Dear Editor, dear referees,

We would like to thank the two anonymous referees for their careful and meticolous reviews, pointing out parts that lacked clarity, weaknesses, redundancies, but at the same time helpfully suggesting ways to overcome the observed deficiencies. The reviews motivated us to substantially revise the manuscript. All reviewer comments were carefully considered and most of them taken into account. Our reply to the individual points in the two reviews (both those accounted for and those where we did not fully agree) are detailed out below. We believe that the manuscript has - thanks to the reviews - undergone a substantial improvement and hope that it is now fitted to be published in The cryosphere.

Sincerely yours,

for the author team: Regula Frauenfelder

Reply to comments made by Anonymous Referee #1 (doi:10.5194/tc-2016-223-RC1)

We thank Anonymous Referee #1 for its critical review and suggestions for improvement. Referee comments indicated as "**RC**:" and the author's reply as "***AR***:".

RC: Rockslides in deglaciated terrain can be caused and triggered by a lot of different processes such as debuttressing, sheeting joint development, seismicity, hydrostatic pressures and also permafrost (McColl, 2012). The authors use a single hypothesis approach to conclude that the rockslide was triggered by permafrost degradation. The effect that ice is present in the landslide scarp and warming occurred indicate that permafrost could be involved. Other potential scenarios such as hydrostatic pressure increase due to snowmelt or static fatigue are also possible and should be discussed.
AR: *The authors acknowledge that an extensive discussion of other potential triggering scenarios is missing. We consider other possible triggering processes as unlikely: debuttressing is not likely since no pre-failure action was identified or noticed; sheeting joint development typically follows the topography, the failure surface was, however, a wedge into the slope face; seismicity: our study area is in a low seismic hazard zone, and there were no corresponding seismic events recorded. Hydrostatic pressure increase due to snowmelt or static fatigue could have played a role. This aspect is now brought into the discussion.*

RC: Furthermore, the effect of permafrost on rock stability or permafrost degradation on rock instability is insufficiently understood. According to the authors, the shear plane is located in depths up to 40 m, therefore, the scenario that fracture ice failed as described by Davies et al. (2001) as a trigger of the rockslide is unlikely. Krautblatter et al. (2013) demonstrated that fracture ice cannot influence stability in depths more than 20 m due to the overburden pressure of the rock mass. They provide two other processes that control rock stability: (1) intact rock bridges and (2) rock-rock-contacts. Permafrost increases the uniaxial and tensile strength of these rock bridges and rock-rock-contacts.
A warming of permafrost would decrease both tensile and uniaxial strength and could trigger rockslope failure (Krautblatter et al., 2013). For a detailed discussion on the interaction between thermal and mechanical processes see Draebing et al. (2014). These authors also provide information on the seasonal

timing of rockslope failures and long-term development of rock stability which you should incorporate in your discussion.

AR: *We agree with this comment and have, therefore, added the main findings (relevant for this paper) of the studies by Krautblatter et al. (2013) and Draebing et al. (2014) into the discussion.*

RC: The mechanical aspect of this paper is poorly addressed, process understanding does not reflect current knowledge, methods are not well introduced and results are not critically discussed.

AR: *The paper has been updated to include reference to state-of-practice papers that correspond to the methods conducted in this research. The discontinuity orientation extraction and mapping was completed using standard techniques (according references are now given) for data of this type. It is not suitable for automated feature mapping. The criticality of the discontinuity orientations is not the subject of this paper, nor is it essential in the authors' opinion. Numerous authors have widely published on this topic, we are simply using the extracted information.*

RC: Landslide terms should be used for clarification.

AR: *The authors have updated the text describing the landslide as a rock avalanche based on the terminology proposed by Hungr et al. (2014).*

RC: The authors derived three post-failure TLS scans but they did not discuss how they derived the estimated volume of 500,000 m3. One TLS scan should be suited for the estimation; the follow-up TLS can provide information on subsequent rockfalls which is not the objective of this paper. In addition, the processing should be described in more detail.

AR: *This is now addressed in the text and references have been added. We agree that the monitoring of subsequent rockfall was not the objective of this paper. However, it was an objective of the monitoring program and therefore we suggest to leave the according information (i.e. that the site was scanned in three year) in the paper.*

RC: The authors describe problems with data holes during the processing of the DEM. Can you provide an error estimation and information about resolution and accuracy of your DEM? This is important to estimate the quality of your fracture mapping which is insufficiently described. Fracture determination is an complicated task according to Abellan et al. (2014) and should be described in full detail.

AR: *The holes in the DEM do not have any impact on the accuracy of the structural measurements, these aspects are not correlated. The accuracy of the measurements from LiDAR data is well documented, and references have been added. There is no need for this to be explained in the text. Abellan et al. (2014) – which Lato is a co-author – is a review paper written for general audiences. The details of converting vector orientations to dip and dip direction requires careful calculation, but in our view, our paper is not the place for it to be discussed as the methodology has been widely published over the past ten years.*

RC: The authors conclude that the bedding surface is steeper that the friction angle which is an important information. Unfortunately, the presentation of this important information is insufficient in the figures and result section.

AR: *This is briefly, but clearly described in Section 3.1 and on Figure 5. The focus of this paper is not the extraction of discontinuity measurements. The authors feel that the descriptions and figures provided are sufficient given the context and focus of this paper. There are numerous references supplied that outline the methodologies of extracting this kind of information from LiDAR data, is the reader is inclined to read more they can refer to the added references.*

RC: Furthermore, the authors should discuss critically the influence of permafrost on increasing the internal angle of friction as described by Krautblatter et al. (2013), thus, this is the link to the thermal regime you monitored and modelled.
***AR:** The main findings (relevant for this paper) of the study by Krautblatter et al. (2013) are now added.*

RC: To estimate the thermal influence, the authors monitored near-surface rock and soil temperatures and used a 2D model to model rock temperature. For this purpose, they used three different temperature loggers. Resolution and accuracy of the logger types should be introduced and the influence on temperature records quantified.
***AR:** We introduced information about resolution and accuracy for the loggers in the text: "The absolute accuracy for all three logger types is ± 0.1 °C. The resolution of M-Log5W, UTL-3 and UTL-1 is 0.01, 0.02 and 0.27 °C respectively. The M-Log5W loggers and the UTL-3 loggers were programmed to measure every 30 minutes, while the UTL-1 loggers (having smaller memory capacities than the other two employed logger types) measured every two hours. Based on experiences from long-term permafrost monitoring programs in Norway using such loggers (e.g., Isaksen et al. 2011; Gisnås et al. 2016) we claim that the bias on the accuracy of the temperature measurements introduced with the given setups is negligible."*

RC: The location of loggers is introduced, however, further information on altitude, aspect, distance to rock ledges and slope angle is required to estimate the influence of topography and snow cover. Up to now, there is temperature information on rockslopes in Norway measured by 3 data loggers by Hipp et al. (2014). The information of this manuscript can provide new interesting data on rock temperatures in this environment. Hipp et al. (2014) showed that near-surface rock temperature distribution is different to the European Alps. The authors should discuss in more detail this difference and potential causes. Please include the influence of snow cover on the thermal and mechanical regime in steep rockwalls as recently addressed by Haberkorn et al. (2015a; 2015b) or Draebing et al. (2016).
***AR:** We included a new table (Table 2) with elevation, aspect and snow conditions for the logger sites. A comparison of aspect dependency at our site as compared to sites in the European Alps has been added to our discussion. We also added some text related to the influence of snow cover on the thermal and mechanical regime in rockwalls, referring to important recent literature in the field (e.g., Haberkorn et al., 2015a, b; Magnin et al., 2015).*

RC: In the next step, the authors used this information to model ground temperatures. They used the CryoGrid2 model which provides a resolution of 1 km2 (Westermann et al., 2013). Consequently, the Polvartinden rockslide is presented by two pixels. Subsurface material is derived from a geological map and is till and colluvium according to the authors. Therefore, CryoGrid 2 is not suited to model rockwall temperatures, thus, resolution is too coarse and subsurface material different than bedrock.
***AR:** We agree that CryoGrid2 may not be the best choice to model rock wall temperatures. However, as written in the Methods section the larger area around the failure zone, outside the steepest slope, is characterized by a combination of small vertical rock outcrops and undulating slopes with an established soil and snow cover. Consequently, temperature loggers were installed in both types of terrain, i.e., in both vertical rock outcrops and within soil material of the more gentle slopes. To be representative for the latter type of terrain it was desirable to look further into the temperature development in the ground in areas close to the release area with a more established snow and soil cover. To this aim, the CryoGrid2 model (cf. Westermann et al., 2013) was used. In our opinion the model is very well suited to study long-*

term changes in ground temperatures representative for the same elevation as the failure zone for sites with more gentle slopes and a developed soil cover and where snow accumulates.

We have tried to make these points clearer in the current version of the manuscript, for example by explaining them better in the method chapter and by highlighting them in the results section and in the discussion.

RC: Models such as CryoGrid2D used by Myhra et al. (2016) or modeling approaches by Noetzli et al. (2007) are better suited. The latter approach is also used in this paper but input data, model parameters such as chosen thermal conductivity, ice content and porosity, or data processing is not introduced as well as resulting effects on modelling results are not quantified or discussed.

***AR:** As written in section 2.3.2 the applied model by Noetzli et al. (2007) was used to get insights into the present subsurface temperature field of Polvartinden and is valid for areas that are assumed not to be influenced by a snow cover, i.e., the steep rock-faces of Polvartinden. In our paper we present subsurface properties in Table 1 that were obtained from representative sites nearby (cf. Lilleøren et al., 2012). The model applied has been widely presented and discussed in earlier literature, which we refer to in the text. We do not see the need to go into detail about data processing, etc. of this model in this paper.*

RC: Comments and suggestions in the supplement.
***AR:** We were impressed by the level of detail and meticulousness of the comments and suggestions in the supplement. Some of them are answered by our replies to the review comments (above and below), all others were carefully considered and amendments made were we agreed with the comments/suggestions.*

Reply to comments made by Anonymous Referee #2 (doi:10.5194/tc-2016-223-RC2)

We thank Anonymous Referee #2 for its critical review and suggestions for improvement. Referee comments indicated as "**RC**:", author reply as "***AR***:".

RC: 1. The manuscript is lengthy and reads like a report rather than a scientific paper. Methods and results are widely inter-mixed.
***AR:** We do not agree with this comment and reviewer #1 did not comment something in this direction. We have, therefore, not altered the structure of the manuscript. If a change of structure is suggested by the editor, we will of course do that.*

RC: 2. Observations: There has been employed 14 loggers around the mountain, in the end 9 of them were used. However, for the reader it is difficult to see where the loggers are placed, in relation to possible snow cover and topographic aspect. You should give a table of logger description, inkl. elevation, aspect etc. Fig. 3b is not useful within this respect; please give a map rather than an image.
***AR:** We included a new table (Table 2) with elevation, aspect and snow conditions for the logger sites. Figure 3b has been updated, now featuring a map as base layer instead of an aerial photo.*

RC: 3. Setting and geomechnical mapping: A setting chapter is lacking, the info is part of the introduction. For readers not particular well-known in the area, I would suggest to provide general geophysiographic setting of the area, including general climate parameters and the regional distribution of permafrost. I do not see the point of the kinematic analysis here. You could simply describe that in the setting chapter as background for the site.

***AR:** See above. We have not changed the structure of the paper at this stage.*
The kinematic analysis was an objective of the overall monitoring program and is important in order to increase our understanding of the possible failure mechanisms that led to the Signaldalen rock avalanche. We believe that the now improved discussion of the mechanical and thermal controls on rock stability (triggered by the comments by reviewer #1) justifies the inclusion of the kinematic results.

RC: 4. CG2 model: A major part of your conclusions are based on the results of the CG2 modelling. First of all, the description of the model, its principles etc must be given even if details are explained in another publication. The same is valid for the reasoning of the parameter choice.

***AR**: We agree that the description of the model, its principles and the reasoning of the parameter choice must be given, so we have added according information to section 2.3.1.*

RC: Last not least, snow is of course here a problem. What snow cover have you assumed? Are there any observations? I understand that the forcing is based on gridded data? What is the relation between the gridded data and e.g. a long met series from one of the met stations nearby?

***AR**: We added the following text in 2.3.1. to make this clearer: "The model is forced by operational gridded (1 x 1 km) air temperature (Mohr, 2009; Tveito et al., 2000) and snow-depth (Engeset et al., 2004; Saloranta, 2012). Snow cover data is based on the seNorge snow model that uses gridded observations of daily temperature and precipitation as its input forcing, and simulates, among others, snow water equivalent (SWE), snow depth (SD), and the snow bulk density (ρ) (Saloranta, 2012). The gridded air temperature for our site is mainly driven by the nearby Skibotn and Rihpojavri weather stations which were validated against our local measurements (see section 3.2). The gridded precipitation for our site is mainly driven by observed precipitation at Skibotn weather station. Since we have no observations of snow cover and owing to the large spatiotemporal variability of snow conditions in our alpine study area, the snow simulations from the seNorge model provide probably the best estimate of the spatial average 1 x 1 km snow conditions for our site (cf. Saloranta, 2012)."*

In addition, according to Westermann et al. (2013) the thermal conductivity of the snow is the largest source of uncertainty in CG2, thus a low (LC) and a high conductivity (HC) scenario run of CG2 are used as a confining range for the true conditions.

RC: A major problem is of course the lack of validation of the model. I understand that there is no borehole at the study site for ground temperature validation. But you could check the modelled ground surface temperatures against some of your loggers? As the snow cover and subsurface parameters are very uncertain and not validated, and the model is certainly based on heat conduction, the model mirrors of course the air temperature forcing. So, it is not an independent support of the findings from the long-term air temperature analysis, which should be discussed somehow.

***AR**: We now checked the modelled ground surface temperatures (for 0.1m depth) against our soil temperature loggers for the first year of measurements when CG2 data were available and added a new figure (Fig. 8) visualizing the results. The following text (going with the new Fig. 8) has been added:*

"A one-year comparison of observed daily SST at sites with an established soil and snow cover and the modelled SST from the CG2 model is shown in Figure 8. As seen in Figure 8a the SST observed in the valley floor is in good agreement (R^2 = 0.86 to 0.88) with the CG2 model results for a grid cell with approximately the same elevation as the valley floor. The Signaldalen valley is dominated by mountain birch forest and is characterized by an open forest layer with heather and lichen species dominating the forest floor layer. This type of forest causes snow to accumulate and insulates the ground against strong cooling (cf. Isaksen et al., 2008; Farbrot et al., 2013), an effect which can be seen in our CG2 model results. In the mountains (Fig. 8b) there is a somewhat weaker correlation (R^2 = 0.69 to 0.82) between observed and modelled data. The MASST are about 1 to 1.5°C lower than in the CG2 model for the corresponding grid cell with similar elevation. The warm bias of the simulations during winter is mainly explained by the different snow conditions assumed in the model: while our observational sites in the mountains are affected by considerably strong snow cover variations induced by wind drift, the snow model used as input in the CG2-model generally overestimates snow depths in high mountain areas (cf. Saloranta, 2012). This is also in line with results of an equilibrium permafrost model used by Gisnås et al. (2013) who found a better agreement between the CG2-model and their validation data when the snow depth in the snow model data was reduced by 30 % for areas above the tree line. During summer, on the other hand, our observed SST in the mountains are well reproduced by the CG2 simulations."

RC: 5. 3D model: With the use of the 3D model I had some problems. The authors acknowledge that there are large uncertainties about the aspect-dependency of ground surface temperatures. The basics of the aspect dependency are related to the measurement array, which partly was influenced by snow cover etc. The author found an aspect dependency of c. 1◦C, which might be something between 150 and 200 m in elevation given certain lapse rates. They show fig. 11a, with a polynomial fit to 8 points, which is not statistically sound (e.g. what are the p-values for this fit).

***AR:** We agree that a polynomial fit to eight points may not be statistically sound. However, our curve fitting must be seen as an attempt to capture the main patterns in our data, rather than to deep-dive into questions about the probability for a given statistical model. Our results should be seen as tentative estimates (as already emphasized in our discussion). The authors feel that the figure, descriptions and discussions provided are reasonable and in line with previous studies using similar approaches, also suffering of a limited number of data points (e.g. Hipp et al. (2014) and Gruber et al. (2004) used 5 and 14 rock-wall temperature data loggers in their studies, respectively).*

RC: These data were used to force the 3D model, and the authors choose to show a slice from a north-south oriented section. There are several problems here: (1) The isolines become more or less horizontal, so the plot does not give much new information in relation to the analysis of the data loggers or the 1D model. (2) The uncertainty is very high here, which the authors also mention, so I wonder if there is any justification of this analysis, beside showing a nice figure, and that such analysis are possible? (3) What was the initialization used? I do not understand page 7, line 15 etc. Does the model now show more or less the same as the 1D model, and is its use scientifically justified here? Especially in the light that also snow cover was neglected. Ok maybe for vertical rock walls, but is this a good approximation for your study site?

AR*: Permafrost is a subsurface phenomenon mainly driven by the state and changes of the surface temperatures. We do not claim to provide a real model of the subsurface temperature field (which would require many more data points as well as a statistically sound analyses of the factors influencing the spatial distribution), but rather to schematically illustrate the general pattern of the subsurface temperatures in a mountain such as the Polarvatiden and to what extent it is still influenced by smaller differences between in MAGST at the different mountain sides. Our main point with using this model was a rough visualization of the permafrost body and the general pattern of the subsurface temperature field in the mountain. This cannot be achieved with a 1D-profile. The model illustrates the extent of the permafrost body within the mountain and in the larger area of the starting zone.*

The text for the initialization and the transient model run has been adapted. The model was initialized for the temperature conditions at 1900 (it is assumed that due to the smaller size of the mountain no earlier surface temperature variations are significant for the subsurface thermal regime, cf. also Noetzli and Gruber, 2009). Then a transient simulation was run with the upper boundary condition (that is the distributed MAGST) linearly increasing by 0.55 °C until the current situation.

RC: 6. Conclusions: The conclusions are a bit thin. I agree that you have indications for a thermal trigger. But only state what you can justify with your observations and/or models. I do not understand what the passage of activity in the slide contributes here? You may omit that.
AR: *We slightly expanded the conclusions, however, we believe that conclusions should be kept short and concise.*
We do not understand the comment about the "passage of activity".

RC: Introduction: see comments above. I suggest to make a setting chapter in addition
AR: *See above. We have so far not altered the structure of the paper.*

RC: p.4, First three paragraphs: Delete, not necessary with a summary first, and details afterwards. Include in the detailed method description.
AR: *See above. We have so far not altered the structure of the paper*

RC: p. 5, l. 4-10: This is typical setting, move,
AR: *See above. We have so far not altered the structure of the paper*

RC: p. 5: l. 11-20: The laser-scan is not necessary here. Figure 4 is not understandable at all, at least not for me, and you can simply write that repeated laser scan analyses did not reveal large movements after the event. In the suggested setting chapter.
AR: *We agree that the figure was poor and have made an improved, updated version.*

RC: p. 5, l. 25 ff: Give resolution/precision of loggers.
AR: *This has now been added: "The absolute accuracy for all three logger types is ± 0.1 °C. The resolution of M-Log5W, UTL-3 and UTL-1 is 0.01, 0.02 and 0.27 °C respectively. The M-Log5W loggers and the UTL-3*

loggers were programmed to measure every 30 minutes, while the UTL-1 loggers (having smaller memory capacities than the other two employed logger types) measured every two hours"

RC: p. 6, l. 11-15: Delete whole paragraph, nothing new.
AR: *Agreed. The paragraph has been deleted.*

RC: p. 6/7: About the models, see comments above. Especially p 7, l16 is problematic. How many places in your study area are really not covered by snow?
AR: *See our reply above and new information about snow cover included in the new Table 2.*

RC: p. 8, l. 1-16: I am not an expert her, but what is the point of this in relation to your hypothesis, e.g. triggering may have been related to permafrost thaw?
AR: *Triggered by the comments by reviewer #1 we have considerably increased the discussion of the mechanical and thermal controls on rock stability. We hope that our hypothesis that triggering might have been related to permafrost warming is now more clearly justified.*

RC: p. 8, l 25: What is the rationale behind to use always a 1a-mean?
AR: *For several of the analysis presented annual mean values were calculated to identify variations in mean annual ground surface temperatures and mean air temperature. This ensures easier comparison between the monitoring sites and makes it easier to identify local maxima and minima as well as trends (cf. Isaksen et al., 2011). This argumentation is now included in the revised manuscript.*

RC: p. 8, l 28: Reference to an EGU abstract is not acceptable for international journals as the data cannot be reproduced. So either show the data, or give them in an appendix or make a figure to include here.
AR: *We omitted the references to the EGU abstract.*

RC: p. 9, 1. Paragr. : This is important for your reasoning, so you may show that somehow.
AR: *See above, we included a new figure (Fig. 8) and commented in detail on the findings visualized in Fig. 8.*

RC: p. 10: Why did you choose 10 m depth for the CG2 model?
AR: *Westermann et al. (2013) found 10 m to be an appropriate depth to study long-term changes in ground temperatures but avoiding dominance of near-surface high-frequency temperature variations.*

Ground thermal and geomechanical conditions in a permafrost-affected high-latitude ~~rockslide~~rock avalanche site (Polvartinden, ~~Northern~~northern Norway)

Regula Frauenfelder[1], Ketil Isaksen[2], Jeannette Noetzli[3], Matthew J. Lato[4]

[1]Norwegian Geotechnical Institute NGI, Oslo, 0806 Norway
[2]The Norwegian Meteorological Institute, Oslo, 0313, Norway
[3]Formerly University of Zurich, Zurich, 8057, Switzerland, now at: WSL-Institute for Snow and Avalanche Research SLF, Davos, 7260, Switzerland
[4]Formerly NGI, now at: BGC Engineering Inc., Ottawa ON, Canada

Correspondence to: R. Frauenfelder (rf@ngi.no)

Abstract. ~~In~~On June 26, 2008, a ~~rockslide~~rock avalanche detached in the northeast facing slope of Polvartinden, a high-alpine mountain in Signaldalen, ~~Northern~~northern Norway. Here, we report on the observed and modelled past and present near-surface temperature regime close to the failure zone, as well as on a subsequent simulation of the subsurface temperature regime, and on initial geomechanical mapping based on laser scanning. The volume of the ~~rockslide~~rock avalanche was estimated to be approximately 500'000 m^3. The depth to the actual failure surface was found to range from 40 m at the back of the failure zone to 0 m at its toe. Visible in-situ ice was observed in the failure zone just after the ~~rockslide.~~rock avalanche. Between September 2009 and August 2013 ground surface temperatures were measured with miniature temperature data loggers at fourteen different localities close to the original failure zone along the northern ridge of Polvartinden, and in the valley floor. The results from these measurements and from a basic three-dimensional heat conduction model suggest that the lower altitudinal limit of permafrost at present is at 600-650 m a.s.l, which corresponds to the upper limit of the failure zone. A coupling of our in-situ data with regional climate data since 1958 suggests a general gradual warming and that a period with highest mean near surface temperatures on record ended four months before the Signaldalen ~~rockslide~~rock avalanche detached. A comparison with a transient permafrost model run at 10 m depth, representative for areas where snow accumulates, strengthen this findings, which are also in congruence with measurements in nearby permafrost boreholes. It is likely that permafrost in and near the failure zone is presently subject to degradation. This degradation, in combination with the extreme warm year antecedent to the rock failure, is seen to have played an important role in the detaching of the Signaldalen ~~rockslide~~rock avalanche.

1 Introduction

In the morning of June 26th, 2008 a ~~rockslide~~rock avalanche (cf. Hungr et al., 2014 for nomenclature) detached from the northeast facing slope of Polvartinden, a 1275 m high mountain in Signaldalen, Troms county, ~~Northern~~northern Norway (located at 69°10'18"N/19°57'47"E; Fig. 1). The ~~rockslide~~rock avalanche endangered two farms and several recreation cabins, and it ~~permanently~~ destroyed a considerable amount of livestock pastures. Just after the ~~slide~~ event~~, experts from the Norwegian Geotechnical Institute were summoned to the slide location by request of the municipality and~~ first reconnaissance work was carried out, including failure zone assessment by means of visual inspection from a helicopter. During this visual inspection, a few hours after the event, in-situ ice was observed in the failure zone at several meters depth (Fig. ~~2).~~2), i.e., clearly indicating permafrost. This observation, together with the absence of any typical pre-weather conditions (such as intensive rainfall or snow melt) that could have triggered the ~~slide~~rock avalanche, led to the hypothesis that warming or degrading of permafrost could have played a role in the timing and the magnitude of the event. ~~Yet, the financial means of the municipality did not allow for a temperature monitoring at that time. First a~~A year after the event, a temporary temperature monitoring network ~~could be~~was put in place ~~within the framework of an R&D cooperation between the Norwegian Geotechnical Institute~~ and ~~the Norwegian Meteorological Institute. Within this framework~~ repeated surveying of the failure zone and the adjacent slopes by means of terrestrial laser scanning was ~~carried out between 2009 and 2013~~initialized.

Scientists have become increasingly aware that atmospheric warming has a potential impact on mountain permafrost (e.g., Haeberli, et al., 2010; Harris et al~~.~~., 2009; Jin et al., 2000; Marchenko et al., 2007; Stoffel and Huggel, 2017). It seems that mountain systems are especially sensitive to a changing climate due to feedback effects in connection with snow cover, albedo and heat budgets, which amplify the alterations caused by climate change (Gobiet et al., 2014; Vavrus 2007; Wang et al., 2014). Several studies from the European Alps show that temperatures have increased twice as much as the global average since around 1900. There is also ~~exists~~ increasing evidence ~~– many from studies from the European Alps –~~ that there exists a link between ~~rock fall~~rockfall magnitude and frequency, and timing and depth of permafrost degradation, the latter ranging from seasonal increase of active-layer depths to long-term, deep-seated warming of the permafrost body as a response to atmospheric temperature rise (e.g., Gruber and Haeberli, 2007~~).~~; Ravanel and Deline, 2010). Fischer et al. (2012) collected published material on large ~~rockslide~~rock avalanche and rock avalanche events (volumes between 10^5 and 10^7 m^3), as well as on more frequent small-volume rockfall events. Their study concludes that such events occur worldwide in periglacial environments, and that many of the reviewed studies suggest that the reported events may be related to ongoing and/or past changes in permafrost and glacierization. Corresponding events have, for example, been documented for the European Alps (e.g., Barla et al., 2000; Crosta et al., 2004; Gruber et al., 2004; Cola, 2005; Deline et al~~.~~., 2011; Oppikofer et al., 2008; Ravanel and Deline, 2010; Phillips et al., 2016), Canada and Alaska (e.g~~.~~., Evans and Clague, 1994; Geertsema et al., 2006), the Caucasus (e.g~~.~~., Haeberli et al., 2004) and for New Zealand (e.g. Allen et al., 2011).

In mainland Norway (i.e., excluding Svalbard), permafrost occurs mainly in the central mountain chains and in higher latitude areas, such as the counties of Troms and Finnmark (Gisnås et al., 2016). Early studies by, e.g., King (1986), Ødegård et al. (1996), and Etzelmüller et al. (2003) have shown that permafrost is discontinuous in the higher mountains of central southern and eastern Norway. More recent studies (e.g., Farbrot et al., 2011) yield that permafrost temperatures in general are between -3 and 0~~-~~°C in the higher mountains of southern Norway and warming and degrading permafrost have been documented to occur at sites with cold permafrost, marginal permafrost and deep seasonal frost (Isaksen et al., 2011). It has, therefore, to be expected that the permafrost is particularly vulnerable to climate change (Christiansen et al., 2010; Farbrot et al., 2011). Large-scale rock-slope failures pose a significant geohazard in the fjords and valleys of western and northern Norway (Blikra et al., 2006). A study by Blikra and Christiansen (2014) on the interactions between permafrost and ~~rockslide~~rock avalanche deformation at the Nordnesfjellet site in northern Norway (location indicated in Fig. 1) shows that systematic seasonal changes – thought to be controlled by sporadic permafrost occurrence – ~~-~~considerably affect the shear strength of the sliding planes (Blikra et al., 2015).

Only few studies are available on the distribution of mountain permafrost and its thermal state in high-relief areas in northern Norway. Due to increased awareness of the potential role of permafrost degradation to landslide risk, a permafrost and climate monitoring programme was initiated in 2002, along climate and altitudinal transects in Troms and Finnmark, the two northernmost counties of mainland Norway (Isaksen et al., ~~2004; Isaksen~~2008; Farbrot et al., ~~2008~~2013). In 2003, a permafrost programme was~~ also~~ launched in the Gaissane Mountains in northernmost county Finnmark (Farbrot et al., 2008). In the mountains of Troms two 30–32 m deep boreholes were drilled in 2004 at altitudes of 786–850 m ~~a.s.l. (Isaksen~~asl (Farbrot et al., ~~2011b~~2013). These studies found permafrost to be warm, but widespread in alpine areas in northern Norway.~~ Some of the most hazardous rockslide areas in Norway are found in the fjord districts of Troms, where several unstable rock slopes exist (Blikra et al., 2006).~~ The existing monitoring network in northern Norway was more extensively instrumented and extended with new boreholes in Troms and Finnmark during the third International Polar Year 2007–2009 (cf. Christiansen et al., 2010~~). Data from these boreholes are freely available in the Norwegian Permafrost Database (NORPERM) developed during the IPY period (~~; Juliussen et al., 2010~~) and were extensively analysed by~~). Through extensive analyses of these borehole data, Farbrot et al. (2013~~). They~~) found that the combined effects of snow depth and vegetation cover are the two most critical factors for the existence of permafrost in northern Norway. ~~Farbrot et al. (2013)~~They also concluded that the depth of seasonal frost or active layer in areas underlain by exposed bedrock was more than 10 m at several sites and was amongst the deepest reported in the international literature~~. Since the end of the Little Ice Age (LIA),~~, and that the altitude of the lower limit of permafrost has probably increased around 200–300 m since the end of the Little Ice Age (LIA) (Farbrot et al., 2013). This was supported in a recent study by Myhra et al. (2015) who modelled permafrost distribution and long-term thermal changes in, among others, a steep rock wall at Revdalsfjellet which is situated close to the unstable and extensively monitored Nordnesfjellet (Blikra and Christiansen, 2014) and 25 km from our Signaldalen site. For the same mountain Myhra et al. (2015) modelled a bedrock warming at 20 and 50 m depth of about 1.0 and 0.5~~-~~°C, respectively, from the end of the LIA and ~~with~~using a present lower limit of permafrost of about 650 m ~~a.s.l.~~asl From the development of a new Nordic permafrost map there exist new modelling results, which confirm ongoing degradation of mountain permafrost in the fjord areas in Troms (Gisnås et al., 2016).

In this paper, we present analyses of a four-year temperature data series from near-surface temperature loggers, subsequent temperature regime modelling, and basic geomechanical mapping in Signaldalen, northern Norway, at a high-latitude, high-relief ~~rockslide~~rock avalanche site. The main objective of our study is to increase our understanding of the permafrost temperature regime at the site and what role the ground temperatures might have played in the release of the ~~slide.~~rock avalanche. Further, we aim to contribute to the understanding of the influence of solar radiation on near surface temperatures in different aspects of steep mountain walls at high-latitudes.

2 Methods

In order to map the thickness and the geometrical characteristics of the failure zone, the release area was repeatedly surveyed by terrestrial laser scanning~~.~~ (TLS). This allowed also for monitoring of eventual continued or ~~fresh~~new movement in the slope.

In-situ measurements of ground surface temperatures and subsequent temperature modelling was employed to get insight into the past- and present temperature regime at and near Polvartinden peak.

The larger area around the failure zone, outside the steepest slope, is characterized by a combination of small vertical rock outcrops and undulating slopes with an established soil cover. Snow cover and snow depth therefore varies considerably in these two different types of topographies. Consequently, temperature loggers were installed in both types of terrain, i.e~~.~~., in both vertical rock outcrops (Fig. 3a; rock surface temperature (RST)) and within soil material of the more gentle slopes~~.~~ (soil surface temperature (SST)). The locations of the temperature logger placements (Fig. 3b) were also selected in order to cover different aspects.

To obtain a robust estimate of how past and present ground temperatures had evolved prior to the failure, i.e~~.~~., in the long- and in the short-term, two different approaches were used: Firstly, the ~~ground surface temperatures~~RST in selected rock outcrops were linked to long-term changes in regional air temperature by comparing their statistical relation with temperature data from the official weather stations Skibotn (5 m asl) and Ripojavri (502 m asl); for locations see Figure 1. We found that the linkage between those two datasets was strongest for rock outcrops with little or no snow, i.e~~.~~., with a fairly direct link to the atmospheric conditions. Therefore, and secondly, it was desirable to look further into the temperature development in the ground in more gentle slope areas close to the release area with a more established snow and soil cover. To this aim, the CryoGrid2 model (cf. Westermann et al., 2013) was used. Since Polvartinden is an alpine peak, with three-dimensional effects that affect the ground temperature field inside the mountain, the lower permafrost limit will vary according to different aspects. In order to better understand the present subsurface temperature field of Polvartinden and to get a better idea of how the lower permafrost limit is located with respect to the release area of the ~~slide~~rock avalanche, a stationary three-dimensional transient heat conduction model (cf. Noetzli et al., 2007a, b) was used in addition. It was natural to install loggers in different compass directions to study the differences in surface temperature as a result of aspect dependency. As input data for the analyses of aspect dependency robust estimates of air temperature were needed for each of the installed rock wall localities. This in order to calculate the temperature difference (ΔT) between calculated air temperature and measured rock face temperatures (surface offset).

2.1 Geomechanical mapping

In the autumns of 2010, 2011 and 2013, the ~~slide~~ failure zone of the rock avalanche was imaged with an HD Optech ~~IIRIS~~ILRIS terrestrial laser scanner (Optech Incorporated, Canada). In addition, the failure zone and its surroundings were imaged with a Giga-Pan camera in autumn 2013. The purpose of the laser scanning data collection was to develop a digital terrain model in order to enable the quantification of the extent and volume of the failure zone~~ and the slide area.~~. The data needed to be of significant resolution to enable temporal modelling of the area and identify zones ~~within the slide area~~ that show differential movement.

The failure zone is located approximately 600 m above the valley bottom, on the western side (facing east). The altitude of the failure zone in combination with the slope of the mountain side resulted in the failure zone being located approximately 800–1100 m away from the nearest possible scanning locations. To produce a 3D terrain model with a minimal amount of occlusion data was collected from three independent locations (identified in ~~Figure 3~~Fig. 3b). Each location was selected because it provided an excellent line-of-sight to the failure zone~~.~~ and reduced the bias in extraction of discontinuity orientations in accordance with Lato et al. (2010). Data was collected at varying resolutions. Low resolution data was collected to enable the visualization and modelling of a large area of the mountainside. Higher resolution data was collected at specific zones near the failure zone and its surrounding areas to enable a geomechanical interpretation. The orientation of structural discontinuities were manually identified and orientation vectors were converted to dip and dip direction measurements (Lato et al., 2012). The structural measurements were extracted from a 3D surface mesh was generated from the overlapping point cloud datasets.

~~Generally, Optech Ilris data would be parsed into pf formats and imported directly into the software used, i.e. into PolyWorks. However, in this instance that method failed. For an unknown reason the normal information collected by the scanner, processed by the Parsing software and interpolated by PolyWorks failed. This meant that the data had to be parsed as an XYZI file and manually interpolated by Polyworks. Once the polygonal model was created, the resultant model was viewed, cleaned, and edited where necessary. During this process the most fundamental change to the model was the filling of data holes using regularized spline interpolation. The data holes were filled primarily for visualization purposes. The extremely irregular surface~~

~~as well as large amounts of snow and ice present resulted in the development of a mesh with numerous larger data holes. All major data holes were filled except for one large area that was covered with ice during the time of the first scanning. The area is approximately 100 m wide and could not be filled with any accuracy.~~

Calculating the volume of the failed mass ~~could only be done~~was completed based on interpolation and estimation. Since there does not exist a sufficiently high resolution 3D image of the area before the failure, an interpolated surface had to be used to estimate the volume. The volume estimation was completed using both PolyWorks and ArcGIS software packages. The volume of the 2008 rock avalanche was computed using the 2012 TLS data by delineating a pre-failure surface based on adjacent slope topography bounding the scarp. The differential volume between the 2012 TLS data and the pre-failure surface was computed using standard 3-dimensional techniques, as presented, e.g., by Lato et al. (2014).

2.2 Ground surface and air-temperature measurements

Ground surface temperatures were measured at fourteen locations employing miniature temperature data loggers of three different types: M-Log5W loggers (GeoPrecision GmbH, Germany), redesigned UTL-1 loggers (University of Bern and University of Zurich, Switzerland, cf. Gruber et al., 2003) and UTL-3 loggers (GEOTEST and WSL Institute for Snow and Avalanche Research SLF, Switzerland). We followed the installation setup described by Gruber et al. (2003), hence measuring near-surface rock temperatures at 10 cm depth (Fig. 3a). The absolute accuracy for all three logger types is ± 0.1°C. The resolution of M-Log5W, UTL-3 and UTL-1 is 0.01, 0.02 and 0.27°C respectively. The M-Log5W loggers and the UTL-3 loggers were programmed to measure every 30 minutes, while the UTL-1 loggers (having smaller memory capacities than the other two employed logger types) measured every two hours. Based on experiences from long-term permafrost monitoring programs in Norway using such loggers (e.g., Isaksen et al., 2011; Gisnås et al., 2016) we claim that the bias on the accuracy of the temperature measurements introduced with the given setups is negligible.

Installation sites for RST were chosen based on the availability of near-vertical rock outcrops, their closeness to the original failure zone and, last but not least, the accessibility of the identified locations. Where possible, a vertical distance of several meters to the flat terrain was chosen, however, this was not possible at all of the sites, which has some implications on the interpretation of the results~~.~~ (cf. chapter 4). The fourteen measurement sites finally chosen are plotted on the map shown in Fig. 3b; they are located along the NNW-ridge of Polvartinden and in the valley ground.

Measurements were ongoing from September 2009 to August 2013 for 10 out of 14 loggers. For R09 and R10 only data for 2009–2011 exists, while R12 and R13 where first installed in 2011, thus yielding data for 2011 to 2013. Nine ~~temperature~~ RST-loggers (type M-Log5W and redesigned UTL-1; identified as R# loggers in text and figures) were installed in vertical rock faces on rock outcrops and along small cliffs with different aspects, while five SST-loggers (standard UTL-1/UTL-3; identified as S# loggers in text and figures) were placed directly into the soil at ca. 10 cm depth, in order to measure ~~ground~~

~~surface temperatures.~~SST. One additional logger was placed in a cairn in the valley floor to monitor air temperature. For several of the analyses presented annual mean values were calculated to identify variations in mean annual rock- and soil surface temperature (MARST and MASST respectively) and mean air temperature (MAAT). This ensures easier comparison between the monitoring sites and makes it easier to identify local maxima and minima as well as trends (cf. Isaksen et al., 2011).

~~Regional air temperature data were obtained from the two meteorological stations Skibotn (5 m a.s.l.) and Rihpojavri (502 m a.s.l.) (for locations see Fig. 1), being the most representative weather stations for our study area. Originally, two more loggers where installed on the Polvartinden ridge, one to measure air temperature, one for soil temperature monitoring. Unfortunately, these two loggers – located spatially close to each other – were subject to theft already during the first measurement year.~~

2.3 Ground surface temperature modelling

2.3.1. Long-term changes in ground temperatures

To study long-term changes in ground temperatures representative for the same elevation as the failure zone, but for sites with a developed soil cover and where snow accumulates, we used data series from the transient permafrost model CryoGrid2 (CG2; cf. Westermann et al., 2013). The physical basis and operational details of CG2 are documented in Westermann et al. (2013), and only a brief overview over the model properties is given here. CG2 calculates ground temperatures according to Fourier's law of conductive heat transfer in the soil and in the snowpack to determine the evolution of ground temperature over time. Thus, CG2 can deliver the transient response of ground temperatures to a changing climate. In addition to conductive heat transfer, the change of internal energy and temperature in the ground is determined by the latent heat generated/consumed by soil freezing/thawing. Subsurface movement of water is not included and only heat flow in the vertical direction is considered, thus solving an effective 1D problem and neglecting lateral heat flow between neighboring cells. This is justified for grid cell sizes considerably larger than the extent of the vertical modelling domain (cf. Westermann et al., 2013).

The model is forced by operational gridded (1 x 1 km) air temperature (Mohr, 2009; Tveito et al., 2000) and snow-depth (Engeset et al., 2004; Saloranta, 2012). ~~A grid cell covering Polvartinden and similar in elevation (665 m a.s.l.) as the failure zone was selected.~~Snow cover data is based on the seNorge snow model (www.senorge.no) that uses gridded observations of daily temperature and precipitation as its input forcing, and simulates, among others, snow water equivalent (SWE), snow depth (SD), and the snow bulk density (ρ) (Saloranta, 2012). The gridded air temperature for our site is mainly driven by the nearby Skibotn and Rihpojavri weather stations which were validated against our local measurements (see section 3.2). The gridded precipitation for our site is mainly driven by observed precipitation at Skibotn weather station. Since we have no observations of snow cover and owing to the large spatiotemporal variability of snow conditions in our alpine study area, the snow simulations from the seNorge model provide probably the best estimate of the spatial average 1 x 1 km snow conditions for our site (cf., Saloranta, 2012). A grid cell covering Polvartinden and similar in elevation (665 m asl) as the failure zone was selected. In addition a grid cell covering the valley floor was selected for validation against our observations. The model parameters for the lower boundary condition and for ground properties were chosen as in Westermann et al. (2013). The surface geology was based on the major surface sediment classification by the Norwegian Geological Survey (NGU, 2010; Thoresen, 1990). For our study ~~site~~sites and the selected grid ~~cell~~cells, the sediment stratigraphy was classified as till and coarse colluvium (class 11 according to the sediment map by NGU, 2010) and followed default settings in CG2 with volumetric fractions of the soil constituents and soil type for each layer as given in Westermann et al. (2013). An interval of snow thermal conductivity (k_{snow}) was regarded as parameter uncertainty by Westermann et al. (2013), and was used as a confining range for the true conditions as a *low* (LC, k_{snow} = 0.3 Wm^{-1} K^{-1}) and a *high* (HC, k_{snow} = 0.5 Wm^{-1} K^{-1}) conductivity scenario run of CG2. For validation, one year of daily SST data from the CG2-model was compared with observed SST-data from our valley and mountain sites. To study long-term changes in ground temperatures and to avoid dominance of near-surface high-frequency temperature variations we selected 10 m depth as an appropriate depth. For more details on the CG2 model, please refer to Westermann et al. (2013).

2.3.2. ~~Present subsurface~~Subsurface temperature field

~~To~~We employed a transient three-dimensional transient heat conduction model to get insights into the general pattern of the present subsurface temperature field of Polvartinden~~ we employed a stationary three-dimensional transient heat conduction model~~. The applied model is discussed in detail in Noetzli et al. (2007a, b) and Noetzli and Gruber (2009~~). Where possible, this stationary~~) and is basically a finite element model including Fourier's law and phase change via an apparent heat capacity (cf., Mottaghy and Rath, 2006). Where possible, this three-dimensional transient heat conduction model was fed with local datasets. The geometry of the model was based on a 10 m digital terrain model for the surface topography of the entire mountain and a 1000 m rectangular box for the subsurface mass. Values defining the subsurface properties (heat capacity, thermal conductivity, porosity) were obtained from representative sites nearby (cf. Lilleøren et al., 2012) and assumed to be uniform (cf. Table 1). The geothermal heat flux as the lower boundary condition was set to 60 mW m^{-2} (Slagstad et al., 2009). The upper boundary condition was set as a fixed temperature with annual mean values from the distributed near-surface rock ~~face temperature measurements (MAGST). The~~ temperatures (MARST). Distributed MARST was calculated based on the aspect dependency ~~was based on~~of MARST. For this, the fit curve from the first two years of data (2009/10–2010/11, cf. Fig. ~~11a). MAGST~~12a) was calculated. MARST changes in the steeper parts of the mountain were assumed to have roughly followed changes in mean annual air temperature (MAAT). The regional LIA glacier maximum is suggested to have occurred about 1900–1910 (e.g., Ballantyne, 1990; Bakke and others, 2005). Therefore, the time-dependent simulation was started with a steady state temperature field for the MARST in 1900, and ~~assuming~~we assumed a linear MAAT increase ~~from 1900 to the start of our measurements of~~ of 0.55~~ ~~°C until the end of the simulation (Lilleøren et al., 2012~~).~~), i.e. the year of the rock fall event. Our results are valid for areas that are assumed not to be influenced by a snow cover, i.e., the steep rock-faces of

Polvartinden. The main source of uncertainty for the three-dimensional modelling is related to the extrapolation of the ~~MAGST~~MARST.

2.3.3. Lapse rates

To study the surface offset and local influences on air temperature lapse rates for our ~~ground surface temperature~~RST locations, we studied the inter-annual variability from 2009-2013 in monthly mean lapse rates. In the absence of local air temperature measurements at higher elevation (similar to the rock wall loggers), the monthly lapse rates were calculated based on the nearest mountain weather station located at Rihpojavri (502 m ~~a.s.l.,~~asl, Fig. ~~1~~1a) and our local air temperature measurement site in Signaldalen valley (65 m ~~a.s.l.).~~asl).

3 Results

3.1 Geomechanical characteristics

Based on the laser scanning results, the volume of the ~~rockslide~~rock avalanche was estimated to be approximately ~~500’000~~500 000 m^3. These results confirm earlier estimates suggested during the emergency response work initiated directly after the event NGI (2008).

The depth to the~~ actual~~ failure surface was found to range from 40 m at the back of the failure zone to 0 m at its toe, with the failure zone being a complex wedge with increasing depth of the plane of failure from the front to the back (Fig. 4). Understanding the principle mechanism behind the failure involves a kinematic evaluation of the failure scarp. ~~Figure 5 illustrates an equal area stereonet. Plotted on the stereonet are the poles to the bedding surface. The bedding planes have been identified as the failure surface from the terrestrial laser scanning data and from interpretation of the GigaPan photography.~~ The extraction of the orientation of the bedding surface and the orientation of the natural slope (“Pre Failure Surface”) are determined directly through measurements using the laser scanning data. The discontinuity ordination data are plotted on a stereonet in Figure 5. The bedding planes have been identified as the failure surface from the terrestrial laser scanning data and from interpretation of the GigaPan photography. The green line in Figure 5 represents the orientation of the natural slope surface, and the green circle represents the corresponding daylight window. The white cone depicts an estimated friction surface of 30°. Poles that are contained outside of the white circle but within the green circle are kinematically unstable and represent potential sliding failure planes~~. This~~ (Goodman, 1995). The stereonet demonstrates that the bedding surface orientation is steeper than the estimated friction angle~~,~~ but shallow enough to daylight with respect to the slope face. The bedding surface meets, therewith, the kinematic requirements of a sliding failure posing a potential rockfall hazard (see e.g., Hasler et al., 2012).

The surficial change of the exposed rock mass between 2011 and 2013 is mapped through the comparison of laser scanning data collected at different points in time. The maximum size of blocks released between 2011 and 2013 range from 1 m^3 to 10 m^3. In summary, the repeated laser scanning measurements between 2009, 2011 and 2013 showed little to no ~~rock fall~~rockfall activity, both within the 2008 failure zone~~,~~ and in the adjacent rock slopes.

3.2 Measured mean annual ground surface temperatures

Measured ~~Ground Surface Temperature (GST)~~RST and SST data series for the four-year measurements period were smoothed with a 365-day running mean filter (Fig. 6). There is a slightly higher variability for ~~air temperature (AirT)~~MAAT than ~~GST~~for MARST and MASST (at ca 10 cm depth), but the overall correlation is high. ~~Rock~~MARST and ~~soil temperatures~~MASST measured during 2009–2013 ~~show mean annual ground surface temperatures (MAGST)~~are between -1.4~~ ~~°C ~~(coldest) ~~and +2.2 °C~~ (warmest),~~, with the highest temperatures recorded in the measurement year April 2011 to March 2012 (Logger R05-E). For the vertical rock face sites, the lowest ~~MAGST~~MARST was recorded at the north-facing site in a 365-days period between September 2009 and September 2010 (Logger R06-N).

The large inter-annual variability found in our temperature measurement series is in congruence with general climate conditions in Troms and is confirmed by measurements in nearby mountain slopes (~~Isaksen~~Farbrot et al., ~~2011a~~2013). For the monitoring periods 2009–2013, average, minimum and maximum ~~mean annual air temperatures (~~MAAT~~)~~ at the nearby Skibotn meteorological station (27 km to the NNE from Signaldalen) were 0.0~~ ~~°C, 1.2°C below and 1.5°C above the MAAT for the normal period of 1981–2010, respectively.

During 2010/2011 some of the sites were clearly influenced by a snow cover. (cf. Table 2). Based on an analysis of wind direction, wind speed and total snow accumulation at nearby weather stations (among others Skibotn), we assume this difference in snow cover to be caused by inter-annual differences in prevailing wind direction and preferential snow deposition.

We found a very high correlation between our air temperature measurements in the valley bottom (2009–2013) and air temperature data covering the same period from the nearest two meteorological stations at Skibotn (R^2 = 0.99; Fig. 7a) and Rihpojavri (R^2 = 0.97; Fig. 7b). Furthermore, we found a very high correlation between our local air temperature measurements in the Signaldalen valley floor and the rock wall loggers R05 (R^2 = 0.99; Fig. 7c) and R06 (R^2 = 0.98; Fig. 7d). The temperature series from loggers R05 and R06 were chosen because they seem least influenced by snow and they are also the "warmest" and "coldest" loggers, respectively, in the four-year measurement period. Temperatures at logger site R05 are about 1.3~~ ~~°C higher than at the north-facing series at logger site R06. This good correlation allowed us for using the meteorological data from Skibotn for modelling of the long-term evolution of ~~GST~~MARST at R05 and R06 for evaluating the potential permafrost distribution near the original failure zone on Polvartinden and recent ~~ground temperature~~RST changes by coupling our in-situ surface temperature data with regional and large-scale climate data.

3.3 Modelled mean annual ground temperatures

In order to be able to link our short temperature rock wall series to the greater regional climate development, we compare it to synthetic series based on the regressions presented in Figure 7 and the ~~transient permafrost model~~ CG2 ~~that calculates ground temperatures according to conductive heat transfer in the soil and in the snowpack (Westermann et al. 2013).~~simulations. The CG2 model results are representative for our soil temperature sites in the valley floor and in flatter areas at ca. 665 m ~~a s.l.~~asl (which is roughly the altitude of the failure zone), where snow accumulates~~. –~~ and where our soil temperature loggers were installed.

A one-year comparison of observed daily SST at sites with an established soil and snow cover and the modelled SST from the CG2 model is shown in Figure 8. As seen in Figure 8a the SST observed in the valley floor is in good agreement (R^2 = 0.86 to 0.88) with the CG2 model results for a grid cell with approximately the same elevation as the valley floor. The Signaldalen valley is dominated by mountain birch forest and is characterized by an open forest layer with heather and lichen species dominating the forest floor layer. This type of forest causes snow to accumulate and insulates the ground against strong cooling (cf. Isaksen et al., 2008; Farbrot et al., 2013), an effect which can be seen in our CG2 model results. In the mountains (Fig. 8b) there is a somewhat weaker correlation (R^2 = 0.69 to 0.82) between observed and modelled data. The MASST are about 1 to 1.5°C lower than in the CG2 model for the corresponding grid cell with similar elevation. The warm bias of the simulations during winter is mainly explained by the different snow conditions assumed in the model: while our observational sites in the mountains are affected by considerably strong snow cover variations induced by wind drift, the snow model used as input in the CG2-model generally overestimates snow depths in high mountain areas (cf. Saloranta, 2012). This is also in line with results of an equilibrium permafrost model used by Gisnås et al. (2013) who found a better agreement between the CG2-model and their validation data when the snow depth in the snow model data was reduced by 30% for areas above the tree line. During summer, on the other hand, our observed SST in the mountains are well reproduced by the CG2 simulations.

The coupling of our coldest in-situ ~~GST~~RST data from R06 with the climate data from Skibotn since 1958 (cf. Fig. 7d) suggests that the highest ~~MAGST~~MARST on record was 1.1 °C and occurred in the 12-month period between March 2007 and February 2008, i.e., ending four months before the Signaldalen ~~rockslide~~rock avalanche detached (Fig. ~~8a~~9a). A comparison with the CG2 model data run at 10 m depth for the same period suggests a gradual warming and degradation of the permafrost and supports our synthetic series with the warmest period occurring just a few months prior to the failure. Figure ~~8b~~9b shows the recent 10-year period for the synthetic series and the four-year series of the in-situ ~~GST~~MARST data for the ~~coldest~~lowest (R06) and the ~~warmest~~highest (R05) rock wall temperatures, covering the range of the measured rock wall temperatures. The figure also shows the two CG2 model runs. According to Westermann et al. (2013) the thermal conductivity of the snow is the largest source of uncertainty in CG2, thus a low (LC) and a high conductivity (HC) scenario run of CG2 are used for the last

10 years as a confining range for the true conditions. Note that the synthetic series are slightly warmer than the ~~GST~~ series for R06 during the first two years. The deviation is likely to be due to a thin snow cover that partly covered the rock-face/outcrop particularly the second winter of the monitoring campaign.

Since 1958, the CG2 results clearly indicate warming. Coincidentally, the period Apr 2011 to Mar 2012 was as warm as the previous record from the early 1990s. On the other hand, our measurements took place during the coldest period since 1988. Our measurement period covers, in other words, most of the temperature regime that can be expected in this region within a multi-decadal perspective. The depth of the fracture zone varies between 20 to 40 m (Fig. 4). Temperature penetration from the surface to such depths typically takes between one (20 m) to two years (40 m) (cf. e.g., Gruber et al., 2004a). There is, therefore, a good temporal link between the maximum ground temperature at 20-40 m ~~deep~~depth (through at least the last 50-60 years) and the actual timing of the ~~slide~~rock avalanche release.

The general pattern of the transient 3D-temperature field in the mountain is illustrated with a 2D transection from South to North in Figure 12. The curve of the isotherms inside the mountain stems from the topography and more pronounced the steeper the topography is (Noetzli et al., 2007). Since the difference in MARST is only in the order of 1.5°C, the isotherms are a little inclined towards the colder mountain flank, however, the main temperature change in the subsurface is experienced with a change in altitude (unlike in e.g. very steep peaks in the Alps, where the main temperature chance is experienced with the position between different warm mountain flanks). Based on this schematic sketch of the subsurface temperatures, the permafrost body underlies all of the steeper part of the mountain and is several hundred meters thick at some places. In the area of the rock fall starting zone more shallow permafrost is simulated.

3.4 Local lapse rate

As shown in Figure 7b our regression analysis shows a high correlation between the nearest mountain weather station located at Rihpojavri (502 m ~~a.s.l.,~~asl, Fig. 1) and our local air temperature measurement site in Signaldalen valley (65 m ~~a.s.l.,~~asl, Fig. 3b). Figure 9 reveals an annual median lapse rate of 6.1 °C km^{-1}, but with substantial seasonal and inter-annual variability. Lapse rates are smallest (4–6 °C km^{-1}) in late-summer to early autumn, and largest (7–10 °C km^{-1}) in spring. We see the strongest gradients in spring (May).

The monthly temperature difference (ΔT) between calculated air temperature and measured rock face temperatures are shown in Figure ~~10~~11. For the sites R03, R04 and R12, periods during which the loggers obviously had been covered by snow (cf. Table 2) were omitted. Monthly values were calculated in the same way as used for the lapse-rate calculations (cf. section 3.4, Fig. ~~9~~10). The results in Figure ~~10~~11 show a clear aspect dependency with a slightly lower permafrost limit in northern exposition as opposed to slopes exposed to the South. The figure also shows the year-to-year variations in the order of ±0.5 to

1.0 °C. There is a clear seasonal dependency, with ΔT near 0 °C or even negative during autumn and early winter, and largest ΔT (1.5 to 5 °C) in late spring and early summer.

Figure ~~11a~~12a shows the annual temperature difference (ΔT) between calculated air temperature and measured rock face temperatures and aspect dependency as derived from the data. The points represent the mean values and the black line is the best polynomial fit to the data ($R^2 = 0.84$). The grey lines show the polynomial fit to the interquartile range of ΔT. The ΔT for logger R06, which is the logger facing most towards north (10°), is +0.6 °C as compared to air temperature, while the two loggers facing most towards south (R05, 90° and R09, 208°) show both a ΔT of +1.7 °C. We, thus, find an aspect difference between north and south facing loggers of 1.1 °C. Figure ~~11b~~12b shows the subsurface temperature field as modelled with the stationary three-dimensional transient heat conduction model (Noetzli et al., 2007a, b) based on rock wall temperature data from 2009–2011. The figure shows a slice transecting Polvartinden mountain from South (left) to North (right).

4 Discussion

Many studies related to mountain climate and permafrost research operate with uniform and/or constant lapse rates of 6.0 or 6.55 °C km^{-1}, justifying this by these values being the typical theoretical adiabatic lapse rate. However, theoretical adiabatic lapse rate can vary considerably (from 3 to 9 °C km^{-1} for surface conditions at mid-latitudes) due to its dependency on pressure and temperature (Minder et al., 2010). In valley bottoms temperature inversions and cold air pooling can affect lapse rates (e.g., Rolland, 2003), and channelled flow over mountain passes can result in large local temperature anomalies (e.g., Steenburgh et al., 1997).

Our study suggest that the high lapse rate values we find in Signaldalen in spring are caused by the fact that the snow cover normally is depleted in the valley bottom in late spring, while the higher areas and the northerly exposed moderately steep mountain slopes still exhibit an extensive snow cover. These results are in accordance with international literature (e.g., Minder et al., 2010) and are likely applicable to other mountainous areas in northern Norway. Other studies have also shown that seasonal cycles in lapse rates have similar amplitudes to those found in our study, but that the phasing of the seasonality varies (Bolstad et al., 1998; Rolland, 2003; Tang and Fang, 2006; Blandford et al., 2008; Gardner et al., 2009). They also highlight the importance of local air temperature measurements in experimental observational networks to reduce uncertainty. Our results clearly support these findings.

Our measured ~~MAGST~~MARST and MASST values from the period 2009–2013 indicate warm/marginal permafrost at all temperature logger sites ~~except at~~ facing north-northeast, i.e., in the ~~valley site (SD02, 65 m a.s.l.) and at a wet, mossy depression site on~~same aspect as the ~~lower part of the ridge (S13).~~failure zone. Our results yield an estimated mean lower limit of permafrost at around 600-650 m ~~a.s.l.;~~asl; this value is in agreement with earlier estimates in the inner fjord- and valley areas of Troms (~~Isaksen et al. 2011a;~~ Farbrot et al., 2013; Gisnås et al., 2016) and coincides with the upper limit of the failure zone. Since all rock wall loggers are installed at small cliffs (rather than in vertical, large rock walls) the snow close and/or atop these cliffs can attenuate normal winter cooling and, thus, affect the results during the winter months. This is visible in Figure ~~10~~11, where some of the NE-facing loggers exhibit a clearly higher winter temperatures than what would be expected when compared with air temperature. ~~Some of the~~The influence of the snow cover on the rock thermal regime has been studied in steep rock walls (Haberkorn et al., 2015a, b; Hasler et al., 2011; Magnin et al., 2015). The highly variable spatial and temporal distribution of the snow cover strongly influences the ground thermal regime of steep rock faces (Haberkorn et al., 2015a, b; Magnin et al., 2015). Haberkorn et al. (2015a) found that snow depths exceeding 0.2 m were enough to have an insulating effect on steep, bare bedrock. Such amounts are likely to accumulate in steep, high rock walls with a certain degree of surface roughness. As snow reduces ground heat loss in winter, it has an overall warming effect on both N- and S-facing rock walls despite the fact that it provides protection from solar radiation in early summer (Haberkorn et al., 2017). However, in moderately inclined (45–70°) sun-exposed rock walls Hasler et al. (2011) suggest a reduction of MARST of up to 3°C compared to estimates in near-vertical, compact rock due to snow persistence during the months with most intense radiation. For our study some of the soil logger sites feature moss or thin vegetation (like the mountainside otherwise) which also affects the temperature. Together, the loggers (including the soil loggers) encompass the variation of the snow and surface conditions present around the Signaldalen rock ~~slide~~avalanche site.

The long-term development in the annual mean temperatures from the instrumental period (the late 1800-century) in northern Norway can be split into four periods: a cold period in the beginning, a period referred to as “early 20th century warming” culminating in the 1930s, a period with cooling from the 1930s to the 1960s, and finally the “recent warming” from the 1960s to present (Hanssen-Bauer et al., 2015). Regional climate data since 1958 from Skibotn weather station suggest a general warming of the greater Signaldalen area. This is in agreement with the general atmospheric warming (Hanssen-Bauer et al., 2015) and observed permafrost temperature rise (Isaksen et al., 2007) and the long-term permafrost degradation (Farbrot et al., 2013) in northern-Norway, with an observed peak during the ~~International Polar Year~~period 2007–2009 (Christiansen et al., 2011; Romanovsky et al., 2016). The CG2 model results suggest (not shown) an increase of the lower permafrost limit for snow covered sites from ca. 600 m ~~a.s.l.~~asl in the 1960s to about 800 m ~~a.s.l.~~asl between 2000–2010 (Westermann, personal communication, 2015). Our modelled data (Fig. ~~8~~9) and observed ground temperature data from the nearby Guolasjávri permafrost borehole (Fig. ~~12~~13, see location in Fig. 1) suggest that the highest mean near surface temperatures on record occurred in the period between March 2007 and February 2008, thus ending only a few months before the Signaldalen ~~rockslide~~rock avalanche detached.

According to a study by Fischer et al. (2012) on potential triggering factors at 56 historical ~~rockslide~~rock avalanche and ~~rock fall~~rockfall events in the Alps, it seems to be the marginal permafrost zones where most of the recent changes concerning ice content and hydrology have taken place; parameters that are seen as having an important influence on slope stability also by, e.g., Allen and Huggel (2013), Deline et al. (2011) and Fischer et al. (2013). In laboratory studies, Davies et al. (2001) demonstrated that the shear strength of an ice-bonded rock discontinuity significantly reduces with warming and that the minimum shear stress is reached between ca. -0.5 and 0 ~~°C.~~ °C. Krautblatter et al. (2013) showed that fracture ice influences stability down to a depth of approximately 20 m, below that depth, the overburden pressure of the rock mass is becoming too high. According to their study, intact rock bridges and rock-rock-contacts exert additional control on rock stability. While permafrost increases the uniaxial and tensile strength of such rock bridges and rock-rock-contacts, warming of permafrost decreases these strengths and could, thus, trigger rock slope failures. In degrading permafrost, rock-mechanical properties may control early stages of destabilization and become more important for higher normal stress, i.e. higher magnitudes of rock slope failure. Ice-mechanical properties outbalance the importance of rock-mechanical components after the deformation accelerates and are more relevant for smaller magnitudes (Krautblatter et al., 2013). In early summer, the combined effect of hydrostatic and cryostatic pressure can cause a peak in shear force exceeding high frozen shear resistance (Draebing et al., 2014).

Analysing temperature conditions prior to 144 past rockfall events in the Swiss Alps and the French Mont Blanc massive, Lüthi et al. (2015) recently showed that small to medium-sized rockfalls (with volumes up to 100’000 m^3) mainly occurred during short-term periods of unusually high temperatures, whereas larger high-elevation rock slope failures occur all year-round. Hasler et al. (2011) showed that local warming of cold permafrost may be induced by advection and the related erosion of cleft ice, and that permafrost degradation through thermal advection by running water can rapidly lead to the development of deep thaw corridors along fracture zones and potentially destabilise much larger volumes of rock than through thermal conduction on similar timescales. Draebing et al. (2016) found that the rock mechanical regime also was snow-controlled in permafrost rock slopes at Steintaelli in the Swiss Alps. They found that during snow-free periods, high-frequency thermal expansion and contraction occurred. Rock temperature locally dropped to -10°C, resulting in thermal contraction of the rock slopes. Snow cover insulation maintained temperatures in the frost-cracking window and favoured ice segregation. Such repetitive occurrence destabilises the rock slope and can potentially lead to failure (Draebing et al., 2016).

~~In Scandinavia, the amount of direct observations about~~A recent synopsis of MARST at 34 locations within the PERMOS network in Switzerland (PERMOS, 2016) shows that MARST are generally higher than MAAT. In south-exposed near-vertical locations this difference amounts to up to 10°C, in north exposed locations MARST is only slightly higher than MAAT (PERMOS, 2016). Similar values were also measured in steep alpine rock walls at the Matterhorn and the Jungfraujoch in Switzerland (by Hasler et al., 2011) and at the Aiguille du Midi in France (by Magnin et al., 2015). In Scandinavia, the amount of studies using direct observations to explore the influence of solar radiation on near surface temperatures in different aspects of steep mountain walls is limited so far. In Jotunheimen in southern Norway north-facing rock wall surfaces were on average less than 1°C warmer than the surrounding air temperature, while MARST in more radiation exposed rock walls was up to 4°C higher than MAAT (Hipp et al., 2014). Generally, the influence of direct solar radiation is less pronounced at high latitudes than in mid-latitude ~~mountain ranges, a fact which can be observed in our modelling results of the subsurface temperature field (Fig. 11b).~~mountains. Our results indicate an altitudinal difference of ~~several tens of meters~~ roughly 200–250 m between northerly and southerly aspect~~,~~ (cf. Fig. 12b), as compared to ~~several hundred~~1000–1500 meters (or up to 8–10 °C) for some sites in~~, e.g.,~~ the Swiss Alps (~~cf. Gruber et al., 2004~~PERMOS, 2016). Due to our small sample size these results should be seen as tentative estimates. Also, the potential effect of the midnight sun has ~~been~~ not been looked at. In addition, it has to be noted that we lack loggers exposed directly to the South. Based on the shape of the polynomial fit curve (Fig. ~~11a~~12a), and on what is known from other studies (e.g., Gruber et al., 2004a; ~~Gruber et al., 2004a;~~ Noetzli and Gruber, 2009; Magnin et al., 2015; PERMOS, 2016), the difference between northerly and southerly aspects is probably 0.2 to 0.4~~-~~°C higher than indicated by our measurements. This would yield an absolute difference between "warmest" and "coldest" aspects of approximately 1.3 to 1.5~~-~~°C, which still is considerably lower than for mid-latitude mountain ranges~~.~~ and about 1.5°C lower that were reported by Hipp et al. (2014) ~~report comparably small aspect dependencies of the lower permafrost limit~~ for a location in ~~Southern~~southern Norway, thus supporting strongly decreasing aspect dependency with increasing latitude. Smaller differences in MARST between North and South-facing rock walls eventually result in less pronounced 3D effects of the subsurface temperature field as illustrated with the modelling results of the subsurface temperature field (Fig. 12b). For alpine peaks with a triangular geometry, isotherms can be nearly vertical in the uppermost part of mountain peaks in the European Alps (cf. Noetzli et al., 2007b) and temperatures change mainly with the position between North and South-facing slopes. Here, isotherms are only little inclined, and the main change in subsurface temperatures is experienced with changing altitude and not changing exposition.

5 Conclusions

Analyses based on four-year ~~ground~~rock- and soil surface temperature series suggest warm/marginal permafrost at several of the investigated sites in Signaldalen, and yield an estimated mean lower limit of permafrost at around 600-650 m ~~a.s.l.,~~asl, an altitude which coincides with the upper limit of the failure zone. Regional climate data since 1958 and nearby borehole data suggest a general warming and that the highest mean near surface temperatures on record occurred some months before the Signaldalen ~~rockslide~~rock avalanche detached. These findings are supported by model results of the transient permafrost model CG2. Our results give also new insights into aspect dependency of mountain permafrost in northern Scandinavia, a subject that so far has been little explored. We found an absolute difference in ground surface temperatures of approximately 1.3 to 1.5°C between south- and north exposed slopes.

The volume calculation based on terrestrial laser scanning data show that the depth to the actual failure surface was found to range from 40 m at the back to 0 m at the toe. The repeated laser scanning between 2009 and 2013 show little to no activity in both the 2008 failure zone and the adjacent rock slopes. Considering that temperature penetration to, e.g., 15–20 m depth in frozen rock typically takes one year it is likely that changing rock-/ice-temperatures due to the general warming and in response to the extreme warm previous year have played an important role in the detaching of the Signaldalen ~~slide~~rock avalanche.

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Table and Figures

Table ~~1.~~1. Values for subsurface properties as used in the stationary three-dimensional transient heat conduction modelling.

Subsurface property	Value (Value range)
Heat capacity	850 J kg^{-1} K^{-1}
Thermal conductivity	2-2.5 Wm^{-1} K^{-1}
Density	2800-2900 kg m^{-3}
Water content	0.2-1 %

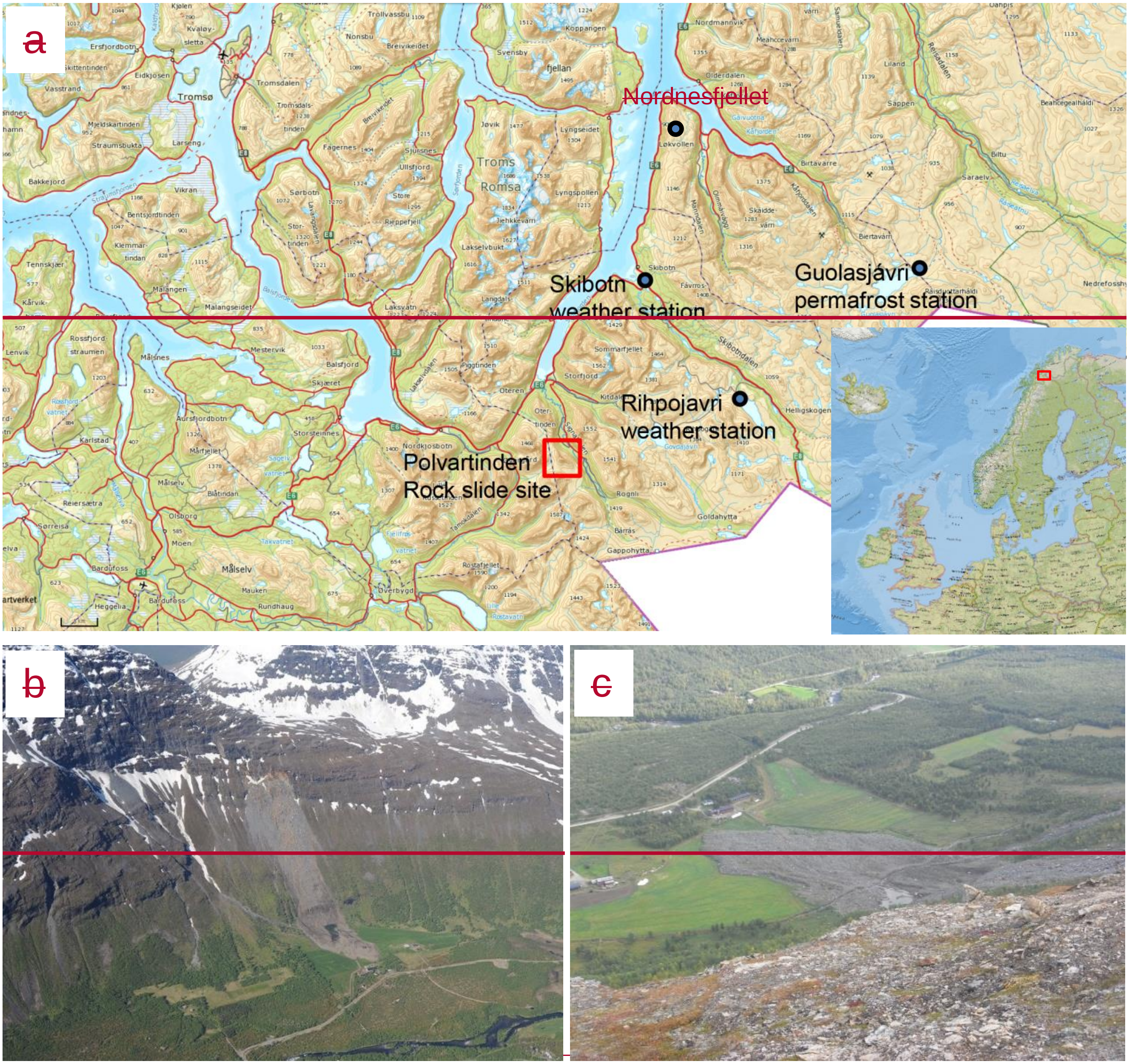


Table 2. Rock wall and soil temperature loggers installed in Polvartinden and Signaldalen. Snow cover = estimated winter snow cover based on a simple evaluation of day-to-day temperature variability and standard deviation compared to reference logger R05, divided into three categories: I) snow cover mostly absent, II) snow cover over short periods, III) snow cover over longer periods.

Site	Elevation (m asl)	Aspect (°)	Snow cover
A01-SD	65	-	-
R03-NE	640	30	III

R04-NW	657	290	III
R05-E	630	90	I
R06-N	632	10	I-II
R09-SW	646	208	I
R10-NE	598	40	I
R12-NE	668	45	II-III
R14-NNE	608	35	II
S02-SD	64	-	III
S02	664	-	I-II
S07	626	-	II
S08	648	-	I-II
S11	540	-	II
S13	643	-	III

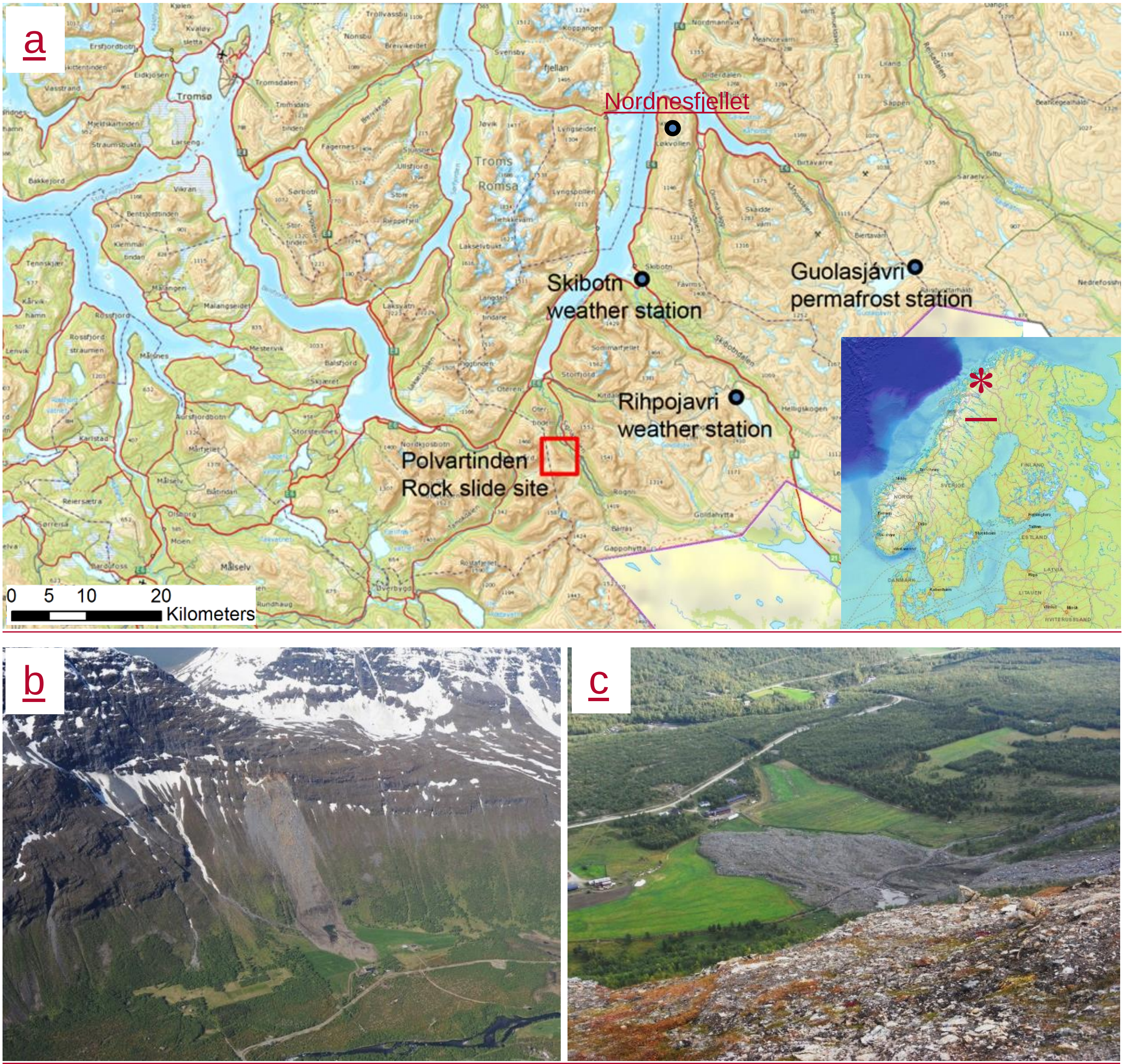


Figure ~~1.~~1. a) Key map showing the location of the Polvartinden ~~rockslide~~rock avalanche site, the two weather stations and the permafrost station used in this study (map source: Copyright © by Norwegian Mapping Authority/Statens kartverk); b) Signaldalen ~~rockslide~~rock avalanche as seen from helicopter on 28.7.2009 (Photograph by courtesy of Gunnar Kristiansen, NVE); c) runout area of the Signaldalen ~~rockslide~~rock avalanche in September 2011 (Photograph by courtesy of Gunilla Kaiser).

Figure ~~2.~~2. Visible in-situ ice (encircled areas) observed in the ~~rockslide~~rock avalanche failure zone on June 26th, 2008. The photograph was taken a few hours after the event (Photograph by courtesy of Kjetil Brattlien, NGI).

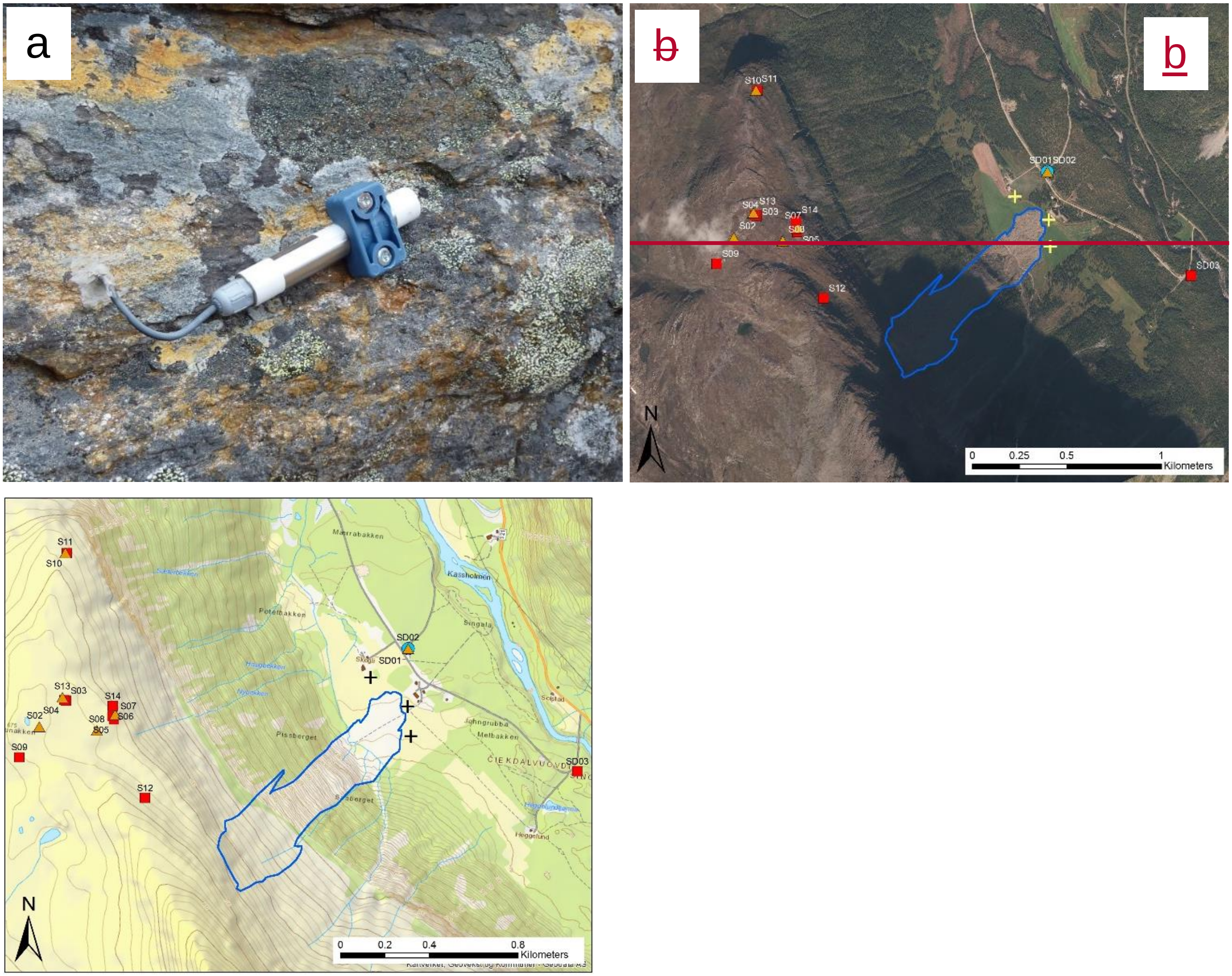
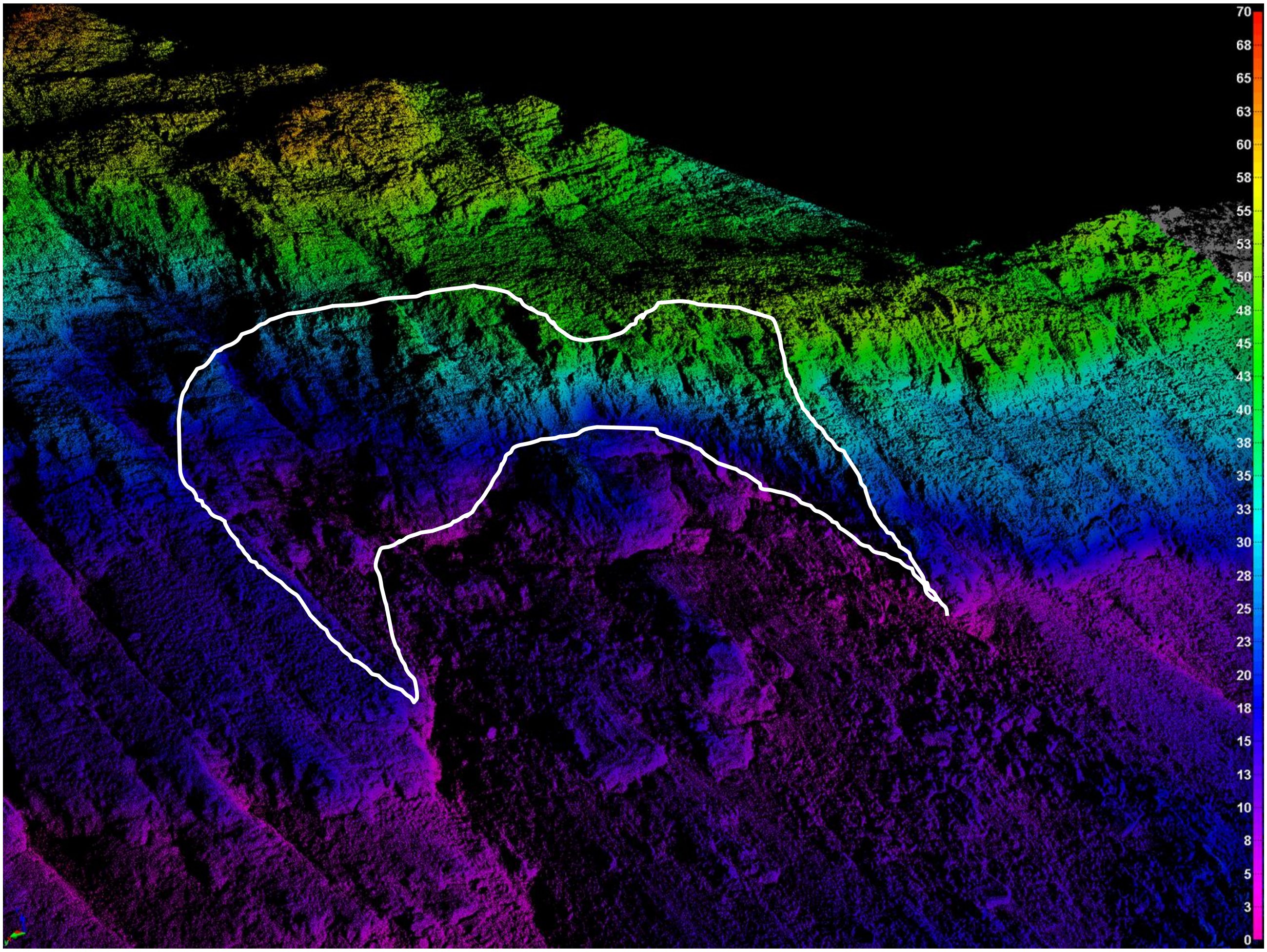


Figure ~~3.~~3. a) Typical logger installation setup; b) Map showing temperature measurement locations in vertical rock faces (red squares), within soil material (yellow triangles) and within a stone cairn (one blue circle in the valley bottom), the three ~~yellow~~black crosses mark the laser scanning locations. ~~Slide~~Rock avalanche outline in blue. ~~Background: Aerial photograph (~~Map source: Copyright © by Norwegian Mapping Authority/Statens kartverk~~)~~.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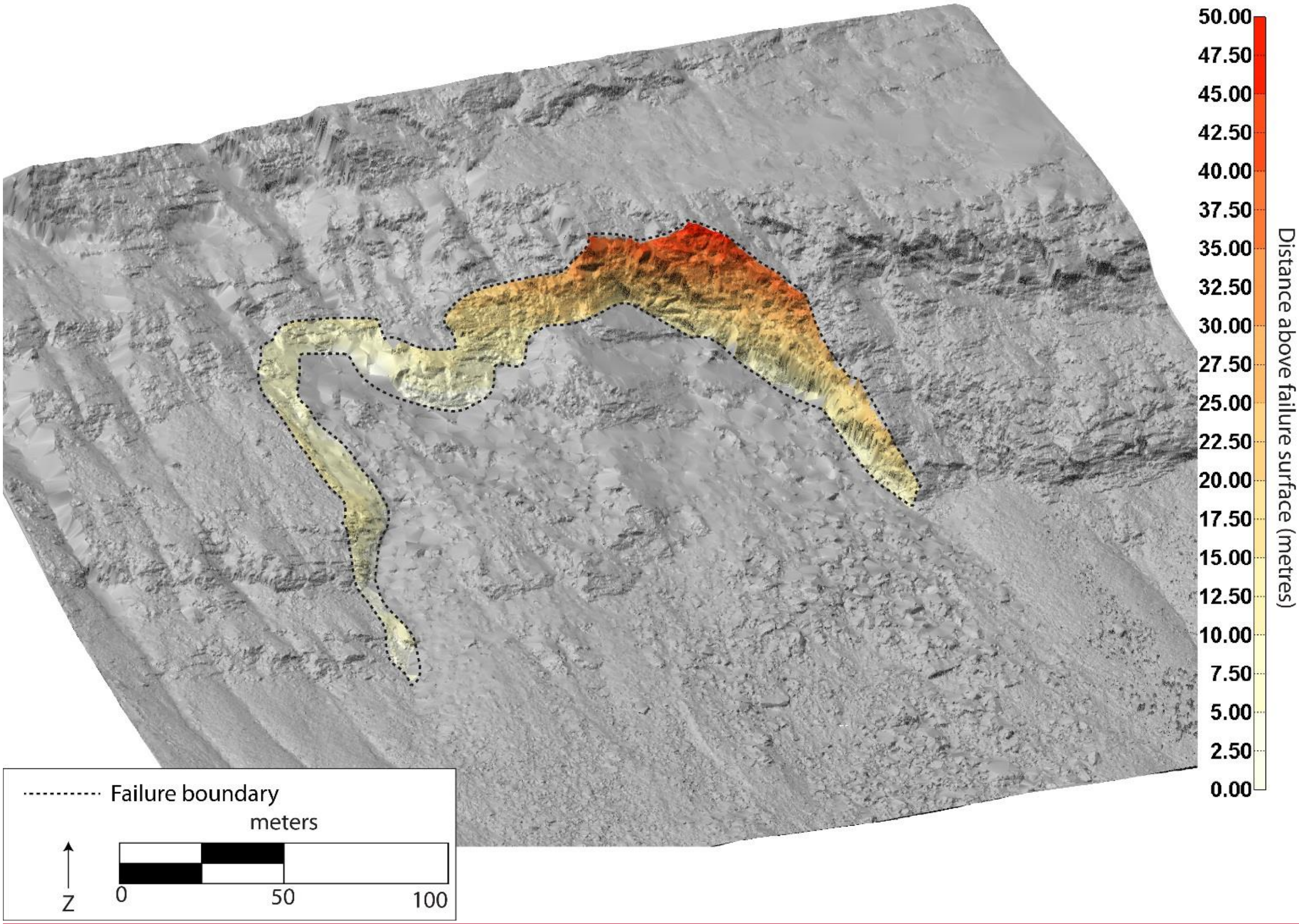


Figure ~~4.~~4. The failure zone of the Signaldalen rock ~~slide~~avalanche (outlined ~~in white~~with black stippled line) is a complex wedge much deeper to the plane of failure at its back than at the front. The colours indicate the distance in metres to the plane of failure (~~in purple~~from white = 0 m to red = 50 m).

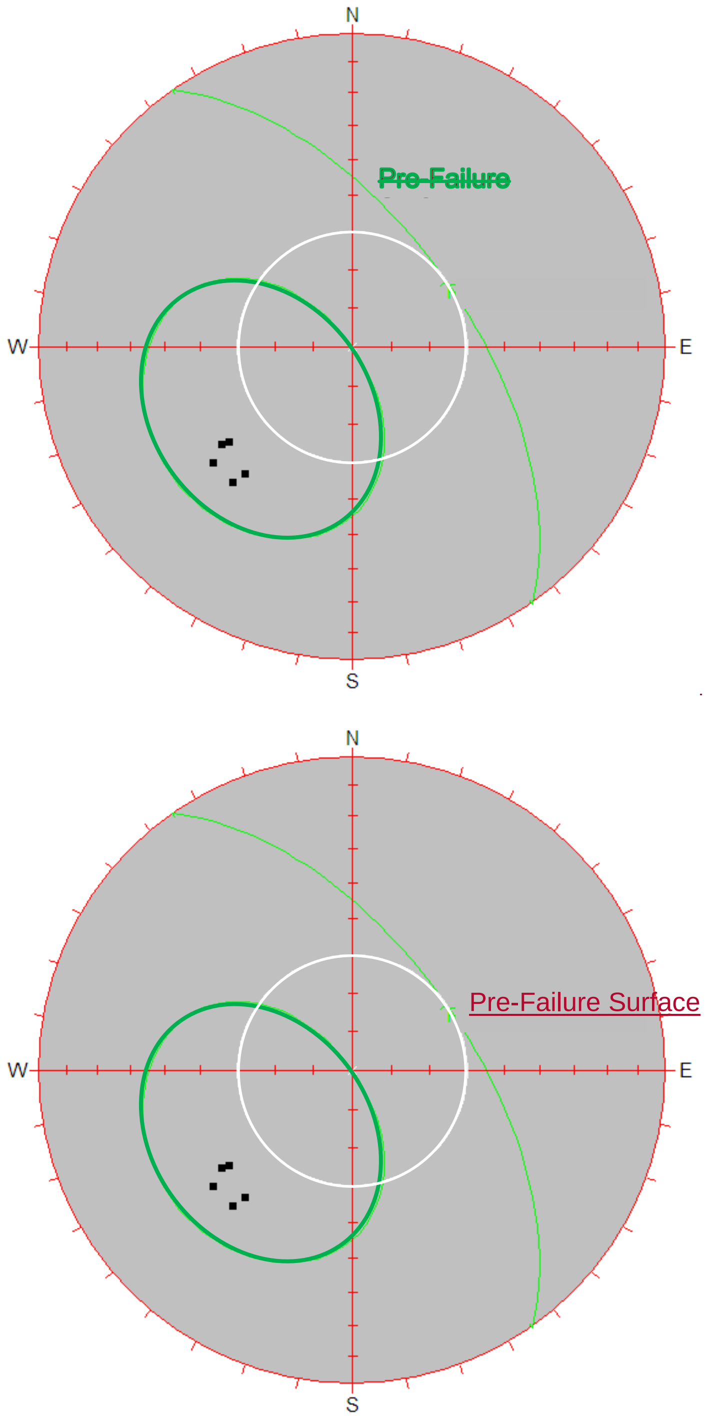


Figure ~~5.~~5. Kinematic analysis of the bedding planes for the sliding failure. The green line represents the orientation of the natural slope surface before failure, the white circle represents an estimated friction cone of 30° and the green cone represents the sliding daylight window for the associated pre-failure surface~~.~~, the poles to the bedding surface are plotted as black markers.

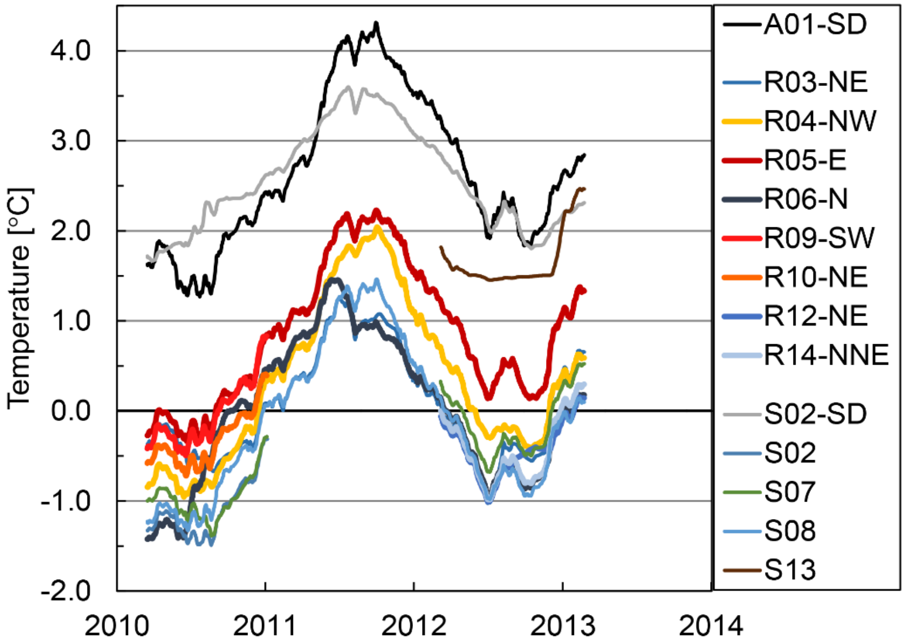


Figure ~~6.~~6. Mean annual ~~ground~~rock- and soil surface temperatures (all except A01-SD) and air temperature (A01-SD) during the period Sept. 2009 to Aug. 2013 shown as simple moving 365-days average for all sites. To ensure that the temperature variability was not shifted in time, the mean values were centered (an equal number of days on either side of the mean value). A01-SD is air temperature in Signaldalen (SD), R03 to R14 are the rock face loggers and S02 to S13 are the soil temperature loggers.

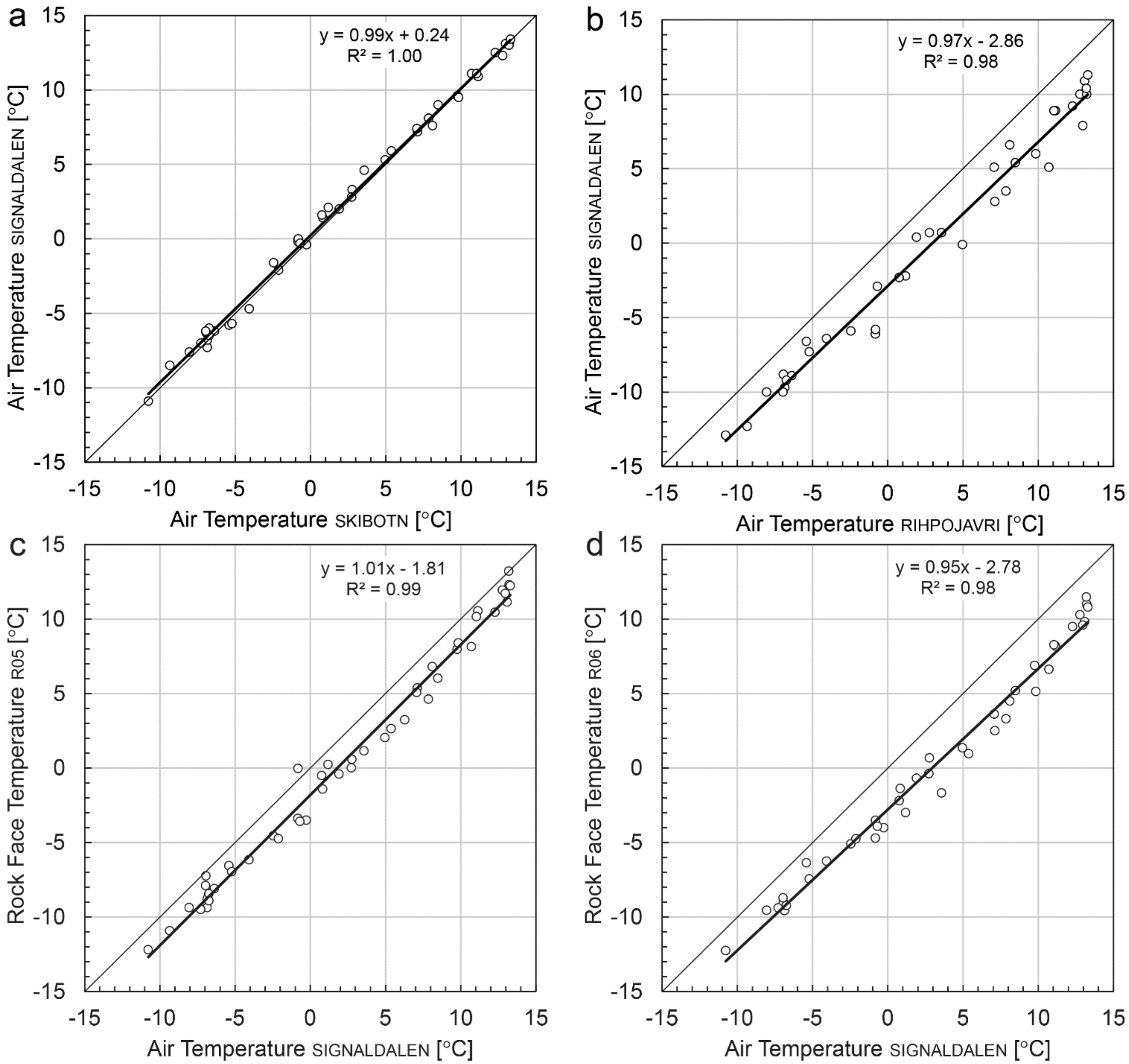


Figure ~~7.~~7. Scatter diagrams showing relations between monthly mean temperature for the main air temperature and rock face temperature series, including linear regression lines. (a) Relation between air temperature in Signaldalen and at meteorological station at Skibotn; (b) relation between air temperature in Signaldalen and air temperature measured at Rihpojavri weather station (25 km from Signaldalen); (c) and (d) relation between rock face temperature at logger site R05, respectively logger site R06 and air temperature in Signaldalen.

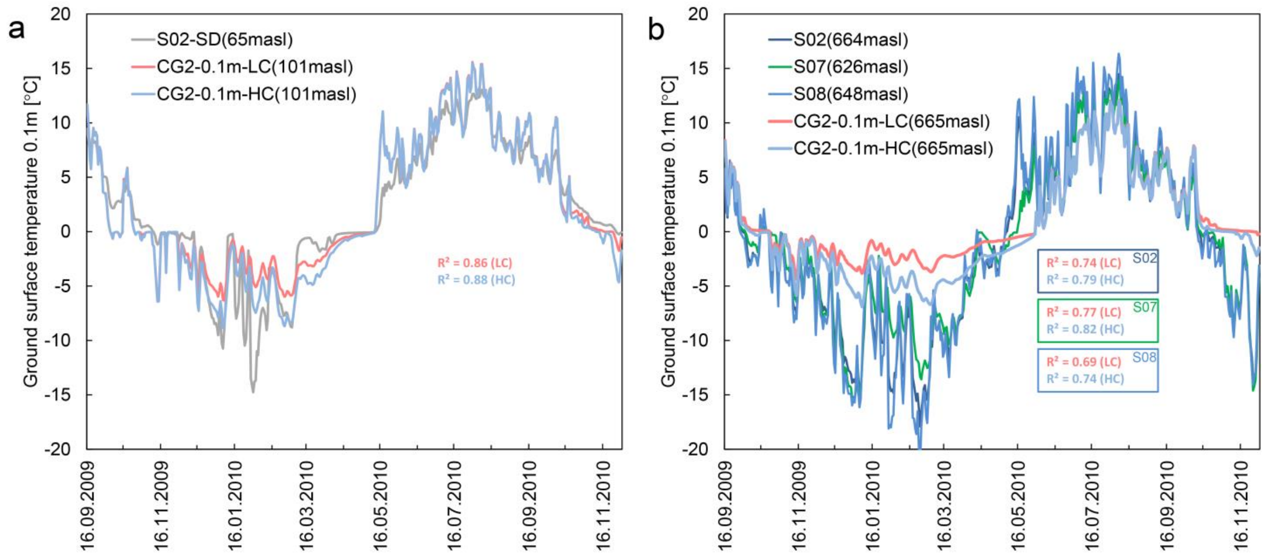


Figure 8. One year of observed soil surface temperatures (SST) at sites with an established soil and snow cover compared with the modelled SST from the transient permafrost model CryoGrid2 (CG2) for high (HC) and low (LC) thermal conductivity of the snow: a) time series of SST at site S02 – placed next to the air temperature measurement location in the valley floor – compared to the modelled CG2 series for the actual grid cell; b) SST series for the soil sites close to the release area (S02, S07 and S08) compared to the corresponding grid cell with similar elevation as the release area (i.e., 665 m asl).

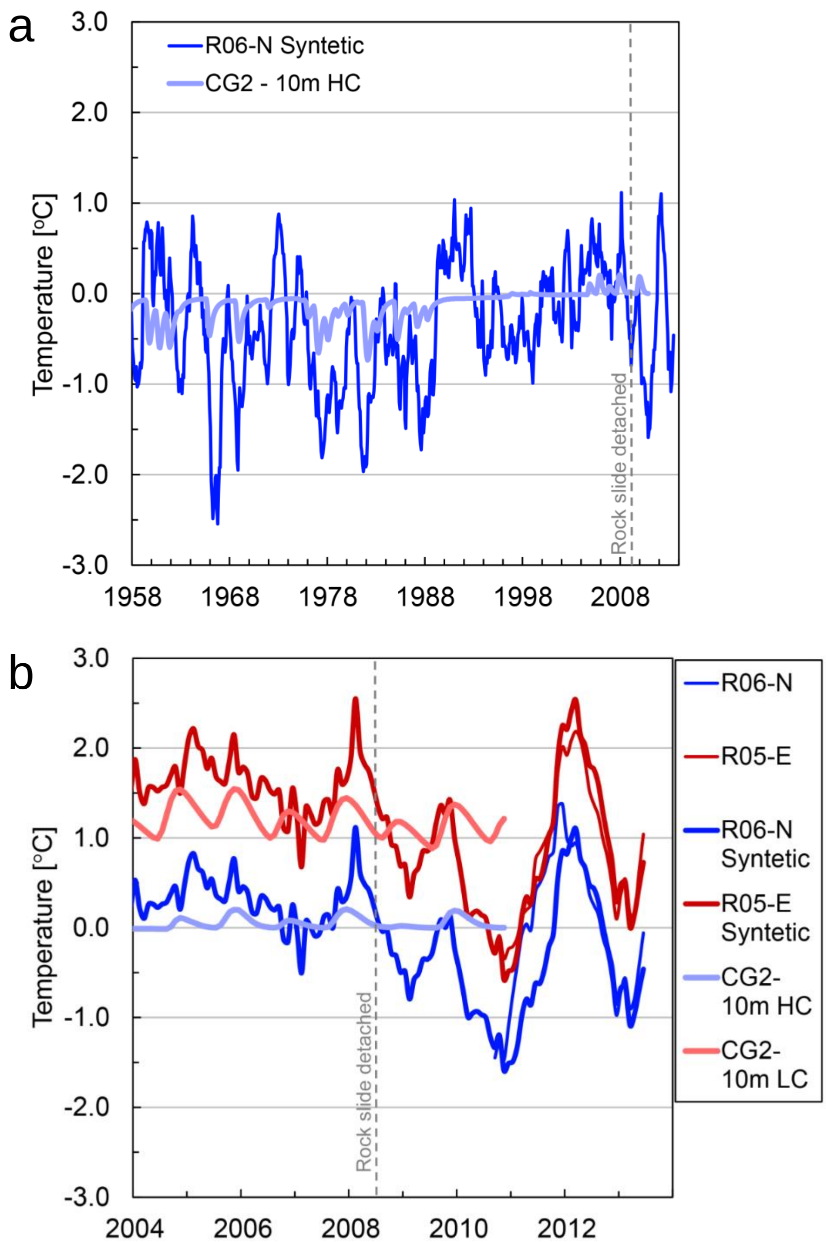


Figure ~~8.~~9. a) Synthetic series since 1958 of the coldest rock wall site (R06) based on the regression presented in Figure 7. Also shown is the transient permafrost model CryoGrid2 (CG2) run that calculates ground temperatures at 10 m depth according to conductive heat transfer in the soil and in the snowpack (Westermann et al., 2013). The CG2 model results are representative for areas at approximately 665 m ~~a s.l.~~asl with slope gradients allowing for snow accumulation. The grey dotted line shows when the Signaldalen rock ~~slide~~avalanche detached~~.~~; b) The recent 10-year period for the synthetic series overlaid on the respective in-situ ~~GST~~RST data for the lowest (R06, blue lines) and highest (R05, red lines) rock wall temperatures. Also shown are the CG2 model data runs at 10 m depth for High (HC) and Low (LC) thermal conductivity of the snow.

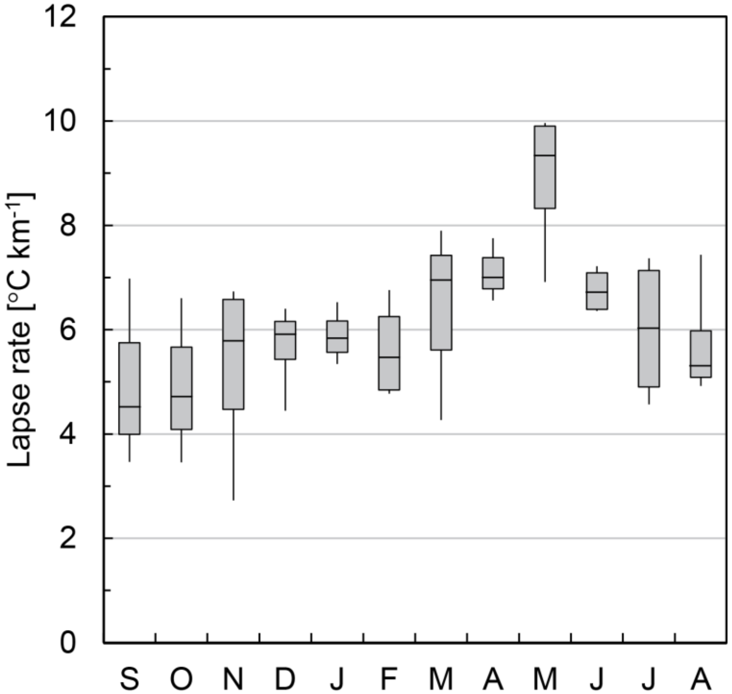


Figure ~~9.~~10. Inter-annual variability from 2009-2013 in monthly-mean lapse rates based on air temperature data from Rihpojavri (502 m ~~a.s.l.)~~asl) and Signaldalen (65 m ~~a.s.l.)~~asl) from 2009-2013. Boxes show the interquartile range of the month's lapse rate, horizontal lines inside the boxes show the median values, and the whiskers show the full range of the data.

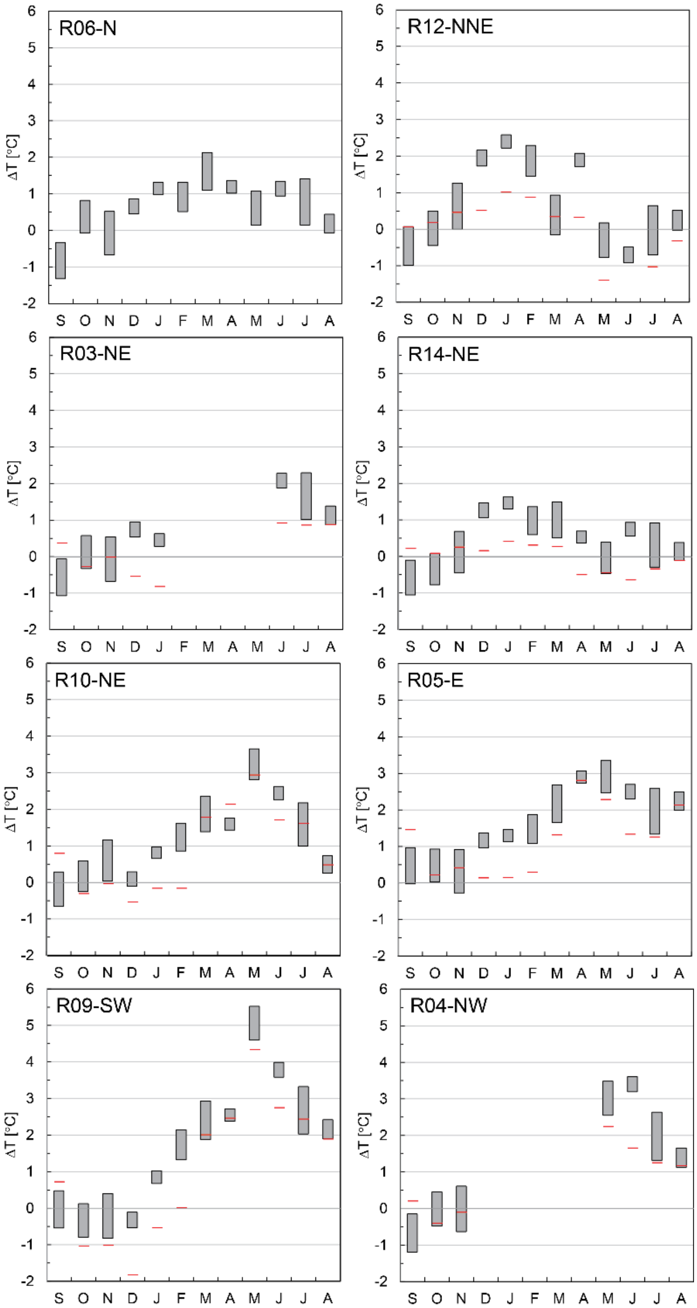
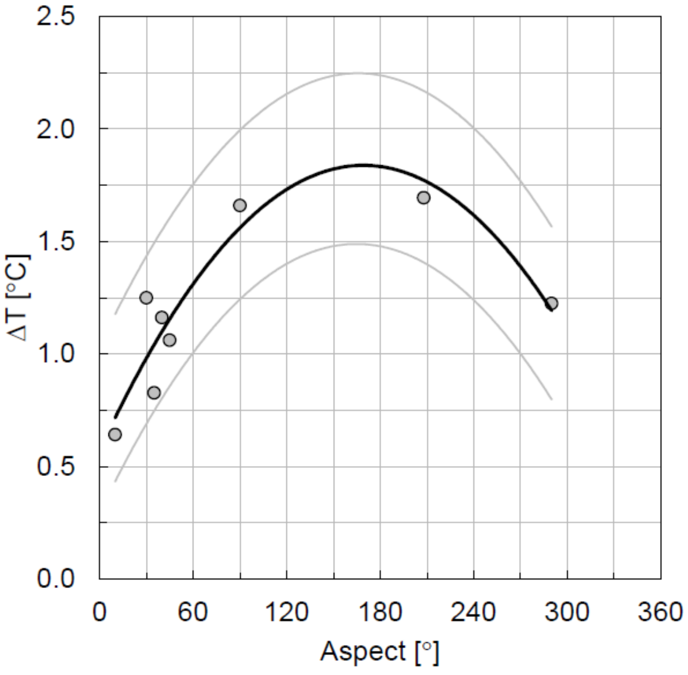


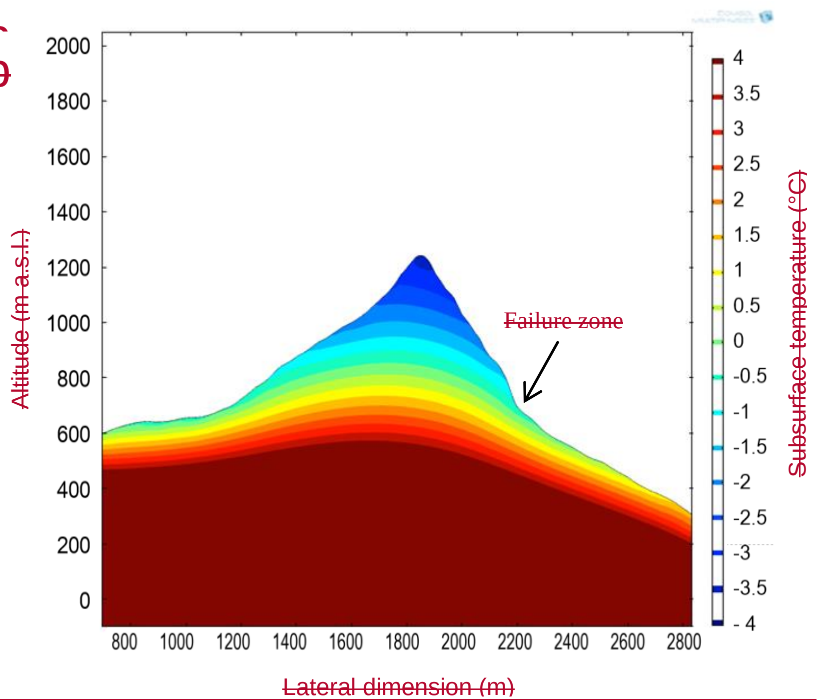
Figure ~~10.~~11. Monthly temperature difference (ΔT) between calculated air temperature and measured rock face temperatures from 2009-2013. Calculated air temperature is based on air temperature from Signaldalen (65 m ~~a.s.l.)~~asl) and lapse rates shown in Figure 9. Boxes show the interquartile range of ΔT. The red horizontal lines show the monthly-mean temperature difference between the respective loggers and the north-facing (i.e., the coldest) logger R06. Loggers R03, R04 and R12 were snow covered during winter and data was removed for snow-influenced months. For R09 and R10 only data for 2009-2011 exist, for R12 and R14 only for 2011-2013.

a



ΔT [°C]
2.5
2.0
1.5
1.0
0.5
0.0
0
60
120
180
240
300
360
Aspect [°]

~~b~~



~~Altitude (m a.s.l.)~~
2000
1800
1600
1400
1200
1000
800
600
400
200
0
800
1000
1200
1400
1600
1800
2000
2200
2400
2600
2800
~~Failure zone~~
4
3.5
3
2.5
2
1.5
1
0.5
0
-0.5
-1
-1.5
-2
-2.5
-3
-3.5
- 4
~~Subsurface temperature (°C)~~
~~Lateral dimension (m)~~

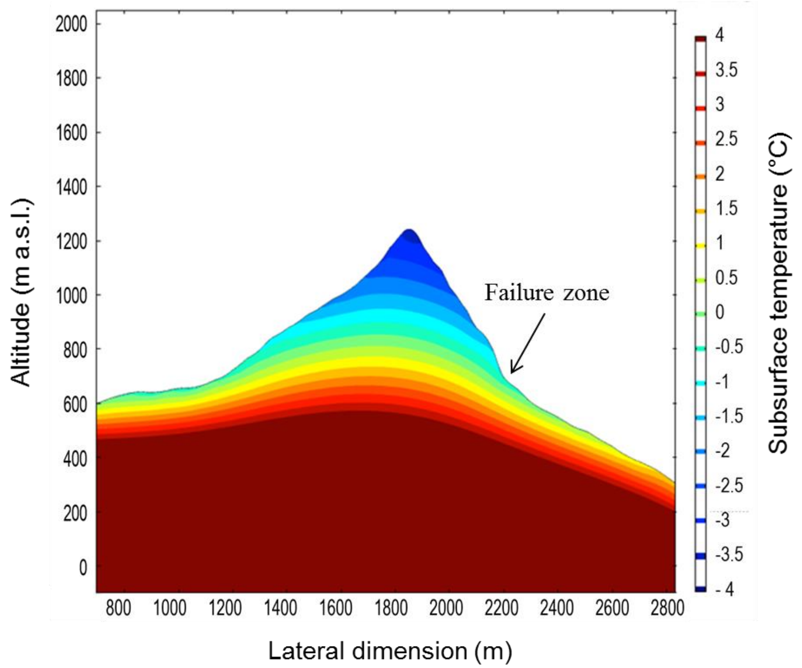


Figure ~~11.~~12. a) Annual temperature difference (ΔT) between calculated air temperature and measured rock face temperatures and aspect dependency as derived from the rock wall temperature data. The points represent the mean values and the black line is the best polynomial fit to the data (R^2=0.84). The grey lines show the polynomial fit to the interquartile range of ΔT~~.~~; b) 3D transient heat modelling of the subsurface temperature field based on values identified in a). Figure shows a slice transecting the mountain from South (left) to North (right).

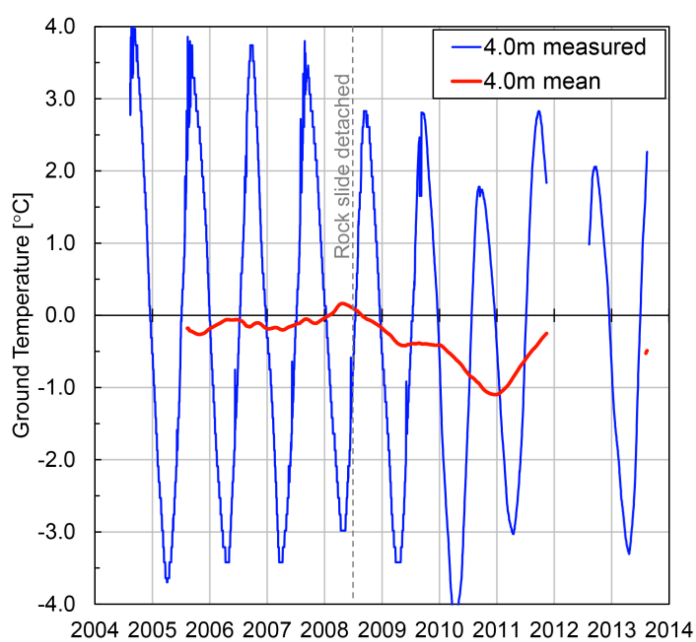


Figure ~~12.~~13. Ground temperature time-series at 4 m depth from a nearby permafrost borehole (Guolasjávri (Gu-B-1~~,~~), 786 m ~~a.s.l.;~~asl; more details can be found in Farbrot et al., 2013). The blue line shows the measured daily values and the red line shows a simple moving 365-

days average, i.e. the unweighted mean of the previous 365 days. The grey dotted line indicates the release date of the Signaldalen ~~rockslide~~rock avalanche.