

# TC-2021-380-review

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March 2022

## 1 General comments

**Originality:** The study makes use of key datasets at Abramov glacier to perform a novel modelling study to reconstruct 52 years of mass balance with concurrent process understanding. While this in itself is novel and the glacier is unusually rich in data for the region, a more explicit statement of the scientific motivation and purpose of the paper as well as critical discussion of how it might contribute to wider glacier process understanding in this region would be welcome. The contribution of modelled subsurface firn conditions is certainly valuable for improving density assumptions for geodetic mass balance estimates and for understanding the evolving potential internal accumulation capacity of the glacier over time. The paper concludes that the effects of warming on glacier mass balance are partially mitigated by increased accumulation, which may be relevant at a regional scale, and highlights that the buffering effect of internal refreezing is diminishing over time.

**Scientific quality:** The purpose of the paper could be more clearly stated with reference to what the key outputs are. For example saying that a process model allows you to build on the existing field firn study, and that the climate drivers of glacier mass change remain contested in this region, and can be investigated with a process-based model. We acknowledge that the parameter calibration is performed with rigorous manual calibration, yet by overlooking potential measurement uncertainties, as well as not using a multi-objective simulation/parameter evaluation this is not state of the art. Could you thus expand on the performance of the calibration? It would be good to justify your choice of sample number and location for the calibration datasets, and the treatment of uncertainties could benefit from recognition of a parameter equifinality issue (Rounce et al., 2020) in the optimisation. Rather than reporting only a single best-fit solution, incorporating a sensitivity analysis and its related uncertainties would be more robust. Indeed your comparison of the two sets of forcing data highlights the sensitivity of the modelled mass balance to the forcing data which demonstrates the value of a more comprehensive sensitivity study. This deserves even more emphasis given that the only difference between the two forcing data sets resides

in the timespan 1980-1998, the period for which you calculated the biases and should hence be closest to the observations.

Given the number of pre-processing steps in the forcing data and selection of model parameters, the impact of uncertainty in these choices is poorly quantified, making the model output and performance difficult to interpret, as we cannot be clear about the quality of the forcing data compared to reality. It may be helpful to show plots of the forcing data in the supplementary material as a start. This would also help addressing the difference between the two forcing data sets. Relating to the question of the overarching goal of this study: if the goal is to draw out the connection between the climate forcing and glacier response, investigations on the annual time series of mass balance alongside climatological properties - for example by adding climatological information into Table 3 or attempting to understand the causes of positive mass balance years in the timeseries - could be a starting point.

The timeframe of this study allows for an in-depth investigation on the drivers of glacier response also on the decadal scale, as continuously highlighted inter-decadal differences in glacier response exist. Therefore, it could be worthwhile to investigate whether there exist any breakpoints in the climatological data that could explain the simulated decadal variations in the mass balance (Table 3).

Relevance: Key findings are that negative mass balance of -0.27m w.e./year persists for the period 1968/1969-2019/2020 alongside a loss of firn pore space causing a reduction of internal accumulation. Despite increasing air temperatures, no acceleration of glacier-wide mass loss was found over time as a result of increasing precipitation rates which appears to exert a strong control over annual mass balance at this glacier. This is of interest given that knowledge of precipitation in this region is quite poor and is thought to contribute to anomalous glacier behaviour in this region..

Presentation: (1) We suggest restructuring so that the model description comes before the description of the datasets - this will help guide the reader to know the constraints on the construction of the forcing data. If this move is made, we suggest headings as follows under a 'methods' section: study site, model description, forcing data, calibration data, validation data. (2) Please consider also the best way to order Sections 3.5 and 3.4 as they describe parameterisations applied before the application/analysis of the modelled results as well as bias corrections based on whole model simulations. (3) Please revisit the naming of your sites: you have sites 1 and 2 in the accumulation zone, but it would be helpful to call your ablation zone site, site 3 to be consistent in the naming convention and also to show its location on the maps. Relatedly, please can you add a statement as to how representative these chosen sites are for mean ablation zone/accumulation zone conditions. (4) Please show which stake locations/data are used for calibration of which for validation in one of the map figures as its currently unclear if the locations for calibration are identical to the locations of validation but just a different subset of the data. (5) As

currently presented the figures and tables are often far from the associated text which could be improved for the final layout, some suggested changes to figures are included in the specific comments, and some English edits suggested in an annotated pdf.

## 2 Specific comments

L13: Is this heterogenous behaviour really only in the last decade?

L18: Is this increase progressive, or stepwise, and what is its magnitude (to help readers not so familiar with the region)?

L25: It may be valuable to specify that the advantage of a physically based model is that it is expected to be more suitable for projecting processes into unknown climates, compared to a temperature index model calibrated for one period in time applied within a non-stationary climate projection. The additional caution of course is that many physically based models include stationary parameterizations.

L28: Please specify which processes and/or which models include/exclude them, or maybe better still just delete the sentence starting "Important ..." as it's really not needed here

L34: Suggest "In the data poor HMA, ..." or maybe better still just delete the sentence starting "In the ..." as it's really not needed given the following sentence

L44: Suggest restructuring of this closing paragraph to more explicitly state some goals of the study. First state the problem/motivation (e.g. why do we care about this glacier/timeseries/region/... need for a continuous record in this poorly understood and anomalously behaving part of the mountain cryosphere?) and how doing this modelling study delivers a novel and useful solution (e.g. to evaluate decadal trends in surface energy balance fluxes, firn evolution ... and their connection to the forcing climate conditions, attempt more process understanding of previously suggested claims?)

L48: Delete "These in situ data, which are unique for the region, allow us to apply a rather complex model with relatively high data requirements."

L50: Delete the last sentence.

L64-69: Give time period of Barundun long term study; and state the long term mass balance of both studies.

L69: Where is this point site on the glacier?

L70: You mention a regime shift in the 1970s - please return to this point in the discussion and in the evidence from your modelled study, does your model show this increase, what are its effects on the glacier scale mass balance and firn development?

L76: Careful how you use this term reference glacier surface - it has a specific meaning as explained in Huss 2012, but I am not sure you mean this.

L78: The creation of the 1968 DEM needs more clarification on how you did it and how reasonable it is. Why do you use only 2 DEMs and not intervening satellite DEMs (e.g. SRMT, ASTER), to examine the trend from timestep to timestep? Would it be possible to use coarser, but more frequent dDEMs to check the reasonableness of the linear backwards extrapolation of surface height change over time, e.g. the ones mentioned from Barandun et al. (2015)?

L83: Was this weather station manually monitored? Is the non-digitized data a good candidate for <https://www.zooniverse.org/projects/edh/weather-rescue/>

L85: Suggest rephrasing to "Here, we use data from January 1968 until December 1998 for the following parameters: daily average air pressure, windspeed, relative humidity, temperature and cloud cover, as well as daily precipitation sum, daily minimum temperature and cloud cover and daily maximum temperature."

L89: Can you say something about expected undercatch based on any other measurements in the region or on the likely mountain undercatch for the device used? To my knowledge snow undercatch can be ~50 percent in mountain regions depending on the sensor so it would be good to mention here.

L91: Why 3 hour timesteps, when the ERA5 data is available hourly? If this is because the model needs this time resolution add this to the model description which should precede this section. Did you consider better ways of reconstructing the daily cycle e.g. by using the daily cycle of the ERA variables? For example it seems inconsistent that you introduce a daily cycle for cloud but then not for precipitation ... these choices need to be justified and their implications considered in the discussion section.

Section 2.3.2: It seems that you don't actually make much use of the recent weather station - is that correct? Just used in the pressure parameterisation and the radiation parameterization? a. Consider mentioning explicitly the use of each dataset in the analysis.

L107: Need some more explanation of the occurrence of missing data, what percentage is typically missing and what happens to the NaN data periods?



L110: What do you mean by an IQ range filter? Can you specify the scale of this here please? With this and the subsequent double application of the std filter it sounds like you substantially reduce your dataset variability range in a more or less arbitrary fashion before you begin with any physically based QC? Can you explain why?

L116: Other publications show the way for some more rigorous quality control e.g. could correct for snow over SW sensor, using the albedo readings, and for freezing of station using non-variant wind direction, and for RH<sub>100</sub> it is commonly reset to 100 according to many standard instrument manuals (e.g. Campbell scientific). RH values above 100% are usually set to 100%. Also you should report how many/what percentage of the total readings are filtered out (by each of the filtering steps)? Potentially include the whole quality controlled data series in the supplement.

L122: Could you please just mention why you did not use the extended ERA5 1950-1979 dataset. It might be still preliminary, but did you explore using it at all?

L124: Possibly expand on what TopoSCALE does - you input upper air field and a DEM, and get out radiation modifications based on the DEM, which all other fields remain the same for the given pressure level? For readers unfamiliar with this product it would be worth explaining also why you want to do this to your data first. As a side note, how does TopoSCALE compare to TopoCLIM now available? <https://gmd.copernicus.org/articles/15/1753/2022/> Can you also show the ERA5 gridcell location on your map please?

L128: Its not clear why you need cloud cover fraction but presumably for the SEB model, so perhaps state this at the start of the section, or better still put the model description before the data description so that its clear to the reader what is needed to force the model. Does the clear sky value come from ERA5 + TopoSCALE? Why usage of cloud cover fraction instead of readily available longwave radiation from ERA5?

L129: You may want to add units for completeness here.

L135: Are the parameter values developed for a particular site, or globally valid?

L140: "Even after downscaling (accounting for resolution diff.)..." here are you referring to what TopoSCALE does, without an explanation of what TopoSCALE is it is not clear to the uninitiated reader at this stage that it is a downscaling scheme, by which I understand it changes the resolution of the reanalysis data?.

L142: Explain why you aggregate to monthly for the bias correction when your data are available at different timescales, these steps seem confusing. Furthermore, you assume that biases 1980-1998 are the same as in recent decades, is

that likely? Why didn't you for example check the other station data to correct biases. Is the height difference between the two weather station locations handled by TopoSCALE?

L147: After correction using monthly data, cloud cover fraction in summer showed bad fit, therefore set cloud cover to 0 for days without precipitation and to 1 if precipitation is above threshold, this seems flawed as cloudy days could occur without precipitation, would it not be better to correct this with long-wave from ERA5 data? Or at least explain and justify these seemingly arbitrary choices - perhaps it is appropriate for this site, but please make a case for this here, using data or citations.

L150: So what is the timescale of these two forcing datasets? Original ERA5 data is hourly, ERA5 TopoSCALE bias corrected data is monthly, and is this dataset 3 hourly. Here please state the purpose of this duplicate dataset ... what do you gain from running the model with two datasets. I think it tells you something about the sensitivity of your results to the forcing data, but please explain this clearly here, and what it tests and does not test..

L150: Can you also show a comparison of the overlapping period of these two datasets, alongside the available weather station data so we can see how well they duplicate each other over periods of overlap - this is important to convince the reader of the homogeneity and robustness of the forcing data derived in this way. Probably fine in the supplement.

L165: Density assumptions including a constant elevation threshold, could you please justify this, especially as your abstract implies the firn is changing density, and presumably changing in spatial extent as well? You highlight in the text that this is a major source of error so it might be nice to consider simple uncertainty analyses based around output sensitivity to varying these assumptions. Also why is the firn line 4200m a.s.l., is this based on satellite data or the stake data, please justify this choice in the text.

L174: Please add in the text that March is selected as the end of the winter seasons before the snowpack is modified by melt and that September is the end of the hydrological year at this site, or otherwise justify these with your measured data, or citations.

L175: What is this selection of 19 stakes for - later it becomes clear this is for calibration, but check this whole paragraph for what you really want to explain.

Section 3: Moving this section to before the data description would make it much clearer what forcing data is actually needed to drive the model and therefore make it easier to understand the data preparations steps.

Figure 2: It would be useful to refer to this figure more in the data description - e.g. when discussing the two forcing datasets, and also maybe include the borehole temperature data on it?

L197: Does the Svalbard development location mean its particularly suitable for this glacier site? Better than alternative models for example? Certainly refreezing is important in Svalbard glaciers, right?

L201: Could you include a comment on the model performance from these studies - did it do well in comparison to others?

L206: State here which meteorological forcing data are required.

L212: Indication on connotation of fluxes (positive, negative) with respect to surface.

L214: How does the transition occur? Binary or gradual?

L216-208: This modelling of radiation seems odd as I thought this was the point of the TopoSCALE downscaling to give relevant radiative properties for a surface? Please start this section with the required forcing data for this model, and why you chose to not force it with the ERA5 incident radiative fluxes? Are cloud parameters for Svalbard (sea level and maritime) appropriate for Abramov? Can you quantify the better agreement and also state which weather station data was used here?

L223: Longwave radiation is computed in ERA5, why not use it to begin with?

L254: Liquid water instantly distributed following normal distribution, is this a widely used approach and do you know how it behaves in reality? Clearly this is a tricky parameter to know, but are there implications of this assumption?

L280: Does this initial subsurface condition differ greatly from just running the first year multiple times - would this not be a more precise way to evolve the starting conditions rather than a 20 year run? Either way its not known if this has a substantial impact on the results. Were there early borehole temperatures as well?

L283: So 'adjust' here means optimisation of parameters in some way described later on .. are these point scale optimizations and how does that relate to the fact that the glacier outline is updated over time - are all optimization points within the final glacier boundary?

L303: In the comparisons does the SMB include refreezing (so the modelled climatic mass balance cf Cogley et al., 2011) or not? Looks like it from L347, but maybe clarify here?

Table 2: Those parameters identified as optimized in this table are only those optimized through a whole model simulation. This seems a bit misleading as others are also optimized but to other data externally to a full model simulation, perhaps instead just separate literature/optimized values.

L317: Clarify pressure decay from historic average and modern average (October 2011 - 2020) as you mention the modern timeseries was only used from 2014 onwards due to data gaps (L102).

L336: This precipitation correction factor is performed within a SEB model simulation like the albedo optimization or is it simply applied to the precipitation timeseries compared to the glaciological data? If the latter doesn't it belong in the forcing data preparation section? Here do you compare the March accumulation in the monthly snowpits to the March accumulation in the forcing data - if so this could be stated more clearly. Maybe also justify the choice of summer bias correction (1.15) from a single different site, is this representative of a range of sites for this instrument?

L345: "Overall, the mass balance of Abramov glacier is negative for the years from 1968/1969 to 2019/2020." - what is meant here? The mean over the whole period? Give a value or consider giving the cumulative mass balance numerically (which could also be included in Figure for as a line plot on a secondary axis).

L347: Consider how to refer to the results - e.g. rather than listing what is shown in each figure you could focus on the description of what is revealed and cite the figure: e.g. The distributed mean annual mass balance for 1968/1969-2019/2020 is shown in Fig. 3. The mass balance of Abramov glacier is predominantly negative for the years from 1968/1969 to 2019/2020, despite part of this being taken up by glacier retreat, and shows no significant trend in annual mass balances ( $+0.0002\text{mw.e. a}^{-1}$ ,  $p\text{-value}=0.979$ ). The most negative modelled mass balances occur at the start of the period, while the two decades between 1978 and 1998 are characterised by almost balanced mass budget, thereafter becoming more negative again (Figure 4). The modelled mass balance gradient shows that accumulation is lowest during the first decade (1968/1969-1977/1978) and highest during the last modelled decade (2008/2009-2017/2018), and ablation is largest during the first decade, followed by the second last decade (1998/1999-2007/2008).

Figure 3: Please highlight the different glacier extents with dates or emphasize the fact that the average over the whole timespan with the adjustment of the glacier area leads to the distinct pattern. It might also be nice to report a long term mean ELA in this or the gradient figure (5b).

Figure 4: It would be nice to show the cumulative modelled mass balance

on a secondary axis.

L357: Again here, avoid just listing what is in the figures, but rather tell us what the figures \*show\*. The title makes it seem like this is part of the model validation but in fact here you also show the model calibration results, right? Suggest moving the calibration part to the end of section 3.5. (e.g. The model calibration was performed on eight snow pits locations at the end of March and annual mass balance measurements from 19 stakes and up to eight snow pit locations for the period 1968/1969-1997/1998; the comparison of optimized model simulations to these in-situ observations used for model calibration shows a stronger RMSE for the annual mass balance, and despite some bias in the linear fit, the datasets generally approach the 1:1 linear fit line (Figure 6).) Then here in this section on validation say: e.g. For model validation we compare the simulated mass balance to a set of point measurements not used in the calibration, which can therefore provide an independent measure of model performance. This comparison is based on 146 stake readings not used in model calibration for 1968/1969-1997/1998 (Figure 7a) and all available point mass balance measurements for 2011/2012-2019/2020 (Figure 7b). In both cases the comparison scatter plots cluster around the unity fit line, but inherit the bias tendencies seen in the calibration (Figure 6)) Consider colouring the points in Figure 6 by decade or year as then you might also reveal periods of better/worse model calibration, which could be interesting.

Figure 6: The different axis labelling in (b) might create a false impression, please homogenize them

Figure 7: So the performance isn't any better in the measurement period (a) compared to downscaled ERA5 period (b)? This could also be rephrased to the fact that both periods show good agreement, but given the impact of the two different forcing data sets, I'm a bit hesitant to jump to that conclusion. Further: Could you maybe indicate the number of observations here just to be clear. The scatterplot also doesn't give us an indication of the point or location of the measurements. Generally, I would have preferred to clearly see the points on a map maybe with a highlighting of the calibration sites and also evaluate how you chose the calibration measurement. For the ablation stakes you use less than 10% to calibrate - where does this number come from?

Figure 8: Depending on the direction of the analysis, the decadal analysis could be further complemented with monthly or seasonal plots.

Table 4: Here annual trends, but earlier figure showed decadal MB, would be interesting to see if there are certain breakpoints in the timeseries corresponding to external forcing (disturbances in atmospheric forcing etc.)?

L371: Interesting that you have decreasing RH but increasing precipitation? Does need explanation?

L396: the correspondence in Fig 13 seems really good to me, perhaps comment in line here that these biases are accompanying a generally good agreement, as evidenced in the figure? Is this match in line with expectations from modelled densification?

L403: Do you return to explain this cooling trend in the discussion?

Figure 11 and 12 could be a bit larger.

L421: Can you add the value you find in the text here to facilitate comparison with the other studies for which you do quote the values.

Figure 15: Its cool to see these data which have not been widely included in previous Abramov studies. What is the added value of this study compared to the others, how do the different studies compare to each other? (Please check the data of the Barandun paper - its labelled 2015 in legend and 2021 in caption). This should be mentioned in the discussion. For example, due to the updating of the glacier extent and glacier elevation, the results of this study are more positive compared to the assumption of static glacier extents in the other studies? You could also include your model simulation results from the recent period and the more recent mass balances studies to complete the picture (e.g. given in Table 6 Barandun et al., 2018 and Barandun et al., 2021 Zenodo data set). This would also open the possibility to compare your modelled results to geodetic mass balances for the glacier available e.g. Hugonnet et al, 2021 and/or Miles et al., 2021 which includes an ASTER based elevation change rate based on results of Brun et al., 2017 (<https://doi.org/10.5281/zenodo.3843292>).

L472: Move these first 2 sentences to the conclusions, as they seem a nice summary of what you have already said in this discussion section.

L485: This statement of performance is related to the validation against point measurements being better for the main run than the alternative run? Where do we see this in the results shown? Or is it just in the supplement. Suggest moving this statement to the part on uncertainty due to forcing data as it deserves more full discussion.

L490: Ideally this section would provide a formal uncertainty assessment. In your situation, as there are many unquantified errors it would be good to include a sensitivity analysis. If this is at all feasible, we strongly encourage you to add this, otherwise please justify why not. You can then describe your approach at end of this paragraph: " As the sources of uncertainty are diverse and sometimes not readily quantifiable, we here discuss the general sources and likely implications, rather than producing a formal uncertainty assessment for the modelled output."

L512: What is the implication of these large differences? Does that mean that the more recent timeframe is wrongly estimated? How can this happen if the bias correction is also based on this period? To get an overview over this, plots of the different forcing datasets (at least in Supplement) could help.

L513: Where have you shown that the standard run is 'better' than the alternative run? Can you here add some quantification on the goodness of fit of the main and alternative run? For example in Figure S5 it would be good to see a comparisons with the actual measured MB for this period and report on which forcing data version matches most closely This is of major importance as, if the alternative forcing produces a very poor match, then it calls into question the validity of the simulated mass balance from 1998 onwards, right? Also consider if this is not better in the main paper rather than the supplement?

### 3 References

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# Long-term firn and mass balance modelling for Abramov glacier, Pamir Alay

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**Abstract.** Several regional studies identified heterogeneous mass changes in western High Mountain Asia over the last decade. Causes for these mass change patterns are still not fully understood. Modelling the physical interactions between glacier surface and atmosphere over several decades can provide insight into relevant processes. Such model applications, however, have data needs which are usually not met in these data scarce regions. Unique glaciological and meteorological data exist for the Abramov glacier in the Pamir Alay range. In this study, we use weather station measurements in combination with downscaled reanalysis data to force a coupled surface energy balance–multilayer subsurface model for Abramov glacier for 52 years. Available *in situ* data are used for model calibration and validation. We find an overall negative mass balance of  $-0.27 \text{ m.w.e. a}^{-1}$  for 1968/1969–2019/2020 and a loss of firn pore space causing a reduction of internal accumulation. Despite increasing air temperatures, we do not find an acceleration of glacier-wide mass loss over time. Such an acceleration is compensated by increasing precipitation rates ( $+0.0022 \text{ m.w.e. a}^{-1}$ , significant at a 90% confidence level). Our results indicate a significant correlation between annual mass balance and precipitation ( $R^2=0.72$ ).

## 1 Introduction

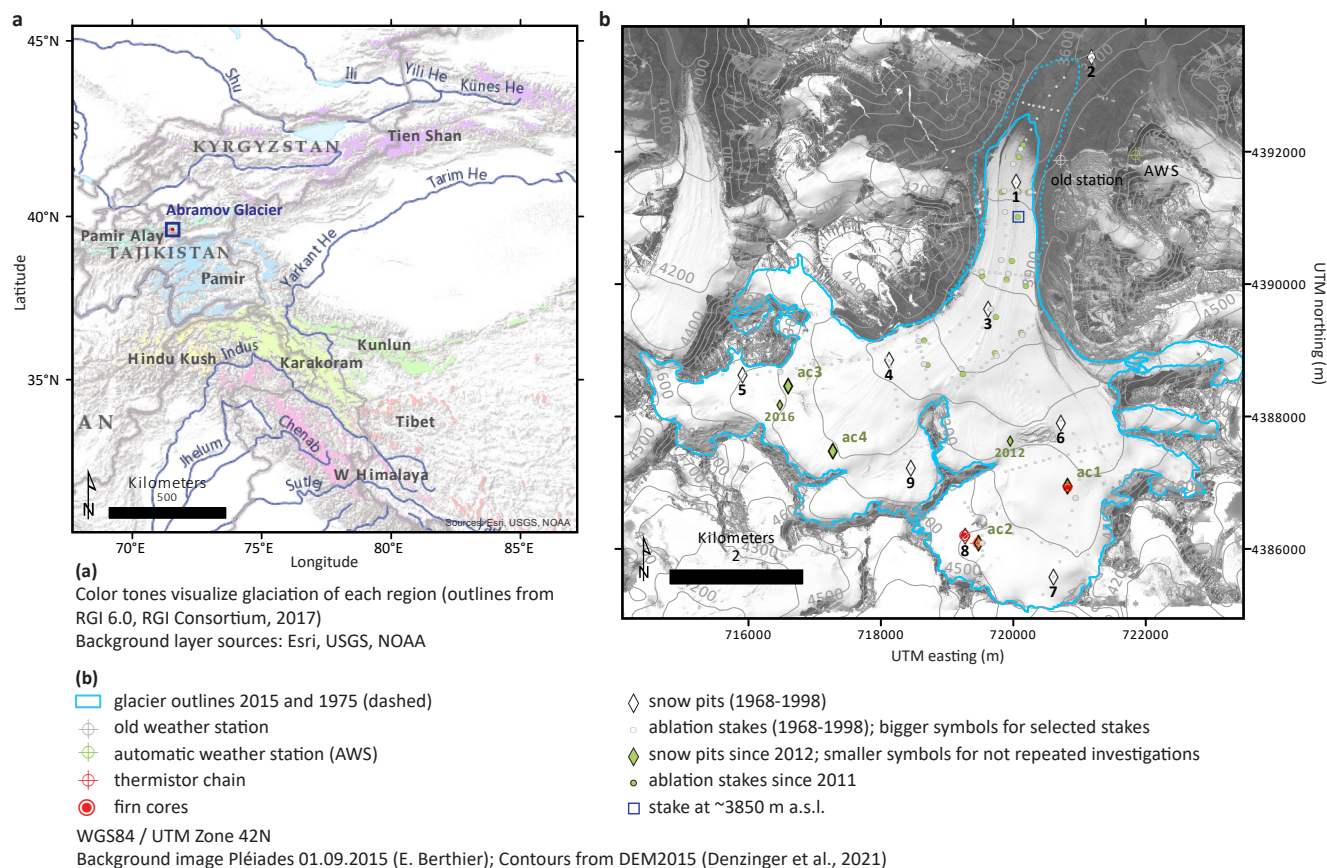
Spatially heterogeneous mass changes of glaciers in High Mountain Asia (HMA) during the last decade have been detected by several regional studies (e.g. Kääb et al., 2012; Brun et al., 2017; Shean et al., 2020; Jakob et al., 2021). Topographical effects in combination with precipitation increases are suggested as reasons for balanced or positive mass changes for numerous glaciers in the Karakoram, Kunlun Shan, Pamir, Pamir Alay, and Tibetan Plateau subregions (Miles et al., 2021). Whereas reliable precipitation data from *in situ* measurements are very scarce for the region (Pohl et al., 2015), the analysis of the gridded Global Precipitation Climatology Project over thirty years has indicated a precipitation increase in the Western part of HMA due to large-scale atmospheric circulation patterns (Yao et al., 2012). Based on regional climate model data, glacier modelling and moisture tracking, De Kok et al. (2020) conclude that changes in irrigation patterns and climate are responsible for the identified mass balance patterns in HMA.



Including *in situ* data and investigating processes at a local scale over several decades can be helpful in better understanding the influence of atmospheric conditions on glacier mass changes (Mölg et al., 2012; Zhu et al., 2020). Mass balance models of varying complexity have been applied to investigate the mass balance response of mountain glaciers to climate (e.g. Klok and Oerlemans, 2002; Pellicciotti et al., 2009; Sicart et al., 2011). Models solving the energy balance at the glacier surface are more physically based and therefore considered more suitable for longer time periods than temperature-index parametrisations (Hock and Holmgren, 2005). Energy balance models are, however, only applicable if sufficient data are available to generate a complete climate forcing and to calibrate uncertain model parameters. Important processes in the accumulation zone, which acts as a buffer against mass loss due to refreezing and water storage, are not included into surface energy balance models. Several studies have applied energy balance models coupled to multi-layer snow models to simulate refreezing processes within the snow and firn as well as heat conduction which are relevant for the glacier mass and energy balance (e.g. Reijmer and Hock, 2008; Huintjes et al., 2015b). Simulating the physical connection between the atmosphere and the glacier provides insights into the climatic control of glacier mass gain or loss (Mölg and Hardy, 2004).

~~In the overall data scarce HMA, only a few glaciers have been modelled with (distributed) energy balance models which have high data requirements and even fewer studies have applied coupled energy balance subsurface models.~~ The availability of historical data for the Abramov glacier located in the Pamir Alay provides a unique opportunity for detailed modelling over longer periods than previous studies did for HMA glaciers: Kayastha et al. (1999) applied a point-scale energy balance model to Glacier AXOIO in the Nepalese Himalaya. Azam et al. (2014) modelled the point energy balance of Chhota Shigri Glacier glacier in the Western Himalaya. Several studies focus on glaciers and ice caps located on the Tibetan Plateau (Mölg et al., 2012; Zhang et al., 2013; Huintjes et al., 2015b, a, 2016). Zhu et al. (2020) applied a surface energy balance model to Muji Glacier, located in the north-eastern Pamir. Except for Huintjes et al. (2016), who used a coupled snowpack and ice surface energy and mass balance model (COSIMA) to reconstruct the climate on the Tibetan Plateau during the little ice age, only relative short periods up to one decade have been investigated.

Here, we apply a coupled surface energy balance – multilayer subsurface model (van Pelt et al., 2012, 2019) for a glacier located in the western part of HMA. We simulate 52 years of firn and mass balance evolution of the Abramov glacier which is located in the Pamir Alay region, Kyrgyzstan (Fig. 1). For this temperate valley-type glacier, detailed glaciological as well as meteorological measurements exist (Kislov, 1982; Pertziger, 1996; Schöne et al., 2013; Hoelzle et al., 2017; Kronenberg et al., 2021a). ~~These *in situ* data, which are unique for the region, allow us to apply a rather complex model with relatively high data requirements.~~ The overall aim of this study is to model the energy and mass fluxes for a period of five decades for Abramov glacier in order to better understand its firn and mass balance evolution as a response to climatic conditions. ~~According to our knowledge, this study represents the first long-term application of a model with this degree of complexity for this data scarce region.~~



**Figure 1.** (a) Overview map showing the location of Abramov Glacier in the Pamir Alay (indicated in blue). (b) Map of Abramov Glacier and surroundings showing the location of weather stations and the mass balance observation network. (Own maps; layer sources are indicated in the legend)

## 2 Study site and data

### 2.1 Abramov glacier

55 Abramov glacier ( $39.50^{\circ}\text{N}$ ,  $71.55^{\circ}\text{E}$ ) is located in the Pamir Alay (north-western Pamir) in Kyrgyzstan (Fig. 1). The north facing valley-type glacier spans an elevation range from 3650-5000 m a.s.l. and covers an area of  $24\text{ km}^2$  (in 2015). Since 1975, the glacier retreated by about 1 km (Barandun et al., 2015). ~~An advance of the glacier tongue was reported for~~ 1972 and 1973 (e.g. Glazyrin et al., 1993). Mean annual air temperature (1968-1998) measured at 3837 m a.s.l. next to the glacier tongue was  $-4.1^{\circ}\text{C}$  and annual precipitation sums for the same period are  $750\text{ mm a}^{-1}$ . Abramov glacier has temperate firn conditions

60 (Kislov, 1977) and cold subsurface conditions were measured in the ablation area throughout the year (Kislov et al., 1977). Liquid water was observed at depths of about 8.5 to 13 m around 4400 m a.s.l. (Glazyrin et al., 1977; Kislov, 1982). The glacier



~~had one of the most detailed and comprehensive mass-balance time series that ended abruptly in 1999. Thanks to international efforts, mass-balance measurements were~~ re-initiated in 2011 (Hoelzle et al., 2017).

Barandun et al. (2015) ~~have~~ previously calculated the long term mass balance of Abramov glacier by applying a mass balance model and calibrating it to *in situ* measurements. ~~Their model is based on a simplified energy balance approach by Oerlemans (2001).~~ Denzinger et al. (2021) computed the geodetic mass balance based on aerial imagery from 1975 and satellite stereo pairs from 2015. Both studies found an overall negative mass balance. According to Barandun et al. (2021), Abramov glacier has a similar mass balance ~~than~~ other glaciers in the Pamir Alay. ~~Based on *in situ* data,~~ Kronenberg et al. (2021a) identified a trend towards a net accumulation increase for a point site on Abramov glacier, suggesting the accumulation regime ~~changed~~ since the 1970s.

## 2.2 Topographical data

In this study we use digital elevation models (DEM) with an initial resolution of 4 m. ~~The DEMs are derived from historical aerial photographs from 1975 (DEM1975) and from three Pléiades stereo pairs from 2015 (DEM2015) provided by Denzinger et al. (2021). Glacier outlines are available for 1975 and 2015 from Denzinger et al. (2021) and for 1986, 1998 and 2005 from~~ Barandun et al. (2015). For the DEM2015, Stainbank (2018) manually removed artefacts and created a gap-filled DEM with 25 m resolution which we use as a reference DEM. Furthermore, we use the 1975 glacier area as the reference glacier surface. We use later glacier extents (one per decade) to estimate the glacier area change (reduction with respect to 1975) over time. Based on these data we calculate an annual height change grid, an elevation grid for 1968 (DEM1968) assuming the same linear elevation trend as between 1975 and 2015, and a reference glacier mask (mask1975) with a resolution of 100 m and an extent of 10.7 km × 10.7 km (107×107 grid points).

## 2.3 Weather station data

### 2.3.1 Original weather station data

The original Abramov glacier weather station was located on a moraine next to the glacier tongue at 3837 m a.s.l. and was operational from October 1967 until summer 1999. The meteorological data are published in Pertziger (1996) who also compiled the data in digital format (daily resolution). Here, we use data from January 1968 until December 1998 for the following parameters: daily average, minimum and maximum air temperature, air pressure (daily averages), relative humidity (daily averages), wind speed (daily averages), precipitation (daily sums) and cloud cover observations (daily minimum and daily average). Most meteorological parameters were recorded eight times a day, however, ~~to our knowledge, these raw data have not been archived and~~ only daily values were available for this study. Precipitation measurements were not corrected for undercatch. More details about the station data can be found in the supplementary material (Table S1).

Based on these data, we created three-hourly value for each variable. We use a scaled sine function to calculate three-hourly air temperatures. The scaling factor is determined for each day based on measured minimum and maximum temperatures and daily averages of the three-hourly time series correspond to reported average air temperature measurements. For air pressure



and relative humidity the mean value is applied throughout the daily cycle. Daily precipitation sums are divided by eight to obtain three-hourly data. During the melt season, convection is a main driver of cloudiness and cloud formation mainly takes place along the mountain ridges (Suslov and Krenke, 1980). Consequently, we assume cloud cover to be lower in the morning hours and lower for large areas of lower parts of the glacier than observed averages. We assign observed daily minimum cloud cover to the first four timesteps and daily average cloud cover for the rest of the day.

### 2.3.2 Automatic weather station data

Since August 2011, an automatic weather station (AWS) is operating at 4100 m a.s.l. about 1 km from the glacier tongue (Fig. 1) and the site of the historical meteorological observations. ~~Please refer to~~ Schöne et al. (2013) for technical information including sensor specifications. Long data gaps and outliers are abundant during the first operational year and we therefore only use data ~~recorded~~ from January 2014 onwards. In contrast to the original Abramov glacier weather station, only liquid precipitation is measured at the station, which is very rare even in summer. Hence, no precipitation data are available. Here, we use air temperature, air pressure, relative humidity, and incoming short- and longwave radiation.

We fix a clock error and apply filtering (as described below) to the raw measurements which are registered 12 times an hour before calculating three-hourly and daily averages of several parameters as described in Stainbank (2018). At least 18 measurements are necessary (50 % of the total measurements) for three-hourly averages, otherwise we set the value to Not a Number (NaN). Daily averages are only calculated if data is available for each three-hourly time step of a day.

For air temperature ~~and~~ pressure we first remove extreme outliers by the application of an inter quartile range filter. Second, we calculate three-hourly mean and standard deviation for the entire time series and remove values that are more than three standard deviations from the three-hourly mean. A visual inspection of the time series indicated that for a few short periods, some outliers were not detected by the automatic filters and therefore removed manually before rerunning the second filter. For relative humidity we applied an upper (100 %) and a lower threshold (0 %), setting values below or above to NaN. The four component net radiation sensor delivers incoming short- and longwave radiation which are used here. Furthermore, the measurements may be affected by sensor freezing. We do not specifically correct for these effects, but apply some general filters. Incoming shortwave radiation measurements below  $0 \text{ W m}^{-2}$  are set to  $0 \text{ W m}^{-2}$  and measurements above  $1500 \text{ W m}^{-2}$  are excluded. ~~Incoming longwave radiation is processed using the same approach as air temperature and pressure.~~

## 2.4 Gridded climate data

### 2.4.1 TopoSCALE ERA5 data

We use hourly output from ERA5 reanalysis by ECMWF Hersbach et al. (2020) for the grid point located nearest to Abramov glacier for 1980-2020. Data from the two Abramov weather stations is completely independent from the ERA5 data set, as the stations are not used during the assimilation procedure (Personal communication from H. Hersbach, ECMWF 2021). Air temperature, air pressure, relative humidity, precipitation, global and clear sky radiation are downscaled using TopoSCALE





125 (Fiddes and Gruber, 2014) which performs a 3D interpolation of upper air fields stored on pressure levels and corrects radiation fields according to topographical setting (e.g. slope, aspect, shading, sky view factor).

#### 2.4.2 Cloud fraction calculated from TopoSCALE ERA5 data

Cloud cover fraction is not directly available from ERA5. We therefore, calculate the TopoSCALE cloud fraction from the TopoSCALE global  $I$  and the clear-sky radiation  $I_{cs}$ . In a first step, the cloud transmissivity  $\tau_{cl}$  is calculated following Klok  
 130 and Oerlemans (2002):

$$\tau_{cl} = I/I_{cs} \quad (1)$$

We use the three-hourly average values of  $I$  and  $I_{cs}$  to calculate  $\tau_{cl}$  for each time step for which  $I$  and  $I_{cl}$  are both above 0. We use the expression by Greuell et al. (1997):

$$\tau_{cl} = 1.00 - c1 \times n - c2 \times n^2 \quad (2)$$

135 to calculate the cloud cover  $n$  from the cloud transmissivity  $\tau_{cl}$ . The parameters  $c1=0.128$  and  $c2=0.346$  are adopted from van Pelt et al. (2012). Night time cloud cover  $n$  values are linearly interpolated from neighboring, non-missing values.

#### 2.4.3 Bias correction of TopoSCALE ERA5 data and creation of a continuous data set

ERA5 is a global reanalysis product and its quality is dependent on density of assimilated observations, which varies globally. Central Asia in general and mountainous regions of the Pamir specifically, are relatively data poor and therefore the reanalysis  
 140 is less well constrained as compared to data rich regions such as Europe, for example. Therefore, even after downscaling (accounting for resolution differences between the model grid and point of interest) we can expect there to be residual biases, which we address with the following procedure. We aggregate TopoSCALE data to monthly averages to compare them to data from the original weather station. Biases are calculated for monthly air temperature, pressure and wind speed for the period 1980-1998 and then used to correct TopoSCALE air temperature, pressure and wind speed for 1980-2020. Monthly  
 145 average ratios between monthly aggregated station measurements and TopoSCALE data are calculated for precipitation (sums) and cloud fraction (averages) and relative humidity (averages) for the period 1980-1998 and used to correct the TopoSCALE time series for 1980-2020. The resulting cloud fraction time series for summer months (July to September) shows a reduced amplitude compared to the station time series for 1968-98. We use the precipitation time series to correct for this by setting the cloud cover to 0 for days without precipitation and to 1 if precipitation is above a threshold value. Monthly averages of the  
 150 final cloud fraction time series correspond to observed values for the years for which measurements are available. We combine the three-hourly observed data from 1968 to 1998 with the bias corrected TopoSCALE ERA5 data for the years 1999-2020 to obtain a final data set for 1968-2020. An alternative data set is created using historical measurements for the period 1968-1979 and TopoSCALE ERA5 data for the years 1980-2020. The both data sets thus differ for the period 1980-1998.



## 2.5 Glaciological data

### 155 2.5.1 Monthly point mass balance data 1967-1998

Monthly mass balance was measured on Abramov glacier from the beginning of the hydrological year 1967/1968 until summer 1999. Winter and annual mass balance data are published in Pertziger (1996) who ~~also~~ compiled the data in digital format (monthly resolution, totalling 42961 stake and 2179 snow pit measurements). Observations are reported with local coordinates, which we transferred to WGS84 / UTM Zone 42N as described in Kronenberg et al. (2021a). Monthly snow height and density  
 160 measurements respective to the previous summer horizon are available from up to nine snow pit locations. Eight snow pits are located on the glacier surface and one snow pit is located just below the glacier tongue in the Koksuy valley (Fig. 1b). A total of 165 stakes installed along 13 profiles were used to measure monthly surface height changes since the beginning of the hydrological year (1 October). For stakes in the ablation area, negative height changes are converted to water equivalents (w.e.) using a density of ice  $900 \text{ kg m}^{-3}$ . For stakes above the firn line, a firn density of  $610 \text{ kg m}^{-3}$  is used (Kislov, 1977; Pertziger,  
 165 1996). Here, we use a constant elevation threshold of 4200 m a.s.l. to select either the firn or ice density for conversion into ablation in w.e. Pertziger (1996) converted positive stake readings to w.e. either using snow densities measured at nearby snow pits or a density of  $550 \text{ kg m}^{-3}$ . If measured densities were high, they assumed a maximum density of  $610 \text{ kg m}^{-3}$ . Here, we consistently use a density of  $550 \text{ kg m}^{-3}$  to convert positive stake readings to w.e. According to Pertziger (1996), the main source of error in mass balance point measurements are errors in density estimates. Another source of uncertainty is related  
 170 to the time of observation. Exact dates are reported for stake readings. For data from snow pits, we do not know the exact measurement dates as only months are reported. Furthermore, the reference date or exact age of the reference surface in the snow pit are unknown due to the stratigraphic nature of snow pit measurements. We assume these dates to correspond to the beginning of the hydrological year until the end of the investigated month.

We mainly use snow pit measurements from March and annual mass balance measurements from September including data  
 175 from up to eight snow pits and 165 stakes per year for 1967/1968-1997/1998. A set of 19 stakes is selected based on the location of current point observations (see following paragraph). Data from the remaining 146 stakes are used separately for validation.

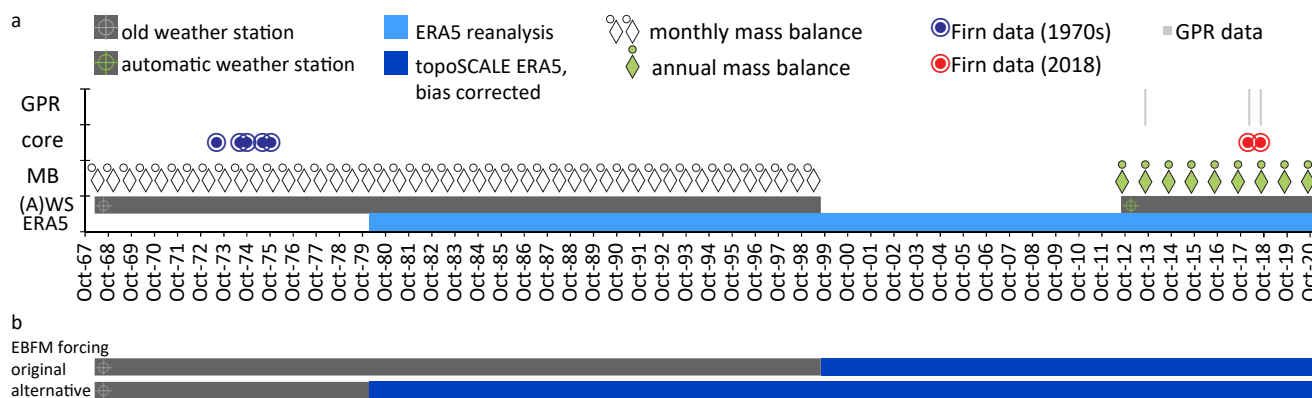
### 2.5.2 Annual point mass balance data 2011-2020

Annual mass balance measurements from up to 20 points on the glacier are available since 2011/2012 (Barandun et al., 2015; Hoelzle et al., 2017). The new monitoring network consists of 16 stakes in the ablation area and up to four snow pits in the  
 180 accumulation area (Fig. 1). Whereas the stake locations were roughly kept constant over time, the locations of the snow pits varied during the first years. Field visits took place once a year during July/August and exact observation dates are available. Only ablation stake readings from snow free locations are considered and the density of ice ( $900 \text{ kg m}^{-3}$ ) is used to calculate floating-date mass balances between the two observation dates. Due to the stratigraphic nature of snow pit observations, the beginning of the accumulation season is not known and we assume the last summer surface to have formed at the beginning  
 185 of the hydrological year. Point accumulation values are thus calculated for the period from the 1 October until the observation date using the snow density and depth measurement of each snow pit.



## 2.6 Firn profile data

Density profiles dating back to the 1970s are available from deep firn pits that were located in the eastern branch of the accumulation area at  $\sim 4400$  m a.s.l. and  $\sim 4250$  m a.s.l. (Kislov, 1982). In 2018, firn cores were drilled at similar and nearby locations. We refer to Kronenberg et al. (2021a) for a detailed description of legacy and recent firn profiles. In addition to density measurements, continuous subsurface temperature measurements from four thermistors located in a borehole at  $\sim 4380$  m a.s.l. are available from February 2018 until April 2020.



**Figure 2.** Temporal overview of data available for this study. The availability of different *in situ* and gridded datasets (ERA5 reanalysis) is shown in a. Panel b summarizes which data are used for forcing the model. We mainly show results using the original forcing. A sensitivity model run was performed using the alternative forcing. Results are shown in the supplementary material

## 3 Energy balance and firn model (EBFM)

We simulate the mass balance evolution of Abramov glacier using a coupled surface energy balance–multilayer subsurface model by van Pelt et al. (2012). The surface energy balance model was developed following Klok and Oerlemans (2002) and the subsurface model based on SOMARS by Greuell and Konzmann (1994). The original model was first employed to simulate the mass balance of Nordenskiöldbreen located on Svalbard. A parameterized water percolation routine was introduced by Marchenko et al. (2017) and the albedo decay scheme was updated based on the work by Bougamont et al. (2005). The model has participated in the firn meltwater Retention Model Intercomparison Project (RetMIP) as "UppsalaUniDeepPerc" (Vandecrux et al., 2020) and it has been applied to several other glaciers located in the Arctic and the Alps (e.g. van Pelt and Kohler, 2015; van Pelt et al., 2021; Mattea et al., 2021). Here, we use the model version described by van Pelt et al. (2019) and we include an updated parametrisation for seasonal snow densification after van Kampenhout et al. (2017). As the model has been previously described in detail, only a short model overview is given in the following. For full details the reader is referred to van Pelt et al. (2019) and preceding studies.





### 205 3.1 Surface energy balance model

The surface energy balance model is forced by meteorological input data and calculates the energy fluxes which contribute to the surface energy budget:

$$Q_{melt} = SW_{net} + LW_{net} + Q_{sens} + Q_{lat} + Q_{sub}, \quad (3)$$

with the total energy available for melting  $Q_{melt}$ , the net shortwave radiation  $SW_{net}$ , the net longwave radiation  $LW_{net}$ ,  
 210 the turbulent sensible and latent heat fluxes  $Q_{sens}$  and  $Q_{lat}$  and the heat flux into the subsurface  $Q_{sub}$ . The model iteratively solves Eq. 3 to find the surface temperature, which is the only unknown and cannot be larger than  $0^{\circ}\text{C}$ .

Surface ablation either occurs in form of melt or as sublimation when the latent heat flux is negative. Surface accumulation occurs either in form of solid precipitation, which is calculated based on air temperature and precipitation forcing using a transition from rain to snow around a threshold temperature ( $T_{sr} \pm 1 \text{ K}$ ), or in form of riming. In case of a snow or firn surface,  
 215 liquid water originating from surface melt, rain or condensation is added to the subsurface (see below) and in case of an ice surface leaving the system as runoff. For the entire modelling period, the incoming shortwave radiation  $SW_{in}$  is simulated as described in Klok and Oerlemans (2002) and van Pelt et al. (2012) and accounts for grid aspect and shading by surrounding terrain. The attenuation by clouds ( $\tau_{cl}$ ) is calculated using Eq. 2 with parameters determined for Nordenskiöldbreen, Svalbard, by van Pelt et al. (2012). We also tested the parameter values determined by Greuell et al. (1997) for an Alpine glacier. Using  
 220 the values determined for Svalbard, however, yielded a better agreement between modelled  $SW_{in}$  and AWS measurements. The outgoing shortwave radiation  $SW_{out}$  depends on the surface albedo. In case snow is present, the albedo  $\alpha_{snow}$  is a function of the snow temperature, wetness and age, as described in van Pelt et al. (2019). In absence of snow the ice albedo  $\alpha_{ice}$  applies. The computation of the longwave radiation follows the Stefan-Boltzmann law. The sky emissivity depends on cloud cover, air temperature and humidity following relations in Konzelmann et al. (1994) and Greuell and Konzelmann (1994).

225 The turbulent sensible and latent heat fluxes  $Q_{sens}$  and  $Q_{lat}$  are calculated for large-scale atmospheric conditions following Oerlemans and Grisogono (2002). The equations are available in Klok and Oerlemans (2002). The subsurface heat flux  $Q_{sub}$  is calculated based on the modelled temperature and conductivity of the subsurface. To obtain  $Q_{sub}$ , a linear gradient through the two uppermost layers is applied (see below). The heat supplied by liquid precipitation is neglected.

### 3.2 Subsurface model

230 The subsurface model computes the temporal evolution of subsurface temperature, density and liquid water content for discrete layers. The temperature depends on heat conduction (vertical diffusive heat flux) and refreezing of percolating melt, rain water and condensed moisture (van Pelt et al., 2012). The expressions for heat capacity and effective conductivity are adopted from Sturm et al. (1997) and Yen (1981). Heat (and mass) advection is accounted for by describing vertical layer movement on a Lagrangian grid. The penetration of shortwave radiation into the subsurface and therefore also subsurface melting are neglected.  
 235 We use an updated subsurface densification routine compared to previous EBFM applications. For layer densities above  $\rho_{firn}$



$= 500 \text{ kg m}^{-3}$ , the in van Pelt et al. (2012) described parametrisation based on the gravitational densification by Arthern et al. (2010) and modified after Ligtenberg et al. (2011) is applied. For layers with a density below  $\rho_{firm} = 500 \text{ kg m}^{-3}$  a newly introduced fresh snow densification parametrisation following van Kampenhout et al. (2017) is used. Seasonal snow densifies due to destructive metamorphism and compaction by overburden pressure. Van Kampenhout et al. (2017) furthermore include a  
 240 snow densification due to drifting snow for densities below  $350 \text{ kg m}^{-3}$ . Snow drift is not included here, as we are not focusing on dynamics of fresh snow. In order to increase numerical efficiency, the fresh snow  $\rho_{fresh}$  density is set to  $350 \text{ kg m}^{-3}$ . The densification due to destructive snow metamorphism is depth independent but varies according to the layer temperature  $T$ :

$$\frac{\partial \rho}{\partial t} = c_{dm3} c_{dm2} c_{dm1} \exp(-c_{dm4}(T_0 - T)), \quad (4)$$

with the constants  $c_{dm3} = 2.777 \times 10^{-6} \text{ s}^{-1}$ ,  $c_{dm4} = 0.04 \text{ K}^{-1}$ ,  $c_{dm2} = 1$  ( $c_{dm2} = 2$ ) in case of absence (presence) of liquid water  
 245 and a tapering constant  $c_{dm1} = 1$  in the range of  $\rho \in [0, \rho_{max}]$  and exponentially decreasing above  $\rho_{max}$ . The densification due to overburden pressure  $P$  ( $\text{kg m}^{-3}$ ) depends on a viscosity coefficient  $\eta$  ( $\text{kg s}^{-1} \text{ m}^{-3}$ )

$$\frac{\partial \rho}{\partial t} = \frac{P}{\eta}, \quad (5)$$

The viscosity coefficient  $\eta$  is calculated following van Kampenhout et al. (2017) who slightly simplified the expression of Vionnet et al. (2012):

$$250 \quad \eta = f_1 \times 4 \times \eta_0 \frac{\rho}{c_\eta} \exp(a_\eta(T_0 - T) + b_\eta \rho), \quad (6)$$

with a correction factor accounting for the liquid water content  $f_1$ ,  $\eta_0 = 7.62237 \times 10^6 \text{ kg s}^{-1} \text{ m}^{-3}$ ,  $a_\eta = 0.1 \text{ K}^{-1}$ ,  $b_\eta = 0.023 \text{ m}^3 \text{ kg}^{-1}$  and  $c_\eta = 358 \text{ kg m}^{-3}$ .

The subsurface water within the firm column originates from percolating melt water, rain or condensed moisture. Preferential percolation is parametrised as described by Marchenko et al. (2017). Liquid water is instantly distributed along the depth axis  
 255 following a normal distribution until a maximum depth  $z_{lim}$  unless it reaches an impermeable ice layer before. Refreezing of percolating water raises subsurface temperatures and densities until the melting point or the density of ice is reached. A small amount of liquid water is stored as irreducible water held by capillary forces and the remaining water percolates further downwards until an impermeable ice layer is reached where it forms a slush layer. The maximum irreducible water content is calculated from the porosity (ratio between pore space and the total volume of the snow/firm layer) following Schneider and  
 260 Jansson (2004). Below  $z_{lim}$ , percolation occurs non-preferentially following a bucket-scheme. Water moves to the next underlying layer if the refreezing capacity is eliminated (layer density or temperature reach density of ice or melting temperature) and the maximum irreducible water content is reached. Surface runoff happens instantaneous and occurs when bare ice is at the surface.

### 3.3 Model set up and initialisation

265 The surface energy balance and subsurface profiles are updated with a temporal resolution of three hours. We use a grid of  $107 \times 107$  grid points with a horizontal resolution of 100 m, only 2654 of these grid points are assigned to the glacier using the



glacier outlines from 1975. For the initial elevation information we use the DEM1968 and update elevation information for each 3 h time step based on the linearly downscaled annual height change grid. For each 3-h time step, we derive topographical parameters used for the computation of incoming solar radiation as described in van Pelt et al. (2012). While modelling, the glacier surface is assumed to be constant and glacier grid points are classified as glacier throughout the modelling period. Distributed characteristics are thus computed for a fixed reference glacier surface and spatial varying elevation in time. We later account for a reduction of glacier surface when analysing the model output as described below.

The subsurface modelling domain consists of 100 vertical layers extending down to a depth of about 35 m below the surface in the accumulation area. The initial layer thickness at the surface is 0.1 m and at layer numbers 15, 25 and 35 (corresponding to initial depths of 1.5, 3.5 and 7.5 m). Layer thickness reduces with time due to gravitational densification of snow and firn. The subsurface layers move along the depth axes to respond to mass gain or loss at the surface. Due to accumulation, the thickness of the uppermost layer can increase until a certain threshold thickness is reached, additional accumulation leads to the creation of a new layer and the removal of the lowermost layer of the modelling domain. In the EBFM, horizontal mass and energy fluxes are neglected and mass and energy exchange between grid cells is only possible along the vertical depth axes.

To initialize subsurface conditions the model is run twice over the period 1968-98 using the three-hourly weather station data to force the model. The first iteration is started with identical subsurface conditions throughout the glacier with a vertical grid consisting of temperate ice (273.15 K). The second initialisation run is started from the final stage of a first run.

We perform several model runs for selected grid points corresponding to the locations of selected ablation stakes and snow pits indicated in Fig. 1 to adjust model parameters (as described below) using the final combined forcing consisting of station (1968-1998) and bias corrected TopoSCALE ERA5 data (1999-2020). Thereafter, we perform a final distributed model run for the period from 1 January 1968 until 31 December 2020 using the same forcing. An additional distributed run is performed using the alternative data set with a shorter period of station measurements (1968-1979) to assess the model forcing sensitivity.

### 3.4 Analysis of model output and mass balance calculation

We calculate the climatic mass balance as the sum of the surface and the internal mass balance for hydrological years (1 October-30 September) (Cogley et al., 2011). The surface mass balance is the result of accumulation (+) and ablation (-) at the surface including precipitation (+), moisture exchange (+/-), mass loss through runoff (-) and refreezing above the previous summer surface (+). The internal mass balance accounts for re-freezing and storage of liquid water below the previous summer surface.

While the model grid elevation is updated for each time step, the modelling extent is kept constant using the glacier mask from 1975 for the entire simulation period. After modelling, the glacier wide mass balance and other results are calculated for decade-wise updated glacier extents. Until the end of the hydrological year 1978, the entire model output is analysed corresponding to the glacier area of 1975. For the next ten hydrological years, a glacier mask corresponding to the glacier area from 1986 is used and output outside this domain is not considered. Masks based on outlines from 1998, 2005 and 2015 are used for 1988/1989-1997/1998, 1998/1999-2007/2008 and 2008/2009-2019/2020.



300 The equilibrium line altitude ELA is calculated as the mean elevation of grid points with a mass balance equal to 0 m w.e. at the end of a the hydrological year. Grid points with a negative mass balance value at the end of the hydrological year are used to calculate ablation gradients, which is the linear relation between elevation and modelled ablation.

For comparison to *in situ* point measurements, the modelled surface mass balance of the grid point nearest to the stake/snow pit location is used. The daily model output is aggregated at the end of the month of interest with respect to the beginning  
 305 of the hydrological year (1 October) to compare to accumulation observations. For comparison to ablation observations, data is accumulated until the reported observation date (section 2.5.1). For *in situ* observations since 2011, the stake installation dates are earlier in the melt season and model output is extracted for exact periods between stake installation and stake reading (section 2.5.2).

### 3.5 Parameter selection and calibration

310 In order to reflect the local conditions of Abramov glacier we adjusted several model parameters based on data available from *in situ* measurements and other studies from HMA. We selected the calibration parameters based on their relevance for our site and the existence of data. The order of calibration is chosen considering the dependence of parameters on calibration of other parameters. First the incoming radiation parameters are estimated. Thereafter, we optimise accumulation parameters and finally parameters affecting summer melt. Model parameters different from values used by van Pelt et al. (2019) are summarized in  
 315 Table 2, the last column indicates whether a parameter was optimised (y/n).

We extrapolate the meteorological forcing over the glacier surface by means of linear and temporally constant elevation gradients. For Abramov glacier, Kislov (1982) reported a temperature lapse rate of  $dT/dz = -0.005 \text{ K m}^{-1}$  which we adopt here. The precipitation lapse rate  $dPrec/dz$  is calculated from March snow pit measurements for 1969-98 (243 points). The monthly mass balance measurements available for 1969-98 indicate that no melt occurred before April and therefore, the  
 320 March snow pit measurements are unaffected by melt. The pressure decay parameter  $dPres$  is calculated from average air pressure measurements at 3837 m a.s.l. (January 1968 - December 1998) and 4100 m a.s.l. (October 2011 - September 2020). The gradient of relative humidity  $dRH/dz = 0.01\% \text{ m}^{-1}$  is adopted from Huintjes et al. (2016), who calculated it from High Asia Refined Analysis data (Maussion et al., 2014) for two Tibetan glaciers. The chosen value is corroborated by measured relative humidity on Abramov glacier for June and July (Suslov and Krenke, 1980). The calculation of the vertical lapse rate  
 325 of potential temperature  $\gamma$  is adopted from Stainbank (2018).

*In situ* measurements from the AWS for 2014-2020 as well as literature values are used to estimate incoming radiation parameters. The aerosol transmission constant  $k_{aer}$  in the  $SW_{in}$  formulation is determined using the linear relation between  $k_{aer}$  and the elevation proposed by Klok and Oerlemans (2002). We estimate  $k_{aer} = 0.98$  for the median glacier elevation (4300 m a.s.l.). The  $LW_{in}$  parameter  $b = 0.43$  is determined based on daily AWS data (T, RH and  $LW_{in}$ ) for manually identified  
 330 clear sky conditions. The value is similar to the value found by Klok and Oerlemans (2002) for Morteratsch glacier. The second longwave radiation parameter  $\epsilon_{cl}$  is set to the default value of 0.96.

We further use *in situ* mass balance measurements from eight snow pits located on the glacier and data from a selection of 19 ablation stakes for the period 1969-1998 (Table 1) for calibration. With these data we manually optimise a set of accumulation



and ablation parameters. Final parameters are determined by minimizing bias between modelled surface mass balances for grid  
 335 cells corresponding to point locations with *in situ* measurements. We use precipitation correction factors in order of debiasing  
 the precipitation forcing. The correction mainly compensates for undercatch, but also accounts for all other biases present in the  
 precipitation forcing. We consider March accumulation measurements to be least affected by melt and use them to calibrate the  
 precipitation bias correction factor  $Prec_{c-w}$  applied for October-June. During summer months, the precipitation undercatch is  
 assumed to be reduced compared to autumn, winter and spring and we adopt a value of  $Prec_{c-s}=1.15$  determined for Alpine  
 340 locations in Switzerland from Sevruck (1985). The bias between modelled and measured June and July snow pit measurements  
 is reduced by a two-parameter exploration for  $\alpha_{fresh}$  and  $t_{wet}^*$ . And finally,  $\alpha_{ice}$  is optimised using September surface mass  
 balance measurements (snow pits and ablation stakes).

Parameters in the subsurface model are default values except for the critical density of destructive metamorphism  $\rho_{max}$   
 which is optimised to obtain a better fit between modelled and measured subsurface densities at  $\sim 4410$  m a.s.l.

**Table 1.** Number of point measurements used for calibration of surface energy balance parameters (cf. Table 2). Accumulation measurements  
 were performed at eight sites on the glacier surface and all available measurements are used for the selected months. We use 19 out of 165  
 ablation stakes for calibration. Stakes are selected to correspond to current observation sites

type	period	use	number of data points
March snow pit measurements	1969-1998	calibration $Prec_{c-w}$	225
June snow pit measurements	1969-1998	calibration $\alpha_{fresh}, t_{wet}^*$	203
July snow pit measurements	1969-1998	calibration $\alpha_{fresh}, t_{wet}^*$	141
annual mass balance measurements	1969-1998	calibration $\alpha_{ice}$	532

## 345 4 Results

### 4.1 Long term mass balance

The distributed mean annual mass balance for 1968/1969-2019/2020 is shown in Fig. 3. ~~Table 3 lists modelled glacier-wide~~  
 climatic mass balances and internal accumulation ~~for different periods; annual climatic mass balances~~ are shown in Fig. 4.  
 Overall, the mass balance of Abramov glacier is negative for the years from 1968/1969 to 2019/2020. We do not find a  
 350 significant trend in annual mass balances ( $+0.0002$  m w.e.  $a^{-1}$ , p-value=0.979). The most negative mass balances are modelled  
 for the beginning of the modelling period. The two decades between 1978 and 1998 are characterised by almost balanced mass  
 budget. Mass balances are more negative after the end of historical investigations in 1998.

The elevation distribution of mass balance ~~for different decades is shown in Fig. 5b.~~ The accumulation is lowest during the  
 first decade (1968/1969-1977/1978) and highest during the last modelled decade (2008/2009-2017/2018). Ablation is highest  
 355 during the first decade, followed by the second last decade (1998/1999-2007/2008).



**Table 2.** List of EBFM parameter choices with references. For optimised parameters (opt), the initial value is given in brackets.

	parameter	unit	value (initial)	reference	opt
$dT/dz$	temperature lapse rate	$K m^{-1}$	-0.005	Kislov (1982)	n
$dPrec/dz$	precipitation lapse rate (factor)	$m^{-1}$	0.0013	calculated from in situ data	n
$dPres$	Pressure decay parameter	-	$-1.45 \cdot 10^{-4}$	calculated from in situ data	n
$dRH/dz$	relative humidity gradient	$\% m^{-1}$	0.01	Huintjes et al. (2016)	n
$\gamma$	potential temperature lapse rate	$K m^{-1}$	0.0055	from in situ and pseudostation data	n
$Prec_{c-w}$	precipitation bias correction winter (factor)	-	1.85 (2)	-	y
$Prec_{c-s}$	precipitation bias correction summer (factor)	-	1.15	Sevruk (1985)	n
$\alpha_{ice}$	ice albedo	-	0.23 (0.3)	Mölg et al. (2012)	y
$\alpha_{fresh}$	fresh snow albedo	-	0.81 (0.85)	Mölg et al. (2012)	y
$t_{wet}^*$	albedo decay time scale wet snow	days	7 (15)	van Pelt et al. (2019)	y
$P_{th}$	snowfall threshold to reset to $\alpha_{fresh}$	$m \text{ w.e. } s^{-1}$	$3.5d-8$	Zhu et al. (2020)	n
$C_b$	background turbulent exchange coefficient	-	0.0037	Klok and Oerlemans (2002)	n
$c1$	cloud transmissivity parameter	-	0.128	van Pelt et al. (2012)	n
$c2$	cloud transmissivity parameter	-	0.346	van Pelt et al. (2012)	n
$k_{aer}$	aerosol transmission constant	-	0.982	Klok and Oerlemans (2002)	n
$b$	clear sky emissivity parameter	-	0.433	calculated from in situ data	n
$\epsilon_{cl}$	overcast emissivity	-	0.960	default value	n
$\lambda$	optical thickness empirical constant	-	3.00	Smith (1966)	n
$z_{lim}$	preferential percolation depth	m	6.0	default value	n
$\rho_{max}$	critical density destructive metamorphism	$kg m^{-3}$	300 (200)	-	y
$\rho_{fresh}$	fresh snow density	$kg m^{-3}$	350	Klok and Oerlemans (2002)	n

## 4.2 Point surface mass balance versus measurements

Figs. 6 and 7 compare modelled surface mass balance to *in situ* measurements. Fig. 6 shows the final calibration results for eight snow pit locations at the end of March (panel a) and annual mass balance measurements from 19 stakes and up to eight snow pit locations (panel b). Both Fig. 6a and b contain data for the period 1968/1969-1997/1998. A different set of point measurements is used for model validation shown in Fig. 7. In Fig. 7a, an additional set of annual mass balance observations from 146 stake locations for 1968/1969-1997/1998 (equal to the calibration period, stakes not used for calibration) is compared to modelled surface mass balances. In Fig. 7b, the model output is compared to mass balance measurements for 2011/2012-2019/2020.



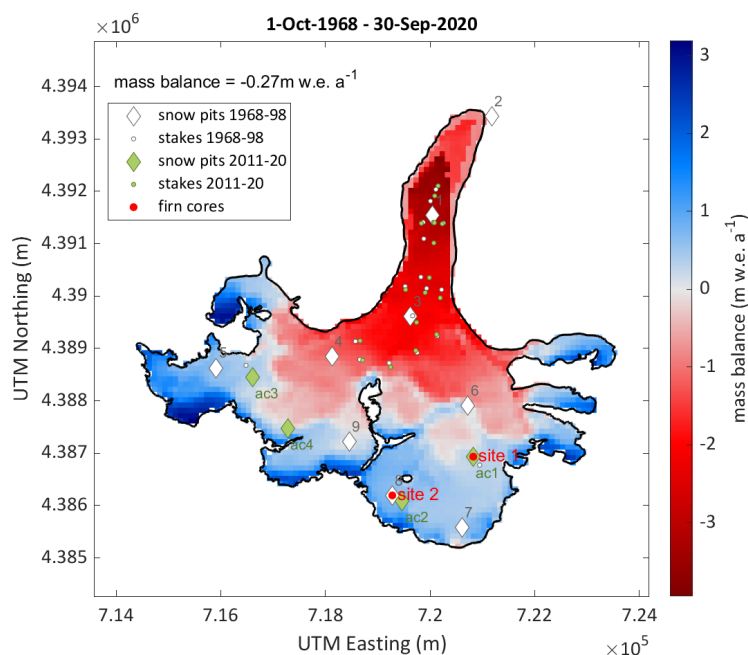
**Table 3.** Mean glacier-wide climatic mass balance, internal accumulation Equilibrium Line Altitude ELA and ablation gradients for each decade and different periods used for comparison with other studies. The periods are hydrological years (e.g. 1968/1969-1977/1978 refers to 1 October 1968 - 30 September 1978) unless precise dates are specified. The mean glacier surface used for the mass balance calculation is also indicated

period	climatic mass balance m w.e. a <sup>-1</sup>	internal accumulation m w.e. a <sup>-1</sup>	ELA m a.s.l.	ablation gradient m w.e. (m a.s.l.) <sup>-1</sup>	mean glacier surface km <sup>2</sup>
1968/1969-1977/1978	-0.63	0.11	4250	0.0077	26.54
1978/1979-1987/1988	-0.04	0.1	4170	0.0078	25.78
1988/1989-1997/1998	-0.08	0.13	4180	0.0083	25.24
1998/1999-2007/2008	-0.33	0.12	4220	0.0076	25.16
2008/2009-2017/2018	-0.23	0.1	4222	0.0066	25.06
1968/1969-1997/1998	-0.25	0.12	4200	0.008	25.85
1968/1969-2013/2014	-0.26	0.12	4207	0.0077	25.6
1971/1972-1993/1994	-0.26	0.11	4200	0.0079	25.87
1975/1976-2014/2015	-0.26	0.11	4208	0.0076	25.42
15 Jul 1975-1 Sep 2015	-0.30	0.11	4208	0.0076	25.42
1998/1999-2019/2020	-0.31	0.11	4225	0.007	25.11
2011/2012-2019/2020	-0.38	0.08	4248	0.0065	25.06
1968/1969-2019/2020	-0.27	0.11	4211	0.0076	25.54

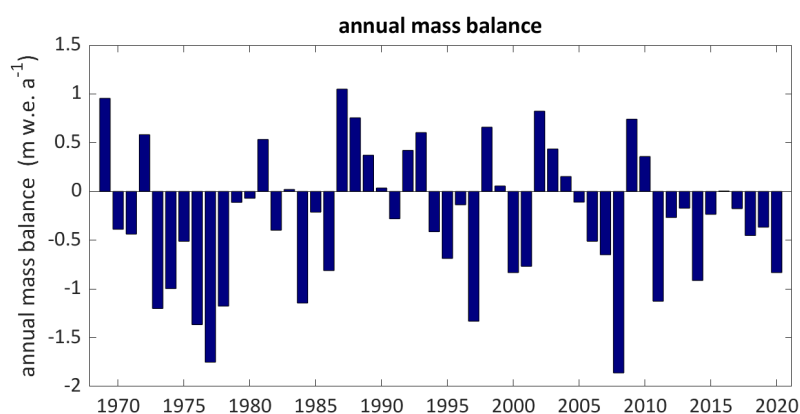
### 4.3 Variation of energy fluxes and climate variables

Figs. 8a and b show mean modelled energy fluxes per decade for a grid point in the ablation area and site 2 in the accumulation area. The only negative heat flux at both sites is the latent heat flux, which becomes less negative over time. It increases for both sites by  $+0.05 \text{ W m}^{-2} \text{ a}^{-1}$  (p-value = 0.000). In the ablation area, the total melt energy is lowest during the second modelled decade (1978/1979-1987/1988, Fig. 8a) and is more homogeneous in the accumulation area (Fig. 8b). In the ablation area, the net radiation flux ( $R_{\text{net}}=27.6 \text{ W m}^{-2}$ ) clearly dominates over the sensible heat flux ( $Q_{\text{sens}}=18.4 \text{ W m}^{-2}$ ) at the beginning of the modelling period. The sensible heat flux increases in the course of time ( $+0.06 \text{ W m}^{-2} \text{ a}^{-1}$ , p-value = 0.001) and almost equals the net radiation during the last modelled decade ( $Q_{\text{sens}}=20 \text{ W m}^{-2}$  and  $R_{\text{net}}=23.8 \text{ W m}^{-2}$ , Fig. 8a). In the accumulation area, the sensible heat flux always dominates over the net radiation and also increases with time by  $+0.04 \text{ W m}^{-2} \text{ a}^{-1}$ , p-value = 0.003 (Fig. 8b). The net radiation mainly varies due to a reduction in incoming shortwave radiation, which is highest at the beginning of the modelling period and decreases over time (ablation area:  $-0.4 \text{ W m}^{-2} \text{ a}^{-1}$  p-value = 0.000; accumulation area:  $-0.3 \text{ W m}^{-2} \text{ a}^{-1}$  p-value = 0.000). At the same time a lower, but significant increase of the incoming longwave radiation is found for the ablation area (Table 4). Mean decadal energy fluxes are given in more detail in Table S2 in the supplementary material.





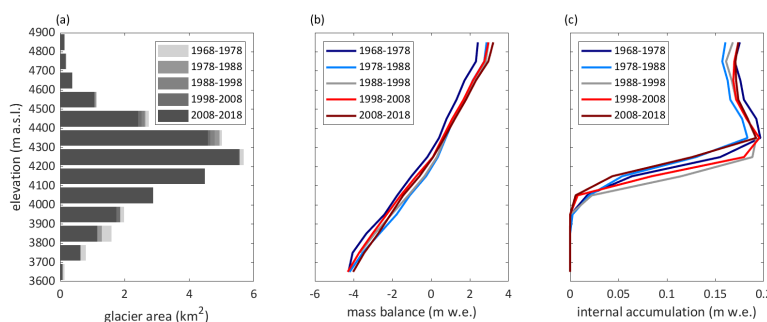
**Figure 3.** Modelled mean annual distributed mass balance for updated glacier extents for the period from 1 October 1968 to 30 September 2020. Note, that the mean annual mass balance for the entire period and updated glacier surfaces is shown. Values are thus reduced on the glacier tongue, where the glacier area reduced over time. Furthermore, the location of point observations used for calibration and validation are indicated with symbols, further details are shown in the legend.



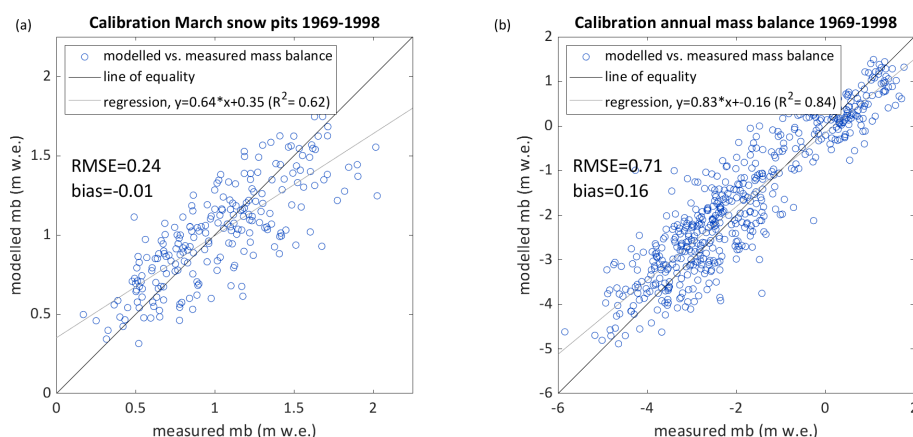
**Figure 4.** Modelled mean annual mass balance for updated glacier extents.

Significant increasing trends for model forcing air temperature (mean over hydrological years) and mean summer air temperatures are found for 1968/1969-2019/2020 (Table 4). Trends are also significant for air pressure (increase), relative humidity (decrease), cloud cover fraction (decrease) and precipitation sums (increase) (Table 4).





**Figure 5.** Modelled mass balance and internal accumulation versus elevation. The elevation distribution of the glacier area for different glacier extents are shown in panel panel (a). In panel (b) the climatic mass balance is plotted versus the elevation for the modelled decades. In panel (c) the internal accumulation is plotted versus elevation. The periods are hydrological years (e.g. 1968-1978 refers to 1 October 1968 - 30 September 1978).

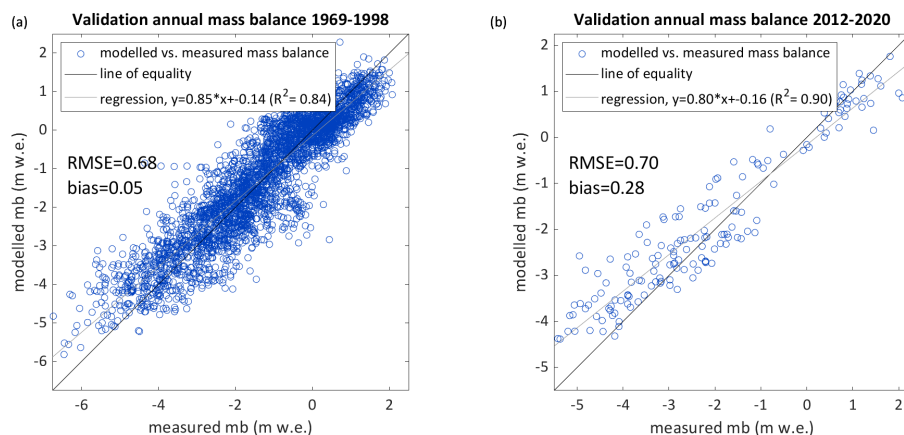


**Figure 6.** Model parameter calibration: Measured versus modelled surface mass balance for end of March (a) and September (b) for point measurements which are used for model calibration (1968/1969-1997/1998). The final calibration is shown. The bias between modelled and measured w.e. for snow pit measurements at the end of March (a) is minimised by optimising the precipitation correction factor. The bias between modelled and measured September mass balance measurements is optimised by tuning the albedo of ice.

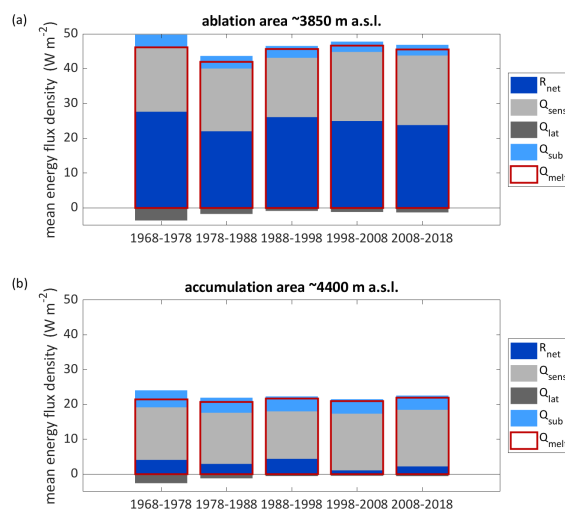
~~In figure 9, annual mass balances are plotted against summer temperatures and precipitation sums over the hydrological year.~~

#### 4.4 Internal accumulation

The average glacier-wide modelled internal accumulation is  $0.11 \text{ m w.e. a}^{-1}$  (Fig. 10). Internal accumulation appears in large parts of the accumulation area and is more pronounced in the orographic right accumulation area. ~~In Fig. 5c the internal accumulation is plotted against elevation for different decades.~~ Highest internal accumulation rates are modelled for the years 1968/1969-1977/1978 between 4400 and 4500 m a.s.l. The decades 1988/1989-1997/1998 and 1998/1999-2007/2008 are char-



**Figure 7.** Model validation: Measured versus modelled annual surface mass balance for an independent set of point measurements from 1968/1969-1997/1998 (a) and from 2011/2012-2019/2020 (b).



**Figure 8.** Mean modelled energy fluxes per decade for a grid point located in the ablation area at  $\sim 3850$  m a.s.l. (a) and in the accumulation area at  $\sim 4400$  m a.s.l. (b). The point locations are indicated in Fig.1.  $R_{net}$  is the net radiation,  $Q_{sens}$  the sensible and  $Q_{lat}$  the latent heat flux,  $Q_{lat}$  the heat flux from/into the subsurface and  $Q_{melt}$  the total energy available for melt. The periods are hydrological years (e.g. 1968-1978 refers to 1 October 1968 - 30 September 1978).

acterised by higher internal accumulation rates at lower elevations around 4250 m a.s.l. Lowest internal accumulation rates are modelled for the second and the last decade.



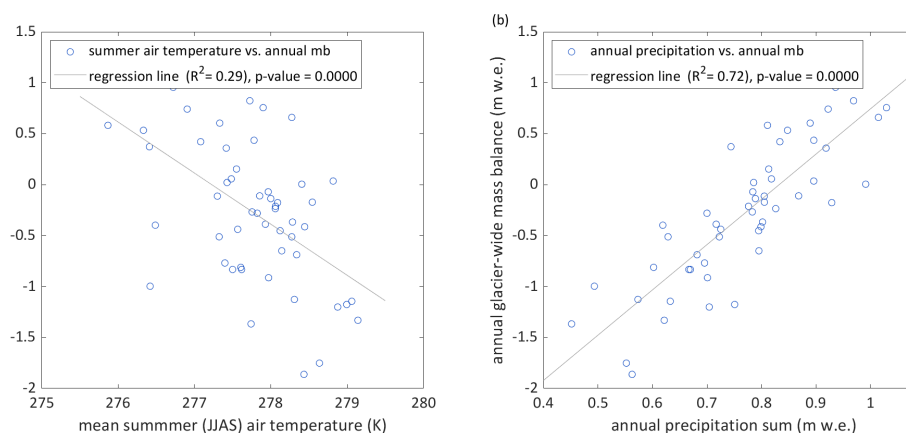
**Table 4.** Trends and p-values for climate variables and energy balance components for the period from the 1 October 1968 until the 30 September 2020 (hydrological years) are listed for a grid point in the ablation area at ~3850 m a.s.l. Values in brackets refer to site 2 located in the accumulation area at ~4400 m a.s.l. y/n stand for significant or not significant at a 90% confidence level. The point location of both grid points are indicated in Fig.1. The trends for glacier-wide mass balance, glacier wide internal accumulation and original climate forcing (for the elevation of the weather station at 3837 m a.s.l.; 1 October 1968 - 30 September 2020) are also given

variable	trend	unit	p-value	significant
mean annual air temperature	+0.0295 (+0.0236)	K a <sup>-1</sup>	0.000 (0.000)	y (y)
mean summer (JJAS) air temperature	+0.0208 (+0.0149)	K a <sup>-1</sup>	0.003 (0.003)	y (y)
annual precipitation sum	+0.0018 (+0.0068)	m w.e. a <sup>-1</sup>	0.421 (0.076)	n (y)
mean annual incoming shortwave radiation	-0.4297 (-0.2815)	W m <sup>-2</sup> a <sup>-1</sup>	0.000 (0.000)	y (y)
mean annual outgoing shortwave radiation	-0.3989 (-0.2362)	W m <sup>-2</sup> a <sup>-1</sup>	0.000 (0.000)	y (y)
mean annual incoming longwave radiation	+0.0702 (+0.0432)	W m <sup>-2</sup> a <sup>-1</sup>	0.074 (0.251)	y (n)
mean annual outgoing longwave radiation	+0.0507 (+0.0388)	W m <sup>-2</sup> a <sup>-1</sup>	0.005 (0.029)	y (y)
mean annual sensible heat flux	+0.0550 (+0.0417)	W m <sup>-2</sup> a <sup>-1</sup>	0.001 (0.003)	y (y)
mean annual latent heat flux	+0.0515 (+0.0473)	W m <sup>-2</sup> a <sup>-1</sup>	0.000 (0.000)	y (y)
mean annual heat flux from the subsurface	-0.0202 (-0.0153)	W m <sup>-2</sup> a <sup>-1</sup>	0.000 (0.001)	y (y)
mean annual glacier wide mass balance	+0.0002	m w.e. a <sup>-1</sup>	0.979	n
mean annual internal accumulation	-0.0003	m w.e. a <sup>-1</sup>	0.159	n
mean annual air temperature 3837 m a.s.l.	+0.0222	K a <sup>-1</sup>	0.000	y
mean summer (JJAS) air temperature 3837 m a.s.l.	+0.0136	K a <sup>-1</sup>	0.047	y
annual precipitation sum 3837 m a.s.l.	+0.0022	m w.e. a <sup>-1</sup>	0.074	y
mean annual cloud cover fraction 3837 m a.s.l.	-0.0012	-	0.002	y
mean annual relative humidity 3837 m a.s.l.	-0.1240	% a <sup>-1</sup>	0.002	y
mean annual air pressure 3837 m a.s.l.	+1.6274	Pa a <sup>-1</sup>	0.008	y

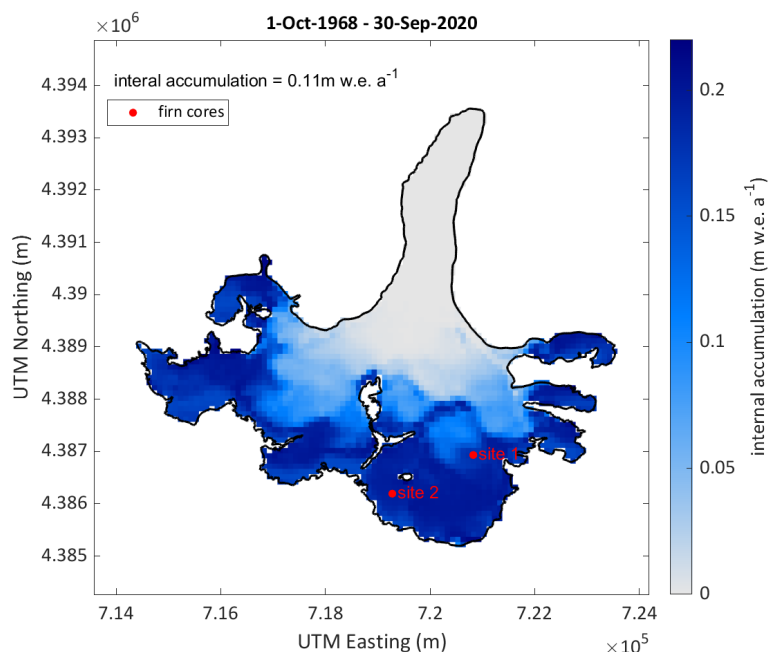
#### 4.5 Firn evolution

~~Modelled subsurface densities and temperatures are shown for two sites in Figs. 11 and 12. In Figs. 13 and 14, modelled subsurface densities and temperatures are shown together with *in situ* measurements.~~

Modelled subsurface densities are higher at ~4250 m a.s.l. (Fig. 11a-c) than at ~4400 m a.s.l. (Fig. 12a-c) and increase over time at both sites. At site 1, a significant increasing trend of subsurface densities for the depth range of 0-10 m is found for 1968/1969-2019/2020 (+1.14 kg m<sup>-3</sup> a<sup>-1</sup>, p-value = 0.014), whereas the trend is significant at site 2 when the first two decades are excluded (+1.53 kg m<sup>-3</sup> a<sup>-1</sup>, p-value = 0.005 for 1988/1989-2019/2020). At the lower site (~4250 m a.s.l., Fig. 11a-c) densities at depth reach the density of ice. Modelled firn densities are compared to measurements for four different dates in Fig. 13. The mean biases between modelled and measured densities for the depth covered by measurements are -18.9 kg m<sup>-3</sup> for June 1973, +52.6 kg m<sup>-3</sup> for June 1974, +20.4 kg m<sup>-3</sup> for June 1975 and -23.8 kg m<sup>-3</sup> for August 2018.



**Figure 9.** Climate variables at the station location (3837 m a.s.l.) versus modelled mass balance: Mean summer air temperature versus annual mass balance (a) and annual precipitation sums versus mean annual mass balance.



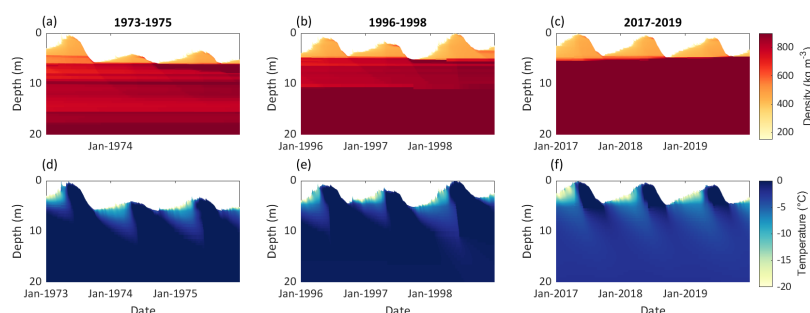
**Figure 10.** Map of mean annual internal accumulation for the mass balance years 1968/1969-2019/2020. Location of firn core drill sites used for validation are indicated with red dots.

For early years, the modelled subsurface temperatures indicate temperate firn conditions and propagation of winter cooling down to depths of about 10 m in the accumulation area at  $\sim 4250$  m a.s.l. (Fig. 11d). In later years, the cold content propagates

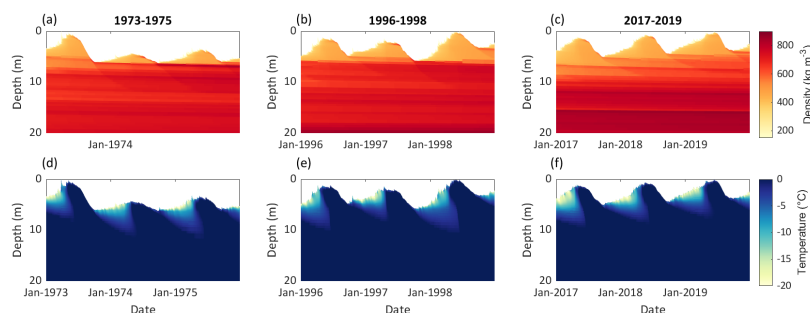


to greater depths and overall cold subsurface conditions are modelled for most recent years (Fig. 11e-f). For the depth range of 0-10 m, the subsurface cooling trend is  $-0.036 \text{ K a}^{-1}$  (p-value=0.000 for 1968/1969-2019/2020).

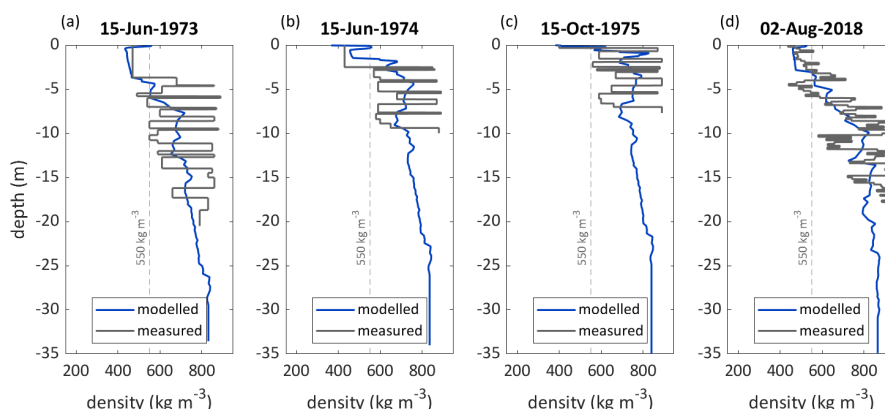
The modelled firn temperatures indicate temperate firn conditions and propagation of winter cooling down to depths of about 10 m in the accumulation area at  $\sim 4400 \text{ m a.s.l.}$  (Fig. 12d-f). Also for site 2, an overall cooling trend is identified ( $-0.0041 \text{ K a}^{-1}$  p-value=0.043 for 1968/1969-2019/2020). In Fig. 14 modelled firn temperatures are compared to thermistor measurements from spring 2018. Measured as well as modelled temperatures remain temperate at depths greater than about 10 m. In March (Fig. 14 a,b), modelled temperatures near the surface correspond well with observations. At a depth of about 7 m, the modelled temperatures are a bit higher. In April (shown in Fig. 14 c), modelled firn temperatures are warmer than measurements also for the uppermost thermistor location. At a depth of about 7 m minimum subsurface temperatures are recorded for 6 June 2018. On this date, modelled firn temperatures are temperate except for a small zone around 7 m depth (Fig. 14 d).



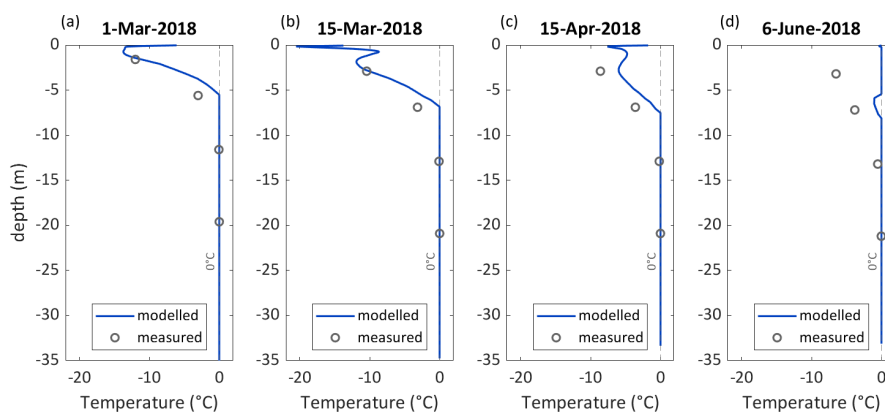
**Figure 11.** Modelled subsurface conditions at site 1 ( $\sim 4250 \text{ m a.s.l.}$ ) for three selected periods: 1973-1975 (panels a and d), 1996-1998 (b,e) and 2017-2019 (c,f). Panels a, b and c show the subsurface density and panels e, d and f the subsurface temperature. The location of the site is indicated in Fig.1 b



**Figure 12.** Modelled subsurface conditions at site 2 ( $\sim 4400 \text{ m a.s.l.}$ ) for three selected periods: 1973-1975 (panels a and d), 1996-1998 (b,e) and 2017-2019 (c,f). Panels a, b and c show the subsurface density and panels e, d and f the subsurface temperature. The location of the site is indicated in Fig.1 b



**Figure 13.** Modelled and measured subsurface densities for site 2 (~4400 m a.s.l.). Measured subsurface densities from the 1970s are digitised from figures 2.1 and 2.2 in Kislov (1982) (a-c) and own measurements are shown for 2018 (d). The location of the site is indicated in Fig. 1b.



**Figure 14.** Modelled and measured subsurface temperature for a station located nearby site 2 (~4400 m a.s.l.). The location of the site is indicated in Fig. 1b.

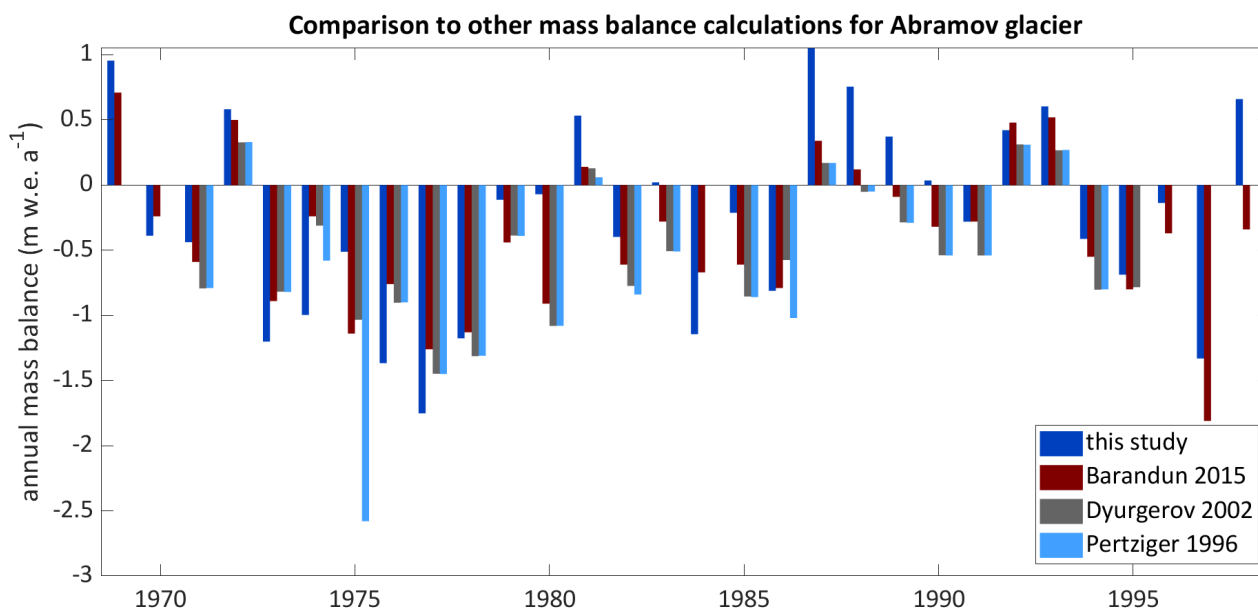
#### 410 4.6 Results of additional model run with alternative forcing

An additional distributed run (hereafter named alternative run) is performed using the alternative data set with a shorter period of station measurements to assess the model forcing sensitivity. The original model forcing consists of station measurements for 1968-1998 and bias corrected TopoSCALE ERA5 for 1999-2020, whereas the additional model run is forced by station measurements for 1968-1979 and corrected TopoSCALE ERA5 for 1980-2020. Results from this additional run yield a more negative mass balance than the original data set of  $-0.34 \text{ m w.e. a}^{-1}$  for 1968-2020. For the years 1980-1998, the alternative forcing yields a balance of  $-0.23 \text{ m w.e. a}^{-1}$ , whereas the balance was almost balanced using the original forcing ( $-0.05 \text{ m w.e. a}^{-1}$ ). The original forcing produces a higher internal accumulation and more positive mass balances for most of the glacier surface. Figures of the alternative model run are shown in the supplementary material (Figs. S2-S6).

## 5 Discussion

### 5.1 Long-term mass balance and firn evolution

The modelled long-term mass balance indicates an overall mass loss of Abramov glacier for the period 1968/1969-2019/2020 which is ~~generally~~ in agreement with the findings of other recent studies. Barandun et al. (2015) and Denzinger et al. (2021) found somewhat more negative mass balances. Barandun et al. (2015) estimate the mass balance for 1967/1968-2013/2014 as  $-0.44 \pm 0.10$  m w.e.  $a^{-1}$ . Their estimate is based on the application of a calibrated surface mass balance model and a simple approximation of internal accumulation as well as basal ablation which together contribute  $+0.07$  m w.e.  $a^{-1}$ . In Fig. 15 annual mass balances of different studies are compared to our results: Previous analysis of annual mass balance data for 1971-1994 yielded more negative mass changes (Barandun et al. (2015):  $-0.39$  m w.e.  $a^{-1}$ , Dyurgerov (2002):  $-0.50$  m w.e.  $a^{-1}$ , and Pertziger (1996):  $-0.61$  m w.e.  $a^{-1}$ ) than the EBFM ( $-0.26$  m w.e.  $a^{-1}$ ). Denzinger et al. (2021) calculate the geodetic mass balance as  $-0.38 \pm 0.12$  m w.e.  $a^{-1}$  for 15 July 1975 until 1 September 2015. Our result for the same time period ( $-0.30$  m w.e.  $a^{-1}$ ) is within their calculated uncertainties.



**Figure 15.** Comparison of annual mass balance results to previously published values for Abramov glacier (Pertziger, 1996; Dyurgerov, 2002; Barandun et al., 2021)

We do ~~neither~~ find a significant trend in the evolution of annual mass balances ( $+0.0002$  m w.e.  $a^{-1}$ , p-value = 0.979), nor in the reduction of mean annual internal accumulation ( $-0.0003$  m w.e.  $a^{-1}$ , p-value = 0.159) for the period 1968/1969-2019/2020.



But significant trends are found for all variables used as model forcing (Table 4). Different changes may thus compensate for each other. Comparing annual mass balances with annual precipitation sums and summer air temperatures shows that annual mass balances are stronger related to precipitation sums than to summer air temperatures (Fig. 9). The effect of the substantial warming ( $+0.0222 \text{ K a}^{-1}$ ,  $p\text{-value} = 0.000$ ) seems thus to be attenuated by increasing precipitation ( $+0.0022 \text{ m w.e. a}^{-1}$ ,  $p\text{-value} = 0.074$ ). The increase of net accumulation in the accumulation area (Fig. 5b) may thus be related to an increase of solid precipitation (Tables S2 and 4). Barandun et al. (2015) also find an increase of accumulation based on their modelling results. An accumulation increase is as well in agreement with Kronenberg et al. (2021a), who identified a net accumulation increase for a point location at  $\sim 4400 \text{ m a.s.l.}$  Abramov glacier might be affected by the same changes in precipitation patterns that could be partly responsible for the mass balance anomaly in western HMA (e.g. Miles et al., 2021). Barandun et al. (2015) furthermore describe a tendency towards more ablation during recent decades. We find high ablation rates for the first decade and for 1998/1999-2007/2008 (Fig. 5b). In contrast to previous studies (e.g. Dyurgerov and Dwyer, 2001), we do not find a steepening of ablation gradients. The ablation gradient is higher for 1968/1969-1997/1998 ( $0.0080 \text{ m w.e. (m a.s.l.)}^{-1}$ ) than for 1998/1999-2019/2020 ( $0.0070 \text{ m w.e. (m a.s.l.)}^{-1}$ ). Miles et al. (2021) calculate an ablation gradient of  $0.0084 \text{ m w.e. (m a.s.l.)}^{-1}$  for 2012-2016. Whereas their ablation gradient is higher, their ELA of  $4163 \text{ m a.s.l.}$  is located lower than the here modelled ELA (Table 3).

We find a relevant contribution of refreezing below the last summer horizon to the overall mass balance. As visible in Fig. 5c and from data in Table 3, internal accumulation also evolves in the course of time. Important changes are visible for the lower accumulation area up to  $\sim 4300 \text{ m a.s.l.}$ , where the internal accumulation strongly increases with elevation and lowest values are modelled for the last decade (Fig. 5c). Site 1, for which the subsurface evolution of density is shown in Fig. 11a-c, is located within this zone. For this point location, the subsurface density reaches the density of ice what hinders internal accumulation. An increase of subsurface densities also happens around  $\sim 4400 \text{ m a.s.l.}$  as shown in Fig. 12a-c. The lowest internal accumulation rates are found for the second (1978/1979-1987/1988) and the last (2008/2009-2017/2018) modelled decade. Both of them follow decades with clearly negative mass balances and high refreezing rates (Table 3). This is confirmed by *in situ* measurements from several years during the first decade which indicate negative annual mass balances at several or all point observation sites in the accumulation area (Kislov, 1982; Pertziger, 1996). The model results suggest, that thanks to high accumulation and limited ablation, the firn could recover during the following years allowing for higher internal accumulation rates thereafter. Results from the most recent decade and years, however, indicate that the necessary pore space for refreezing is again reducing. Whereas the subsurface density at  $\sim 4250 \text{ m a.s.l.}$  decreases since the beginning of the modelling period, a significant decrease is also found for  $\sim 4400 \text{ m a.s.l.}$  after 1987/1988. Abramov glacier is thus losing its mass loss buffering capacity through refreezing. The loss of pore space occurs despite the high amounts of solid precipitation (Table S2) responsible for the high accumulation compared to the first decade (Fig. 5b) and affects large areas of the accumulation zone (Fig. 5a). For the upper elevation bands, modelled internal accumulation remains high also for recent decades. With further warming, these zones are, however, expected to also lose pore space in future.

Mass balance observations often neglect internal accumulation or account for the process only in a simplified manner. This applies for mass balance calculation based on glaciological data but also for geodetic mass change estimates. The latter usually





use constant density conversion factors to convert elevation to mass changes. Our results indicate that internal accumulation plays an important role also for glaciers with temperate accumulation areas. Whereas mass change calculations neglecting the internal accumulation may be overestimating the mass loss for periods with available pore space, a smaller overestimation of mass loss may become more likely as glaciers loose pore space.

Our results suggest that the firn of Abramov glacier is currently losing pore space and that the loss of pore space is more advanced in the lower accumulation area. The warm and also cold infiltration zones of glaciers located at higher locations may not yet be affected. Evidence from *in situ* measurements in HMA is hardly available. Lambrecht et al. (2020) expect a strong elevation gradient of melt water refreezing on Fedschenko Glacier, western Pamir. Only at elevations above 5200 m a.s.l. the annual accumulation layers are currently preserved, but evidence of summer melt was found at high elevations of more than 5300 m a.s.l. Based on a sensitivity experiments with a region-wide application of a relatively simple model, Wang et al. (2019) concluded that under warmer conditions, refreezing will increase for continental glaciers and decrease for glaciers located in more humid and warmer environments. In their study, Abramov glacier is located in a region for which they find an increase of refreezing. Our results indicate a decrease of internal accumulation related to the retreat of the firn line. A retreat of the firn line is likely also occurring elsewhere, however, the underlying processes are often not sufficiently included into simple, regional approaches.

## 5.2 Uncertainties and validation

The comparison of modelled and measured surface mass balances (Fig. 7) and especially the comparison of subsurface properties (Figs. 13, 14) show, that the EBFM overall satisfyingly reproduces the observations. Our study furthermore demonstrates, that results with higher confidence can be reproduced for the period for which *in situ* measurements serve as a model forcing. The presented modelling results are influenced by several sources of uncertainties which contribute to deviations between observed and measured data. Especially towards the end of the modelling period, the model underestimates melt rates in the lower part of the ablation area (7b). The modelled ablation gradient may thus be somewhat underestimated for recent decades. In the following we discuss how these uncertainties may affect the presented results.

Several uncertainties are related to the model setup. The spatial vertical and horizontal as well as temporal resolutions are too coarse to resolve all the observed variations and to reproduce all relevant processes. Consequences of the horizontal resolutions are for example that several point observations shown in Figs. 6 and 7 are located within one modelled grid cell. The model is unable to fully reproduce the spatial heterogeneity of *in situ* observations. This also applies for subsurface conditions. On Abramov glacier highly variable firn conditions and accumulation rates are found based on firn cores and ground penetrating radar data (Kronenberg et al., 2021a). This study showed that annual accumulation, measured in firn cores, can vary by a factor of 1.5 over a horizontal distance of 250 m. The model satisfyingly reproduces the bulk densities measured at site 2 (Fig. 13), however the lower measured densities at nearby drill sites (c4381 and c4382 in Kronenberg et al. (2021a)) are not reproduced satisfactorily as the model does not simulate the horizontal small scale variability of accumulation and does furthermore not consider fine scale processes such as wind redistribution of snow. Another uncertainty related to the model set up is the use of a linearly updated glacier surface elevations and decade-wise changing glacier outlines. These approximations of topographical



changes may be responsible for underestimations (overestimations) of melt rates on the glacier tongue if surface elevations are to high (to low). The initial conditions ~~derived from initialisation runs using the first thirty year of climate input~~ are a further source of uncertainties. It is very likely, that the atmospheric conditions were different from the used data set prior to the modelling period. Nevertheless, modelled subsurface conditions agree well with measurements during early years of the modelling period (Fig. 13a-c).

An important source of uncertainties is related to the model forcing. While station data is available for the first thirty years, data from a gridded climate product (ERA5) has to be used for the remaining modelling period. To homogenize the input, the downscaled ERA5 data is bias corrected to match with monthly averages of observations for the overlapping period. Despite these corrections, the differences in both data sets affect the model output as evident from the results of the alternative model run which yielded a more negative mass balance for the period for which both data sets are available (alternative forcing:  $-0.23 \text{ m w.e. a}^{-1}$  original forcing  $-0.05 \text{ m w.e. a}^{-1}$ ). Uncertainties in the model forcing may thus be a further reason for the misfits between measured and modelled surface mass balances.

The parametrisations used in modelling energy balance and subsurface conditions are based on simplifications causing further uncertainties. The snow albedo parametrisation serves as an example. Snow albedo is re-set to fresh snow albedo after each snowfall event greater than  $3 \text{ mm w.e. d}^{-1}$ . Thereby, also the albedo decay scheme starts again with the value for fresh snow. During field visits in summer, we observed the deposition of fresh snow on a longer exposed snow surface with a visibly reduced albedo. The fresh snow then melted away within hours to days again exposing the darker snow. The used parametrisation is not able to reproduce this evolution. This simplification may be compensated by the calibration of  $t_{wet}^* = 7$  days, which is more suitable to represent conditions in Central Asia than the higher default value of 15 days used in the Arctic (Bougamont et al., 2005; van Pelt et al., 2019). As previously discussed in van Pelt and Kohler (2015), the model is not able to reproduce the small scale density variability of the subsurface (Fig. 13) as several processes such as wind crust formation, modelling of firn/snow grains or local vertical pooling of melt water on existing high density layers are not included.

Further model uncertainties are related to the choice of parameters. Thanks to the exceptional data availability for Abramov glacier, several model parameters can be constrained. These data, however, are only available for point locations. We use spatially and temporally constant parameter values for periods with and without *in situ* measurements. This necessary simplification has implications. We use temporally constant albedo parameters, whereas a darkening of the glacier surface is likely (Sarangi et al., 2019). An underestimation of recent ablation rates (Fig. 7b) may be related to albedo decreases (Schmale et al., 2017) which are not considered due to a lack of respective calibration data.

Our study shows, that long-term *in situ* measurements are of great value for simulating the long-term evolution of Abramov glacier with a coupled surface energy balance–multilayer subsurface model. The comprehensive model output complements the exceptional observational data set for this glacier. The combination of both data sets provides an opportunity to discuss processes on a high spatial and temporal resolution for a period of more than five decades which is unprecedented for the data sparse HMA.



## 535 6 Conclusions

In this study, we apply a distributed coupled surface energy balance and firn model to simulate 52 years of mass balance and firn evolution of Abramov glacier, Pamir Alay. The model is forced with weather station and downscaled reanalysis data. The modelled surface mass balance as well as subsurface conditions agree well with *in situ* measurements for the beginning of the modelling period and recent years. We find an overall negative mass balance of  $-0.27 \text{ m w.e. a}^{-1}$  for 1968/1969-2019/2020, which is somewhat less negative than the mass balance determined by previous studies. The first modelled decade 1968/1969-1977/1978 is characterized by the most negative mass balance and is followed by two decades of almost balanced conditions. More recent years are again characterised by clearly negative conditions. Our results indicate a loss of pore space and reduction of internal accumulation on Abramov glacier. The correlation of annual mass balance with annual precipitation sums ( $R^2=0.72$ ,  $p\text{-value} = 0.000$ ) is stronger than the correlation with summer air temperature ( $R^2=0.29$ ,  $p\text{-value} = 0.000$ ). Increasing precipitation rates have thus compensated for increasing air temperatures, preventing an acceleration of mass loss for Abramov glacier during the last five decades. To our knowledge, this is the first application of a model of similar complexity for such a long time period for a glacier in HMA and may thus provide valuable insights into processes of this data scarce region.

*Code and data availability.* The EBFM code, the model forcing and topographical grid used in this study are available at <https://doi.org/10.5281/zenodo.5773796> (Kronenberg et al., 2021b). Due to their large volume, modelled grids are available on request. The majority of in situ mass balance data used for calibration and validation are available from the World Glacier Monitoring Service ([doi.org/10.5904/wgms-fog-2021-05](https://doi.org/10.5904/wgms-fog-2021-05)) (WGMS, 2021). Additional mass balance data at monthly resolution and for recent years are also available on request. Measurements of the automatic weather station can be downloaded from [http://178.217.169.232/sdss/index.php?&page=measure\\_page](http://178.217.169.232/sdss/index.php?&page=measure_page) (Schöne et al, 2013).

*Author contributions.* MK performed the analysis and wrote the paper. WvP supplied the EBFM model code and support to use it. MK, HM and MH took place in several field campaigns acquiring *in situ* data. JF applied TopoSCALE to downscale the ERA5 data. FP provided the historical meteorological and mass balance data as well as extensive clarification on the data set. All authors participated in the discussion of the results.

*Competing interests.* The authors declare that they have no conflict of interest.

*Acknowledgements.* We acknowledge all the people who have been involved in the collection of field data on Abramov glacier. Especially we would like to thank to Erlan Azsivov, Ruslan Kenzhebaev Bolot Moldobekov and Ryskul Usubaliev from the Central Asian Institute for Applied Geosciences (CAIAG) and Sultanbek Belekov and Iurii Novomlintsev from KyrgyzHydromet for organizing the mass balance measurements on Abramov glacier during the last years and to Martina Barandun and Tomas Saks for their continuous efforts in coordinating



the monitoring activities. We greatly acknowledge the help of Stanislav Kutuzov, Ivan Lavrentiev, Alex Merkushkin, Yuri Tarasov, François Valla and Andrey Yakovlev for providing access to literature and data published in Russian and complementary information. We thank  
565 Enrico Mattea for the valuable exchange about the model application and also for sharing his code used for modelling experiments. This study was financed by the Swiss National Science Foundation (SNSF), grant 200021\_169453 and the Project CICADA (Cryospheric Climate Services for improved Adaptation, contract no. 81049674 between the Swiss Agency for Development and Cooperation and the University of Fribourg).



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