, L18707 (2007).

6 Diaz, H. F. & Bradley, R. S. Temperature variations during the last century at high elevation sites. *Climatic Change* **36**, 253-279 (1997).

7 Gilbert, A. & Vincent, C. Atmospheric temperature changes over the 20th century at very high elevations in the European Alps from englacial temperatures. *Geophysical Research Letters* **40**, 2102-2108, doi:10.1002/grl.50401 (2013).

8 Giorgi, F., Hurrell, J., Marinucci, M. & Beniston, M. Elevation dependency of the surface climate change signal: a model study. *Journal of Climate* **10**, 288-296 (1997).

9 Hu, Y., Maskey, S. & Uhlenbrook, S. Expected changes in future temperature extremes and their elevation dependency over the Yellow River source region. *Hydrology and Earth System Sciences* **17**, 2501-2514 (2013).

10 Hu, Z., Zhang, C., Hu, Q. S. & Tian, H. Temperature Changes in Central Asia from 1979 to 2011 Based on Multiple Datasets. *Journal of Climate* **27**, 1143-1167, doi:DOI: 10.1175/JCLI-D-13-00064.1 (2014).

11 Im, E.-S. & Ahn, J.-B. On the elevation dependency of present-day climate and future change over Korea from a high resolution regional climate simulation. *Journal of the Meteorological Society of Japan* **89**, 89-100 (2011).

12 Li, Z. *et al.* Spatial and temporal trends of temperature and precipitation during 1960–2008 at the Hengduan Mountains, China. *Quaternary International* **236**, 127-142 (2011).

13 Liu, X. & Chen, B. Climatic warming in the Tibetan Plateau during recent decades. *International Journal of Climatology* **20**, 1729-1742 (2000).

14 Liu, X., Cheng, Z., Yan, L. & Yin, Z. Elevation dependency of recent and future minimum surface air temperature trends in the Tibetan Plateau and its surroundings. *Global and Planetary Change* **68**, 164-174 (2009).

15 Lu, A., Kang, S., Li, Z. & Theakstone, W. Altitude effects of climatic variation on Tibetan Plateau and its vicinities. *Journal of Earth Science* **21**, 189-198 (2010).

16 McGuire, C. R., Nufio, C. R., Bowers, M. D. & Guralnick, R. P. Elevation-dependent temperature trends in the Rocky Mountain Front Range: changes over a 56-and 20-year record. *PloS one* **7**, e44370 (2012).

17 Ohmura, A. Enhanced temperature variability in high-altitude climate change. *Theoretical and Applied Climatology* **110**, 499-508 (2012).

18 Pepin, N. & Lundquist, J. Temperature trends at high elevations: patterns across the globe. *Geophysical research letters* **35**, 1-L14701 (2008).

19 Qin, J., Yang, K., Liang, S. & Guo, X. The altitudinal dependence of recent rapid warming over the Tibetan Plateau. *Climatic Change* **97**, 321-327 (2009).

20 Rangwala, I., Barsugli, J., Cozzetto, K., Neff, J. & Prairie, J. Mid-21st century projections in temperature extremes in the southern Colorado Rocky Mountains from regional climate models. *Climate Dynamics* **39**, 1823-1840 (2012).

21 Rangwala, I., Miller, J., Russell, G. & Xu, M. Warming in the Tibetan Plateau: Possible influences of the changes in surface water vapor. *Geophysical research letters* **36**, L06703 (2009).

22 Rangwala, I. & Miller, J. R. Twentieth Century Temperature Trends in Colorado's San Juan Mountains. *Arctic, Antarctic, and Alpine Research* **42**, 89-97 (2010).

23 Seidel, D. & Free, M. Comparison of lower-tropospheric temperature climatologies and trends at low and high elevation radiosonde sites. *Climatic Change* **59**, 53-74 (2003).

24 Shrestha, A., Wake, C., Mayewski, P. & Dibb, J. Maximum temperature trends in the Himalaya and its vicinity: an analysis based on temperature records from Nepal for the period 1971–94. *Journal of Climate* **12**, 2775-2786 (1999).

25 Tang, G. & Arnone, J. A. Trends in surface air temperature and temperature extremes in the Great Basin during the 20th century from ground-based observations. *Journal of Geophysical Research: Atmospheres* **118**, 3579-3589 (2013).

26 Tao, J. *et al.* Elevation-dependent temperature change in the Qinghai–Xizang Plateau grassland during the past decade. *Theoretical and Applied Climatology* **117**, 61-71 (2013).

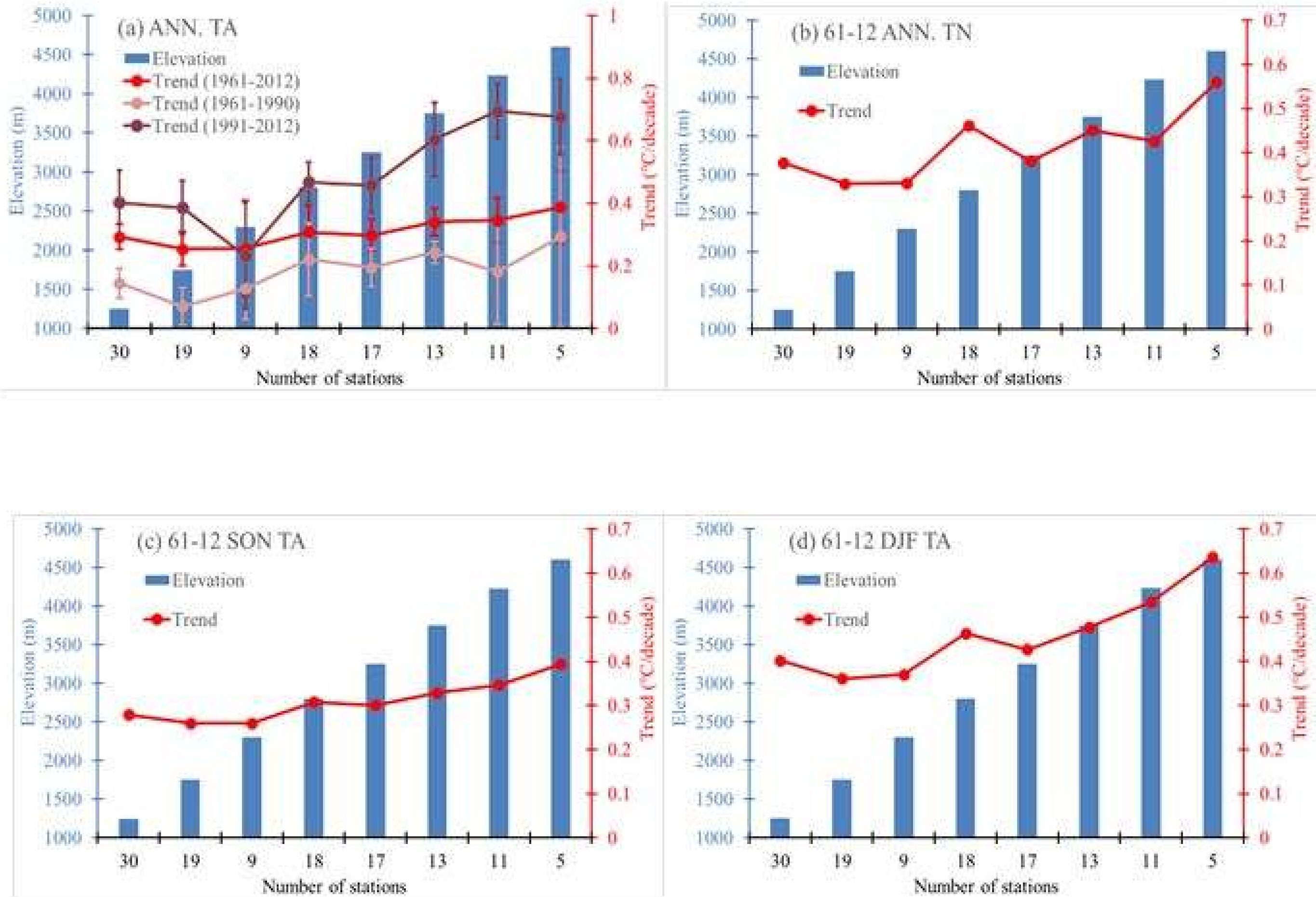
27 Vuille, M. & Bradley, R. Mean annual temperature trends and their vertical structure in the tropical Andes. *Geophys. Res. Lett* **27**, 3885–3888 (2000).

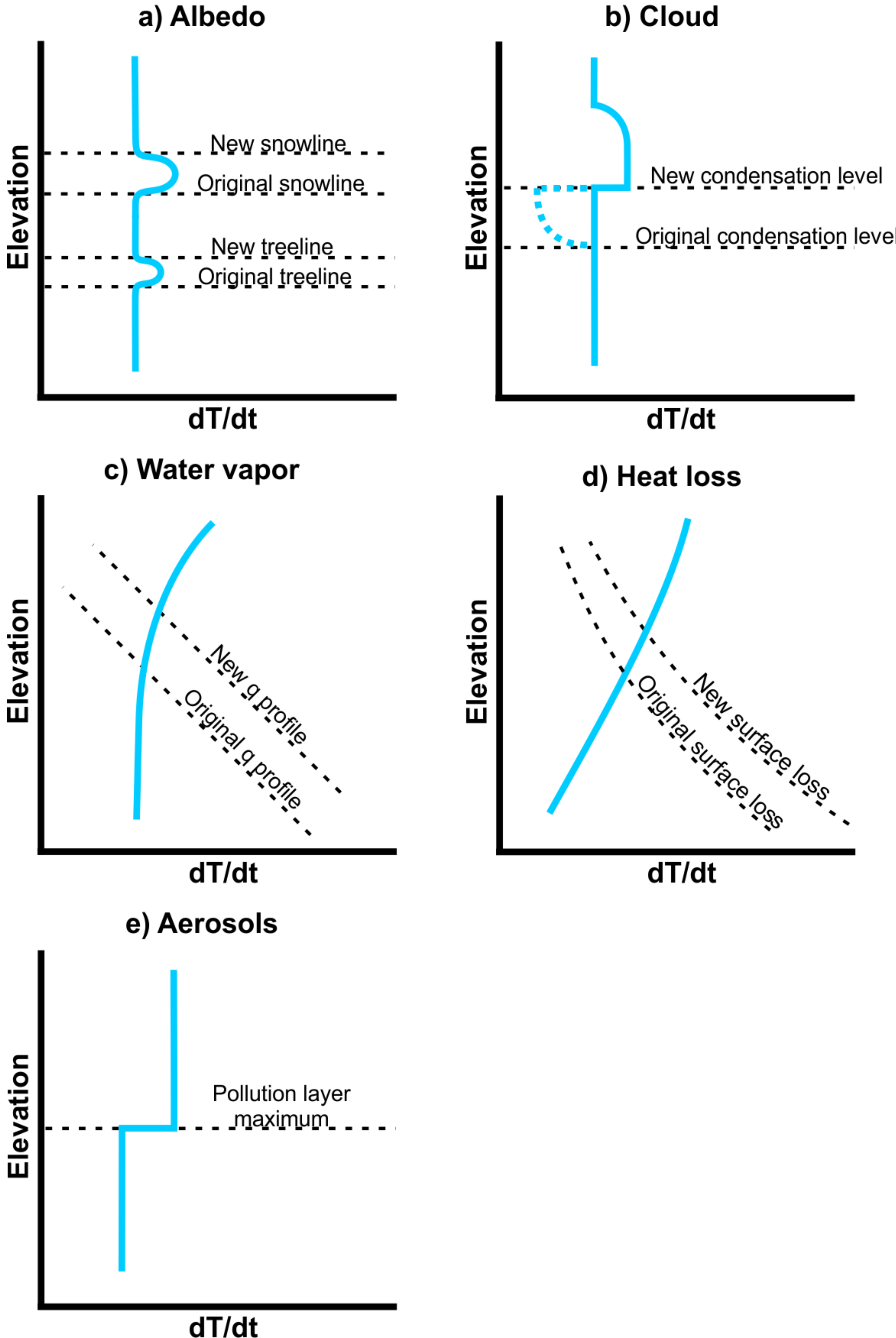
28 Vuille, M., Bradley, R., Werner, M. & Keimig, F. 20th century climate change in the tropical Andes: observations and model results. *Climatic Change* **59**, 75-99 (2003).

29 Wang, Q., Fan, X. & Wang, M. Recent warming amplification over high elevation regions across the globe. *Climate Dynamics* **43**, 87-101 (2014).

30 You, Q. *et al.* Relationship between temperature trend magnitude, elevation and mean temperature in the Tibetan Plateau from homogenized surface stations and reanalysis data. *Global and Planetary Change* (2010).

31 Zubler, E. M. *et al.* Localized climate change scenarios of mean temperature and precipitation over Switzerland. *Climatic Change* **125**, 237-252 (2014).


(a) ANN. TA
Elevation
Trend (1961-2012)
Trend (1961-1990)
Trend (1991-2012)
Elevation (m)
Trend (°C/decade)
Number of stations
30
19
9
18
17
13
11
5
(b) 61-12 ANN. TN
Elevation
Trend
Elevation (m)
Trend (°C/decade)
Number of stations
(c) 61-12 SON TA
Elevation
Trend
Elevation (m)
Trend (°C/decade)
Number of stations
(d) 61-12 DJF TA
Elevation
Trend
Elevation (m)
Trend (°C/decade)
Number of stations


a) Albedo
New snowline
Original snowline
New treeline
Original treeline
Elevation
dT/dt
b) Cloud
New condensation level
Original condensation level
Elevation
dT/dt
c) Water vapor
New q profile
Original q profile
Elevation
dT/dt
d) Heat loss
New surface loss
Original surface loss
Elevation
dT/dt
e) Aerosols
Pollution layer
maximum
Elevation
dT/dt