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Permafrost model sensitivity to seasonal climatic changes and extreme events in mountainous regions

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Abstract
Climate models project considerable ranges and uncertainties in future climatic changes. To assess the potential impacts of climatic changes on mountain permafrost within these ranges of uncertainty, this study presents a sensitivity analysis using a permafrost process model combined with climate input based on delta-change approaches. Delta values comprise a multitude of coupled air temperature and precipitation changes to analyse long-term, seasonal and seasonal extreme changes on a typical low-ice content mountain permafrost location in the Swiss Alps. The results show that seasonal changes in autumn (SON) have the largest impact on the near-surface permafrost thermal regime in the model, and lowest impacts in winter (DJF). For most of the variability, snow cover duration and timing are the most important factors, whereas maximum snow height only plays a secondary role unless maximum snow heights are very small. At least for the low-ice content site of this study, extreme events have only short-term effects and have less impact on permafrost than long-term air temperature trends.

Keywords: permafrost modelling, climate change, climate extreme, snow cover duration, seasonal changes

1. Introduction

Permafrost as a thermal state of the polar and mountainous subsurface has shown increasing temperatures during the past decades (e.g. Brown *et al* 2010, Romanovsky *et al* 2010, Vieira *et al* 2010), which can at least partly be attributed to observed changes in the atmosphere (Harris *et al* 2003). Concerns of increasing permafrost temperatures in mountain permafrost are mostly related to stability issues of steep slopes and infrastructures or potential future water resources (Harris *et al* 2009, Bommer *et al* 2010).

The response of permafrost temperatures to changes in the atmospheric conditions is non-linear and depends on various factors such as the subsurface composition, ice content or the timing and duration of the seasonal snow cover, which temporarily decouples the ground from the atmosphere (Zhang *et al* 2001, Schneider *et al* 2012, Gubler *et al* 2013, Langer *et al* 2013). Improving the understanding of the sensitivity of permafrost to climatic changes and climatic extreme events has been the objectives of various studies during the past years (e.g. Lawrence *et al* 2008, Lütschg *et al* 2008, Slater and Lawrence 2013, Scherler *et al* 2013, Westermann *et al* 2011). Salzmann *et al* (2007a, 2007b) demonstrated the added value of using output from regional climate models (RCMs) for mountain permafrost modelling and Scherler *et al* (2013) showed that the COUP model, a coupled heat and mass transfer model (Jansson 2012), combined with output from RCMs can be a suitable model setup for in-depth investigations of climate–permafrost interactions.

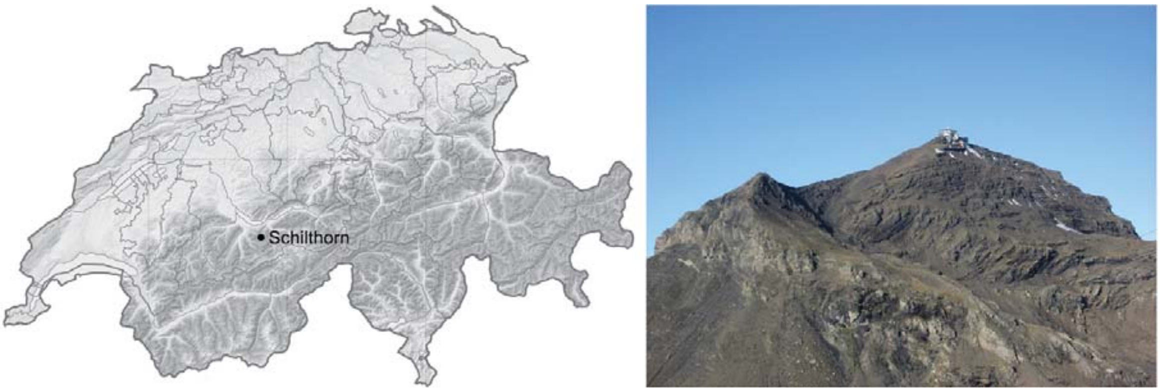

Figure 1. Location and picture of the study site: Schilthorn, Bernese Alps, Northern Swiss Alps (2900 m asl). Photo: J Noetzli.

In general, climatic extremes are supposed to have particular adverse impacts (IPCC 2012). However, due to the 'rareness by definition' (statistically) of extreme events and the still relatively coarse resolution of climate models, impact analyses of extreme events are hampered. Nevertheless, their relevance for mountain permafrost on short timescales has been proven through observations. Anomalously warm and dry seasons had clearly observably impacts on the active layer thickness (ALT) (Central Europe 2003; e.g. Gruber *et al* 2004, Hilbich *et al* 2008; Svalbard 2006; e.g. Isaksen *et al* 2007). The effective sensitivity of permafrost on climatic extreme events on the long-term, however, is not fully clear yet.

In this contribution, the sensitivity of a soil model (COUP model; see section 2) at a typical high mountain permafrost site (Schilthorn, Swiss Alps) is analysed by an extensive delta-change approach. This approach consists of applying a multitude of pairs of delta values for air temperature (by addition) and for precipitation (by multiplication) reflecting potential changes in the two most important climate variables. Due to the rigid and theoretical character of this approach, it represents an optimal way to assess the sensitivity. It is not restricted to projected changes as in Scherler *et al* (2013) but to various possible future changes in permafrost as the uncertainty range of climate model scenarios is large. This modelling approach also permits to test single parameter variation to better understand the impact and its persistence of any given climate change.

2. Field site, model, data and methods

Model simulations were conducted for an alpine permafrost site, the Schilthorn, Bernese Alps, Northern Swiss Alps (2900 m asl) (figure 1). This site is representative for low-ice content mountain permafrost. Low-ice content permafrost sites with a thermal regime close to the melting point are studied because their sensitivity to climate change is supposed to be largest, in contrast to high-ice content rock glaciers, whose thermal regime react much slower due to the necessary latent heat for ice melting (Scherler *et al* 2013). These highly sensitive sites are of particular interest in the Alps regarding slope stability issues. The Schilthorn has been chosen for this study as this site has one of the longest data records in Alpine permafrost research including (micro)meteorological measurements, borehole temperatures measurements down to 100 m, soil moisture and continuous electrical resistivity tomography (ERT) measurements as an indicator for spatial freeze and thaw processes (Hoelzle and Gruber 2008, Hilbich *et al* 2011). COUP model simulations for Schilthorn were already conducted by Scherler *et al* (2010, 2013) and Engelhardt *et al* (2010).

The one-dimensional COUP model, as constructed in this study, is composed of 50 vertical layers with increasing thickness with depth: from 5 cm in the uppermost layers to 5 m in the lowermost ones. The maximal investigation depth is 70 m. Details and governing equations of the model and further details are given in the appendix.

Validation experiments using the micrometeorological data (as input) and observations of borehole temperatures at Schilthorn were successfully conducted for a 10-year period and are published in Scherler *et al* (2010, 2013). The various climate sensitivity experiments of this study are shown in figure 2 and are conducted each with a 30-year model spin-up. A reference run REF (figures 2 and 3) has been driven with air temperature and precipitation daily time series from 30 years observations at the MeteoSwiss station of Mürren (1638 m asl, 1 km distance) with a correction for the altitude difference (Stocker-Mittaz *et al* 2002, Scherler *et al* 2010). This reference run with present-day conditions is used as a base for assessing the potential effect of different delta changes. The climate scenarios are provided by the EU-ENSEMBLES program (van der Linden and Mitchell 2009) and based on 14 RCMs driven by different GCMs and forced by the A1B SRES scenarios. The RCM experiments cover the period 1951–2099 and have been carried out at a horizontal resolution of about 25 km. The scenario ranges for temperature (T) and precipitation (P) used in this study were prepared by the CH2011 project (CH2011 2011).

For the sensitivity experiments of this study the delta-change approach has been applied for long-term and seasonal changes as well as for changes in seasonal extremes. The approach consists of applying a delta value of temperature (ΔT) and/or precipitation (ΔP) time series onto the REF. For the sensitivity to annual changes (ANN), the delta values range from -20% to $+20\%$ for precipitation and from 0 to

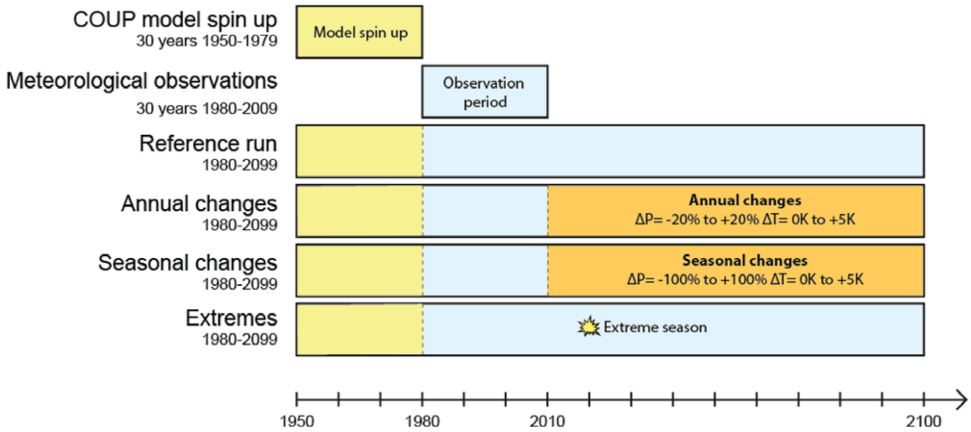

Figure 2. Scheme of the simulation setup for this study.

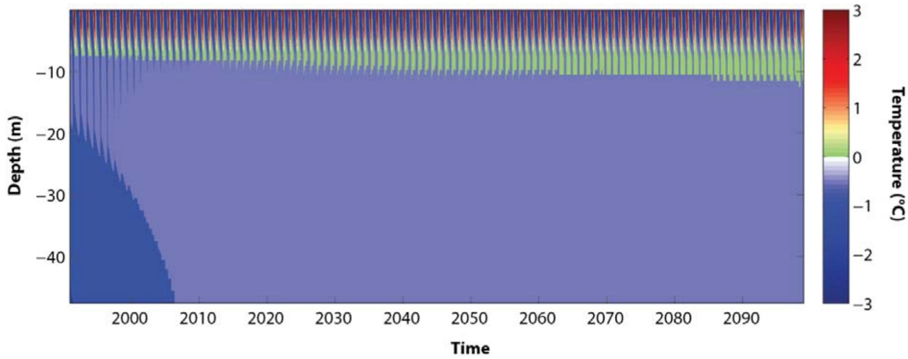

Figure 3. Simulated ground thermal regime evolution of the reference run for Schilthorn based on meteorological observations of the period 1980–2010.

+5 K for air temperature. Annual delta values are applied for all years from 2010 to 2099.

The procedure is similar for the sensitivity to seasonal changes (SEA), where the delta values are applied every year to three consecutive months only corresponding to a selected season: winter (DJF), spring (MAM), summer (JJA) and fall (SON). Delta values range from −100% to +100% for precipitation and from 0 to +5 K for air temperature. For the sensitivity to seasonal extreme events (EXT), the hottest, coldest, driest and wettest conditions with a return period of 30 years have been identified in the observational data set for Mürren. Their deviation from 30-years median air temperature and precipitation served as seasonal extreme delta values. Every combination (hot and dry JJA, cold and wet MAM, etc) has been tested in the model by a single application in 2020 (arbitrary) of the extreme seasonal delta values. Corresponding changes in ALT and persistence of the anomaly were analysed.

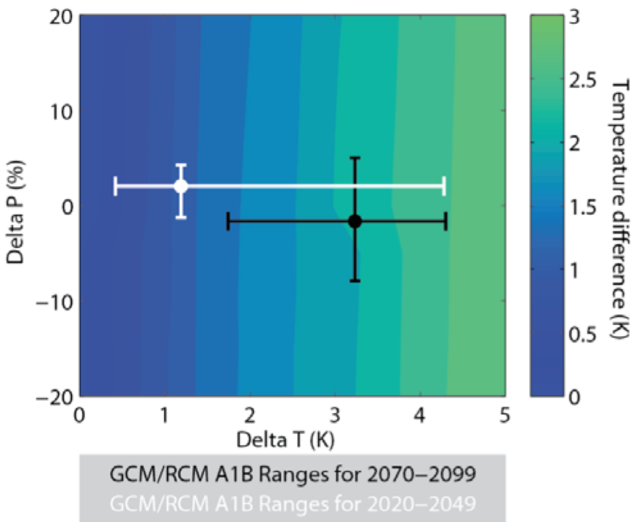

Figure 4. Difference in mean annual soil temperature at 5 m depth at the end of the century between ANN (i.e. the application of an annual constant ΔT and ΔP) and the reference run (ANN–REF). Range of GCM/RCM projections (scenario A1B) for two time periods are indicated by the white/black crosses.

3. Results

3.1. Sensitivity to annual changes

In a first step, constant air temperature and/or precipitation changes over the entire year were applied as described in section 2 (ANN). Figure 4 shows the resulting differences of mean annual temperatures at 5 m depth at the end of the century between ANN and REF. Generally, permafrost reacts to an increase in air temperature in a non-trivial and spatially heterogeneous manner due to the insulating influence of snow cover and an inhomogeneous distribution of surface and subsurface parameters (such as thermal conductivity or ice content). A warming of 5 K in air temperature will not be followed by an equivalent warming in the ground but by a ~3 K warming. With respect to the GCM/RCM projected range of air temperature increase between 1.5 and 4.3 K by the end of the century (indicated by the bars in figure 4), the temperature is expected to rise from 1.5 to 2.4 K at 5 m depth and from 1.3 to 2.1 K at 10 m depth (not shown). All simulations of this study show that the seasonal thaw layer is not able to entirely refreeze as soon as the ALT reaches approximately 12 m.

The vertical pattern of the colours in figure 4 indicates that for Schilthorn conditions, permafrost has a low sensitivity to changes in the amount of precipitation on the long-term, since for a given ΔT, the soil temperature remains almost constant for any ΔP. As the GCM/RCM projected future changes in annual precipitation are expected to be moderate (figure 4), one can assume that the impact of a general increase or decrease of the amount of precipitation on permafrost is negligible in this case. Changes in the seasonal repartition of precipitation are nevertheless important (see below).

3.2. Sensitivity to seasonal changes

The above shown application of constant delta-change values throughout the year is indicative for the general sensitivity of the modelled permafrost site to long-term trends. However, it is important to consider also changes in every single season as the soil model reacts in a different manner depending on the presence, respectively the absence of a snow cover. Snow (especially new snow) has a very low thermal conductivity and is thus a good insulator (French 2007). The presence of a snow cover partly or totally decouples the ground thermal regime from the processes in the atmosphere due to its insulation properties but also because of changes in albedo (Armstrong and Brun 2008). Seasonal air temperature and precipitation changes can therefore impact the permafrost thermal regime in a complex manner (Lütschg *et al* 2008, Engelhardt *et al* 2010).

Figure 5 shows the difference in mean annual soil temperature at 5 m depth at the end of the century between the reference run and the seasonal sensitivity run (SEA–REF). The model shows a low sensitivity to changes in DJF. A strong positive ΔT does not lead to a warming of the soil as a snow cover is already building up from October onwards and changes in DJF only affect the height of the snow cover and not its presence or absence. Even +5 K to the seasonal mean air temperature is not sufficient to prevent the existence of a snow cover: the soil temperature regime keeps being decoupled from atmospheric forcing. The subsurface is cooling only in the case of a strong negative ΔP (e.g. −100%, this means no precipitation during concerned months) when the snow cover (formed in October and November) stays around 10 to 20 cm (cf figure 6(a)) which is not sufficient to fully isolate the ground: thus, more negative winter air temperatures can penetrate and cool the ground. The individual ENSEMBLE GCM/RCM model chains show a large spread for changes in air temperature in winter by the end of the century (+0.8 to +5.1 K) but only moderate changes in precipitation (−15.9 to +19.6%).

A bigger impact on permafrost occurs if the delta values are applied during SON. A strong negative ΔP has a slight cooling effect: the snow does not built up as usual in fall permitting the negative air temperatures in the very early winter to influence the ground thermal regime. Because air temperatures during SON are often close to the transition temperature between snow- and rainfall any positive ΔT will delay the presence of snow until November (figure 6(b)) permitting positive temperatures of September and October to warm up the ground. This is a crucial result as the climate models for the end of the century project a strong warming (+1.6 to +6.3 K) and a large spread of possible change in precipitation (−24.4 to +13.8%) for SON.

For MAM and JJA, the changes in precipitation have opposite effects. In MAM, the warming of the soil is delayed by a positive ΔP. As precipitation usually falls as snow in early spring, the increased precipitation permits the snow cover to last a bit longer and therefore to be less exposed to positive air temperature later in the season. In JJA, the warming of the soil is slightly accelerated by a positive ΔP. The resulting enhanced water infiltration warms the ground through increased heat transport. Even if precipitations would decrease in JJA, it will not be sufficient to prevent the warming of the soil because of a strong air temperature warming projected in the climate models (+1.9 to 6.4 K). In addition, drier soils tend to increase the warming effect due to reduced evaporation and enhanced local land–atmosphere coupling (Seneviratne *et al* 2010).

3.3. Sensitivity to extremes

Even though the long-term evolution of permafrost will be governed by long-term climatic trends, observations have shown that isolated (short-term) extreme events can have a persistent impact on permafrost conditions lasting for several years. Using electrical resistivity data Hilbich *et al* (2008) showed that reduced electrical resistivity values at the Schilthorn site, corresponding to reduced ice contents, persisted for three years after the anomalous hot summer 2003 even though borehole temperatures returned to normal already in 2004. Permafrost degradation may be accelerated stepwise due to isolated extreme anomalies which could be more effective than a small but continuous increase of temperature (Zenklusen Mutter and Phillips 2012). Extreme climatic seasons are expected to happen more frequently in the future (e.g. hot and dry summers, hot and wet winters) (Beniston *et al* 2007, Orlowsky and Seneviratne 2012). It is therefore meaningful to analyse extreme seasonal anomalies to understand how sensitive soil model reacts to such extremes. Figure 7 shows the changes in ALT and the analyses of the duration of these changes to assess the resilience of such extreme events.

In general, changes in ALT and persistence positively correlate: a larger increase of ALT lasts longer. DJF hot/dry,

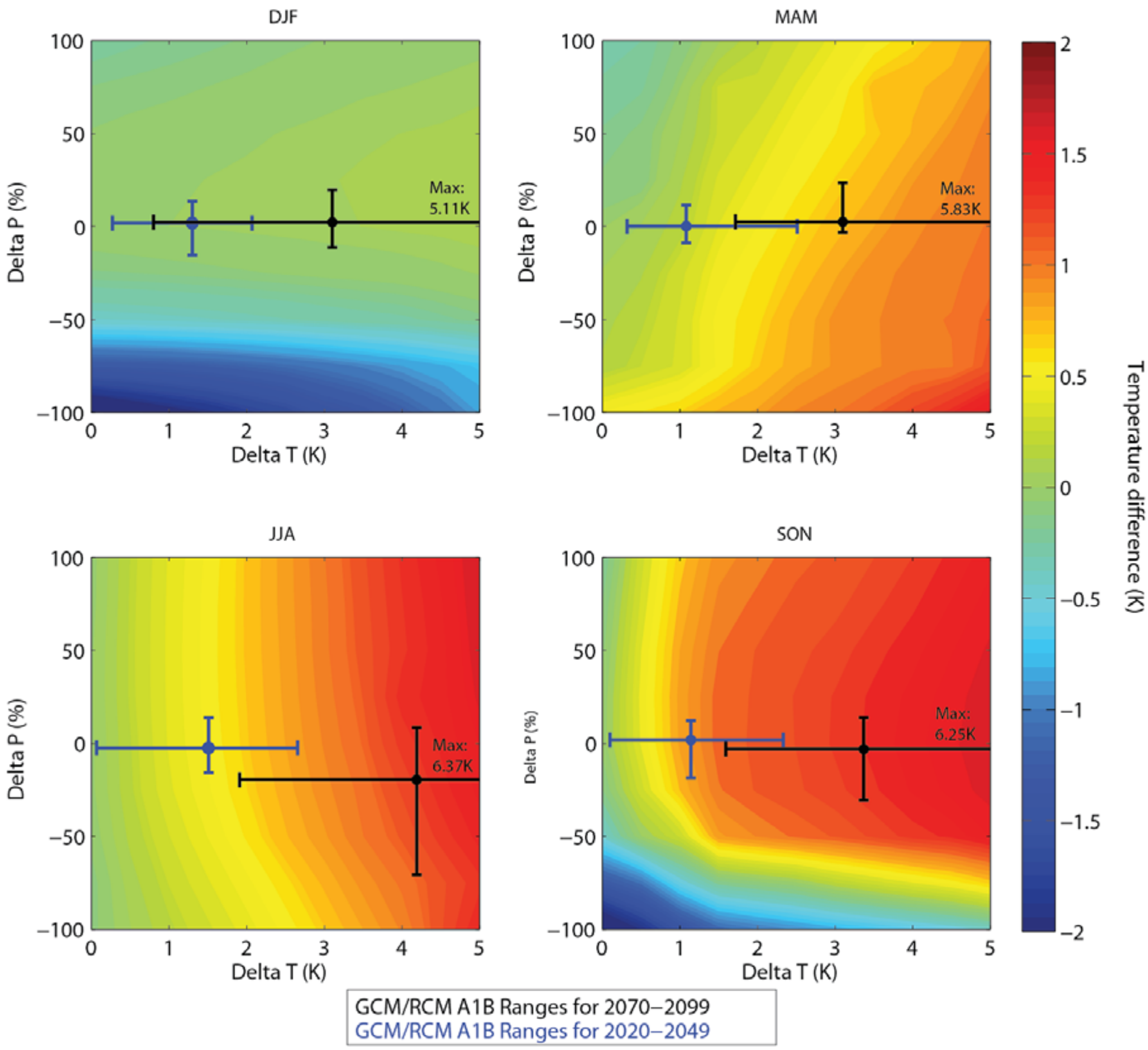

Figure 5. Difference in mean annual soil temperature at 5 m depth at the end of the century between SEA (i.e. the application of a ΔT and a ΔP during a given season: DJF, MAM, JJA and SON, for every year) compared to the reference run (SEA–REF). The bars indicate the range of 10 individual GCM/RCM model chains for A1B scenario for the period 2020–2049 (blue) and for the period 2070–2099 (black).

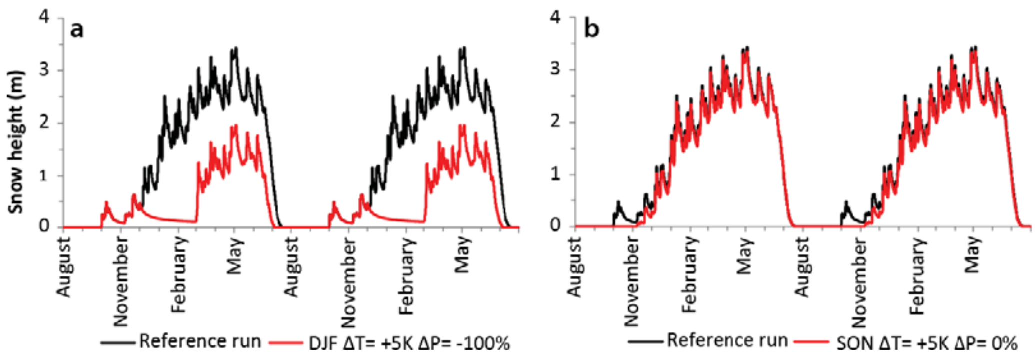

Figure 6. Snow cover evolution for a model year at the end of the century for two scenarios from the SEA experiment compared to the REF: (a) DJF with $\Delta T = +5$ K $\Delta P = -100\%$ and (b) SON with $\Delta T = +5$ K $\Delta P = 0$. In (a) the delta values mostly influence the height of the snow cover whereas the duration of the snow cover is just a few days shorter than in the REF, influencing the permafrost only little. In (b) the delta changes have almost no influence on the height of the snow cover but delay the building of the snow cover of more than 1 month. See the text for more details.

DJF cold/dry and MAM cold/dry have no or almost no effect on permafrost because the snow cover is insulating the ground from atmospheric changes. Cold JJA (wet and dry) are the extreme conditions at which permafrost is aggrading most (-1.49 and -0.99 m change in ALT) because JJA is snow-free and cooler conditions in summer months provide

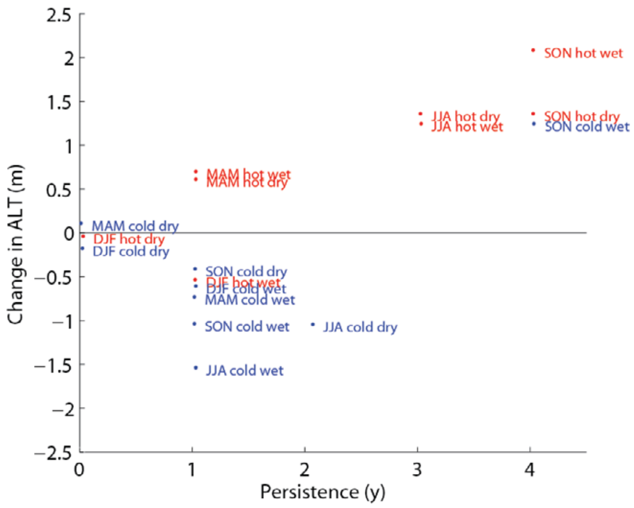

Figure 7. The change in ALT and the persistence of the anomaly for every combination of hot/cold and dry/wet extreme seasons (EXT–REF). In general, changes in ALT and persistence positively correlate. All hot extremes (red) lead to permafrost degradation (i.e. increase in ALT) except DJF hot/wet and all cold extremes (blue) lead to a decrease in ALT except SON cold/wet.

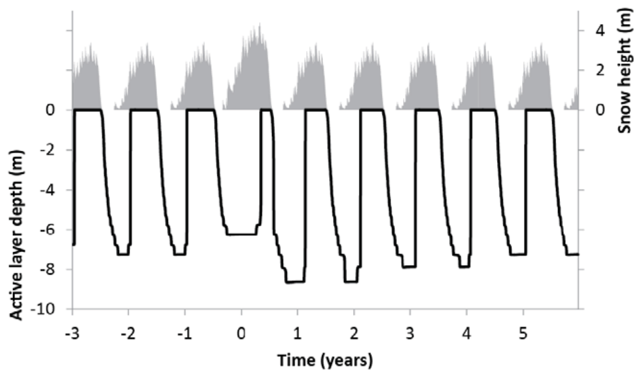

Figure 8. ALT and snow cover before and after the simulated extreme cold and wet SON event (time 0). In the year when the extreme season is applied, the ALT deceases by 0.99 m. During the consecutive years, the ALT increases by 1.37 m and the anomaly lasts for 4 years.

less energy to thaw the still frozen active layer. As for the SEA experiment, the soil model shows a higher sensitivity to extreme SON seasons. Both SON hot/dry and SON hot/wet show a large increase of ALT (+1.40 m and +2.14 m). A special case is the SON cold/wet case: in the year when the anomaly is applied, the ALT decreases by 0.99 m. The following years, the ALT increases by 1.37 m and this subsurface anomaly lasts for 4 years (figure 8). This special case will be discussed below. DJF hot/wet is the only hot extreme that leads to a decrease of ALT. This apparent paradox is due to the wet conditions: in this modified DJF case, the temperatures are still cold enough for precipitation to fall only as snow.

4. Discussion

The model did not show a large sensitivity to mean annual precipitation changes (ANN). Changes within a season due to a high/low ΔP (impacting the snow cover) would be compensated by an opposite change in another season. As the projected long-term trend is small (−7.9 to +5.0% by the end of the century), we can affirm that the effect of changes in annual precipitation is negligible if the seasonal repartition of the total precipitation amount is not changing. As seen for SEA, changes in the seasonal repartition of precipitation can impact permafrost much stronger because it would change the timing and/or the duration of the presence of an insulating snow cover. This is in accordance with earlier studies using COUP (Engelhardt *et al* 2010) and other soil and snow models (e.g. SNOWPACK, Lütschg *et al* 2008; GEOTOP, Gubler *et al* 2013).

In both SEA and EXT, changes in SON have the largest impact on permafrost conditions whereas changes in DJF have smallest. In SON, the seasonal mean air temperature (−0.7 °C) extrapolated from Mürren is close to the transition temperature between snow- and rainfall. Any positive ΔT will delay the presence of snow until November permitting positive temperatures of September and October to warm the ground. In contrast, in DJF the seasonal mean air temperature is strongly negative (−8.5 °C), so that even the projected warming is not sufficient to prevent the presence of a snow cover. As seen from both SEA and EXT, the late building of the snow cover due to warm SON temperatures is much more important for the permafrost evolution than a 1 m-difference in maximal snow height due to hot and dry DJF (cf also figure 6). This finding could be generalized and should be similar for other sites with comparatively large maximum snow thickness (> ∼2 m). For sites with less snow, a reduction of snow fall in DJF would lead to reduced insulation and therefore cooling of the ground.

As GCM/RCM projected SON temperatures show a comparatively large increase, a delayed arrival of the snow cover can be expected with an impact on ALT and possible permafrost degradation (see also Scherler *et al* 2013). According to the mean of all ENSEMBLES model chains for the A1B scenario, the SON air temperature would rise up to 3.4 K by the end of the century. Under these conditions, the ALT would increase up to values of 12 m. From this ALT, which appears as an active layer tipping point for this site, the summer thaw layer would not be able to entirely refreeze during winter. After the ALT reaches 12 m, the thaw horizon would deepen successively, including layers which may not have been thawed for several centuries. This process would even be enhanced and accelerated with an additional warming expected in JJA and MAM and/or in combination with extreme hot events. At Schilthorn and at several other places in the European Alps, the permafrost is several hundred metres deep and projected climate change would not lead to a complete permafrost disappearance at the end of the century (Switzerland; Scherler *et al* 2013; Norway; Hipp *et al* 2012). However, the increased ALT may lead to increased rock falls and debris flows impacting the safety of constructions and infrastructures. This finding is consistent with those from Nötzli *et al* (2007). Alpine sites that are ice-rich are expected to degrade slower because of the supplementary energy needed to melt the ice (Scherler *et al* 2013). In arctic permafrost environments, the climate is changing faster and harsher than in the Alps (Arctic Climate Impact Assessment 2004) but the permafrost degradation is smoothed thank to the insulating properties of the peat frequently occurring in arctic soils (Wisser *et al* 2011). In the Arctic, the issue is not the slope stability, but the release of greenhouse gases stored in permafrost (e.g. McGuire *et al* 2009).

The ANN and SEA experiments confirm most findings and hypotheses from previous model and (short-term) observational studies (PERMOS 2010), but now in a rigorous model approach covering a large parameter space of temperature and precipitation changes. In contrast, the EXT experiment show that non-linearities in the interaction between air temperature, precipitation and subsurface ice/water content may complicate the analysis of permafrost responses to extreme seasonal anomalies. In the case of cold/wet SON, the anomaly induced a thinner active layer in the year of the event but a thicker active layer in the consecutive years (figure 8). September is snow-free in REF with positive mean air temperatures (3.1 °C). It seems thus to be a critical month for the thawing of the active layer. With a cold September, less energy is injected into the ground and the thaw season is one month shorter, leading to the simulated decrease in ALT. In October and November, the snow cover is building up. A cold and wet fall will increase the snow cover height with maximal snow height of 4.36 m in EXT versus 3.28 m in REF. This increased snow cover does not influence much the duration of the snow cover as the main melting of the snow cover occurs usually during the first substantial warm period in spring/early summer. But a higher snow cover provides an additional source of water during the melt season. As in the case of wet JJA in SEA, this water will infiltrate into the ground transporting energy that leads to an increase of ALT in the following year. These kind of non-linearities are supposed to play a large role at all permafrost sites with high maximum snow cover thickness but small ground ice content and/or warm permafrost temperatures which allows the melt water from the snow cover to infiltrate.

The seasonal extremes have been applied to a reference run assuming a constant climate until the end of the century. More realistically, the extremes have to be combined with projected long-term warming trend, which increases the sensitivity to changes in SON compared to REF for the same reasons discussed above (absence/presence of snow), but also to hot DJF anomalies because the combination of the warming trend and the hot extremes would bring the air temperature closer to the transition temperature between snow- and rainfall.

According to CH2011 (2011), an initiative developed by the collaboration of several Swiss climate-related institutions, the occurrence of dry and warm summers will significantly increase in frequency, duration and intensity in Switzerland. Winter cold waves are expected to decrease in frequency and duration but winter intense rainfall events may increase in frequency (Zolina *et al* 2009). These winter rainfall events can degrade permafrost strongly if they are repeated for 3–5 consecutive years (Westermann *et al* 2011). According to Rajczak *et al* (2013), the frequency of extreme precipitation events are expected to increase during spring and fall. However, none of the extreme seasonal scenarios tested in this study show a persistence of more than four years. The permafrost should therefore not be strongly affected by extreme events even if they are going to happen with a higher frequency in the future, as long as the thermal regime of the permafrost site is not close to the ALT tipping point described above. The long-term trend appears to stronger impact the permafrost.

5. Conclusion

The sensitivity of a dedicated simulating soil model (COUP) has been tested regarding future annual and seasonal changes as well as seasonal extremes of air temperature and precipitation at a typical low-ice content mountain permafrost site Schilthorn, Swiss Alps. The following conclusions can be drawn from this study:

- Changes in air temperature show an important impact in all experiments, i.e. annually, seasonally (except for DJF) and regarding extreme events. Hereby, changes in SON show the largest impacts.
- Changes in annual mean precipitation do not show an important impact, while changes in the seasonal repartition of precipitation are more critical due to the insulating properties of snow.
- The critical role of the snow cover for the long-term permafrost evolution is largest for SON where its presence or absence is crucial for the impact on permafrost temperature.
- In general, climatic changes during DJF have smallest effects on the modelled permafrost conditions, at least for permafrost sites with large maximum snow heights.
- The duration of the snow cover and the timing of its arrival are the largest influencing factors whereas the snow height plays a secondary role. Again, this is valid for sites with a comparatively high maximum snow cover thickness. The infiltration of water from rain during the snow-free period or from the snow during the melt season also appears to be an important factor and influences the thermal and hydraulic conductivity.

With the modelling setup of our study we can expect to have assessed a plausible range of potential climatic changes impacts on mountain permafrost for low-ice content high mountain permafrost sites (as on Schilthorn) which can be used as basis for discussion of slope stability issues. Subsurface parameters like ice content or surface parameter like albedo vary from site to site, which in turn shift the sensitivity ranges at the local scale. Hereby, low-ice content (and warm) permafrost sites can be considered as highly sensitive, compared to sites with higher ice contents or colder permafrost temperatures. The overall findings of this study, however, are not affected. A thorough analysis of the site-specific factors for many permafrost sites in the

Swiss Alps is currently being conducted within the Sinergia project TEMPS (The Evaluation of Mountain Permafrost in Switzerland) funded by Swiss National Science Foundation (SNF).

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Appendix

A.1. Model description

The COUP model is a one-dimensional numerical model that couples soil water and heat transfer using the general heat flow equation (Jansson and Karlberg 2004, Jansson 2012):

$$\frac{\delta(CT)}{\delta t} - L_{f\rho}\frac{\delta\Theta_i}{\delta t} = \frac{\delta}{\delta z}\left(k\frac{\delta T}{\mathrm{d}z}\right) - C_w T\frac{\delta q_w}{\delta z} - L_v\frac{\delta q_v}{\delta z} \qquad (1)$$

where C (J K^{-1}) is the heat capacity, T (K) is the soil temperature, L_f (J kg^{-1}) is the latent heat of freezing, ρ (kg m^{-3}) is the density, Θ_i is the volumetric ice content, k (W m^{-1} K^{-1}) is the thermal conductivity and q_w and q_v (kg m^{-2} s^{-1}) are the water and vapour fluxes, respectively.

The model was tested and calibrated for Schilthorn for a 10-year observation period by Scherler *et al* (2013) using on-site meteorological driving variables. In the configuration used for the sensitivity experiments, the model is driven only by air temperature and precipitation in the form of observed daily time series. As radiation data is not available before the year 2000, global radiation, R_{is}, is calculated by the model deduced from the potential global radiation, R_{pris} and the turbidity:

$$R_{is} = R_{pris} \cdot f(\text{turbidity}) \qquad (2)$$

$$R_{pris} = E_0 \cdot 60 \cdot \Delta T_{Max} \cdot a_2 \qquad (3)$$

where E_0 represents the solar constant (1360 W m^{-2}), 60 is the number of second per minute and a_2 is a parameter obtained combining latitude, declination, day-length and air temperature amplitude. The turbidity is a function of the relative duration of sunshine and is calculated by the Ångströms equation (Kuo-nan 2002). Net radiation is given by:

$$R_{net} = 86\,400\sigma(\epsilon_s(T_s + 273.15)^4 - \epsilon_a(T_a + 273.15)^4) \qquad (4)$$

where T_a is the air temperature, T_s is the surface temperature of the soil or the snow, ε_s is the surface emissivity and assumed to be equal to 1 and ε_a is the emissivity of the atmosphere calculated with Konzelmann's equation (Konzelmann *et al* 1994). Relative humidity and wind speed are given as constant parameters.

Input data have a daily time resolution and the internal time step is 6 h to generate diurnal variations. The lower boundary condition is defined by a minimal geothermal heat flux as the borehole is almost isothermal between 20 and 100 m (PERMOS 2010). The snow cover is generated by the model, partitioning precipitation into rain or into snow depending on prescribed air temperature thresholds. Snow melt is controlled by air temperature, global radiation (depending on the snow age, i.e. albedo) and heat flux from the ground.

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