Development of proglacial lakes and evaluation of related outburst hazard at Adygine complex, northern Tien Shan

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Abstract. Formation and development of glacial lakes in mountain regions is one of the consequences of glacier recession. Such lakes may drain partially or completely when the stability of their dam is disturbed or as a consequence of impacts. We presented a case study from Central-Asian mountain range of Tien Shan, a north-oriented tributary Adygine valley, where a glacier retreat resulted in formation of several generations of lakes. The aim of this study was to analyse past development of different types of glacial lakes influenced by the same glacier, to project site's future development, and to evaluate the hazard of individual lakes with an outlook for expected future change. We addressed the problem with a combination of methods, namely bathymetric, geodetic, and geophysical on-site survey, satellite image and digital elevation model (DEM) analysis, and modelling of glacier development. Based on the case of glacial lakes of varied age and type, we demonstrate the significance of glacier ice in lake's development. Lake 3, which is in contact with glacier terminus, has changed rapidly over the last decade, expanding both in area and depth and increasing its volume more than 13 times (7 800 m^3 to 106 000 m^3). Hydrological connections and routing of glacier meltwater proved to be an important factor as well, since most lakes in the region are drained by subsurface channels. Within the hazard evaluation of lakes, we highlighted the importance of field data which can provide crucial information on lake stability. In our case, the understanding of site's hydrological system and its regime helped to categorise Lake 2 into low outburst hazard, while Lake 1 and Lake 3 were labelled as medium hazard lakes. Further development of the site will be driven mainly by rising air temperature and increasingly negative glacier mass balance. All three scenarios predict a significant glacier area decrease by 2050, specifically leaving 73.2 % (A1B), 62.3 % (A2), and 55.6 % (B1) of the 2012 glacier extent. The glacier retreat will be accompanied by changes in glacier runoff with first peak expected around the year 2020.

1 Introduction

Glacial lakes can be seen as gems gleaming in the harsh mountain environment. However, at the same time, they can pose a serious threat to downstream settlements and infrastructure. If we focus on Asian mountain ranges, the now common term GLOF (glacial lake outburst flood), first used for the Himalayan region, is ubiquitous. Mainly due to the development of satellite-based sensor technologies, the areas under scrutiny of investigators have increased from the Himalayas (Benn et al., 2012; Fujita et al., 2009; Shrestha et al., 2010) and Karakoram (Chen et al., 2010; Haemmig et al., 2014) to the Tibetan Plateau (Liu et al., 2014; Wang et al., 2013; Zhang et al., 2017), Pamir (Mergili and Schneider, 2011), and Tien Shan (Bolch et al., 2011; Engel et al., 2012; Narama et al., 2017; Sorg et al., 2012). In the territory of Kyrgyzstan, there are about 2 000 glacial lakes (>1 000 m^2), almost 20 % of them are potentially dangerous, and approximately 15-20 lakes are at risk of sudden drainage each year (Erokhin and Zaginaev, 2016). However, the dry and glacier meltwater-dependent region of Central Asia remains rather in the background of interest.

Some glacial lakes form and drain within a relatively short time (Erokhin et al., 2017). Others exist for years and decades without major change, and lastly, there are lakes in a phase of enlargement. Expansion of glacial lakes broadly correlates with warming temperature trends and negative glacier mass balance patterns (Lei et al., 2012; Mergili et al., 2013; Zhao et al., 2015). In addition, a recent study shows that the expansion of glacial lakes within a dry continental climate regime can be closely related to thermokarst processes as well (Li et al., 2014). This means that lakes enlarge when in contact with a retreating glacier tongue, by melting of ice in the lake basin bed or sides, or by filling due to increased inflow from a glacier. Other factors playing a role in lake evolution include: erosion of surface drainage channel, formation, expansion or blockage of subsurface drainage channels, dam morphology changes, or slope movements adjacent to the lake.

Outbursts of mountain lakes occur almost annually in Kyrgyzstan (in recent years, these were Lake Merzbacher - 2017, Lake Chelektor - 2017, and Aksai - 2015) and given the generally increasing use of mountain valleys, greater losses can be expected as a consequence (Dussaillant et al., 2010). Once an outburst is triggered, a flood (often evolving into hyperconcentrated flow or debris flow) threatens areas in the lower parts of a valley. To lower the risk, an assessment framework (example in GAPHAZ Technical guidance document; GAPHAZ, 2017), should be applied including the following steps: i) identification of potentially dangerous lakes, ii) detailed evaluation of the hazard, and iii) application of mitigating measures. Our study includes the second step, a detailed hazard evaluation of selected lakes identified as potentially dangerous. As needs for field data in such a hazard assessment has been generally acknowledged, we present an assessment based on combination of on-the-spot and remotely-gained data.

Although the study site is not particularly suited for the purpose of glacier retreat monitoring alone, in terms of observing the site in a more complex way, we find it very appropriate. The glacier is considered to be representative of smaller glaciers, which are subjected to relatively fast retreat rates (Narama et al., 2010) and have a generally shorter response time to changes of climatic conditions compared to larger ones (Wang et al., 2014). This allows us to follow a larger development period in a smaller time scale. Also the existence of glacial lakes documenting glacier shrinkage is convenient – they are at various stages of development and could help us to better understand the evolution over time of different types of lakes. Due to the occurrence of lake outbursts in the region, the high elevation of the site, and its proximity to densely populated areas, a hazard assessment strategy with estimate of future change is desirable and of practical relevance.

Based on a detailed study of one particular site, we look at the issue of a glacier development and the associated actively evolving glacier forefield, and its consequences on downstream areas. In order to address the problem of the fast processes in mountain regions, we incorporate the perspectives of future scenarios for the development of the site and related outburst hazard into our study. As our intention is to promote a more complex approach to the issue of hazard assessment, our study includes three particular objectives: i) to analyse past evolution of different types of glacial lakes influenced by the same glacier, ii) to provide an insight into probable future evolution of the site, and iii) to evaluate hazard of individual lakes with an outlook for possible future change. As a consequence of our objectives the structure of the paper is as follows: First, we analyse the past development of the local setting, following both the glacier and the proglacial lakes. Then, we look at the potential future conditions at the site, and finally we evaluate the lakes' outburst hazard and estimate how it can evolve with changing conditions. In our opinion, to analyse glacier retreat and lakes' formation without touching the topic of consequences (present and those in near future) is as incomplete as evaluating outburst hazard without knowing the past evolution of the site.

2 Data and methods

2.1 Study area

The research site is situated in northern Tien Shan, specifically Kyrgyz Ala-Too. The highest peak of this East-West-directed range is Pik Semenova Tienshanskogo, with 4 895 m a.s.l. Although the Kyrgyz Ala-Too is not as highly glaciated as ranges in central Tien Shan, the now glacier-free parts of the valleys show signs of former glacier extent. The main Ala Archa Valley currently has approximately 33 km² of its area covered by glaciers, and is drained into the Chuy River, which is part of a large endorheic basin. The most common lake type of Kyrgyz Ala-Too is a lake formed in an intramorainic depression (Table 1). According to Erokhin (2012), 83.6 % of potentially dangerous lakes are ice-cored moraine-dammed lakes, which include intramorainic and thermokarst lakes with presence of buried ice.

Table 1. Representation of different lake types in Kyrgyz Ala-Too, northern Tien Shan, in 2017. Only lakes with minimum area of 1 500 m^2 were categorised. Based on manual mapping using satellite imagery in Google Earth and refined with field data included in Erokhin and Zaginaev (2016).

Lake type	**Number and share**	
Moraine-dammed	13	14.5 %
Intramorainic depression	48	53.3 %
Rockstep (+ moraine)	17	18.9 %
Landslide dam	8	8.9 %
Ice dam	4	4.4 %
Surface drainage	21	23.3 %
Subsurface drainage	69	76.7 %
Total	**90**	**100 %**

Development of glaciation in the Ala Archa basin is monitored in the long term mainly due to its position in proximity of the Kyrgyz capital, Bishkek, and the popularity of the valley among tourists visiting the National Park. According to Aizen et al.

(2006), there was a reduction of glaciated area in the Ala Archa watershed by -0.29%/yr between 1963 and 1981, and -0.51%/yr between 1981 and 2003 documenting an accelerated retreat by the end of the 20th century. The overall reduction of glaciated area over 60 years (1943-2003) was 15.7 %, which is an above-average value – in Tien Shan, the glacier area shrank by 14.2 % (Aizen et al., 2007a). Farinotti et al. (2015) confirm this trend and Bolch (2015) points out that, although the number of glaciers increased due to disintegration of several glaciers, the total glacierised area was reduced by 18.3±5.0 % between 1964 and 2010. Rising air temperatures (especially since the 1970s) have caused a negative mass balance of most glaciers, which has also led to changes in the hydrological regime of glacial streams (Aizen et al., 1996; Glazirin, 1996; Pieczonka et al., 2013). Sporadic permafrost occurs at elevations above 2 700 m a.s.l., discontinuous at 3 200–3 500 m a.s.l., and continuous above ~3 500 m a.s.l. (Gorbunov et al., 1996).

The Adygine watershed has an area of 39.6 km² and its glacierised area constitutes ~10 % of the basin (3.9 km²). The total elevation differences are up to 2 370 m, the valley below the glacier has an average slope of 10.8 °. The upper part of the Adygine Valley contains large amounts of loose glaciofluvial sediments and is dominated by several generations of moraines up to a distance of 3.5 km from the current glacier terminus. The Adygine complex (42°30’10’’ N, 74°26’20’’ E) closes this 8-km long tributary valley with northern orientation and reaches an elevation of 3 400-4 200 m a.s.l. (Fig. 1). The upper part of the complex involves a polythermal glacier (2.8 km²) (Kovalenko et al., 2014; similar structure described at Sary-Tor glacier by Petrakov et al., 2014), followed by a three-level cascade of glacial lakes that have evolved as a consequence of glacier retreat in the past 50 years. The glacier terminus is currently situated at 3 600 m a.s.l.; the tongue is rather steep and short, emerging from a relatively flat, larger source area. A large part of the glacier is covered with a thin layer of dust and debris, considerably lowering the albedo of its surface. The position of equilibrium line altitude (ELA) on the northern slopes of Kyrgyz Ala-Too is at approx. 3 900 m a.s.l., resulting in an ablation zone covering more than 65 % of the glacier area. The lakes found at the site (Table 2) are of varying age and have different positions in the connected hydrological system draining meltwater down the valley. The most recently formed lakes are close to the glacier terminus, reacting to changes in glacier melt rates sensitively and redistributing the water further downstream, either by surface or subsurface channels. The middle level is represented by the largest lake of the site, Lake 2 (32 000 m²), with a fairly stable annual hydrological regime (Falatkova et al., 2014) supplying water to the lowest part of the cascade, Lake 1. Not being a permanent lake, this intramorainic depression is filled with water only during an ablation season when the rate of incoming meltwater is higher than the capacity of its subsurface drainage.

A Holocene moraine complex at an elevation of 3 450 m a.s.l. forms the lower part of the Adygine complex. It consists of several parts of varying age (Shatravin, 2000), some of them can be identified even based on visual inspection only. The western part is formed by creeping ice-rich debris with a typical relief of arcuate ridges and furrows at its terminal part, ice lenses are present and exposed occasionally (Fig. 1b). From the east, a smaller ice-debris matrix moves under the influence of gravity and adjoins the central landform. The main body is a so called debris-derived rock glacier that formed below the glacier terminus. The landform has a prominent oversteepened front almost 700 m wide and it consists of glacier ice overlain by perennially frozen debris. In its western part, thermokarst processes are manifested in subsidence craters, cracks, and sagging, numerous thermokarst lakes and exposed ice lenses (Fig. 1c) can be found here. The eastern part is considered an older generation with rather flattened surface and stable lakes of lower turbidity.

Climatic conditions in the area are continental, characterised by relatively low precipitation and high annual and daily air temperature fluctuation. This is clearly evident at the Ala Archa station (2 200 m a.s.l.) which reports a 22 °C difference between mean air temperature (MAT) of the warmest and coldest months (13 °C for July, -9° C for January) and mean annual precipitation totals of 450 mm. Some 1 500 m higher, at the Adygine glacier, the MAT for January drops to -22 °C and reaches 5 °C in July; precipitation is estimated at 700-800 mm. On the northern slope of Kyrgyz Ala-Too, winters are characterised by low precipitation totals, whereas the maximum of precipitation is typically reached in late spring and the beginning of summer.

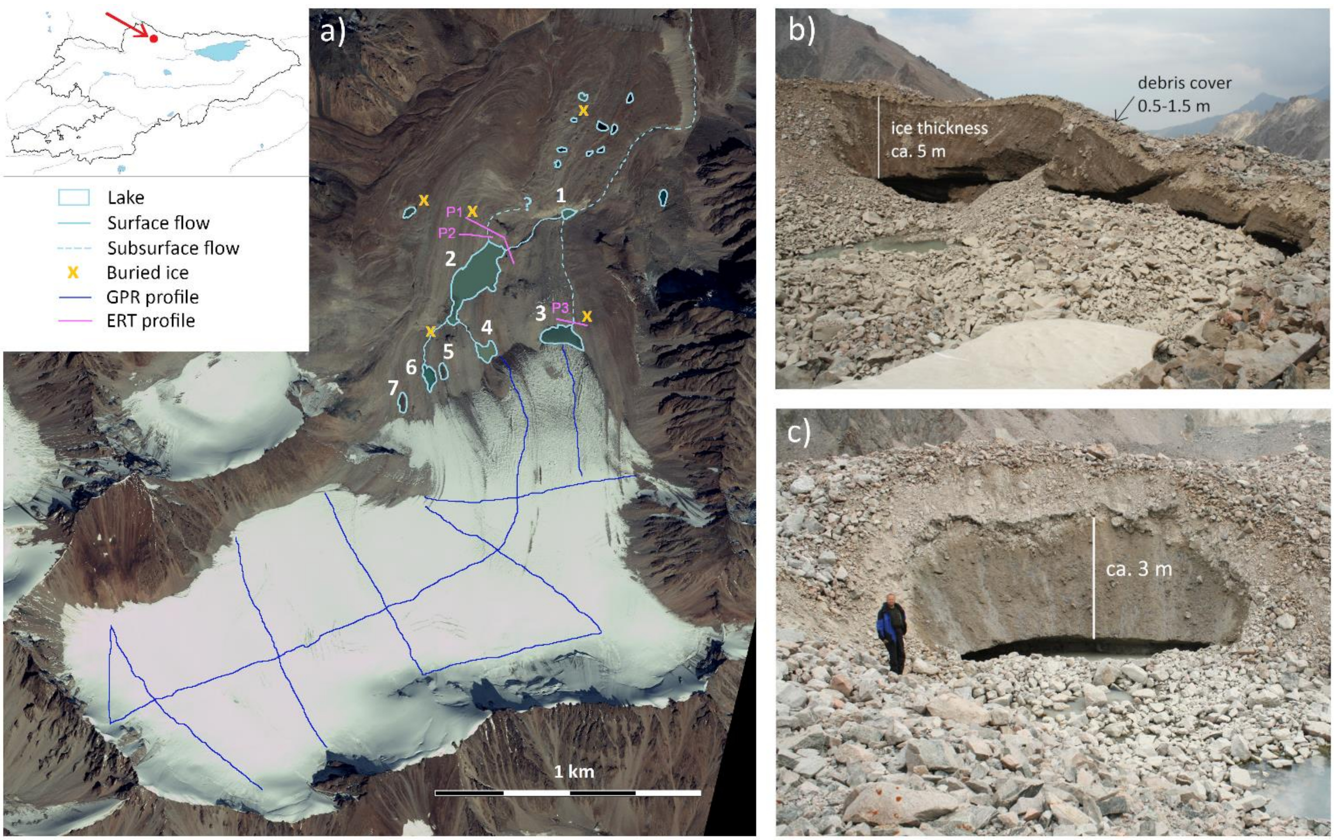

Figure 1. Overview of the Adygine complex (a) and its recent dynamics – exposure of buried ice (b, c). Numbers 1-7 refer to studied lakes, P1-P3 refer to geophysical profiles. Study site marked with an arrow on a map of Kyrgyzstan in upper left corner. The position of b) and c) spots is marked in Fig. 1a. Image source: Worldview-2, 2011; photo: K. Falatkova (2017).

Table 2. Basic parameters of the three largest lakes of the site.

Lake	Max volume [m³]	Max depth [m]	Elevation [m a.s.l.]	Dam type	Freeboard [m]	Buried ice	Drainage
1	15 000 (200 000)*	5	3 450	intramorainic depression	-	yes	subsurface
2	210 000	22	3 540	riegel + moraine	0 - 2.6	yes	surface + subsurface
3	106 000	14	3 580	riegel + moraine	0 - 1	yes	subsurface**

* potential volume of the lake in case of a drainage blocking
** during high water level dam overflow observed, water sinks underground after approx. 5 m

2.2 Field mapping

A research station built at the site provides an adequate background for field measurements. Climatic data and ground temperatures are obtained from an automated weather station situated near the Adygine glacier at 3 560 m a.s.l. The hydrological regime of the lakes is monitored with pressure sensors installed in Lake 1 (2012-2015), Lake 2 (2007-2017), and Lake 3 (2012-2017).

The glacier retreat has been monitored since 2006 during the ablation season by means of a geodetic survey carried out with a Leica TCR 705 total station (infrared beam on a reflective prism – accuracy: 0.005 m). The survey was focused solely on the terminal part of the tongue, where the retreat is most significant and may result in exposure of overdeepenings (potential future lakes). The series is supplemented with glacier limits acquired from aerial images taken in 1988 and 1962 (scale of the survey: 1:38 600, image resolution: 1 m). The lakes' shoreline changes are measured by the same means as the glacier terminus. To capture changes of lake basin and derive the volume of detained water, a repeated bathymetric survey was carried out at selected lakes (Sobr and Jansky, 2016). The depth was recorded in a 2.5 m step at defined profiles, using an echo-sounder mounted on a boat (depth measurement accuracy: 0.1 m). We processed the data obtained and interpolated them into a bathymetric map.

The selected key locations of the site (lakes' surroundings and their dams) were surveyed by electrical resistivity tomography and spontaneous polarisation, in order to detect buried glacier remnants and possible seepage routes. In order to acquire information on glacier thickness and bed topography for adjusting the model of future glacier extent and identifying potential areas for lake formation, ground penetrating radar (GPR) was used. The location of the profiles is shown in Fig. 1a.

The method of electrical resistivity tomography (ERT) consists in measuring the resistivity of the subsurface by means of a number of electrodes located along a profile. Interpretation of measured data was performed using the software Res2Dinv (Loke and Barker, 1995), which serves for 2D inversion of measured resistance data and calculation of 2D cross section under the profile. The resistive section of the observed profile approximates the actual distribution of the individual resistive layers in the depth range given by the maximum spacing of the electrodes. We used the ARS-200E (GF Instruments CR) with 48 electrodes in 4.4 m increments. The maximum range of electrodes was 206.8 m, with a depth range of about 50 m.

Spontaneous polarization method (SP) measures the natural electrical potential of the rock environment; in case of lake dams, the filtering potential that arises from water filtration through the porous environment is tested. Measurements were made with GEOTOR I with offset compensation. Non-polarizable electrodes have been used to minimize the effect of the transient resistance between the measuring electrode and the geological environment. The measurement was carried out in a potential variation where the electrical potential of a given point on the profile is measured relative to the fixed reference point. Measurement step was around 10 m, every 5th point was recorded using GPS.

Ground-penetrating radar (GPR) survey was applied on Adygine Glacier to acquire information on glacier thickness and subglacial topography. GPR was carried out along the central longitudinal profile, five transverse profiles, and three short connecting profiles. A longitudinal profiling was carried out from the glacier terminus to the upper reaches of the ablation zone (3610–3970 m a.s.l.). An unshielded 50MHz RTA (rough terrain antenna) and RAMAC CU-II control unit (MALÅ GeoScience) were used for data collection. The signal acquisition time was set to 2040 ns and scan spacing to 0.1 s. The GPR profiling equipment was man-hauled at a mean speed of ~0.5 m s–1. The collected data were processed and interpreted using the REFLEXW software version 4.5 (Sandmeier, 2017).

2.3 Modelling of glacier evolution

The glacier mass balance model GERM (glacier evolution runoff model) applied in this study is well established for simulating glacier mass balance from climate data based on a sophisticated degree-day approach (Hock, 1999; Huss et al., 2008). In our case, future glacier mass balances of Adygine glacier until 2050 are calculated with GERM, forced by empirically downscaled scenarios data for daily air temperature and precipitation, described below. As glacier mass balance is not measured at Adygine glacier, we used the mass balance data from nearby Golubina glacier (WGMS, 2017) for validating the simulated mass balances of the GERM model against measured mass balances for the period 1981-1994. Mass balances could be validated for both winter and summer balances. Additionally, we could use discharge data from Adygine River for 1960-1987 to validate the modelled discharge of GERM with observed data.

Despite generally low data availability in Central Asia, daily air temperature means and precipitation totals were gathered from 2 stations close to the study area: Alplager (2 130 m a.s.l., 1978-2010) and Baityk (1 580 m a.s.l., 1914-1979). The meteorological station data were used to calibrate and validate the downscaling model (multivariate regression and analogue method, Benestad, 2004; Benestad et al. 2007). Reanalysis data from NCEP/NCAR (Kalnay et al., 1996) were used as predictor fields in order to downscale long term data sets of precipitation and air temperature for the study region. For scenario simulations until 2050 three different SRES scenarios, A1B, B1, and A2 (IPCC, 2007), were performed by the ECHAM5/MPI-

OM coupled atmosphere-ocean model (Röckner et al., 2003). Though the SRES scenario data are meanwhile replaced by the representative concentration pathways (RCP) approach of IPCC (IPCC, 2013), the uncertainty ranges of future large scale temperature and precipitation changes of the two approaches are rather similar and smaller compared to the uncertainty range of climate data coming from further downscaling. We argue, that for our purpose of assessment of future glacier extent, in particular for the future development of glacial lakes, these scenarios are sufficiently realistic.

In order to estimate the potential spots for glacial lake formation, the data on glacier bed topography (Fig. A2) acquired by ground penetrating radar (GPR) were used. The digital elevation model (DEM) data for the glacier surface were consulted from Shuttle Radar Topography Mission (SRTM) with a resolution of 30 m (Farr et al. 2007). The combined information on future glacier extent and topography of the exposed area were analysed in GIS by determining flow directions and identifying sinks (i.e. overdeepenings).

2.4 Hazard assessment

Presented lake outburst hazard assessment is qualitative and was adapted to the regional distinctive features, i.e. dry climate, common occurrence of buried ice, permafrost. We based the assessment procedure on our knowledge gained through fieldwork in the region, focused especially on hydrological systems of proglacial lakes. The specificity of the region is the origin of lakes – the most common type is a lake formed in an intramorainic depression (Table 1). Many lake hazard evaluations are based on cases of moraine dams with a distinct morphometry. However, this type of lake does not occur that frequently in the region, representing only 2 % of the potentially dangerous ones (Erokhin and Zaginaev, 2016). Nevertheless, inspiration is drawn from assessments by ICIMOD (ICIMOD, 2011), Allen et al. (2016), Frey et al. (2010), and Huggel et al. (2004). Hazard assessments based solely on remotely sensed data are appropriate when applied over a large area and as a first step. However, at least for lakes labelled as potentially dangerous, additional evaluation which would include on-site research needs to be carried out.

In this paper, overall hazard is introduced as a combination of a lake’s susceptibility to burst and cause flooding, and the presence of possible triggers that have a capacity to cause an outburst. To assess lake susceptibility, the following parameters were selected and a simple qualitative rating suggested (Table 3):

Lake volume – involves the size of a lake and volume of retained water. Although hazard does not increase proportionally with lake size, a larger volume generally means higher hydrostatic pressure on a dam and greater potential to cause damage. The thresholds are set according to information on GLOF cases in Kyrgyzstan.

Lake type – is primarily connected to the stability of a dam. The material forming a dam (or a depression in which a lake exists) has varying characteristics and behaviour in relation to water. Rockstep is considered the most stable, as water usually has a minor effect on it. However, when covered with loose morainic material left behind a retreating glacier, this part of a dam is then prone to erosion, piping, and is less stable overall. Interaction of a lake with an ice dam results in even higher risks – ice melting due to heat transfer from water, high hydrostatic pressure on the ice dam can lead to its uplift, causing lake drainage.

Another parameter influencing lake stability is *ice contact*. A lake in direct contact with a glacier is considered more dangerous than one at a distance, as it is strongly affected by glacier behaviour. Its evolution and morphological changes can be dynamic as the glacier retreats, advances or disintegrates.

A lake's hydrological regime is represented by one parameter – *drainage type*, with only two options – surface or subsurface drainage. It is the limited (or changing) capacity of a subsurface channel which makes lakes with this type of drainage more dangerous, as such lakes are prone to filling when inflow is increased. Accumulated water can then be released rapidly due to channel enlargement under higher hydrostatic pressure. A surface channel, in comparison, regulates the lake water level naturally and maintains its hydrological balance.

Growth possibility of a lake depends on lake basin characteristics and the presence of ice that can melt. Lakes with terminated expansion possibilities are considered more stable than those which can enlarge for some reason.

Table 3. Lake's susceptibility parameters and rating. The scheme on the right shows possible combinations of parameters' rating and the resulting susceptibility. Note that this simplifying table serves as guidelines to susceptibility estimate and that even medium total susceptibility means a lake is dangerous, may burst and cause damage.

Parameter		**Rating**
Lake volume [m³]	< 50 000	low
	50 000-500 000	medium
	> 500 000	high
Lake/dam type	Rock step	low
	moraine-rock, moraine dam	medium
	intramorainic depression	medium
	ice	high
Lake drainage	surface	low
	subsurface	medium
Glacier contact	no	low
	yes	medium
Growth possible	no	low
	yes	medium

Parameters rating					**Result**
H	H	M	M	M	**H**
H	H	M	M	L	**H**
H	H	M	L	L	**H**
H	M	M	M	M	**H**
H	M	M	M	L	**H**
H	M	M	L	L	**H**
H	M	L	L	L	**H**
H	L	L	L	L	**M**
M	M	M	M	M	**M**
M	M	M	M	L	**M**
M	M	M	L	L	**M**
M	M	L	L	L	**L**
M	L	L	L	L	**L**
L	L	L	L	L	**L**

Determination of the degree to which a lake is susceptible to outburst is followed by assessment of potential triggers. The means to determine whether the trigger has a potential to cause outburst of the specific lake or not, may be distant (VHR satellite image, DEM), as is the case of the first group of triggers (Table 4), or on-site, as is necessary e.g. for inner dam stability or subsurface channel functioning.

The fall of mass into a lake is one of the most common triggers of lake outburst (Emmer and Cochachin, 2013; Ding and Liu, 1992; Falatkova, 2016), resulting in an impact wave which overtops or destabilises the dam. Following a widely used procedure, the distance between a lake and a slope and slope steepness are the parameters determining the trigger potential (Alean, 1985; Fischer et al., 2012; Noetzli et al., 2003). In the case of calving, a lake's contact with a cliff-like glacier terminus is a crucial condition.

Development and stability of a glacial lake can be affected by melting of buried ice and related thermokarst processes. During stagnation of a debris-covered glacier, an ice-cored moraine dam can be formed. These cores present a weak point in a dam, as their ablation changes a dam's inner structure (Richardson and Reynolds, 2000).

Not only landslide dams are prone to failure caused by piping. Cases of GLOF due to piping in moraine dams were observed as well (Xu, 1988). Moraines consist of unconsolidated heterogeneous material and therefore may be prone to seepage and internal erosion (piping) as a conduit grows and a dam is weakened (Awal et al., 2011). Another dam stability-decreasing phenomenon – earthquake, was confirmed as a primary trigger of several lake outbursts (Clague and Evans, 2000; Lliboutry et al., 1977).

In the case of a lake not having a surface drainage and meltwater from a glacier being routed into the lake, there is the potential of significant hydrostatic pressure increase in times of higher melt rates of snow or ice. A sudden air temperature rise may therefore cause disturbance of the lake hydrological balance with possible consequences for dam stability. Increase of hydrostatic pressure may also be caused by blockage of a subsurface channel, which is a very unpredictable and still rather poorly understood phenomenon. Monitoring of changes in lake's morphometry due to frequent sliding of lake walls or the hydrological regime of a lake can serve as a tool for determining channel blockage potential.

Change of lake outburst hazard is a complex situation, resulting from interaction of geomorphological processes and the behaviour of the glacio-hydrological system. Development of englacial and subsurface routing of meltwater is especially crucial, yet difficult to determine, and beyond the scope of this paper. The presented estimate of hazard change is linked to climatic changes (rising MAAT) only, and involves impacts accompanying deglaciation that were observed in other mountain ranges, such as glacier retreat, altered glacier melt rates, also slope instabilities (debuttressing effect), permafrost degradation, or buried ice melting (Allen et al., 2016; Frey et al., 2010; Haeberli et al., 2017). The general assumptions of future climatic conditions, glacier retreat, and altered melt rates are supported by the glacier mass balance model (Sect. 3.3).

Table 4. Triggers with potential to cause outburst and their description.

Trigger		Potential to cause outburst: not present or very low	Potential to cause outburst: present
Fall of mass into lake	Ice avalanche	average slope lake-steep glacier parts* below 17°	average slope lake-steep glacier parts* over 17°
	Landslide/ rockfall	no proximal** slope over 30°	proximal** steep slope - unconsolidated material (over 30°) or rockwall (over 50°)
	Calving	no contact with glacier	contact with glacier (drifting ice blocks, crevassed front)
Inner dam stability	Buried ice melting	no buried ice detected, no surface signs in lake's surroundings	buried ice detected/surface signs observed in lake's proximity
	Seepage/ piping	dam not prone to piping (rockstep, ice dam)	water getting to surface at the airy side of dam, moisture marks at airy side of dam, sediment accumulation near spring
	Earthquake	region with very low seismic activity	considerable seismic activity in the region
Hydrostatic pressure increase	Increased inflow	stable surface outflow regulating lake level, lake not hydrologically connected to glacier	lake without surface drainage, glacier meltwater inflow
	Subsurface channel blockage	lake drained by surface channel	lake drained by subsurface channel, channel varying capacity, previous lake filling-up without increased inflow observed
	Upper lake outburst	no lake (of substantial volume) upstream	presence of potentially dangerous lake upstream

* avalanche starting zone steepness: polythermal glacier over 25°, cold-based glacier over 45° (Alean, 1985)
**average slope trajectory over 14° (Noetzli et al., 2003)

3 Results

3.1 Past glacier development

The terminus position of the investigated glacier has changed significantly since the 1960s. Over the past five decades, the furthermost part of the glacier terminus retreated approx. 630 m (Fig. 2). Adygine glacier became disconnected from its smaller western branch probably during the 1950s. According to an aerial image taken in 1962, the tongue was divided into two parts by a resistant rock outcrop. The larger western part reached the rock riegel at an elevation of approx. 3 540 m a.s.l. (close to the current Lake 2 drainage channel); the narrower eastern tongue bypassed the outcrop reaching 100 m lower, to 3 450 m a.s.l. (currently the position of Lake 1). In the following decades, both parts retreated at an average rate of 8 m a^{-1} in 1962-1988 and 14 m a^{-1} in 1988-2006. Currently, there is no elevation difference between the western and eastern parts of the terminus (3 640 m a.s.l.). In the latest period of observation (2007-17), the terminus retreat rate had similar values to the earlier 1962-1988 phase, slightly over 8 m a^{-1}. A significant recession period in recent years was recorded between 2013 and 2015, when the terminus retreated some 10-15 m per year. Positional differences of retreat rate are linked to variability of glacier bed topography (rock outcrops) and to heat transfer with meltwater (contact with glacial lake).

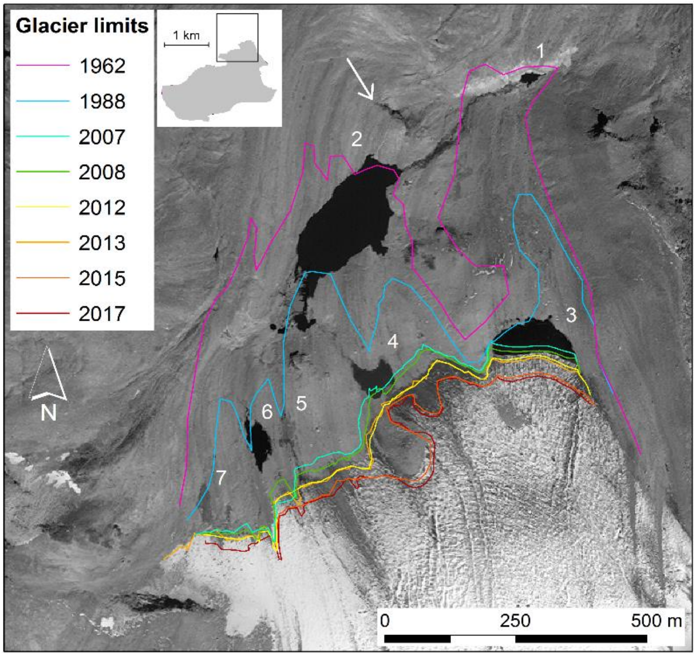

Figure 2. Retreat of the Adygine glacier tongue between 1962 and 2017. Numbers 1-7 represent the proglacial lakes, arrow marks the rock outcrop (riegel). Base image: Worldview-2, 2011.

3.2 Formation and development of the lakes

There are numerous lakes of varying size, type, and stage of development at the study site. They form a three-level cascade and at least some of the lakes are hydrologically connected.

3.2.1 Lake 1

The lowest part of the cascade is formed by about a dozen thermokarst lakes, situated on a large morainic platform at an elevation of approx. 3 450 m. The largest of them (Lake 1; Janský et al., 2010) is situated directly below a rock riegel in a depression (approx. 200 000 m^3), which was revealed when the eastern part of the tongue receded. The exact year when the depression filled with water is not clear. According to our knowledge of the drainage system, it was after Lake 2 developed and started supplying Lake 1 with a larger volume of meltwater (probably during the late 1980s). This intramorainic lake is not permanent, as the inflow exceeds the outflow channel capacity only during an ablation season. The lake level fluctuates significantly during the course of a day (leading to volume changes between 10 000 and 15 000 m^3), caused by limited capacity of subsurface outflow channels. Besides surface water intake, water is probably routed here by seepage from Lake 2 and Lake 3. There have been changes in morphometry of the basin within recent years, as sliding of unconsolidated material from the basin walls was observed. The part close to the lake drainage has changed in particular recently, extending the basin in a north-easterly direction.

3.2.2 Lake 2

Lake 2, the largest lake of the site, began forming in about 1960, after the western glacier tongue retreated behind a rock riegel at 3 540 m a.s.l. Since then, the lake area has extended, together with further terminus recession, reaching 32 700 m^2 in 2005 (first on-site survey). In the second half of the 1990s, the lake lost direct contact with the glacier and, as a result, the lake ceased to grow. Lake volume is regulated by a surface outflow channel eroded in morainic material, covering a rock outcrop. In the last decade, a slight change in the lake area has been recorded, caused by siltation near the inflow. In 2017, the lake area was 30 900 m^2, and the maximal lake depth has slightly decreased as well since the beginning of monitoring (from 22.2 m in 2008 to 21.3 m in 2015), probably also as the result of siltation. The lake volume changed accordingly from 208 000 m^3 (2008) to about 195 000 m^3 (2015). Geophysical sounding confirmed the presence of buried ice in the western part of the lake dam area – the remnants of a glacier tongue formerly bypassing the rock outcrop, now covered by 8-10 m of moraine material (Fig. 3). The overall thickness of ice exceeds instrument depth range, i.e. ice is over 40 m thick. Seepage has been detected at the outflow (probably through fissured rock bar and moraine), and at the western dam part linked to the moraine-bed interface and also likely to the buried ice englacial system.

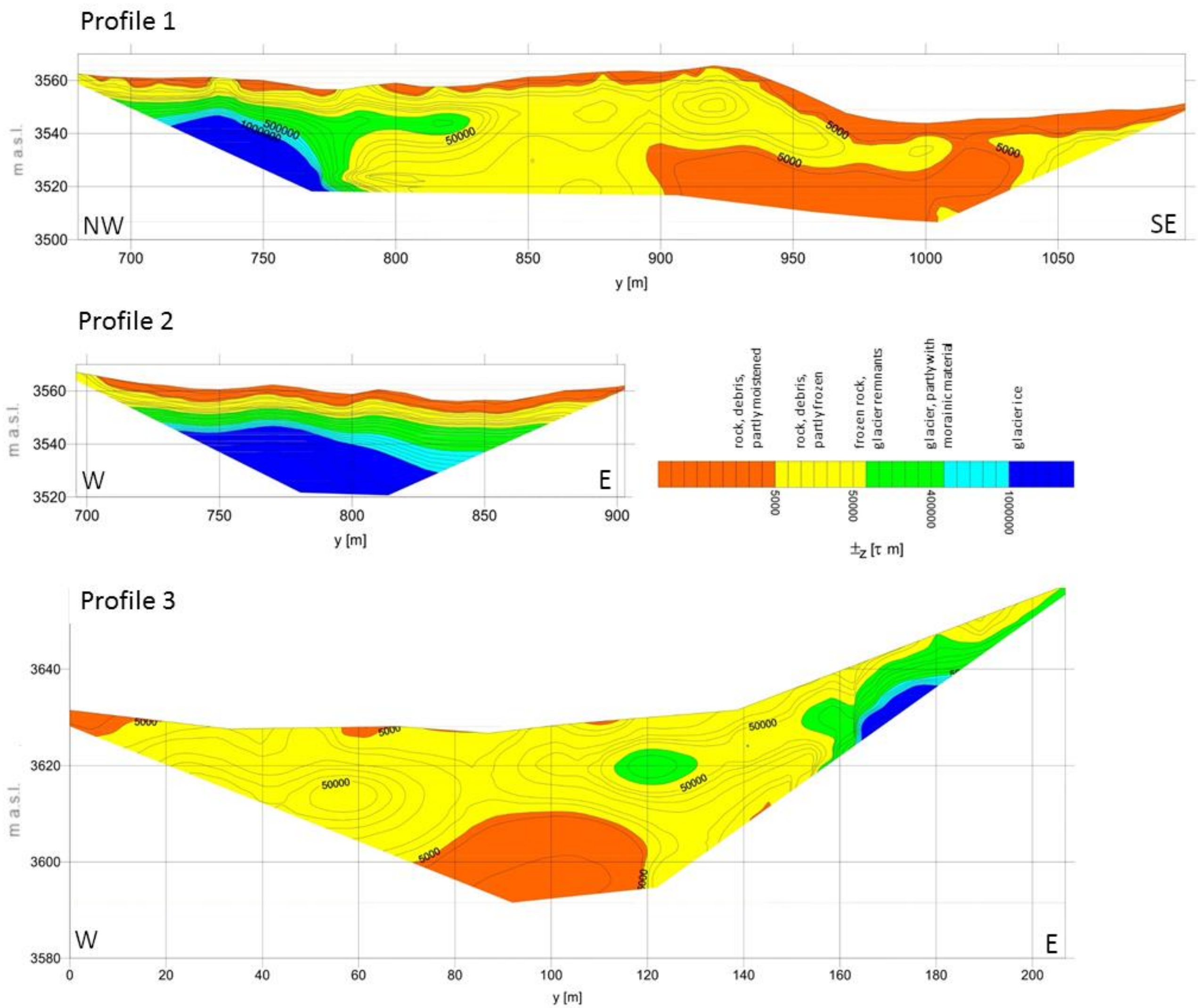

Fig. 3. Electrical resistivity results along profiles 1 and 2 for Lake 2, and profile 3 for Lake 3. For location of the profiles (P1-P3) see Fig. 1a.

3.2.3 Lake 3 and newly emerging lakes

Since 2004, several lakes have formed in the proximity of the retreating glacier terminus, filling rather shallow depressions. The lakes' development is closely linked to the amount and routing of glacier meltwater, as well as to basin stability. Our data suggest the importance of direct contact with the terminus, as only such lakes showed an increase in area. Lake 4 is an example of one which was enlarging slightly only until it lost contact with the glacier in 2011/12. Most lakes formed, enlarged or remained stable over a period of several years, and finally drained by subsurface routes (for detailed information on lakes' morphometric changes see Table A1). An exception is Lake 6, which seemed stable over the monitoring period, only until it suddenly drained in 2015, but was filled again the following year. The recently exposed area after glacier regression is underlain by buried ice lenses (visual inspection, see Fig. 1a). For this reason, the probable explanation of lake sudden drainage is linked to melting of the ice and formation of new drainage routes connected to the lake.

Although these proglacial lakes are very unstable, the hazard they pose is negligible, as only a small amount of water (up to 4 000 m^3 in Lake 4 and Lake 6) can be retained in such shallow depressions. Lake 3 is an exception; the easternmost lake is in direct contact with the glacier and is growing annually. It is situated in a deep basin - in 2007 a bathymetric survey recorded a maximum depth of 3.7 m; in 2017, as the lake enlarged, it was over 14 m. Besides the climate-driven glacier retreat, the lake has been growing due to increased glacier melting, caused by heat transfer between lake water and ice. The lake expanded from 5 710 m^2 in 2007 to 8 830 m^2 in 2012. In the past few years, the growth accelerated, leading to a lake area of 16 020 m^2 in 2017 (Fig. 4). Shortly after its formation, the lake had a partly surface outflow. However, soon the drainage switched to subsurface channels. Currently, during high lake water level, water overflows the dam. However, this minor surface drainage submerges under the surface after 5-6 metres. Geophysical survey (Fig. 3) revealed a similar dam structure to that of the Lake 2 dam – rock outcrop covered by thick moraine forms the western part of the dam, and the eastern part consists of stagnant ice buried by morainic material – remains of the eastern glacier tongue. As the volume of the retained water increases rapidly (7 800 m^3 in 2007, 29 300 m^3 in 2012, and 106 000 m^3 in 2017), this recently insignificant lake has become the centre of attention.

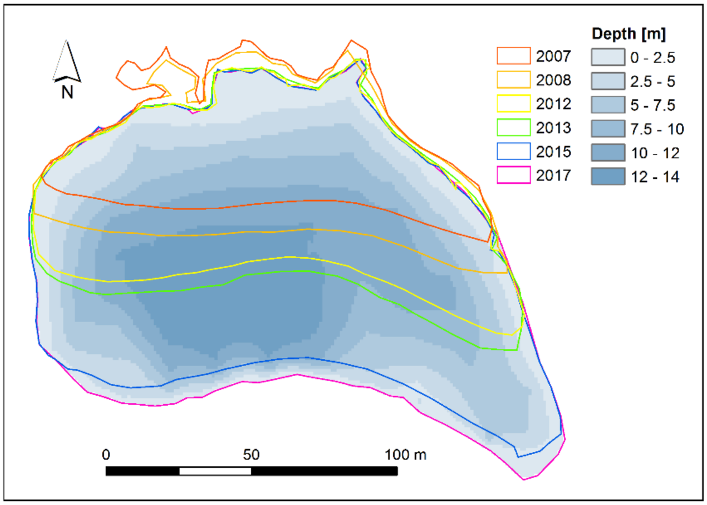

Figure 4. Bathymetric map and area changes of Lake 3 between 2007 and 2017. The lake depth data were gained during field survey carried out in 2017.

Figure 5. Thermokarst lakes of varying age in the moraine complex below Adygine glacier. These small lakes are situated at ~3450 m a.s.l., all within a few hundred meters from each other. Difference in lakes' water turbidity/colour suggests existing (left) or missing (right) recharge from melting buried ice. Photo: K. Falatkova (2017)

At the level of Lake 1 there are several small glacial lakes of varying age (Fig. 5), which are formed by thermokarst processes. As these lakes have only a very small volume and no significant area development was observed, a closer investigation from outburst hazard point of view was not carried out. However, a few lakes situated in the youngest generation of a moraine recently uncovered the basin sides formed by buried glacier ice (Fig. 5, left). These lakes are expected to further develop in future, either enlarge or drain through newly opened subsurface channels.

3.3 Expected future site conditions

According to all scenarios, the rise of air temperature in the area is very likely in future decades, with estimated higher temperature increase in spring and summer and a moderate temperature increase in autumn and winter. Scenarios of annual precipitation shows a reduction under all three scenarios. Especially summer and autumn depicts a decrease of precipitation. Winter and spring displays a minor reduction or even a slight increase of precipitation totals (for winter under A2, for spring under B1). Consequently, future mass balances simulated for the Adygine glacier by the GERM model are negative, and runoff will be altered. Low-pass filtered values of the mass balance of Adygine glacier are negative throughout the simulation period (Fig. 6a). Scenario A1B shows significantly less negative mass balances between 2015 and 2035 compared to the other scenarios. In this period, scenario A2 is characterized by the highest decadal variation with filtered values between -1 300 to -500 mm w.e. After 2030, all scenarios illustrate a significant negative trend indicating that a tipping point for glacier shrinkage will be reached. Scenario A1B shows the steepest decline for this period with a gradual recovery after 2045.

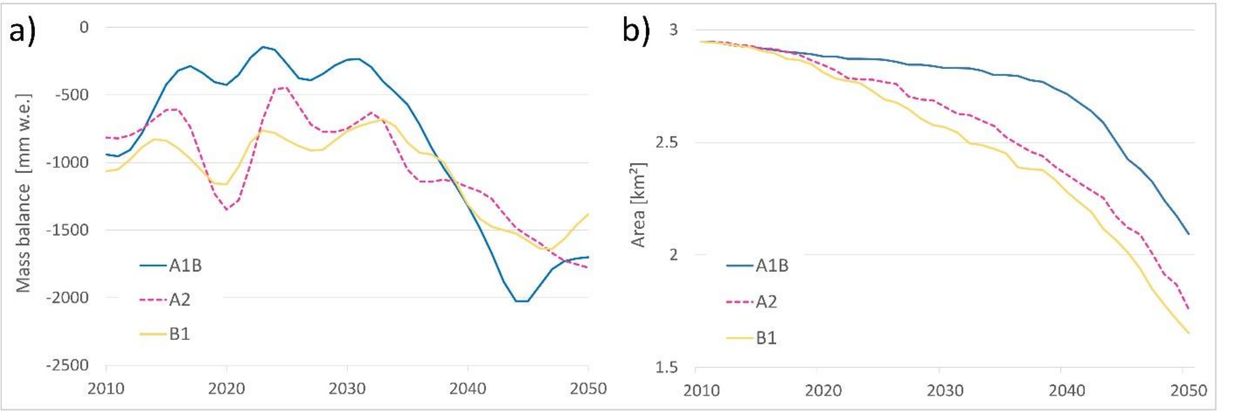

Figure 6. Glacier mass balance (a) and glacier area evolution (b) modelled according to scenarios A1B, A2, and B1 for the period 2010-2050. Data smoothed with 10-year low pass filter.

Due to the negative mass balances (Fig. 6a), all scenarios result in a reduction of the glacier area. Under the scenario A1B, the glacier area is reduced only slightly in the first two decades (2015-2035), if compared to the other two scenarios (Fig. 6b). After 2035, scenario A1B shows accelerated area shrinkage (resulting from the accelerated mass loss). All three scenarios simulate a significant decrease in glacier area by 2050, specifically, leaving 73.2 % (A1B), 62.3 % (A2), and 55.6 % (B1) of the 2012 glacier area. The model simulations indicate (Fig. 7), that under scenario B1 the glacier will disintegrate into several parts, whereas under scenario A2 only one smaller part remains. The modelled glacier retreat implies the potential for new lakes to form in the exposed area. The glacier bed topography (Fig. A2) based on GPR profiling reveals several overdeepenings. Under scenario A1B, three new lakes with total area of ca 0.1 km² can emerge, scenario A2 presents a potential for seven new lakes with a total area of 0.13 km² to form. Scenario B1 exhibits the largest modelled glacier retreat and the exposed terrain has a potential for eleven new lakes with a total area of 0.17 km².

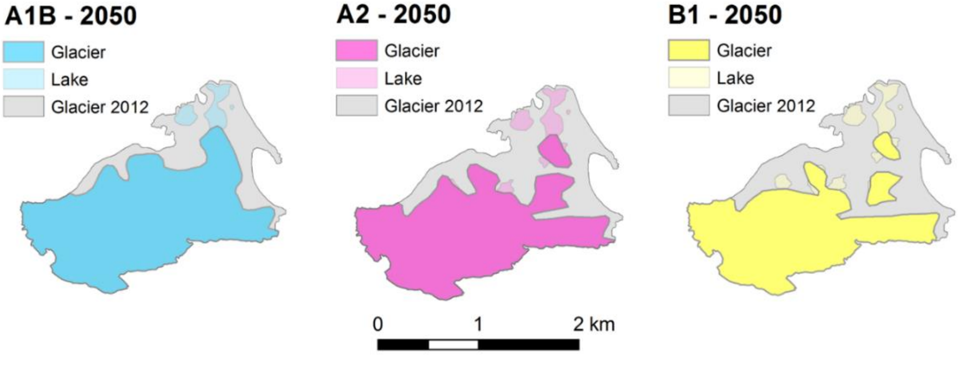

Figure 7. Modelled future glacier extent and potential areas for lake formation for the year 2050 under scenarios A1B, A2, and B1.

As a result of more negative glacier mass balances, total runoff from the Adygine glacier catchment is expected to rise. According to scenarios A1B and A2, the total runoff increases by 19.7 % and 25 % respectively, between 2010 and 2050 (Fig. 8a). However, the scenario B1 shows a rather sensitive reaction of glacier dynamics to climate change, with a clear peak of discharge values around 2020. This is followed by a decline and overall stagnation in the following decade, resulting in no significant trend in the modelled period.

Obviously, runoff for individual months will change in the future. Most pronounced changes can be seen at the beginning and the end of an ablation period, specifically in April, May, and October. Slight increase in runoff is also expected in June and September. However, for the core period of the ablation season, July and August, no significant changes are modelled. Separating total runoff into its two main components - melt from snow and melt from ice, for both today and the future (Fig. 8b), highlights a significant temporal shift for the snow component. The time of peak runoff from snow melt, which is in mid-June for the first decade of 2010-2020, will change to mid-May in the decade of 2040-2050. This will coincide with slightly earlier start of ice melt at the glacier, as the snow cover will deplete earlier as well, thus exposing the bare ice.

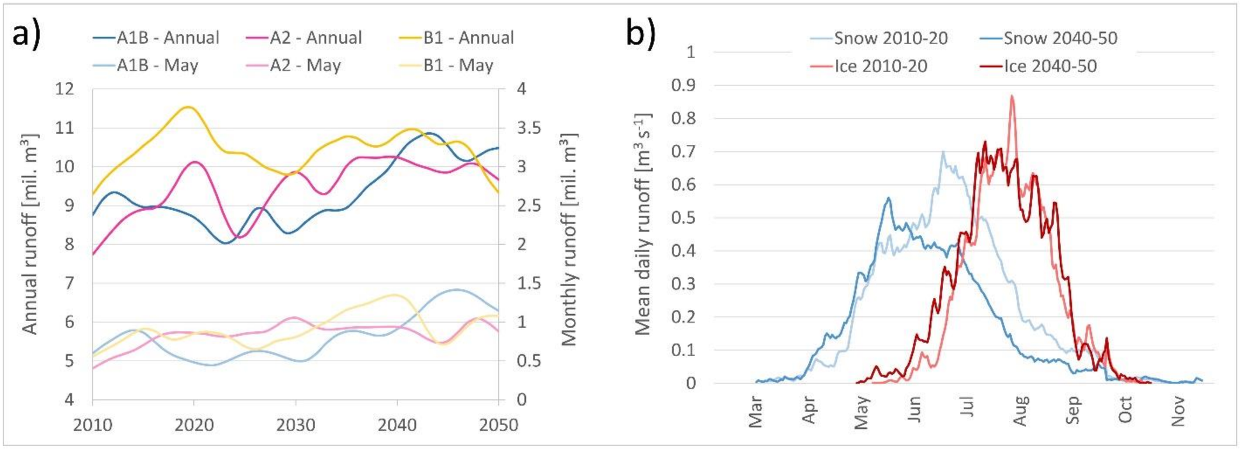

Figure 8. a) Accumulated annual runoff and monthly runoff for May between 2010 and 2050. Data smoothed with 10-year low-pass filter. b) Mean daily runoff (averaged over all scenarios) from snow- and ice-melt for the periods of 2010-20 and 2040-50.

3.4 Hazard assessment and its expected change

Overall susceptibility of the studied lakes is summarised in Table 5. Lakes 1 and 3 were marked as having medium susceptibility, whereas the largest lake of the site, Lake 2, is considered as being less inclined to cause a flood. Several areas were determined as a potential source of mass movement that could reach the lakes (Fig. 9). These are mainly the steep upper part of the glacier with visible deformations and cracks and also the eastern lateral moraine with buried ice. Both surface and subsurface hydrological connections among the lakes show routing of excess water in the case of rapid melting or lake drainage.

As shown by the model projections and also indicated by the position of ELA, the glacier is expected to shrink significantly in the following decades. Sorg et al. (2014) confirm that future hazard is likely to change due to glacier retreat and accompanying signs. South-oriented slopes in particular are most prone to destabilisation, due to rising temperature and the effect of high temperature differences caused by radiation. Our studied site has mostly a northern exposition, however, there are slopes that could be prone to destabilisation.

Table 5. Lakes' susceptibility to cause outburst flood.

Parameters	Area	Dam type	Drainage	Ice contact	Growth possibility	Total
Lake 1	low	medium	medium	low	medium	**medium**
Lake 2	medium	medium	low	low	low	**low**
Lake 3	medium	medium	medium	medium	medium	**medium**

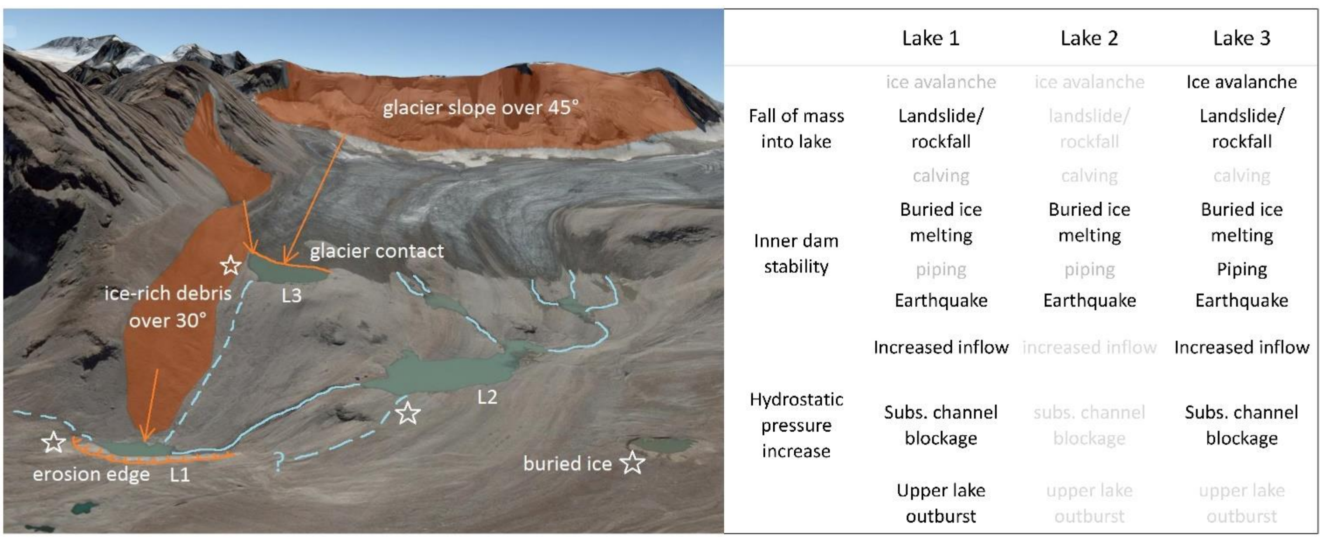

Figure 9. Geomorphological conditions and potential triggers of the lakes' outburst at Adygine complex. View from the north (Google Earth, 2018).

Lake 1 has medium susceptibility to cause flood, mainly due to its origin (intramorainic depression), subsurface drainage, and possibility for enlargement. Outburst can be triggered by a fall of mass from the north-oriented slope of unconsolidated debris which contains buried ice. The weak point is the subsurface drainage channel which leads through the moraine complex.

Changes in capacity and, above all, blockage of such a channel would lead to lake filling up and subsequent outburst. That could also be initiated by outburst or overflow of Lake 3. Hazard is expected to increase in the future, due to intensified melting of buried and exposed ice near the lake basin, and destabilised slopes in its vicinity.

Lake 2 has a low susceptibility to cause flood because of its stable surface drainage, distance from the glacier terminus and associated ending of lake basin development. As there are no proximal unstable slopes (only rock outcrops), outburst is not likely to be triggered by a displacement wave from the fall of mass into the lake. A possible weak point is represented by buried ice and water seepage in the western part of the dam that could lead to changes in subsurface drainage network. Hazard change is expected to be negligible in the future, as surface drainage can deal with possible higher inflow and lake stability will not be threatened by unstable slopes.

Lake 3 has medium susceptibility to cause flood, due to its position in contact with the glacier, rather unstable subsurface drainage and potential to grow both in area and depth. Despite the contact with the glacier tongue, calving is not probable. However, the lake is within reach of an ice avalanche from the steep glacier slope and also landslide from adjacent unstable lateral moraine. The eastern part of the dam is prone to failure, due to the presence of buried ice. Overflow of the dam caused by increased inflow from the glacier may lead to progressive dam erosion. Hazard is expected to increase together with increasing lake volume and hydrostatic pressure on the dam. Also, destabilised slopes in the lake's proximity will threaten the lake stability.

Individually, the lakes present a certain threat. Nevertheless, as these lakes are interconnected, the overall hazard of the site should be considered. In this case, a chain reaction is a very probable scenario. The lowest Lake 1 is subjected to changes in its basin morphometry and thus in subsurface drainage channel functioning. In the case of a rapid increase in drainage from Lake 2, or outburst of Lake 3, this lake would be hit. And it is rather unclear how this still-evolving, subsurface-drained basin would react. In regard to similar cases from the region, the lake filling up with subsequent rapid drainage can be expected.

4 Discussion

4.1 Influence of a glacier on lake's development

Lake 2 is a typical example of a lake formed when a glacier retreated over a rock step. The lake growth was coupled with glacier recession, until the terminus reached a higher elevation than the lake outflow. After the loss of contact, the lake stabilised and its formation ended. Lake 3 has a similar genesis and is currently in the stage of growth. Thanks to intensive thermal exchange between lake water and ice, the glacier terminus recedes by melting and sometimes even by calving in favour of lake area and depth. As a lake loses contact with a glacier tongue, some development can continue due to melting of residual ice in the lake bed or dam. The extent of changes is, however, considerably reduced.

The case of Lake 1, formed in an intramorainic depression, demonstrates the difference in development. The lake's development is not terminated or stabilised after loss of contact with the glacier. Its further evolution is linked to buried ice, its exposure and melting. The depression formed in the 1960s and filled with water two decades later. Yet even today, the lake is subjected to changes of basin morphology and drainage as well. These lakes do not grow to large dimensions like typical moraine-dammed lakes (e.g. Tsho Rolpa and Dig Tsho). However, they can hold sufficient volumes of water and cause an outburst flood. In many cases, such lakes are even non-stationary, filling-up only when subsurface channel gets blocked, often in times of increased inflow due to snow or ice melting (Erokhin et al., 2017). The 2012 outburst from the Teztor Lake, situated in a neighbouring valley at similar altitude as Adygine, serves as an example. When filled, the lake has a volume of 70 000 m^3 and is usually drained by a subsurface channel with a flow rate of a few $m^3 s^{-1}$ (Erokhin et al., 2017). Nevertheless, in 2012 the channel's capacity enlarged, the flood changed into a debris flow with peak discharge estimated at 350 $m^3 s^{-1}$ (at the junction of Teztor and Adygine valleys, ~3 km from the glacier) and 200 000 m^3 of debris was deposited at the fan entering Ala Archa valley (Erokhin et al., 2017). The flood continued further downstream, caused dismay even among the capital's inhabitants and material damage of approx. 100 000 USD (Zaginaev, 2013). It is the often variable subsurface drainage, the common presence of buried glacier remnants, and a steep valley full of loose sediments downstream, which make these lakes potentially dangerous.

4.2 Model uncertainties and future conditions

Huss et al. (2014) describe in detail the uncertainties of glacier and runoff modelling in general. In our case, GERM was calibrated with summer and winter mass balance data of Golubina glacier from 1981 to 1994. Whereas simulated winter mass balances for Golubina show no consistent trend of under- or overestimation, the modelled summer mass balance slightly overestimates the real conditions from 1981 to 1984 and generally underestimates from 1985 to 1994. The performance of the model for simulating discharge is described by the Nash-Sutcliffe efficiency (Nash and Sutcliffe, 1970), which is 0.64 based on monthly mean values of the Adygine catchment for 1960-1987. The biases and the uncertainties in the downscaled data for precipitation over the complex mountain topography remain a weakness and need further improvements in the future. Also, the lack of DEM from 1960 as well as the glacier area from this time attenuates the accuracy of the calibration of GERM simulations.

Both area and volume of glaciers all over Tien Shan are expected to decrease throughout the 21st century (Sorg et al., 2012). Aizen et al. (2007b) as well predict significant glacier degradation linked to ELA shifting to higher altitudes as rising air temperature won't be balanced out by sufficient increase in precipitation totals. Sorg et al. (2014) used the GERM for a glacier and runoff modelling in Chon Kemin valley (Zailiyskiy and Kungey Alatau, Tien Shan), where the glaciers are expected to vanish by 2080 under the more pessimistic scenarios (dry-warm, wet-warm). The future runoff results of Chon Kemin (Sorg et al., 2014) are very similar to those of our study site with 'warm' scenarios expecting peak runoff in 2020s and main change in spring runoff (increase) caused by higher winter precipitation and enhanced snowmelt in spring. A general prediction of runoff change in Tien Shan is possible increase in near future followed by steady decline till the end of the 21st century (Sorg et al., 2012). However, different regions of Tien Shan may exhibit varying responses and uncertainties connected to future runoff modelling are still rather high. An advanced state of glacier degradation is presented by Huss et al. (2016), focusing on small-sized glaciers (< 0.5 km^2) in the Swiss Alps. Runoff from these glaciers has declined since the peak runoff year, which occurred between 1997 and 2004. There is a significant decrease in August runoff which is typical for advanced stage of glacier runoff decline. Based on a comparison with expected future summer runoff from Adygine glacier, we can estimate that our study glacier is currently near the stage of runoff peak.

Our approach to the identification of potential spots for future lake formation is based on glacier bed topography derived from in-situ data. For obvious reasons, it is not possible to apply such method for whole regions. Numerical models are a good substitute and are widely used on large areas, from Peruvian Andes (Colonia et al., 2017), Swiss Alps (Linsbauer et al., 2012), to Himalaya-Karakoram region (Linsbauer et al., 2016). Commonly used model is GlabTop, introduced by Linsbauer et al. (2009) and Paul and Linsbauer (2012), later improved (GlabTop2) by Frey et al. (2014). The model calculates glacier ice thickness based on a DEM, glacier outlines and branch lines; the improved version (GlabTop2) avoids manual delineation of the lines. Similar concept and results (Frey et al., 2014) are obtained from a model presented in Huss and Farinotti (2012). The potential sites with overdeepenings can also be identified, or confirmed, based on three morphological criteria (glacier surface features) introduced by Frey et al. (2010). This is a more laborious approach, which includes manual analysis of DEMs and high-resolution satellite/aerial imagery. If we apply these morphological criteria at the Adygine glacier, one spot with a distinct narrowing and also a change of slope can be identified. This coincides with our results pointing to presence of an overdeepening at the place of current glacier tongue. Nevertheless, as Haeberli et al. (2016a) correctly points out, there are still significant limitations to understanding the principles of the depth erosion by glaciers and thus modelling of glacier bed overdeepenings. Moreover, estimate of debris cover volume left after glacier recession is a challenge as excessive debris can smooth topographic irregularities and lead to formation of outwash plains, and not lakes (Linsbauer et al., 2016). We are aware of limitations of our approach as well. Due to the low resolution of the GPR-derived ice thickness raster, the resulting extent of potential sites for lake formation must be interpreted with caution.

4.3 Lake outburst hazard assessments

The presented evaluation strategy combines knowledge from the field with benefits of remotely gained data – a good insight into the interaction of site elements (a moraine, buried ice, meltwater channels, a glacier, lakes) and a broader spatial and temporal perspective thanks to airborne imagery. Field investigations are necessary to process a realistic assessment of a lake outburst hazard. McKillop and Clague (2007) mention knowledge of seepage, lake bathymetry, and its changes as important information for understanding dam hydraulic conditions. Bolch et al. (2008) highlight subsurface glacier meltwater routing, which we see as extremely important as well, together with buried ice distribution. At the Adygine complex, the ERT technique helped to discover an extensive ice core near Lake 1, buried under a considerably thick (8-10 m) layer of debris. A similar structure was observed, for example, at the moraine dam of Thulagi glacier lake, Himalayas (Pant and Reynolds, 2000). The seepage routes and hydrological processes within moraine dams were also examined with SP and ERT in Thompson et al. (2012) at Miage glacier, Italian Alps. Remote sensing has proved to be a very useful tool in the field of environmental studies and is the basis for many hazard assessments, e.g. Allen et al. (2016), Bolch et al. (2008), Huggel et al. (2004), and McKillop and Clague (2007). Comprehensive hazard assessments elaborated as a basis for wide usage are, for example, GAPHAZ technical document (GAPHAZ, 2017), the strategy prepared by ICIMOD, focused on the high-mountain region of Himalayas-Karakoram (ICIMOD, 2011), and a useful summary of various lake outburst hazard assessments is also provided in Emmer and Vilimek (2013).

For a first, large-scale outburst hazard assessment, the possible triggers are often represented solely by fall of mass into a lake (e.g. Allen et al., 2016). The presented procedure aims to encompass all trigger factors that could have the capacity to lower dam's or lake basin's stability and cause lake drainage. However, there are still some limitations to a full understanding of the varying mechanisms of lake outbursts. We did not include permafrost degradation as a separate trigger, but we consider it one of the main factors which can increase the hazard in future, as presented for the Alps by Haeberli et al. (2017). The lower part of the site is situated at an altitude of 3500 m a.s.l., which is considered the current interface between continuous and discontinuous permafrost in the northern Tien Shan (Marchenko et al., 2007). An earthquake is definitely a phenomenon that can destabilise a structure from unconsolidated material, although e.g. McKillop and Clague (2007) decided not to include it in their hazard assessment due to low spatial variability. Several major earthquakes with epicentres in the vicinity of Kyrgyz Ala-Too were reported in the 20th century. The last case was the M7.5 quake near Suusamyr (ca 50 km from Adygine) in 1992. Also increased inflow due to heavy rainfall was reported to cause moraine dam failure (Yamada, 1998), but such events are not common in the studied region. Heavy rains have been reported to trigger a debris flow (Zaginaev et al., 2016), but have not been linked to a lake outburst.

Glacier retreat in our site's case may not lead to new avalanche-starting zones as described by Haeberli et al. (2017), but the rising MAAT could influence the glacier thermal regime and cause a shift of certain parts from cold- to warm-based, and therefore lower the friction between the glacier bed and steep ice masses. Areas exposed after glacier recession often consist of unconsolidated material which, if steep, may become a potential starting zone of landslide or debris flow. The already exposed steep slopes may be destabilised due to melting of interstitial ice or, where present, degradation of mountain permafrost (Huggel et al., 2010). Formation of new lakes at higher elevations increases the hazard of the site due to a possible cascade effect when even a small volume released could trigger the outburst of a lower-lying lake. Such situations are expected to occur in future and the potentially arising challenges should be addressed in advance (Haeberli et al., 2016b).

We would like to draw attention to the few research papers supported with field data from the region of Central Asia. A well-observed site is Lake Merzbacher and its dam, represented by the Southern Inylchek glacier, with the outburst history described by Glazirin (2010). From the latest examples, Narama et al. (2017) studied supraglacial lakes in the central Tien Shan, focusing on hydrological parameters of these small-sized and often temporary lakes, and Erokhin et al. (2017) compared two outbursts from a non-stationary glacial lake and presented hazards linked to lakes that form in an intramorainic depression. The phenomenon of repeated debris flow activity triggered by glacier meltwater release was studied by Solomina et al. (1994) using lichonometry, and recently by Zaginaev et al. (2016) by dendrochronological methods. Such studies are a very welcome contribution to assist in the improved understanding of GLOF-related processes and influencing factors.

5 Conclusion

The presented paper aims to describe the impact of glacier retreat on its forefield, to highlight the differences in evolution and outburst hazard of proglacial lakes, and to promote more field-based regionally-specified approach to hazard assessment.

Main results concerning the study site:

- The glacier is subjected to relatively fast recession comparable to other glaciers in Tien Shan, since 1962 it has retreated on average 11.5 m per year. By 2050, it’s expected to shrink to 55.6-73.2 % of its 2012 extent.

- As a consequence, glacier runoff will change in following decades. Specifically, the beginning of ablation season is supposed to occur earlier compared to current situation, and the peak runoff from melting snow will very likely shift from mid-June to mid-May, resulting in slightly earlier start of glacier ice melting.

- A three-level cascade of glacial lakes formed in the glacier forefield with the largest Lake 2 detaining volume of about 200 000 m^3. However, due to its stable surface drainage and no contact with glacier tongue it terminated its development in 1990s. Two other lakes have potential to grow, Lake 1 thanks to filling up and melting of buried ice, Lake 3 by further glacier retreat. The latter is currently the most dynamic one, it formed in 2005 and till present it has expanded to volume of 106 000 m^3 and depth of over 14 m.

- Lake 1 and Lake 3 were categorised into medium hazard mainly due to their subsurface drainage and presence of buried ice in their vicinity. Despite the large size, outburst hazard of Lake 2 was assessed as low. Destabilisation of steep slopes, exposure and melting of buried ice, and changes in glacier runoff may further deteriorate stability of Lake 1 and Lake 3 and lead to increase of the outburst hazard in future.

A very similar development of other glacier complexes in this region is expected, runoff peak occurring at present or near future, gradual melting of buried ice, and formation of new lakes. This will result in new outburst hazards that can arise in a relatively short time. One of the main outputs of the study is an analysis of lakes’ development dynamics, which indicate lake’s potential to grow and the related outburst hazard. As we documented, the lake size is not as important a factor in terms of hazard, we would rather emphasize the influence of buried ice (especially in the dry region of Central Asia) and unpredictability of subsurface drainage channels. In terms of future hazard change, we would like to draw attention to the influence of glacier runoff. Expected extension of ablation season and related activity of glacial lakes may lead to increased outburst hazard. The prolonged period of melting season in relation to buried ice will result in higher potential for alteration of hydrological network of drainage channels. Further research should be dedicated to hydrological connections in proglacial morainic environments and interactions of subsurface channels with buried ice, as there is still insufficient understanding of these processes. Such research would indeed increase the reliability of lake hazard assessments.

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

Author contribution. KF, MŠ, and BJ carried out the field work and processed the obtained data. KF designed the hazard assessment, produced the figures, and wrote the manuscript with contribution of AN and WS for the model part. ZE provided ice-thickness data from GPR survey, VB provided results of the ERT and SP surveys. AN and WS ran the glacier evolution model, HH supervised the EURAS-CLIMPACT project and initiated formation of this joint paper. All authors contributed to improvement of the paper.

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Appendices

Table A1. Morphometry of lakes at Adygine site. Lakes 3-9 are the most dynamic ones situated near glacier terminus. A hyphen is used when a measurement was not carried out for the individual lake. Asterisk signifies a value obtained from a satellite image.

	Lake	**1**	**2**	**3**	**4**	**5**	**6**	**7**	**8**	**9**
Area (m^2)	2007 (2005)	3 390	32 700	5 710	3 230	3 230	3 350	1 860	0	0
	2008	-	31 900	7 030	3 680	1 580	3 410	1 910	0	0
	2012	-	-	8 800	5 100	740	3 800	1 280	1 130	0
	2013	-	-	9 870	4 870	0	3 405	0	558	0
	2015	-	31 700	14 880	4 540	0	0	0	0	1 540
	2017	2 700*	31 900*	16 020	4 200*	0	2 700*	0	0	1 150*
Length (m)	2007	109	-	161	90	120	106	85	0	0
	2008	-	364	166	92	94	113	85	0	0
	2012	-	-	168	105	54	113	65	60	0
	2013	-	-	174	106	0	113	0	67	0
	2015	-	390	198	100	0	0	0	0	70
	2017	110*	390*	200	95*	0	90*	0	0	50*
Max depth (m)	2007 (2005)	4.4	21.6	3.8	-	-	-	-	0	0
	2008	-	22.2	-	-	-	-	-	0	0
	2012	-	-	10.3	2.4	-	3	0.5	1.8	0
	2015	-	21.3	12.3	2.4	0	0	0	0	-
	2017	-	-	14	-	0	-	0	0	-
Volume (m^3)	2007 (2005)	6 290	208 000	7 800	-	-	-	-	0	0
	2008	-	206 000	-	-	-	-	-	0	0
	2012	-	-	29 300	3 900	-	3 800	500	900	0
	2015	-	194 000	72 000	3 630	0	0	0	0	-
	2017	-	-	106 000	-	0	-	0	0	-

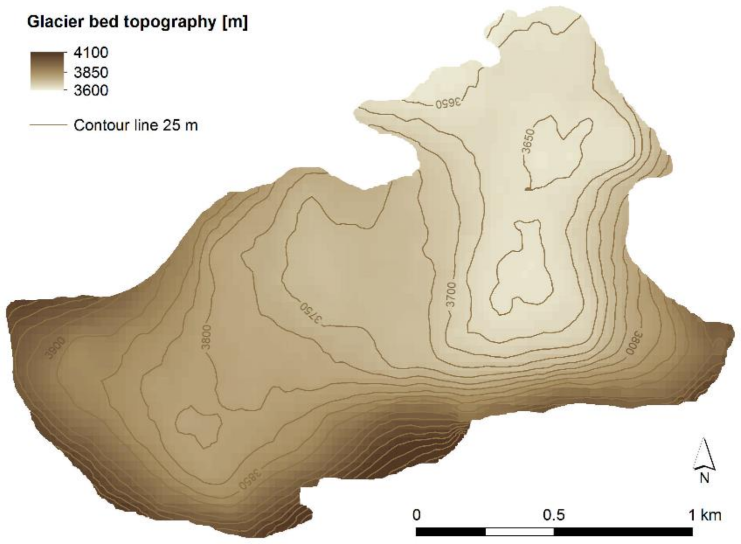

Figure A2. Adygine glacier bed topography derived from the GPR survey.

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