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Dynamics of glacial lakes and hydrological conditions of a glacio-morainic complex
(Adygine, northern Tien Shan)

Dynamika ledovcových jezer a hydrologické poměry glaciálně-morénového komplexu
(Adygine, severní Tien Shan)

Doctoral thesis

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Prohlašuji, že jsem předloženou závěrečnou práci zpracovala samostatně a že jsem uvedla všechny použité informační zdroje a literaturu. Tato práce ani její podstatná část nebyly předloženy k získání jiného nebo stejného akademického titulu.

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Abstract

The thesis deals with hydrological conditions in a proglacial environment, focusing on the development of glacial lakes and the assessment of their susceptibility to outburst. The study site is the Adygine glacier-moraine complex located in the north-facing valley of the Kyrgyz Ridge, northern Tien Shan, Kyrgyzstan, at an altitude of 3400-4200 m a.s.l. In the past 50 years, the receding glacier allowed formation of several lakes, which form a three-level cascade and are fed by glacier meltwater. Below the glacier, there is a complex of several generations of moraines, through which the glacier meltwater is routed downstream. The aims of the work were to evaluate the development of individual lakes, their susceptibility to sudden outburst and possible triggers, to estimate the probable development of the site in the future, to analyse the hydrological regime of the lakes and to obtain basic information on the subsurface flow of water from the site to the stream. For the purposes of assessing the development of the lakes, the data obtained in the field (geodetic surveying of a shore line, bathymetric measurements), as well as satellite and aerial images were used. Fluctuation of lake water level was monitored by pressure sensors and the processing of this data allowed to analyse the hydrological regime of these lakes on a daily, seasonal, and annual scale. For the purposes of assessing the susceptibility of lakes to outburst, a regionally-based approach, using field data and observation together with digital map data, have been developed. The probable further development of the site (glacier retreat, formation of new lakes) was introduced using GERM model outputs. Lastly, the passage of meltwater through subsurface routes in the morainic complex was investigated - the connection between the lower lake and the stream was tested with dye tracing method. Thanks to the observed dye concentrations in the stream it was possible to determine the duration of the water passage as well as significant dilution of the traced water in the drainage system. The connection of small tarns found in the morainic complex to melt water was found by analysing the isotopic composition of their water. Some of the tarns actually had a very similar water composition to the large lakes fed by glacier meltwater, others showed only partial or very little influence of meltwater on their hydrological balance.

Key words: Glacial lake, Proglacial area, Lake outburst, Hydrological regime, Glacial meltwater

Abstrakt

Práce se zabývá hydrologickými poměry v proglaciálním prostředí, se zaměřením na vývoj ledovcových jezer a zhodnocení jejich náchylnosti k průvalu. Studovanou lokalitou je ledovcovo-morénový komplex Adygine, nacházející se v severně orientovaném údolí v pohoří severní Tien Shan, Kyrgyzstán, v nadmořské výšce 3400-4200 m n. m. Ustupující ledovec podmínil za posledních 50 let vznik několika jezer, jež leží ve třech výškových úrovních a mají hydrologické propojení s ledovcem. Pod čelem ledovce se nachází komplex několika generací morén, jímž je tavná voda z ledovce odváděna z lokality. Cíli práce bylo vyhodnotit dosavadní vývoj jednotlivých jezer, jejich náchylnost k náhlému vyprázdnění a možné příčiny, odhadnout pravděpodobný vývoj lokality v budoucnu, analyzovat hydrologický režim jezer a získat bližší informace o podpovrchovém proudění vody z lokality do toku. Pro účely vyhodnocení vývoje jezer byla použita data získaná v terénu (geodetické zaměřování břehové linie, batymetrická měření), ale i satelitní a letecké snímky. Kolísání hladiny jezer bylo sledováno pomocí tlakových čidel a zpracování těchto dat umožnilo analyzovat hydrologický režim těchto jezer v denním, sezónním a ročním měřítku. Pro účely zhodnocení náchylnosti jezer k průvalu byl vytvořen regionálně zaměřený postup využívající data a pozorování z terénu i digitální mapové podklady. Pravděpodobný další vývoj lokality (ústup ledovce, vznik nových jezer) byl představen pomocí výstupů z modelu GERM. Na závěr byl zkoumán průchod tavné vody podpovrchovými cestami v morénovém komplexu - spojení mezi spodním jezerem a stálým ledovcovým tokem bylo testováno pomocí stopovacího barviva. Podle zjištěných koncentrací barviva v toku bylo možné určit dobu průchodu vody i silné naředění označené vody v systému. Napojení malých termokrasových jezírek nacházejících se v morénovém komplexu na tavnou vodu z ledovce bylo zjištěno pomocí analýzy izotopového složení jejich vody. Některá jezírka skutečně vykazovala velmi podobné složení vody jako velká jezera napájená vodou z ledovce, u jiných se prokázal jen částečný nebo velmi malý vliv tavné vody na jejich hydrologickou bilanci.

Klíčová slova: Ledovcové jezero, Proglaciální prostředí, Průval jezera, Hydrologický režim, Ledovcová tavná voda

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1. Introduction and objectives

In the new millennium, the pronounced retreat of mountain glaciers in high-altitude regions of the world (Barry, 2006; Radić et al., 2014; Zemp et al., 2015) and its consequences have been a ubiquitous topic at geoscientific meetings and conferences, countless scientific papers addressing various aspects of the problematics have been published. One of the implications of glacier recession are changes in proglacial hydrological conditions (Yao et al., 2007; Moore et al., 2009; Huss et al., 2010; Bliss et al., 2014) that result from varying meltwater supply and geomorphological changes of the environment. Glacier meltwater is an important component of runoff in glaciated basins, supplying the stream in a summer season it is an indispensable water source in many regions (Bradley et al., 2006; Yao et al., 2007; Akhtar et al., 2008; Sorg et al., 2012). Besides the benefits it brings in terms of fresh water source (agriculture, power generation), the meltwater accumulated in depressions forming glacial lakes can pose a threat to downstream settlements and infrastructure.

Proglacial area is a zone in front of the glacier terminus, formed after glacier tongue receded to higher altitude and left behind accumulations of debris. Its dynamics is connected to presence of permafrost and its degradation, exposure and melting of glacier ice remnants and buried ice, and the effect (erosional, thermal) of water flowing through this environment. As this specific hydrological environment controls meltwater passage from glacier to a stream and thus has potential to alter the basin runoff, deepening the knowledge of hydrological functioning of proglacial areas is essential.

This thesis presents results on research of hydrological conditions at the proglacial area in the Adygine Valley, northern Tien Shan, Kyrgyzstan. Although the research was carried out mainly at one site, the findings are transferable and hopefully bring new insights into the high-mountain (glacier-related) hydrology. The aim of the thesis was to investigate the proglacial lakes dynamics and water flow from glacier to a stream. The results are included within four scientific papers (Section 5.1–5.4) addressing the objectives to fulfil the common aim of the thesis.

The individual objectives were the following:

- To evaluate formation and development of glacial lakes in relation to glacier retreat;

- To monitor and assess hydrological regime of proglacial lakes;

- To construct a region-specified assessment of lake outburst susceptibility;

- To summarize circumstances of GLOF cases in high-mountain Asia;

- To investigate subsurface water passage through a glacio-morainic complex to a stream.

The work summarized in this thesis partly builds upon and thematically supplements an extensive research carried out by a research team of the Department of Physical Geography and Geoecology of Charles University, which was initiated in 2004. The research of dangerous lakes in Kyrgyzstan was conducted in cooperation with Czech geological company Geomin s r.o. and the Kyrgyz Integrated Hydrogeological Expedition of the State Committee for Industry, Energy and Subsoil Use of the Kyrgyz Republic, under two development cooperation projects of the Ministry of the Environment of the Czech Republic. The first one (2004-2007) was entitled "Monitoring of alpine glacial lakes and protection of the population from the catastrophic consequences of floods from moraine dam failures". It was followed by the second one (2007-2010): "Risk analysis and mitigation of the consequences of rupture of alpine lakes". Another project addressing the issue was supported by NATO SPS (2012-2013): "Glacier hazards in Kyrgyzstan: implications for resource development and water security in Central Asia". Within the first years of field work in Kyrgyzstan, the Adygine Valley was selected for further, more detailed survey and monitoring because of the convenient conditions for research of a glacio-morainic complex. Thus, in 2008 a research station was built close to the glacier terminus, on the shore of a large proglacial lake (Fig. 1). The hazard arising from the formation and development of several proglacial lakes was noticed and because the potentially inundated zone included settlements and infrastructure in the main Ala Archa Valley, further evaluation of lake's outburst susceptibility was recommended.

Figure 1. The research station Adygine after a night snowfall. View towards the south-east, 22 July 2013.

2. Scientific background

Glaciers are a major source of water, which transforms the proglacial area and determines its hydrological conditions. In the following chapters, formation and development of several types of glacial lakes is described, with focus on their stability and hazard they may pose. Lake's hydrological regime – one of the influencing factors of a hazard assessment, is presented together with meltwater runoff characteristics.

2.1 Formation and development of glacial lakes

Glacial lakes form as a consequence of glacier retreat, buried ice melting or by filling of a depression by glacier meltwater. Lake formation as well as its further development depends on characteristics of a dam, a basin, and lake's surroundings, which can be summarized by a lake type (Korup and Tweed, 2007). A simple distinction was used, for example, by Huggel et al. (2004), who categorized glacial lakes according to their dam into moraine-dammed, ice-dammed, and dammed by a rock step. There are also more complex typologies, e.g. by Janský et al. (2006), taking into account the process of lake formation besides the material forming a dam. Here, I present formation and factors influencing the development of several glacial lake types. The selection of lake types corresponds to their occurrence in the Tien Shan (Tab. 1).

Table 1. Representation of different lake types in the Kyrgyz Ridge, northern Tien Shan, in 2017. Only lakes with minimum area of 1500 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	Share
Moraine-dammed	13	14.5 %
Rock step (+moraine)	17	18.9 %
Intramorainic depression	48	53.3 %
Ice dam	4	4.4 %
Landslide dam	8	8.9 %
Total	90	100 %
Surface drainage	21	23.3 %
Subsurface drainage	69	76.7 %

Formation of lakes dammed by a moraine is linked to recession and thinning of a glacier tongue. Meltwater starts to gather behind a ridge of morainic material (former frontal moraine) left in a lower part of a valley, downstream of a current glacier terminus position. Lateral moraines blocking tributary valleys or moraine complexes can serve as a barrier as well and detain large amount of water. Depending on a dam material composition, such lakes can be drained by seepage that can gradually lead to formation of underground channels, or by a surface channel eroded into a dam by overflow (Clague and Evans, 2000). Development of these lakes is linked to a glacier – often starting as lakes in contact with glacier terminus, they grow as the terminus recedes further up. They enlarge due to melting of subaerial and subaqueous ice and also by calving of the terminus (Thompson et al., 2012). The lake's main development is completed when it loses the contact. Further changes in lake's area are mostly caused by changes in subsurface channels capacity or an overall stability of the dam (presence of an ice core; Richardson and Reynolds, 2000b).

Similarly to moraine-dammed lakes, formation of lakes dammed by a rock step is linked to retreat of a glacier tongue. In this case, it is the topography of exposed area that determines where a new lake can be formed (Linsbauer et al., 2012). An outcrop of resistant rock, often covered by a layer of morainic material, serves as a barrier behind which meltwater is collected. A depression behind such a step, where a glacial lake forms, is called an overdeepening. A lake enlarges when in contact with receding terminus and finds a steady state after the contact is lost. A dam consisting of a solid rock outcrop is much more stable compared to the one from morainic material, thus little changes in a lake's morphometry are expected after the glacier-related expansion is ended. These lakes are drained by a surface channel situated at the lowest part of the rock step, or by seepage through an eroded material covering the step (at the interface between the two layers; Janský et al., 2006).

Also in the third category of lakes, morainic material forms a lake basin; this type is a lake in an intramorainic (thermokarst) depression (Janský et al., 2009). Both formation and development of these lakes is not as closely linked to glacier and its dynamics as the two previous types. Buried ice and its melting is the main factor here, driving the lake's expansion. Moraine complexes in a proglacial area contain ice blocks or lenses, pieces of disintegrated glacier tongue which was covered with debris and subjected to thinning. Partial exposure of such an ice block leads to its accelerated melting and formation of a depression (Fig. 2). Another means of lake basin creation is subsidence due to internal melting caused by heat transfer from water flowing through a moraine. These lakes are drained by subsurface channels and thus lake filling and emptying depends on their characteristics (location, capacity). In some cases, these basins stay empty most of the time, filling up only occasionally when the draining routes are blocked or the volume of incoming water is significantly increased. Erokhin et al. (2018) call them non-stationary lakes and highlight their dynamic nature – as long as there is ice in the basin surroundings, the lake's development is not terminated and thus it cannot be regarded as stable.

Figure 2. Exposed buried ice in an empty lake basin, Adygine, Kyrgyzstan.

An ice-dammed lake forms when a glacier tongue in a main valley blocks a stream draining a tributary valley or vice versa (Ding and Liu, 1992). In case of a glacier surge, such ice barrier can be of a temporary character. Lakes of this type also form at the margin of a glacier or on its surface ('supraglacial pond'; Benn et al., 2001). It is important to note that formation of some large lakes was initiated as gradual coalescence of several supraglacial ponds (Watanabe et al., 2009). Growth of these lakes is caused by heat transfer at the contact of lake water and the ice forming the basin walls and/or bottom (Sakai et al., 2009). There are several mechanisms allowing drainage of ice-dammed lakes: ice-dam floatation (vertical displacement) leading to subglacial drainage, overflow and incision of a surface channel, drainage along a glacier-valley side interface, and mechanical failure of a dam (Richardson and Reynolds, 2000a). Some of these lakes drain regularly, when the lake level rises to the point at which the ice dam cannot withstand the increased hydrostatic pressure and thus a subglacial channel is opened (Glazirin, 2010; Huss et al., 2007).

Rock glaciers, lobate proglacial landforms with typical arcuate ridges, can also block a stream and impound a lake. Although rock glaciers have seldom been studied in terms of their influence on a river network, the number of lakes formed behind such barrier is not negligible (Blöthe et al., 2018). As Ischuk (2013) argues, these dams blocking a river were quite often misinterpreted as land- or rock slides. Inner structure of these landforms can vary significantly, and so does the hydrological network of drainage channels within the rock glacier body (Brenning, 2005; Rangecroft et al., 2015). A lake dammed by a rock glacier can thus be drained either by channels through the barrier or by an overspill usually and the contact of the lobe front and a valley side.

In general, it is very hard to predict further development of a lake without detailed knowledge of a lake's basin, especially when there is buried ice within lake basin. An important factor in this instance is the position of a lake in relation to the local water table, in other words, whether a lake is above or at the water table. A perched lake (above water table) can be drained suddenly when a subsurface conduit is created. In case of a base-level lake (at the water table), the local water table must have been lowered (e.g. by incision of a spillway) if the lake water level is decreased (Thompson et al., 2012).

2.2 Glacial lake stability and outburst hazard

When we talk about glacial hydrology and, specifically, about glacial lakes, it is impossible not to mention hazard connected to these lakes. A now very well-known abbreviation – GLOF, describes a flood initiated by a glacial lake outburst (a GLOF database was established by Vilímek et al., 2014). The high-altitude position (and therefore high potential energy of the water body), often not stable construction of its natural dam, and presence of steep slopes in its surroundings make a glacial lake potentially dangerous for downstream areas. Since recognition of this hazard, there have been attempts to identify the spots and evaluate the degree of danger that individual lakes represent.

There are numerous methods of GLOF hazard assessment, a good overview of different approaches is presented, for example, in Emmer and Vilímek (2013). The approaches to hazard evaluation vary in several respects, it can be qualitative, using the distinction of low, medium, high level of hazard (Huggel et al., 2004), or quantitative, assigning each parameter a certain weight (McKillop and Clague, 2007a). The approaches may vary in comprehensiveness. It does not refer only to the number of employed parameters/variables (ranging from 2 (O'Connor et al., 2001) to 17 (Emmer and

Vilímek, 2014)), some assessments are elaborated solely for lakes dammed by a moraine (McKillop and Clague, 2007b) while others aspire to be applicable to more glacial lake types (Mergili and Schneider, 2011). The complexity of an assessment is, of course, connected to the type of required input data (only remote sensing-based (Bolch et al., 2008) or including field data) and also depends on whether it is regionally specific. In the latter case, the suggested procedure is adjusted to the region's characteristics in terms of suitable and capturing parameters (according to, e.g. common dam failure mechanism, lake type, presence of ice core in a moraine; Clague and Evans, 2000; Worni et al., 2013; Kougkoulos et al., 2018a) or limitations in form of data availability (Frey et al., 2015).

The full procedure contains three main steps: 1) identification of potentially dangerous lakes (also called a first-order assessment); 2) detailed evaluation of the hazard (at lakes marked within the first step); 3) application of mitigation measures (at lakes posing a threat). The first step is typically an almost fully-automated process run over a large area (e.g. a mountain range), using satellite imagery and a digital elevation model (DEM; Quincey et al., 2007; Rounce et al., 2016). All lakes are identified using a normalized differenced water index (NDWI) and then selected variables are tested to determine the lakes that are potentially dangerous (Bolch et al., 2011). The second step usually includes detailed evaluation of the hazard, incorporating field survey (lake bathymetry, dam inner structure) and/or modelling of the outburst and subsequent flood (map of possibly affected area; Bajracharya et al., 2007; Schneider et al., 2014). To address the problem of downstream impact of a possible outburst, mitigation measures can be implemented in form of engineering solutions of lake stabilization (water level lowering, dam or spillway reinforcement; Reynolds, 1998). To lower the risk, measures lowering the possible impact of GLOF are taken, e.g. installation of an early warning system, demarcation of inundation zones (Hegglin and Huggel, 2008). An example of a comprehensive hazard assessment procedure is a technical guidance document by GAPHAZ (2017) or a strategy prepared by ICIMOD (Ives et al., 2010), focused on the high-mountain region of Himalayas-Karakoram.

Probability of a lake outburst and thus existence of a hazard is determined by lake dam stability and potential for an outburst triggering process (Richardson and Reynolds, 2000a). Whether a lake is susceptible to failure, is commonly determined from dam material (ice, moraine, rock type), dam geometry (width-to-height ratio, freeboard, slope of a downstream side), dam inner structure (ice core, piping/seepage) or lake characteristics (lake area and its changes, depth, volume, drainage type) (Worni et al., 2013; Kougkoulos et al., 2018b). The possible external triggers (Fig. 3) of an outburst include impact of a mass into a lake (landslide, rock fall, ice/snow avalanche, calving glacier terminus), increased inflow (outburst of an upstream lake, heavy rainfall, intense snow/ice melt) or seismic activity (Clague and Evans, 2000; Yamada, 1998). The main mechanisms of a lake outburst are overtopping due to a displacement wave or increased volume, dam failure and drainage by subsurface channels. In the first case, a surface channel may be created by erosion, lead to further incision and lake water level lowering (so called progressive breach; Worni et al., 2014). A collapse of a dam is linked to weakened inner structure (moraine dam degradation due to melting buried ice, Richardson and Reynolds, 2000b) and can be initiated by a subsurface channel opening. A glacial lake can also drain, partially or completely, by subsurface channel opening, leaving the dam intact. This case is relatively common in Tien Shan (Erokhin et al., 2018).

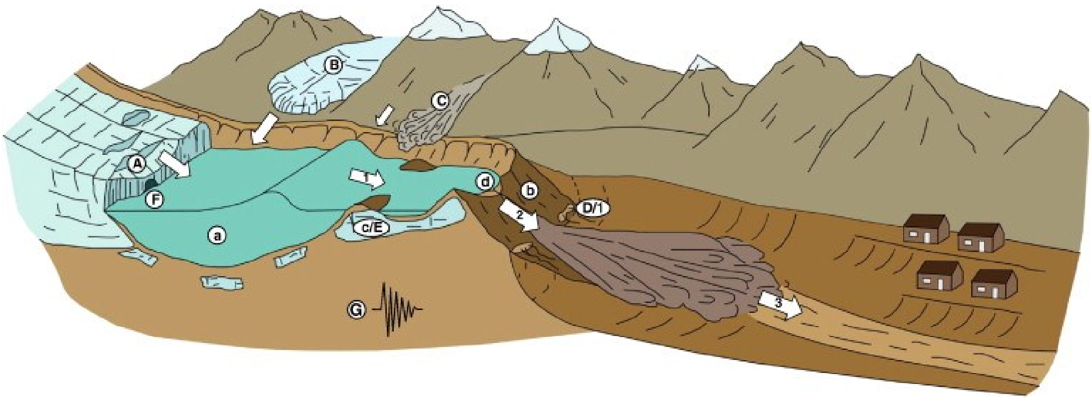

Figure 3. Possible triggers of a glacial lake outburst flood. (Source: Westoby et al., 2014)

With changing climate and expected further glacier retreat, a question of future lake outburst hazard situation arises (Dussaillant et al., 2010). Although it is hard to predict many aspects of a hazard development, some influencing factors can be well estimated or modelled. Possible shifts in temporal and spatial distribution of precipitation and higher mean air temperatures lead not only to glacier tongue recession (Sorg et al., 2012), but also to a change in permafrost distribution and destabilization of slopes (Haeberli et al., 2017). Glacier melting pattern throughout the ablation season can change as well (Immerzeel et al., 2012), having consequences on lakes' hydrological regime (and thus their stability) that are difficult to estimate. In terms of emergence of new lakes in response to glacier retreat, focus has been put on modelling of glacier bed topography and detection of overdeepenings (Linsbauer et al., 2012; Frey et al., 2010). These newly formed lakes can present a new threat but also contribute to higher hazard in connection to downstream-lying ones, creating a precondition for a chain-reaction.

2.3 Glacial hydrological regime

Hydrological regime is commonly assigned to a stream, whereas lakes are described by their hydrological balance. However, in case of lakes with a notable inflow we can describe their hydrological regime on the basis of their water level fluctuation. To determine a lake's regime, the major source of incoming water must be recognized. There are four main water sources that can contribute to a flow – it is glacier ice melt, snow melt, precipitation, and groundwater. In case of a glacial regime, glacier meltwater plays a dominant role; however, the other components can be present as well. The volume and/or share of each source on the flow changes throughout the year (Singh and Singh, 2001). In a flow with a typical glacial regime (if we take an example from the northern hemisphere), the ice melting contributes mainly from June/July to September, supplemented with snow meltwater from late April to June; a certain share comes upon precipitation (annual distribution dependent on a climate zone) and groundwater (minor intra-annual changes; Röthlisberger and Lang, 1987).

A glacial hydrological regime can be described in different temporal scales, following a water level fluctuation during a day, its development in an ablation season, and also throughout a year – i.e. in a daily, seasonal, and annual time frame. A pronounced daily fluctuation is typical for a glacial regime and it has a distinct character (Singh and Singh, 2001). The lowest water level is observed in the morning a certain period of time after the ice melting starts. The water level (or discharge in case of a stream) rises till the afternoon when a peak occurs (a certain time after the sun's culmination, i.e. highest intensity of melting). This time lag depends on a distance of the gauging station (or spot of measurement) from a glacier terminus and timing within an ablation season (Irvine-Fynn et al., 2011). Since the water level reaches its daily maximum, it declines until the next morning. The daily fluctuation occurs only during ablation season (mainly July and August) when glacier ice melts.

There are also typical seasonal features characterizing glacial hydrological regime. Due to a changing share of snow melt in favour of ice melt in the first part of an ablation season, the daily water level fluctuation changes. To be specific, the daily water level peaks become higher because glacier ice has much lower albedo than snow (and it becomes even lower with impurity accumulation) and thus melting is more intensive (Hock, 2005). Moreover, water storage capacity of ice is minor compared to the one of a snow cover. Not only the peak value, but also the amplitude of such fluctuation increases (Huss et al., 2007) because glacier melting is dampened during the night. The third distinct characteristic is an earlier occurrence of daily peak water level (Fig. 4). It is linked to the gradual development of glacier drainage channel network that becomes more effective and routes the meltwater faster to the glacier foreground (Nienow et al., 1998).

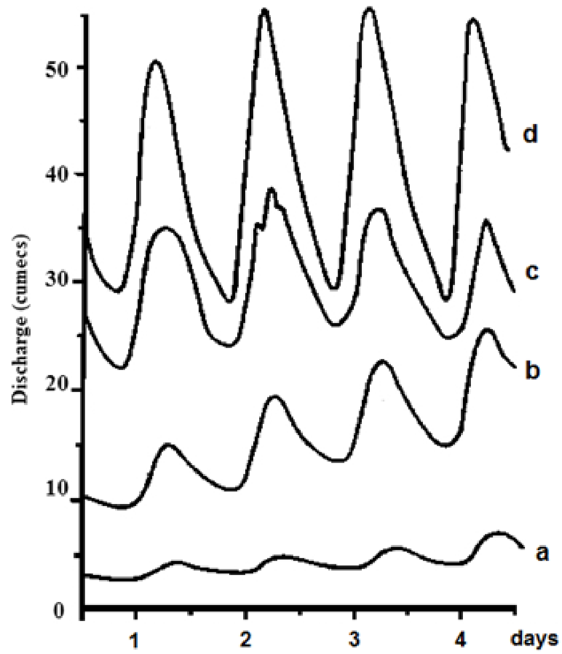

Figure 4. Development of a daily amplitude of a glacier meltwater-fed stream. a) 17–20 May, b) 14–17 June, c) 23–26 June, d) 19–22 July. (Source: Hubbard and Glasser, 2005)

As mentioned before, the main component of glacial regime is glacier meltwater and thus the annual course of water level corresponds to that (Déry et al., 2009; Janský et al., 2011). During the winter, the flow is fed mainly by groundwater so it is rather low without notable daily or monthly changes (Nepal et al., 2016). By the end of April or in May, snow cover accumulated during the cold season starts to melt and the water level rises. The water level fluctuations are often on a several-day basis depending on the air temperature. More pronounced peaks that occur in this period can be linked to rain-on-snow events which cause accelerated melting (Singh et al., 1997). As a snow cover of the lowest part of a glacier melts away, ice is exposed and its meltwater begin to contribute to the total runoff (usually during June). This brings subdued daily fluctuation which gradually develops and becomes more pronounced as the runoff share of ice meltwater increases. September or early October is characterized by subdued daily fluctuation and overall lower water level, followed bya drop to the base level values of a cold season. Summer liquid precipitation plays a certain role reflected in rainfall episodes and temporal increase of water level. In the glaciated basins influenced by summer monsoon (Thayyen and Gergan, 2010), the influence of rainfall is significantly higher (precipitation maximum overlaps with glacier melting maximum). Glacier, however, serves as an important balancing factor concerning runoff. When it comes to larger timescales, the larger the glacier-covered part of the basin, the lower inter-annual variability of runoff there is (Fleming and Clarke, 2005; Moore et al., 2009).

Hydrological regime is not monitored often at glacial lakes, the focus is mainly on lake area changes and its dam. However, lake's hydrological conditions are considered one of the factors influencing lake 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 knowledge gaps in subsurface glacier meltwater routing into a lake. It is often an ice-dammed lake that is studied also from the hydrological perspective as these lakes tend to burst repeatedly. Lake Merzbacher in central Tien Shan, Kyrgyzstan, could serve as an example. The lake is dammed by the southern branch of Inylchek Glacier and its outbursts have been recorded since the beginning of the 20th century (Glazirin, 2010). For that reason, the lake level has been monitored in order to forsee the upcoming outburst. Similar case is Görnersee, an ice-marginal lake situated at the confluence of two tributaries of the Gornergletscher, Switzerland. The annual regime of both lakes is very similar: they start to fill-up at the beginning of an ablation season and then drain abruptly by subglacial channels within several days (Huss et al., 2007). The level fluctuation of both Görnersee and Lake Merzbacher exhibit a certain development over the monitored period, leading to earlier timing of water release (Glazirin, 2010; Huss et al., 2007). This could be attributed to earlier onset of a melting season, more intensive melting, or thinning of the ice-dam and changes in its internal structure (Huss et al., 2007).

2.4 Glacier meltwater passage and runoff changes in future

A glacier meltwater component of a basin's total runoff emerges on a surface and within glacier itself, passes through the glacier, then it may be delayed while passing through a proglacial area, to finally form a stream. Glacial drainage systems can be divided into a supraglacial, englacial, and subglacial one. Water from melting of glacier surface merges into streams that route the water off the glacier body, incising sinuous or even meandering channels (Cuffey and Paterson, 2010). The surface water system is connected with subglacial system by moulins and crevasses that route meltwater into the glacier (Fountain and Walder, 1998). The inner hydrological system could be described by concepts of channelized (Röthlisberger, 1972; Nye, 1976) and distributed drainage (Flowers and Clarke, 2002), seasonal changes between the two (Nienow et al., 1998), a fast (arborescent) or slow (non-arborescent) drainage system (linked-cavity network; Shreve, 1972; Fountain et al., 2005), where water flows according to hydraulic potential but its movement is also driven by differences in channels' water pressure (Mathews, 1964). The complexity of water flow through a glacier is well summarized in a paper by Walder (2010).

Proglacial area can have a substantial influence on glacial stream discharge characteristics. Meltwater can be delayed on the surface, when impounded in a depression as a proglacial lake, or when routed through underground channels. Proglacial landforms, such as a moraine complex or a rock glacier, represent a largely permeable environment through which water flows but are considered an important water storage as well (Jones et al., 2018a). Thanks to that, they have a significant impact on basin's hydrology acting as 'a buffer', because their contribution can balance out low flows during a dry season or in arid climate (Croce and Milana, 2002; Rangecroft et al., 2015; Duguay et al., 2015). Hydrological systems of these landforms are thus studied just due to their impact on a stream's discharge (both its volume and timing of peaks; Langston et al., 2011; Jones et al., 2018b). Very useful methods uncovering the hydrological functioning of a such landform include a GPR survey, seismic refraction, tracer tests or analysis of chemical properties of water (Buchli et al., 2013; Langston et al., 2013; Krainer and Mostler, 2002; Penna et al., 2014). The underground flow is influenced by interaction with surface water bodies, presence of buried ice blocks and lenses, position of 0°C isotherm and presence of permafrost (Buchli et al., 2013). Due to the complex inner structure of these landforms (layers and positions of varying permeability), routing of water through the system has dual character

(Winkler et al., 2016) – there is a fast flow closer to surface (flow velocity reaching up to 0.09 $m^3 s^{-1}$; Krainer and Mostler, 2002) and a deeper, slow flow which can delay water within the landform for months or even years (Pauritsch et al., 2017). Due to changing climatic conditions in mountain regions all over the world (Pepin et al., 2015), changes in hydrological functioning of this 'buffer area' can be expected as a result of changing input (i.e. glacier runoff), melting of buried ice blocks or permafrost degradation (Cooper et al., 2011).

Notable hydrological changes in high-altitude areas are expected due to overall warming throughout the 21st century (IPCC, 2014). As a result of more intensive glacier melting and thus long-term negative mass balance, most mountain glaciers are expected to shrink and cause a distinct change in runoff (Huss and Hock, 2018). The most prominent impact is expected at basins with rather dry climatic conditions where glacier meltwater signifies a vital water source (Milner et al., 2017). Also the underground water levels in proglacial area are expected to be affected by glacier retreat (Levy et al., 2015). Most projections of future hydrological response of glaciated catchments are consistent regarding the general development (Lutz et al., 2014; Huss et al., 2010; Hagg et al., 2013) – earlier start of an ablation season and thus runoff from melting snow, followed by earlier onset of exposed glacier ice melting. The peak runoff will therefore shift from July/August to July and later to June (Bliss et al., 2014). The shift of a snowline to higher elevation will cause more precipitation to fall as rain instead of snow and so a basin will have lower snow storage capacity (Nepal et al., 2016). This will result in lowered base flow, which may initially seem like a minor problem, but in long-term (when glacier meltwater contribution to total runoff will be lowered) the insufficient recharge of groundwater may lead to serious trouble for water resource management (Nepal et al., 2016). The impact of the climate changes on hydrology of individual high-mountain regions are presented in numerous scientific papers; here is an example from the Himalayas (Miller et al., 2012), the Tien Shan (Sorg et al., 2012), the Alps (Huss, 2011), the Canadian Rocky mountains (Stahl et al., 2008), or the Andes (Vuille et al., 2008). The overall prolongation of ablation season and variations in glacier runoff pattern will also lead to changes in proglacial lakes' hydrological regime and hardly predictable effect on their stability.

3. Applied methods and data

The first part of this chapter is dedicated to the description of the study site‘s (3.1). The following subsections describe the methods used to investigate the formation and development of proglacial lakes (3.2), hydrological conditions of the site (3.3), and to assess lake outburst susceptibility (3.4).

3.1 The study site

The study site (42°30’10’’ N, 74°26’20’’ E; Fig. 5), a glacier-moraine complex Adygine, consists of a relatively small-sized glacier (area of 2.8 km^2), a three-level cascade of lakes of varying age and type, and a large morainic landform with buried ice. The proglacial lakes formed after recession of the glacier tongue, the first (Lake 2) appearing around 1960 behind a rock outcrop (3540 m a.s.l.), the second (Lake 1) in the late 1980s in a morainic depression (3450 m a.s.l.), and the last generation of lakes has emerged in the proximity of the current terminus position (3600 m a.s.l.) since 2005. All these lakes are fed by glacier meltwater and are interconnected by surface or subsurface channels. The lakes’ development is linked to the glacier (lake growth when in contact with the tongue), meltwater routing, character of a basin and subsurface drainage channels, and melting of buried ice in the lake’s vicinity. On the morainic landform in the lower part of the complex, there are also a number of tarns - small ponds of meltwater without surface inflow or outflow, which formed as a result of thermokarst processes.

After reaching the glacier terminus, the meltwater continues via two main routes: larger part through small proglacial lakes to Lake 2 and by a surface channel to Lake 1; eastern part of the glacier is drained by sub- or englacial channels into Lake 3 and then water flows below surface into Lake 1. As this lake is situated in a deep morainic depression, it is drained solely by subsurface channels. It is therefore the lowest spot where water flows on the surface. The meltwater emerges to the surface forming a stream only after more than 3 km, at an elevation of 2900 m a.s.l.

The site was selected for further research due to several reasons. First, the glacier itself is a good representative of glaciers in the region. Compared to those within the Kyrgyz Ridge, it is actually of medium size (Bolch, 2015). Because of its fast retreat, it is possible to observe dynamic changes of the proglacial area in a relatively short time (60 years). Several generations of moraines, large amounts of perennially frozen creeping debris mixed with buried ice blocks, complete the overall complex appearance of the site, which shows favourable conditions for lake formation. Also, the possibility to study development of different lake types at one site is an advantage. Some lakes are in the early stage of the development, others are stagnant or their basins are filled with sediments. Finally, the site's position increases its importance, especially in terms of creating an outburst hazard assessment. The Adygine valley is a left tributary of the Ala Archa valley, where a well visited National Park was established in 1976. The main valley is then directing all the meltwater through several villages towards the Kyrgyz capital, Bishkek, which is only about 40 km distant.

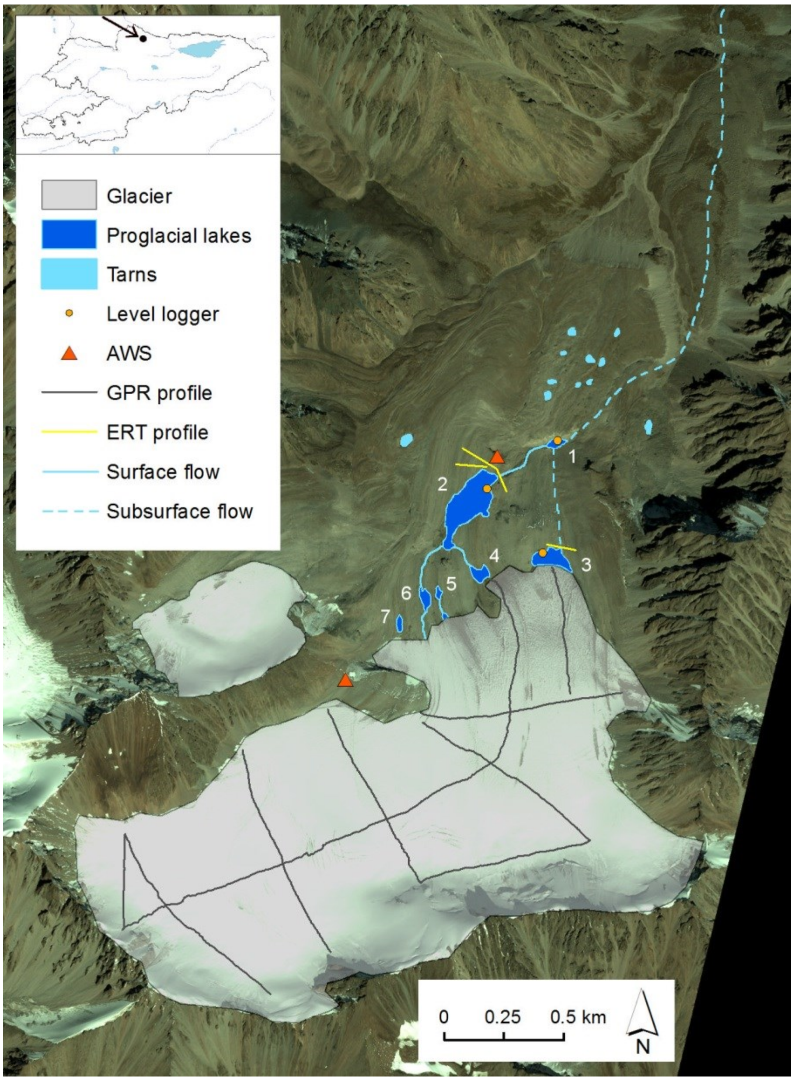

Figure 5. The study site Adygine in the Kyrgyz Ridge, northern Tien Shan. The numbers 1-7 refer to studied proglacial lakes. The site's position marked with an arrow on the map of Kyrgyzstan, upper left corner.

3.2 Formation and development of the lakes

First, the possible extent of the problematics in the study site's region was investigated. In order to assess the number of glacial lakes in the Kyrgyz Ala-Too, freely available multispectral satellite images from USGS portal Earth Explorer were used. After merging the individual images into a mosaic, a normalized difference water index (NDWI = [NIR-blue] / [NIR+blue]) was applied, which is commonly used to detect water bodies (Bolch et al., 2008). Manual corrections of the automatically delineated areas (e.g. misclassification of shadows) improved the result.

In terms of the study site, we analysed the gradual glacier terminus retreat and resulting formation of lakes based on analysis of historical aerial and satellite imagery. The oldest aerial images of the site date back to 1962 (scale of the survey: 1:38 600, image resolution: 1 m) and were acquired from the Integrated Kyrgyz Hydrogeological Expedition under Kyrgyz State Committee for Industry, Energy and Subsoil Use, Bishkek, Kyrgyzstan. The satellite imagery includes freely available Landsat data accessible via USGS portal, but also a purchased VHR image from WorldView-2. The glacier terminus outline and lakes' shoreline were vectorised manually based on individual orthorectified images. The spatial resolution of the images vary from 30 m (Landsat) to 2 m (WorldView-2), however, for the scope of the approximate dating of lakes formation, it is sufficient. The lakes's spatial development and further terminus retreat has been observed in detail since 2007 by means of geodetic surveying. A system of control points was established at the site in order to create reference for the mapping results of the individual years. The total station Leica TCR 705 with a reflective prism was used, accuracy of the measurement is 0.005 m.

As lake development isn't limited to its areal extent, so the lake basins have been surveyed repeatedly as well since 2008. The method of bathymetric measurement is described in detail by Šobr and Česák (2005). We used an echo sounder (Garmin Fishfinder) mounted to a boat and measured lake depths at previously determined profiles with a step of 2.5 or 5 meters (the latter used at the largest lake). The data points were processed in ArcMap (ESRI software) and a suitable interpolation method was applied. With regard to the expected gradual changes in basin morphometry and irregular distribution of data points (due to measurement along profiles), the most suitable method proved to be Kriging. A volumetric curve (showing dependence of lake volume on depth) was created for each lake. Based on the curve and known lake water level (3.3), volumetric changes of lakes could be determined both inter-annually and on a daily scale.

3.3 Hydrology of the site

In order to determine and monitor hydrological regime of the site, pressure sensors (Levelogger M5, Solinst) were installed in the main three lakes which are fed by glacier meltwater. These sensors measure lake water level fluctuation based on differences of absolute pressure (i.e. water (gauge) pressure + atmospheric pressure). To compensate for atmospheric pressure changes, data from a barometer (Barologger Edge, Solinst) installed at the site were used. The lake water level is recorded in 30-minute steps with accuracy of 0.006 m. The longest data series is available for Lake 2 (2007–2017), the other two lakes were monitored in the period of 2012–2015. The data processing and further analysis were carried out in R software. The lake level fluctuation data were supplemented with meteorological data (mainly air temperature) from automated weather stations installed at the site at the altitude of 3550 m and 3700 m a.s.l.

The glacier meltwater flows mainly through subsurface channels in the proglacial area, thus it is problematic to measure the total discharge. The only spot where a discharge measurement is possible is the surface stream leading water from Lake 2 to Lake 1. We measured the flow rate (3–5 August 2012) with a hydraulic propeller (OTT C2) in accordance with ČSN ISO 748 at 19 verticals with a 0.5 m steps. A rating curve was established based on the measured flow rates and the respective water level values. The calculated discharges (for every water level value, i.e. in 0.5-hour steps) served to quantify the water volume which passes through the lake on a daily, monthly, or seasonal scale. Unfortunately, the same could not be carried out for Lake 3, through which meltwater from the eastern part of the glacier's watershed is transported, as both inflow and outflow from the lake are of subsurface character.

Besides the main lakes, small tarns situated on the glacio-morainic landform were studied in terms of their possible linkage to groundwater. These ponds, having no surface inflow or outflow, differ in water colour (turbidity), thus were suspected to have varying level of subsurface meltwater recharge. Water from the tarns was sampled (25 July 2017) in order to have it analysed for ratio of stable isotopes of oxygen (^{18}O) and hydrogen (^{2}H). The analysis was carried out in the Isotopic laboratory in Ceske Budejovice, Academy of Sciences of the Czech Republic.

The resulting values of individual lakes were compared with the ratio found at the main lakes and also in solid precipitation at the site. Similarity to the lakes fed by glacier meltwater would suggest significant share of glacier meltwater recharge in a tarn's balance. Samples enriched in heavy isotopes or having lower D-excess value could, on the contrary, point to notable influence of liquid (summer) precipitation or influence of evaporation (Fig. 6). To check the correctness of the results and to put it in a broader perspective, freely available data from the region were used. Through the Nucleus portal of the International Atomic Energy Agency (IAEA), data from Global network of isotopes in precipitation and in rivers (GNIP and GNIR databases) were employed for comparison.

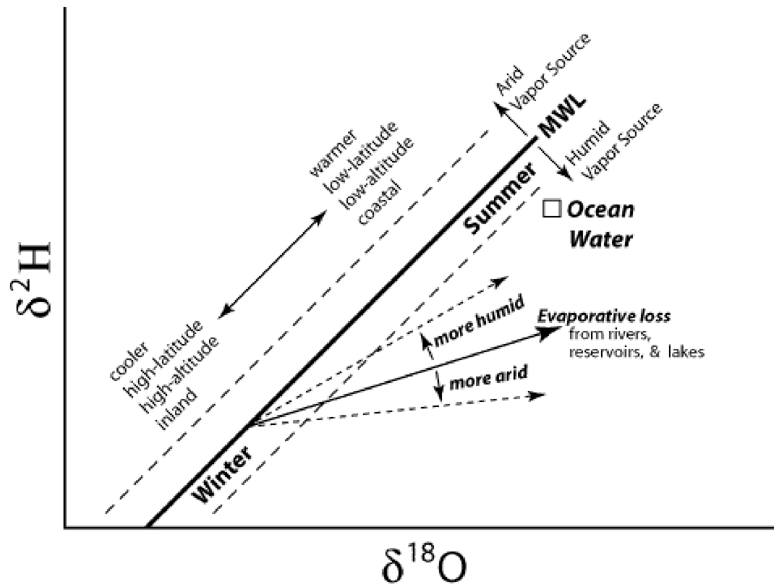

Figure 6. Summary of effects of climate and hydrological processes on isotopic composition of water. (Source: Clark and Fritz, 1997)

At last, to examine the underground meltwater passage through the morainic material to a stream, a dye tracer test was carried out (22–24 July 2017). A fluorescent dye called uranine (Fluorescein Sodium salt) was selected for this purpose as it is non-toxic, readily soluble in water, and it is detectable even in very low concentration (10^{-12} g ml^{-1}). The disadvantages include lowering dye detectability with low pH (fluorescence at 50% of maxima when pH is 6.5, Käss 1998), light sensitivity, and adsorption onto organic matter. In our case, the pH of water is sufficiently high (7.2), significant sorption of the dye is not probable, and the test was planned for the night time. An amount of 3000 g of uranine was injected to Lake 1 because from there on, the meltwater is routed only below the surface. The water samples were collected in 1-hour intervals from a stream, 3100 m downstream from the injection spot (straight-line distance). As a back-up, three passive samplers with granulated charcoal were installed in the stream. Before the test itself, blank samples were collected to establish background concentrations. The samples were analysed in a fluorometer (LS55, Perkin Elmer) with an excitation wavelength of 492 mm, the uranine emission peak was observed at the intensity of 512 mm. The height of the resulting peak was compared to the standard concentrations of 10^{-10}–10^{-12} g ml^{-1}, and a breakthrough curve was plotted.

3.4 Lake outburst susceptibility

The research on lake outburst susceptibility was initiated with a scientific literature search of GLOF cases in Asian high-mountain areas. The cases were categorized according to the outburst causes, mountain ranges, and date of occurrence. Temporal analysis was carried out focusing on GLOF occurrence throughout the 20th century and distribution of the cases within an ablation season.

In order to assess the possible triggers of lake outburst at the site, a combination of repeated field mapping, analysis of DEM and satellite images was applied. The global 1-arcsecond SRTM DEM (resolution of ~30 m) served to assess steepness of slopes (surrounding the lakes) that could be a spot of initiation of a gravitational process such as rock fall, landslide, or snow/ice avalanche. Information on development of the site's geomorphological features were derived from temporal series of satellite images available through Google Earth platform. These include buried ice exposure, erosion of lake banks or changes in surface channel network. Other possible triggers (mainly connected to lake dam stability and hydrological conditions of the site) had to be evaluated according to the results of a field survey. Hydrological methods applied at the site are described in Section 3.3, for examination of inner dam stability, a geophysical survey was carried out in 2008 by a company G Impuls Praha spol. s r.o. Methods of electrical resistivity tomography and spontaneous polarisation were used to investigate presence (and depth) of buried ice and seepage routes leading through a dam.

The parameters characterizing lake's susceptibility to burst and cause flooding were selected according to the regional characteristics and based on knowledge of previous outburst cases in the region. The qualitative assessment scheme of the outburst hazard draws also from several published assessment procedures (Ives et al., 2010; Allen et al., 2016; Frey et al., 2010; and Huggel et al., 2004). The total hazard is introduced as a combination of a lake's susceptibility to burst and presence of possible triggers that have capacity to cause outburst.

Future development of the site is built upon results of glacier evolution model (GERM, Huss et al., 2008). Because of the expected reduction of the glacier area extent in 2050, the topography of the glacier bed was calculated by subtraction of a glacier thickness layer from the DEM. The glacier ice thickness values were obtained from a GPR (ground penetrating radar) survey carried out in 2012 by Dr. Z. Engel. With the known exposed topography, potential spots for formation of new lakes were identified (detection of overdeepenings in ArcMap software). Besides formation of new lakes, there are other factors that could influence the outburst hazard in the future – e.g. changes in glacier runoff which would have an impact on hydrological regime of lakes.

4. Author's contribution statement

1. Temporal analysis of GLOFs in high-mountain regions of Asia and assessment of their causes

- Falátková, 2016, AUC Geographica, WoS 100%

2. Development of proglacial lakes and evaluation of related outburst susceptibility at Adygine ice-debris complex, northern Tien Shan

- Falatkova et al., 2019, Earth Surface Dynamics, IF 3.176 60%

3. Hydrological Regime of Lake Adygine, Tien Shan, Kyrgyzstan

- Falátková et al., 2014, Geografie, IF 0.787 60%

4. Hydrological and isotopic characterisation of proglacial lakes and their connectivity, Adygine glacier-moraine complex, northern Tien Shan.

- Falatkova et al., 2019 (in rev.), Hydrological Sciences Journal, IF 2.061 60%

I confirm the contribution of K. Falátková in the above mentioned publications.

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Prof. RNDr. Bohumír Janský, CSc.

Figure 7. Proglacial area below the Adygine glacier. View from the south-west, July 2012.

5. Selected publications

5.1 GLOF temporal and spatial distribution and its causes

Citation:

Falátková, K. (2016). Temporal analysis of GLOFs in high-mountain regions of Asia and assessment of their causes. AUC Geographica, 51(2), 145-154.

TEMPORAL ANALYSIS OF GLOFS IN HIGH-MOUNTAIN REGIONS OF ASIA AND ASSESSMENT OF THEIR CAUSES

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ABSTRACT

Glacial lake outburst flood (or shortly GLOF) has become a well-known phenomenon, one of natural hazards occurring in glaciated high mountain areas of the world. The aim of this study was to investigate temporal distribution of these events in Asia and to assess causes of lake outbursts. Therefore, a search of scientific literature and reports was carried out resulting in 219 flood cases found. In order to detect possible differences in temporal distribution a group of ice-dammed lakes was detached and compared with the rest. Concerning spatial distribution of GLOFs, it is influenced by availability of scientific literature which is determined by research teams' region interest. Temporal analysis revealed a certain pattern in ice-dammed lake outburst distribution and notable difference between the two lake groups in terms of outburst occurrence within a year. The moraine-dammed lake outbursts were recorded earlier in an ablation season (compared to ice-dammed lakes) which could be attributed to different mechanism of dam failure. Majority of lake outburst causes were included in the category of dynamic causes (e.g. ice avalanche), long-term causes (e.g. dead-ice melting) were less represented. Results of the study imply there can be notable variations of temporal distribution and causes of GLOFs among individual mountain regions even within one continent. Therefore, varying behavior of potentially dangerous lakes should be taken into consideration when, for instance, proposing mitigation measures.

Keywords: GLOF, glacial lake, mountain region, temporal distribution, outburst cause

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1. Introduction

Climate changes and its manifestations linked to mountain glaciation represent one of the most topical issues in the world of geosciences (Bates et al. 2008; Bliss et al. 2014; Li et al. 2007; Rowan et al. 2015; Zhao et al. 2015; Zhou et al. 2010). Faster rate of glacier melting leads to raised summer discharges in glacier-fed streams (Aizen et al. 2007; Wang et al. 2014), overfilling of glacial lake basins and destabilization of moraine dams. These processes may result in the phenomenon called GLOF (= glacial lake outburst flood), which has become a feared natural catastrophic process due to its difficult predictability, high velocity of spreading and often unexpectedly large affected area (Bajracharya and Mool 2009). The main goals of this paper are i) to analyze temporal distribution of recorded GLOFs, and ii) to assess causes of recorded glacial lakes outbursts.

The highest mountain range of the world, Himalayas, provide ideal conditions for the emergence of potential hazards of large proportions due to significant differences in elevations and very steep slopes, the term GLOF was developed for this area (Mool 1995). Climate change affects glaciers whose retreat or degradation results in the formation and development of potentially dangerous lakes (Chen et al. 2010; Komori 2008). These lakes can be of large dimensions and their outburst would cause a flood striking areas several tens to even hundreds of kilometers distant (Richardson and Reynolds 2000).

Other Asian mountain ranges where floods from glacial lake outbursts were recorded include Caucasus (Petrakov et al. 2007; Chernomorets et al. 2007), Pamir (Mergili et al. 2011), Hindu Kush-Karakoram (Gardelle et al. 2011) and Tien Shan (Narama et al. 2010; Janský et al. 2010).

2. Glacial lake outburst flood

Seasonal floods caused by snow melting or torrential rains have affected humans and their livelihood ever since. However, the GLOF, a natural hazard typical for post-LIA era, can be even more destructive – the highest recorded peak discharge was 30,000 $m^3 s^{-1}$ (Richardson and Reynolds 2000).

Recently, a rapid retreat of glaciers was recorded in Himalayas (Bolch et al. 2012; Chen et al. 2007) and other glaciated Asian mountain ranges (Sarikaya et al. 2012; Sorg et al. 2012; Shahgedanova et al. 2014) leading to a formation of new glacial lakes, enlarging of the existing ones and rising of a glacial lake outburst potential (Watanabe et al. 1994; Richardson and Reynolds 2000; Bajracharya and Mool 2009). These floods can reach extremely high flow rates and therefore are able to erode and transport huge amounts of material – up to millions m^3 (Hubbard et al. 2005). Consequently, debris flows reaching distances of as much as 200 km may evolve moving down a valley at higher speed than a flood wave due to double density

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Falátková, K. (2016):
Temporal analysis of GLOFs in high-mountain regions of Asia and assessment of their causes
AUC Geographica, 51, No. 2, pp. 145–154

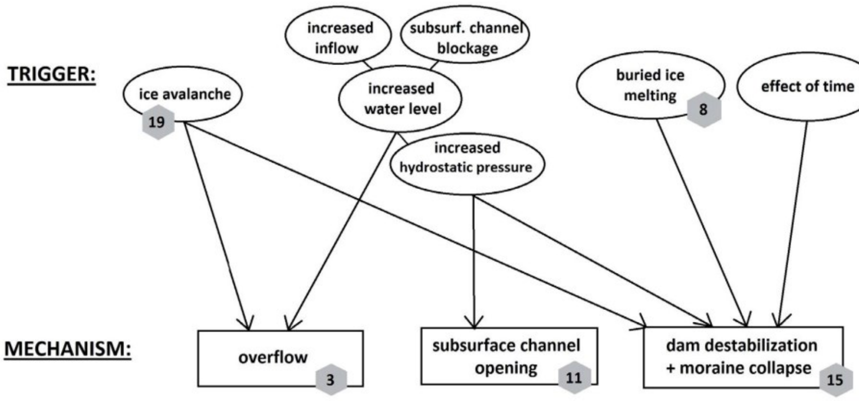

Fig. 1 Relationship of selected outburst triggers and mechanisms with number of cases.

compared to clear water (Richardson and Reynolds 2000). As GLOF is difficult to predict – outburst mechanism is very complex (Kershaw et al. 2005), longitudinal profile of mountain valleys is rather steep and there is often poor or non-existent warning system, material damage can be large and in some cases there could be even many casualties (Lliboutry et al. 1977).

It is important to understand the response of glaciers and glacial lakes to increase of air temperature, to identify potential risks and plan mitigation measures (Bajracharya and Mool 2009; Bennett and Glasser 2009). Remote sensing together with GIS models proved to be a vital tool in assessing risk and defining endangered areas (Bolch et al. 2011; Huggel et al. 2003; Komori 2008; Worni et al. 2012; Pitman et al. 2013).

Flood volume and course depends on many factors including the amount of water released from a lake, height, width and structure of its dam, outburst mechanism, valley shape and available quantity of sediment in the area affected by a flood (Costa and Schuster 1988). One example of an enormous lake outburst flood is an event from 1985, when part of a glacier terminus calved into Dig Tsho Lake, Nepal (Bajracharya and Mool 2009). A displacement wave ran over the dam which failed due to consequent erosion. Resulting flood wave had an initial flow rate of 2,000 $m^3 s^{-1}$ (Vuichard and Zimmermann 1987), Cenderelli and Wohl (2001) indicate even 2,350 $m^3 s^{-1}$. The consequences were noticeable even 90 km below the dam lake (Richardson and Reynolds 2000). Another catastrophic flood of 1994 from outburst of Luggye Tsho Lake was described by Richardson and Reynolds (2000), who claim that the flood wave (over 2 m high) was recorded on a hydrograph at distance greater than 200 kilometers from the source lake.

3. Methods

Total number of 219 cases of glacial lake outburst flood were compiled for this paper based on search of scientific publications and reports. The event parameters were searched as follows: lake's name, date of outburst (year, month, day), cause of outburst (probable trigger or mechanism), mountain range, and lake's coordinates. However, for some of the cases not all the desired information was available. In nine cases the exact year of event is not known, no information on cause of outburst was obtained in 17 cases, and the temporal analysis within a year was based on 128 cases only. The time span of compiled outburst floods begins in 1533 and ends with an event from 2012. Most recorded cases are from the 19th and 20th century, earlier ones are only sporadic.

When analyzing the GLOF cases a distinction is made between two groups – moraine-dammed lakes and ice-dammed lakes. Moraine-dammed lakes drain in most cases once, some do several times. Ice-dammed lakes, on the other hand, are dammed by a glacier blocking a valley; such lakes often form and drain repeatedly. The latter are set aside since the lake formation and consequent outburst are driven by different mechanism and glacier behavior (glacier retreat and degradation vs. glacier advance). And as the outburst is often repeated for decades, the statistics would be significantly influenced – 146 flood cases out of 219 were from ice-dammed lakes. Furthermore, the 146 cases were recorded within only a few localities: Inylchek glacier, Tien Shan (48), several valleys in upper Yarkant basin (24) and upper Indus basin (74), Hindu Kush-Karakoram. Spatial representation of the ice-dammed lake outburst floods is therefore rather unbalanced. The detachment of this group of cases allows to perform a comparative temporal distribution analysis between the two and to reveal differences in occurrence.

For 56 out of 73 cases of moraine-dammed lake outbursts there was information concerning GLOF cause. However, some sources stated an initial trigger of a lake outburst whereas the others mentioned an outburst mechanism. As there are more possible triggers leading to a certain outburst mechanism (Costa and Schuster

Tab. 1 Sources of information on GLOF cases.

		Number of cases with known					
Source	No. of cases	Cause	Day	Month	Year	Time span	Mt range
Gerasimov 1965	1	1	1	1	1	1963	Tien Shan
Gerassimow 1909	1	0	0	0	1	1909	Caucasus
Ives et al. 2010*	34	30	23	24	27	1935–2004	Himalayas
Liu et al. 2013	2	2	2	2	2	1998–2002	Himalayas
Liu et al. 2014	1	1	1	1	1	1988	Himalayas
Mergili et al. 2011	1	1	1	1	1	2002	Pamir
Narama et al. 2009	8	1	8	8	8	1970–1980	Tien Shan
Narama et al. 2010	7	5	6	6	7	1974–2008	Tien Shan
Petrakov et al. 2007	2	1	1	1	2	1993–2006	Caucasus
Petrakov et al. 2012	3	3	1	1	3	1988–2012	Tien Shan
Seinova and Zolotarev 2001	2	2	0	0	2	1958–1959	Caucasus
Wang et al. 2011	10	8	7	9	9	1955–2009	Himalayas
Yesenov and Degovets 1979	1	1	1	1	1	1977	Tien Shan
Glazirin 2010	48	48	37	48	48	1902–2005	Tien Shan
Hewitt and Liu 2010**	95	95	17	25	94	1533–2009	Karakoram
Iturrizaga 2005	3	3	0	0	3	1860–1909	Karakoram
Total	219	202	106	128	210		

* compiled from: Mool et al. 1995, 2001a, 2001b, Yamada 1998, Bajracharya et al. (2008); supplemented with information from: Wang et al. 2011

** supplemented with information from: Iturrizaga 2005

1988), some additional information would be necessary to assess the causes of all 56 events. In Figure 1 relationships of the identified triggers and mechanisms are specified.

Tha GLOF causes can be divided into long-term and dynamic causes according to Emmer and Cochachin (2013). The former include dam failures where an initial external dynamic trigger is absent, the latter are caused by a dynamic event (Yamada 1998).

4. Analysis of GLOFs

Following chapters assess the temporal distribution of outburst flood events and causes of lake outbursts within the high-mountain regions of Asia. Although the number of all outburst flood events is relatively high (219), not all enter the assessment as many lack some piece of information (Table 1).

Throughout the continent of Asia, information on a glacial lake outburst flood was found in following mountain ranges: Caucasus, Pamir, Tien Shan, Karakoram, and Himalayas. Altay and Central range of Kamchatka showed precondition for flood events as well, but no GLOF related publication from these regions was found in scientific literature.

Within Caucasus, information on only a few cases of outbursts were acquired (Petrakov et al. 2012), all of them situated in Elbrus region – surroundings of a glaciated massif of Mt. Elbrus (5,642 m asl). In Pamir, one case of lake outburst was recorded on the territory of Tajikistan (Mergili et al. 2011).

As many of Tien Shan ridges are glaciated, steep valleys and glacier retreat of last decades provide good conditions for lake outburst floods (Bolch 2007; Narama et al. 2010; Petrakov et al. 2012; Yerokhin 2003). However, probably only a minor number of cases were described in scientific literature as this region has long been rather neglected by foreign researchers. Repeatedly drained Lake Merzbacher, dammed by a glacier Inylchek, is an exception as it has been monitored closely for more than a century (Glazirin 2010).

Within the Hindu Kush-Karakoram range, only cases of ice-dammed lake outburst were found (Hewitt and Liu 2010; Iturrizaga 2005). These lakes are situated in upper parts of two basins: Indus and Yarkant, and floods caused by sudden drainage of these lakes have been regularly recorded by local population of downstream villages since 1830s.

A large number of glacial lake outburst floods were recorded in the Himalayas, partly because of the extensiveness of this mountain system and therefore vast glaciated area, but also due to the considerable interest of research teams from all around the world (Benn et al. 2012; Bolch and Kamp 2006; Rana et al. 2000; Richardson and Reynolds 2000; Quincey et al. 2007; Yamada and Sharma 1993). Number of potentially dangerous lakes and GLOFs has been rising in Himalayas since 1930 (Liu

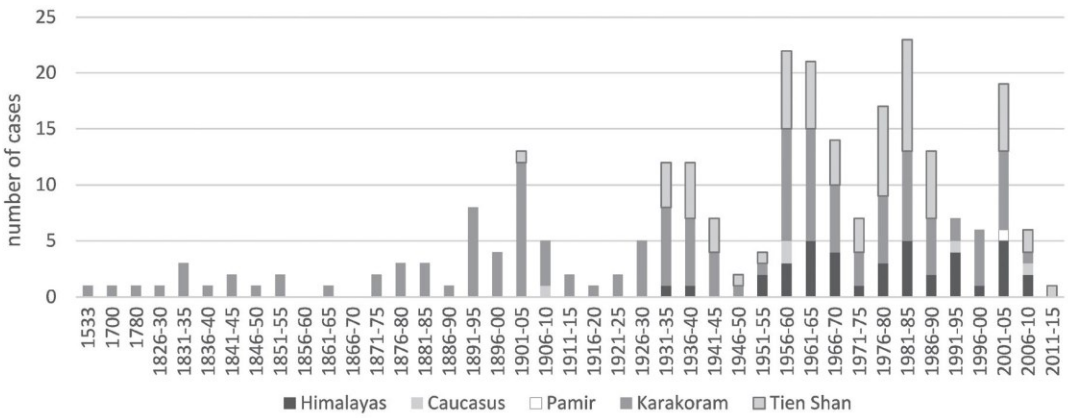

Fig. 2 Temporal distribution of GLOFs according to a mountain range.

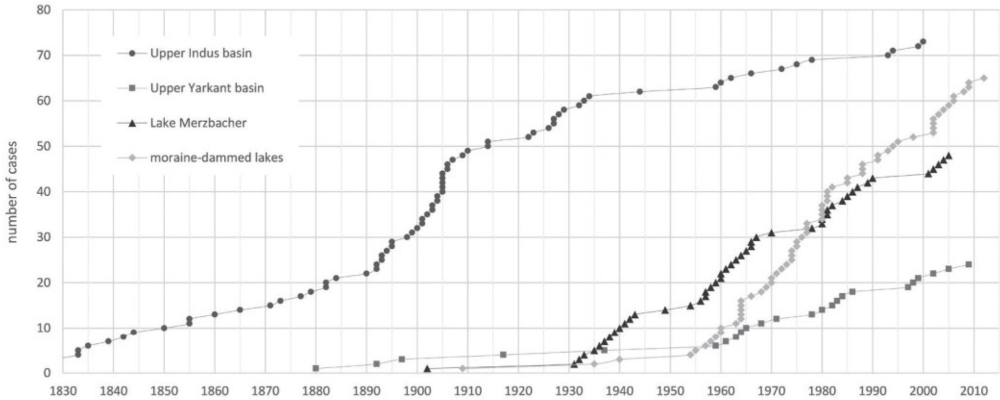

Fig. 3 Cumulative number of GLOFs in Asia divided into 4 categories.

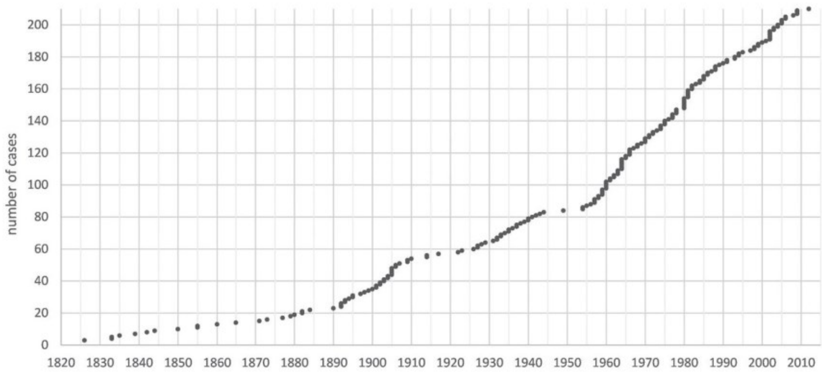

Fig. 4 Cumulative number of all GLOFs in high mountain Asia.

et al. 2013; Bolch et al. 2008), Richardson and Reynolds (2000) report that 33 outburst floods have taken place here until 2000.

4.1 GLOF temporal distribution

Outburst flood events with known year (210 cases) were compiled according to a mountain range where they occurred and classified into five-year segments (Figure 2). Number of recorded cases begins to rise at the end of the 19th century and peaks between 1901 and 1905 when 13 GLOFs occurred. This is followed by a noticeable drop in numbers around 1920. Similar pattern continues further on (peak in the 1930s and drop around the year 1950) with overall higher number of cases since late 1950s. Last significant drop in number of flood cases emerges in 1990s with the average of 1.3 cases per year in all Asian mountain ranges together, which is rather low compared to previous decade (1980s: 2.8 cases/year).

Comparison of the regions in terms of temporal distribution of cases is focused on three ranges with higher number of GLOFs – Himalayas, Karakoram, and Tien Shan. There is consistency among the ranges around the year 1950 when only very few cases occurred. However, the second drop in 1990s is not significant for either Himalayas or Karakoram whereas in Tien Shan not a single outburst flood was recorded. On the other hand, the periods with high numbers of cases are mostly coincident in all three mountain ranges.

An interesting pattern emerges when plotting the data into cumulative number of cases (Figure 3). The data were divided into groups of moraine-dammed lakes and ice-dammed lakes, the latter was further divided by watershed into three parts (Upper Indus basin, Upper Yarkant basin, Lake Merzbacher). All the localities where floods from ice-dammed lakes were recorded show perceptible grouping of cases, so that periods with high and very low number of cases alternate.

In the Indus basin (predominantly in valleys of Shyok, Shimshal, and Hunza) floods were recorded rather regularly during most of the 19th century. However, the first decade od the 20th century was characterized by significantly increased frequency of GLOFs; on average 1.7 cases per year occurred in the region. In contrast, the following period had only two outburst cases between the 1910 and 1922 events. Similar but less pronounced steps follow with rarely any cases recorded in periods 1935–1959 and 1978–1994.

GLOFs in the Yarkant basin also exhibit such pattern – rather short periods of higher and low number of cases alternate. Although the pattern emerged only since 1960s, it seems to be rather regular as well as pattern of outburst cases of Lake Merzbacher. The lake dammed by glacier Inylchek shows periods of almost annual drainage followed by shorter periods with one or no case.

The curve representing cumulative number of moraine-dammed lake events does not exhibit such obvious pattern, although certain periods of lower and higher outburst flood numbers can be found. However, it is not in accordance with the ice-dammed lake cases, except for the time around 1980 when many cases were recorded. That is also apparent in Figure 4 which shows development of cumulative number of all GLOFs in Asia. The blotting effect of moraine-dammed lake cases on the described pattern is confirmed as the alternating periods are visible only until the 1950s when the moraine-dammed GLOFs became frequent.

Distribution of GLOFs within a year was analyzed based on 128 cases with known month of occurrence. The floods are distributed mainly among months characteristic of ablation (June–September), however, there are even few cases which occurred in unusual time of a year (Figure 5). As expected, most outbursts were recorded in August, less in July and September. Slight difference arises due to separation of ice-dammed and moraine-dammed lakes: the former having most cases later in a year (1. August 2. September 3. October) compared to the latter (1. July 2. August 3. June).

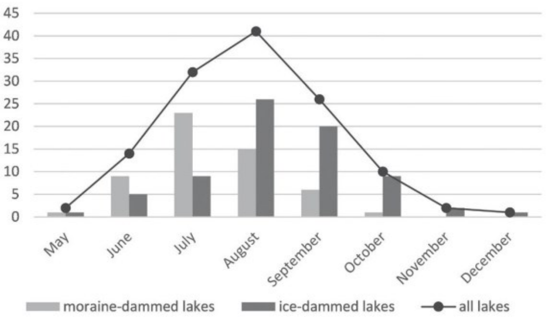

Fig. 5 Monthly distribution of GLOFs with distinction between moraine-dammed and ice-dammed lakes.

4.2 GLOF causes

The cause of GLOF may be difficult to determine as witness is rarely present and evidence may not always indicate to a particular cause with certainty. A total of 202 cases out of 219 were appointed with a cause of a flood, although not all were specific in terms of the outburst trigger.

All ice-dammed lake outbursts, i.e. 72.1% of all cases, were set aside into a category of increased hydrostatic pressure (Zhang 1992). Hewitt and Liu (2010) and Glazirin (2010) describe the mechanism of release of water detained behind the glacier tongue as a consequence of raised hydrostatic pressure which led to partial glacier uplift and opening of drainage channels.

Concerning causes of moraine-dammed lake outbursts, there are only 38 cases with known trigger, 18 with known outburst mechanism, and 17 cases without any information (Figure 6). The most often mentioned cause of lake outburst was ice avalanche falling into a lake (34%). Fall of mass into a lake generates a displacement wave which may either overflow the dam and commence its incision or destabilize the dam and lead to its collapse (Clague and Evans 2000).

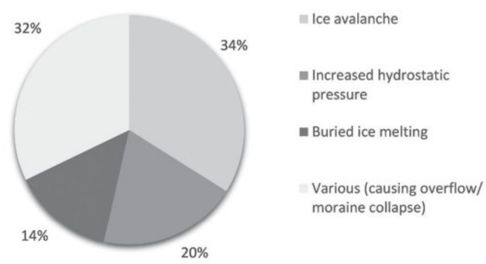

Fig. 6 A percentage share of triggers of floods from moraine-dammed lakes in Asia.

In case a lake does not have a surface outflow, it is sensitive to the amount of inflowing water. Significantly increased inflow (either from rapid snow melt or heavy

Tab. 2 Causes of lake outburst floods according to mountain ranges.

Cause / Mt range	Moraine dam				Ice dam	Total
	Ice avalanche	Increased hydrostatic pressure	Buried ice melting	Various (*resulting in overflow / moraine collapse*)	Increased hydrostatic pressure	
Himalayas	19	/	6	16	/	41
Tien Shan	/	8	1	2	48	59
Caucasus	/	2	1	/	/	3
Pamir	/	1	/	/	/	1
Karakoram	/	/	/	/	97	97
Total	19	11	8	18	145	201

rainfall) causes lake water level to rise together with hydrostatic pressure on a dam. This may lead to subsurface channel opening and lake drainage, which happened in 20% of the cases.

The third specifically mentioned outburst cause was melting of buried ice (14%), which is a part of a moraine damming a lake. The ice melting may disrupt the dam structure and destabilize it to such extent, that it cannot withstand the hydrostatic pressure of detained water and it collapses (Yamada 1998). A moraine dam degraded due to buried ice melting is also more prone to collapse even with a minor trigger (Clague and Evans 2000).

Remaining cases with known mechanism of outburst (32%) could not be classified to causes as both "overflow" and "moraine collapse" are too general and may be a consequence of various triggers.

Based on the knowledge of moraine-dammed lake outburst causes, these can be further divided into long-term and dynamic causes. Ice avalanche belongs among dynamic causes, increased hydrostatic pressure and cases with overflow as an outburst mechanism were also incorporated in this group. The long-term causes include buried ice melting. The ratio of dynamic and long-term causes then makes 33 : 8, with further 15 cases unclassified. The cause of ice-dammed lakes outburst (increased hydrostatic pressure) is considered also dynamic, which means that dynamic causes of lake outburst floods are generally more frequent in Asian high mountain areas.

Deployment of GLOF events in Asia together with the cause of lake outburst are summarized in Table 2. Although there are not enough cases for all the mountain ranges, some interesting differences in terms of outburst causes arise among them. Ice avalanche appears as a relatively common cause of outburst in Himalayas, however, avalanche or other mass movement into a lake was not recorded as a lake outburst cause anywhere else. In Tien Shan, Caucasus and Pamir the lakes drained often by opening of subsurface channels due to incresed hydrostatic pressure whereas in Himalayas this cause did not occur.

5. Discussion

Some of the major glacial lake outburst floods on the territory of Asia were studied in detail, e.g. Luggye Tsho in Bhutan Himalaya (Watanabe and Rothacher 1996), Tam Pokhari in Mt. Everest region (Osti and Egashira 2009) or Lake Zyndan in Tien Shan (Narama et al. 2010). Studies encompassing more lake outbursts include Bajracharya et al. (2008), Narama et al. (2009), Hewitt and Liu (2010), ICIMOD (2011), or Chen et al. (2010). This paper attempted to compile data on all recorded outburst floods in Asia, however, the main obstacle became unbalanced availability of GLOF reports among the mountain regions. Since most cases of GLOF included in this analysis were found in only a few articles dealing with specific locations or time periods, all statistics can be slightly biased due to the uneven spatial (and temporal) distribution of the obtained data. Special case is a group of ice-dammed lakes that are located within few sites and their outbursts are repeated. Although statistics from these data cannot be generalized, they provide interesting insight into the temporal distribution of the outburst floods and a comparison with outbursts from glacial lakes.

Concerning the temporal distribution, a significantly lower number of cases was recorded in 1950s and 1990s, on the contrary, 1960s were a period of very high number of cases. Chen et al. (2010) argue that lake outbursts are closely related to positive anomalous air temperature of a year. Precipitation, Chen et al. (2010) add, plays a role in flood peak discharge value. A certain correspondence was found for Tien Shan as both air temperature and precipitation in 1950s were at their low compared to previous and following decade, in the 1990s the air temperature raised rather slowly from its low at the end of previous decade (Černý et al. 2007). Liu et al. (2014) studied correlation between GLOF events and air temperature in Tibet and confirm that 1960s, 1980s and 2000s were very active periods for GLOFs due to higher temperatures during ablation season but also during accumulation.

Analysis of events distribution within a year found a noticeable difference between the ice- and moraine-dammed lakes with the latter draining earlier in a year. Ice-dammed lakes may react with greater delay as glacier dam uplift requires large amount of water causing sufficiently increased hydrostatic pressure (Glazirin 2010). However, Huss et al. (2007) and Glazirin (2010) both indicate a shift of ice-dammed lakes drainage to earlier time within a year mainly in the second half of the 20th century. It means the diference between the two lake types would not probably be that large if only the latest data were included. Liu et al. (2014) also note that the timing of outburst is influenced by lake's altitude – the higher altitude, the later burst within a season. However, in our case the lake's altitude is not of such importance to influence the outburst timing as the lakes are situated in similar altitude. The drained moraine-dammed lakes lay between 2,500 m and 5,500 m asl. and the ice-dammed ones in an altitude probably between 3,000–4,300 m asl. (the presize location within a valley – the damming glacier – is often unknown).

While frequency of ice-dammed lake outbursts has been significantly lowered since 1930 in Upper Indus basin (Hewitt and Liu 2010), it seems that floods from moraine-dammed lakes began to occur much more often since 1950s. As Hewitt (1982) mentions, it could be associated with general glacier recession which may have opposite effect on the observed lake groups. The overall higher numbers of moraine-dammed lake outbursts could be contributed to both accelerated glacier retreat as well as increased interest of researchers or better accessibility of the records by internet searching (Bajracharya et al. 2007).

Assessment of floods in terms of outburst causes is limited by the fact that some authors reported an outburst mechanism, not a trigger, and so the cause was specified for 38 cases of floods from a moraine-dammed lake. However, conclusions of this paper are relatively in accordance with other GLOF-related studies. Narama et al. (2009) reported buried ice melting and moraine collapse due to headwater erosion of the dam and increased inflow leading to subsurface channel opening as main causes of lake outbursts in northern Tien Shan. Most frequent causes of GLOFs in Tibet are, according to Liu et al. (2013), overflow due to fall of ice into a lake and moraine deformation and collapse due to piping, very similar results are also presented by Emmer and Cochachin (2013).

6. Conclusions

Cases of floods from glacial lake outburst, known by the acronym GLOF, were searched within the territory of Asia. Alpine glaciated areas around the world face this threat mainly due to retreat of glaciers and the subsequent formation of lakes or glacier dynamics generally. Within Asia, cases of flood from moraine-dammed lakes were found in following mountain regions: Caucasus, Pamir, Tien Shan and Himalaya. A large number of floods from lakes dammed by a glacier tongue were recorded in the Hindu Kush-Karakoram mountain range.

Spatial distribution of GLOF cases used in this paper is rather unbalanced. It is probably influenced by availability of reports and publications and the fact that some areas are more favorable for foreign researchers than others. Regarding temporal distribution of found outburst events, significant increase is apparent since 1950s, earlier cases include mostly ice-dammed lake outbursts. Generally, certain periods of higher (1960s, 1980s) and lower (1990s) number of events arise, this pattern is even more obvious for ice-dammed lake cases. Most outbursts occurred within ablation season with peak in August which is consistent with general assumptions. Slight difference was observed between moraine- and ice-dammed lakes as the former tend to drain earlier in a year.

Concerning causes of lake outburst, increased hydrostatic pressure leading to englacial channel opening was appointed to all ice-dammed lake cases (145). The most common cause of moraine-dammed lake outburst is an ice avalanche falling into a lake. Other observed causes include melting of buried ice and increased hydrostatic pressure on a dam due to water level rise. Although the proportion of outburst causes differ among the mountain ranges, dynamic causes constitute the majority of cases.

The main contribution of this paper is an assembly and following comparison of all available GLOFs in high mountain regions of Asia. Unlike other studies, it encompasses both moraine-dammed and glacier-dammed lakes and focuses on differences between the two groups uncovered by temporal analysis of outburst occurrence. Found patterns characterised by alternating periods of high and low number of events could be further analyzed in relation to climate in order to improve our knowledge on link between GLOFs and climatic elements. This is, however, beyond the scope of this paper.

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REFERENCES

AIZEN, V. B., AIZEN, E. M., KUZMICHENOK, V. A. (2007). Glaciers and hydrological changes in the Tien Shan: simulation and prediction. Environmental Research Letters, 2(4), 1–10. http://dx.doi.org/10.1088/1748-9326/2/4/045019

BAJRACHARYA, S. R., MOOL, P. (2009). Glaciers, glacial lakes and glacial lake outburst floods in the Mount Everest region, Nepal. Annals of Glaciology, 50(53), 81–86. http://dx.doi.org/10.3189/172756410790595895

BAJRACHARYA, S. R., MOOL, P. K., SHRESTHA, B. R. (2007). Impact of climate change on Himalayan glaciers and glacial lakes: Case studies on GLOF and associated hazards in Nepal and Bhutan. Kathmandu: International Centre for Integrated Mountain Development.

BAJRACHARYA, S. R., MOOL, P. K., SHRESTHA, B. R. (2008). Global climate change and melting of Himalayan glaciers. Melting glaciers and rising sea levels: Impacts and implications, 28–46.

BATES, B. C., KUNDZEWICS, Z. W., WU, S., PALUTIKOF, J. P. (2008). Climate Change and Water. Technical Paper of the Intergovernmental Panel on Climate Change, IPCC Secretariat, Geneva, 210 pp.

BENN, D. I., BOLCH, T., HANDS, K., GULLEY, J., LUCKMAN, A., NICHOLSON, L. I., QUINCEY, D., THOMPSON, S., TOUMI, R., WISEMAN, S. (2012). Response of debris-covered glaciers in the Mount Everest region to recent warming, and implications for outburst flood hazards. Earth-Science Reviews, 114(1–2), 156–174. http://dx.doi.org/10.1016/j.earscirev.2012.03.008

BENNETT, M. R., GLASSER, N. F. (2009). Glacial geology: ice sheets and landforms. 2nd edition. Chcichester: Wiley-Blackwell, 385 pp.

BLISS, A., HOCK, R., RADIĆ, V. (2014). Global response of glacier runoff to twenty-first century climate change. Journal of Geophysical Research: Earth Surface, 119(4), 717–730. http://dx.doi.org/10.1002/2013jf002931

BOLCH, T. (2007). Climate change and glacier retreat in northern Tien Shan (Kazakhstan/Kyrgyzstan) using remote sensing data. Global and Planetary Change, 56(1), 1–12. http://dx.doi.org/10.1016/j.gloplacha.2006.07.009

BOLCH, T., BUCHROITHNER, M. F., PETERS, J., BAESSLER, M., BAJRACHARYA, S. (2008). Identification of glacier motion and potentially dangerous glacial lakes in the Mt. Everest region/Nepal using spaceborne imagery. Natural Hazards and Earth System Science, 8(6), 1329–1340. http://dx.doi.org/10.5194/nhess-8-1329-2008

BOLCH, T., KAMP, U. (2006). Glacier mapping in high mountains using DEMs, Landsat and ASTER data. Grazer Schriften der Geographie und Raumforschung, 41, 37–48.

BOLCH, T., KULKARNI, A., KÄÄB, A., HUGGEL, C., PAUL, F., COGLEY, J. G., FREY, H., KARGEL, J. S., FUJITA, K., SCHEEL, M., BAJRACHARYA, S., STOFFEL, M. (2012). The state and fate of Himalayan glaciers. Science, 336(6079), 310–314. http://dx.doi.org/10.1126/science.1215828

BOLCH, T., PETERS, J., YEGOROV, A., PRADHAN, B., BUCHROITHNER, M., BLAGOVESHCHENSKY, V. (2011). Identification of potentially dangerous glacial lakes in the nothern Tien Shan. Natural Hazards, 59, 1691–1714. http://dx.doi.org/10.1007/s11069-011-9860-2

CENDERELLI, D. A., WOHL, E.E. (2001). Peak discharge estimates of glacial-lake outburst floods and 'normal' climatic floods in the Mount Everest region, Nepal. Geomorphology, 40(1–2), 57–90. http://dx.doi.org/10.1016/S0169-555X(01)00037-X

ČERNÝ, M., JANSKÝ, B., ŠOBR, M., ENGEL, Z., ČESÁK, J., KOCUM, J., ŽÁČEK, M., YEROKHIN, S. A. Monitoring of high mountain glacial lakes and protection of population against catastrophic impacts of floods caused by moraine lake outbursts. Final report of Development Cooperation Project between Czech and Kyrgyz republic. RP/27/2004, 2007, GEOMIN družstvo, Jihlava, 206 pp.

CHEN, X., CUI, P., LI, Y., YANG, Z., QI, Y. (2007). Changes in glacial lakes and glaciers of post-1986 in the Poiqu River basin, Nyalam, Xizang (Tibet). Geomorphology, 88(3–4), 298–311. http://dx.doi.org/10.1016/j.geomorph.2006.11.012

CHEN, Yaning, XU, C., CHEN, Yapeng, LI, W., LIU, J. (2010). Response of glacial-lake outburst floods to climate change in the Yarkant River basin on northern slope of Karakoram Mountains, China. Quaternary International, 226(1–2), 75–81. http://dx.doi.org/10.1016/j.quaint.2010.01.003

CHERNOMORETS, S. S., PETRAKOV, D. A., KRYLENKO, I. V., KRYLENKO, I. N., TUTUBALINA, O. V., ALEJNIKOV, A. A., TARBEEVA, A. M. (2007). Dynamics of glacial-lake complex Bashkara and evaluation of debris flow formation risk in Adyl-Su valley (Caucasus). Криосфера Земли, 11(1), 72–84. (in Russian)

CLAGUE, J. J., EVANS, S.G. (2000). A review of catastrophic drainage of moraine-dammed lakes in British Columbia. Quaternary Science Reviews, 19, 1763–1783. http://dx.doi.org/10.1016/S0277-3791(00)00090-1

COSTA, J. E., SCHUSTER, R. L. (1988). The formation and failure of natural dams. Geological Society of America Bulletin, 100, 1054–1068. http://dx.doi.org/10.1130/0016-7606(1988)100<1054:TFAFON>2.3.CO;2

EMMER, A., COCHACHIN, A. (2013): The causes and mechanisms of moraine-dammed lake failures in the Cordillera Blanca, North American Cordillera, and Himalayas. AUC Geographica, 48(2), 5–15. http://dx.doi.org/10.14712/23361980.2014.23

GARDELLE, J., ARNAUD, Y., BERTHIER, E. (2011). Contrasted evolution of glacial lakes along the Hindu Kush Himalaya mountain range between 1990 and 2009. Global and Planetary Change, 75(1–2), 47–55. http://dx.doi.org/10.1016/j.gloplacha.2010.10.003

GERASSIMOW, A. P. (1909). O proryve lednikovogo ozera na NO sklone Elbrusa [About glacial lake outburst at the north-east slope of Mt. Elbrus]. Izvestiya Geolkoma, 28(7), 156–160. (in Russian)

GERASIMOV, V.A. (1965). Issykskaia katastrofa 1963 g. i otrazhenie ee in geomorfoogii doliny r. Issyk [The Issyk catastrophe in 1963 and its effect on geomorphology of the Issyk River valley]. Akademiia Nauk SSSR, Izvestiia Vsesoiuznogo, Geograficheskogo Obshchestva, 97(6), 541–547. (in Russian)

GLAZIRIN, G. E. (2010). A century of investigations on outbursts of the ice-dammed Lake Merzbacher (Central Tien Shan). Austrian Journal of Earth Sciences, 103(2), 171–179.

HEWITT, K. (1982). Natural dams and outburst floods of the Karakoram Himalaya. In Hydrological Aspects of Alpine and High Mountain Areas (Proceedings of the Exeter Symposium, July 1982), 259–269.

HEWITT, K., LIU, J. (2010). Ice-dammed lakes and outburst floods, Karakoram Himalaya: historical perspectives on emerging threats. Physical Geography, 31(6), 528–551. http://dx.doi.org/10.2747/0272-3646.31.6.528

HUBBARD, B., HEALD, A., REYNOLDS, J. M., QUINCEY, D., RICHARDSON, S. D., ZAPATA, M. L., SANTILLAN, N. P., HAMBREY, M. J. (2005). Impact of a rock avalanche on a moraine-dammed proglacial lake: Laguna Safuna Alta, Cordillera Blanca, Peru. Earth Surface Processes and Landforms, 30, 1251–1264. http://dx.doi.org/10.1002/esp.1198

HUGGEL, C., KÄÄB, A., HAEBERLI, W., KRUMMENACHER, B. (2003). Regional-scale GIS-models for assessment of hazards from glacier lake outbursts: evaluation and application in the Swiss Alps. Natural Hazards and Earth System Science, 3(6), 647–662. http://dx.doi.org/10.5194/nhess-3-647-2003

HUSS, M., BAUDER, A., WERDER, M., FUNK, M., HOCK, R. (2007). Glacier-dammed lake outburst events of Gornersee,

Switzerland. Journal of Glaciology, 53 (181), 189–200. http://dx.doi.org/10.3189/172756507782202784
ICIMOD (2011). Glacial lakes and glacial lake outburst floods in Nepal. International Centre for Integrated Mountain Development, Kathmandu: ICIMOD, 99 pp.
ITURRIZAGA, L. (2005). New observations on present and prehistorical glacier-dammed lakes in the Shimshal valley (Karakoram Mountains). Journal of Asian Earth Sciences, 25(4), 545–555. http://dx.doi.org/10.1016/j.jseaes.2004.04.011
IVES, J. D., SHRESTHA, R. B., MOOL, P. K. (2010). Formation of glacial lakes in the Hindu Kush-Himalayas and GLOF risk assessment. Kathmandu: ICIMOD.
JANSKÝ, B., ŠOBR, M., ENGEL, Z. (2010). Outburst flood hazard: Case studies from the Tien-Shan Mountains, Kyrgyzstan. Limnologica, 40, 358–364. http://dx.doi.org/10.1016/j.limno.2009.11.013
KERSHAW, J. A., CLAGUE, J. J., EVANS, S. G. (2005). Geomorphic and sedimentological signature of a two-phase outburst flood from moraine-dammed Queen Bess Lake, British Columbia, Canada. Earth Surface Processes and Landforms, 30, 1–25. http://dx.doi.org/10.1002/esp.1122
KOMORI, J. (2008). Recent expansions of glacial lakes in the Bhutan Himalayas. Quaternary International, 184, 177–186. http://dx.doi.org/10.1016/j.quaint.2007.09.012
LI, Z. Q., SHEN, Y. P., WANG, F. T., LI, H. L., DONG, Z. W., WANG, W. B., & WANG, L. (2007). Response of glacier melting to climate change-take Urumqi Glacier No. 1 as an example. Journal of Glaciology and Geocryology, 3.
LIU, J., TANG, C., CHENG, Z. (2013). The Two Main Mechanisms of Glacier Lake Outburst Flood in Tibet, China. Journal of Mountain Science, 10(2), 239–248. http://dx.doi.org/10.1007/s11629-013-2517-8
LIU, J. J., CHENG, Z. L., SU, P. C. (2014). The relationship between air temperature fluctuation and Glacial Lake Outburst Floods in Tibet, China. Quaternary International, 321, 78–87. http://dx.doi.org/10.1016/j.quaint.2013.11.023
LLIBOUTRY, I., MORALES, B., PAUTRE, A., SCHNEIDER, B. (1977). Glaciological problems set by the control of dangerous lakes in Cordillera Blanca, Peru. I: Historical failure of morainic dams, their causes and prevention. Journal of Glaciology, 18, 239–254.
MERGILI, M., SCHNEIDER, J. F. (2011). Regional-scale analysis of lake outburst hazards in the southwestern Pamir, Tajikistan, based on remote sensing and GIS. Natural Hazrads and Earth System Sciences, 11, 1447–1462. http://dx.doi.org/10.5194/nhess-11-1447-2011
MERGILI, M., SCHNEIDER, D., WORNI, R., SCHNEIDER, J. (2011). Glacial lake outburst floods in the Pamir of Tajikistan: challenges in prediction and modelling. In 5th international conference on debris-flow hazards mitigation: mechanics, prediction and assessment, University of Padova, Italy, pp. 14–17.
MOOL, P. K. (1995). Glacier lake outburst flood in Nepal. Journal of Nepal Geological Society, 11, 273–280.
MOOL, P. K., BAJRACHARYA, S. R., JOSHI, S. P. (2001a). Inventory of glaciers, glacial lakes, and glacial lake outburst floods: Monitoring and early warning systems in the Hindu Kush-Himalayan regions – Nepal. Kathmandu: ICIMOD.
MOOL, P. K., WANGDA, D., BAJRACHARYA, S. R., KUNZANG, K., GURUNG, D. R., JOSHI, S. P. (2001b). Inventory of glaciers, glacial lakes, and glacial lake outburst floods: Monitoring and early warning systems in the Hindu Kush-Himalayan region – Bhutan. Kathmandu: ICIMOD.
NARAMA, C., DUISHONAKUNOV, M., KÄÄB, A., DAIYROV, M., ABDRAKHMATOV, K. (2010). The 24 July 2008 outburst flood at the western Zyndan glacier lake and recent regional changes in glacier lakes of the Teskey Ala-Too range, Tien Shan, Kyrgyzstan. Natural Hazards and Earth System Sciences, 10, 647–659. http://dx.doi.org/10.5194/nhess-10-647-2010
NARAMA, C., SEVERSKIY, I., YEGOROV, A. (2009). Current State of Glacier Changes, Glacial Lakes, and Outburst Floods in the Ile Ala-Tau and Kungöy Ala-Too Ranges, Northern Tien Shan Mountains. Geographical Studies, 84(1), 22–32. http://dx.doi.org/10.7886/hgs.84.22
OSTI, R., EGASHIRA, S. (2009). Hydrodynamic characteristics of the Tam Pokhari Glacial Lake outburst flood in the Mt. Everest region, Nepal. Hydrological Processes, 23(20), 2943–2955. http://dx.doi.org/10.1002/hyp.7405
PETRAKOV, D. A., KRYLENKO, I. V., CHERNOMORETS, S. S., TUTUBALINA, O. V., KRYLENKO, I. N., SHAKHMINA, M. S. (2007). Debris flow hazard of glacial lakes in the Central Caucasus. In Chen, Y., Major, S. (eds.). Debris-Flow Hazards Mitigation: Mechanics, Prediction, and Assessment. Netherlands: Millpress, 703–714.
PETRAKOV, D.A., TUTUBALINA, O.V., ALEINIKOV, A.A., CHERNOMORETS, S.S., EVANS, S.G., KIDYAEVA, V.M., KRYLENKO, I.N., NORIN, S.V., SHAKHMINA, M.S., SEYNOVA, I.B. (2012). Monitoring of Bashkara Glacier lakes (Central Caucasus, Russia) and modelling of their potential outburst. Natural Hazards, 61(3), 1293–1316. http://dx.doi.org/10.1007/s11069-011-9983-5
PITMAN, E.B., PATRA, A.K., KUMAR, D., NISHIMURA, K., KOMORI, J. (2013). Two phase simulations of glacier lake outburst flows. Journal of Computational Science, 4, 71–79. http://dx.doi.org/10.1016/j.jocs.2012.04.007
QUINCEY, D. J., RICHARDSON, S., LUCKMAN, A., LUCAS, R., REYNOLDS, J., HAMBREY, M., GLASSER, N.F. (2007). Early recognition of glacial lake hazards in the Himalaya using remote sensing datasets. Global and Planetary Change, 56, 137–152. http://dx.doi.org/10.1016/j.gloplacha.2006.07.013
RANA, B., SHRESTHA, A. B., REYNOLDS, J. M., ARYAL, R. (2000). Hazard assessment of the Tsho Rolpa Glacier Lake and ongoing. Journal of Nepal Geological Society, 22, 563–570.
RICHARDSON, S. D., REYNOLDS, J. M. (2000). An overview on glacial hazards in the Himalayas. Quaternary International, 65–66, 31–47. http://dx.doi.org/10.1016/S1040-6182(99)00035-X
ROWAN, A. V., EGHOLM, D. L., QUINCEY, D. J., GLASSER, N. F. (2015). Modelling the feedbacks between mass balance, ice flow and debris transport to predict the response to climate change of debris-covered glaciers in the Himalaya. Earth and Planetary Science Letters, 430, 427–438. http://dx.doi.org/10.1016/j.epsl.2015.09.004
SARIKAYA, M. A., BISHOP, M. P., SHRODER, J. F., OLSENHOLLER, J. A. (2012). Space-based observations of Eastern Hindu Kush glaciers between 1976 and 2007, Afghanistan and Pakistan. Remote Sensing Letters, 3(1), 77–84. http://dx.doi.org/10.1080/01431161.2010.536181
SEINOVA, I. B., ZOLOTAREV, E. A. (2001). Glaciers and debris flows in the Elbrus region (evolution of glaciation and debris flow activity). Nauchny Mir, Moscow. (in Russian)
SHAHGEDANOVA, M., NOSENKO, G., KUTUZOV, S., ROTOTAEVA, O., KHROMOVA, T. (2014). Deglaciation of the Caucasus Mountains, Russia/Georgia, in the 21st century observed with ASTER satellite imagery and aerial photography. The Cryosphere, 8(6), 2367–2379. http://dx.doi.org/10.5194/tc-8-2367-2014
SORG, A., BOLCH, T., STOFFEL, M., SOLOMINA, O., BENISTON, M. (2012). Climate change impacts on glaciers and runoff

in Tien Shan (Central Asia). Nature Climate Change, 2(10), 725–731. http://dx.doi.org/10.1038/nclimate1592
VUICHARD, D., ZIMMERMANN, M. (1987). The 1985 catastrophic drainage of a moraine-dammed lake, Khumbu Himal, Nepal: cause and consequences. Mountain Research and Development, 91–110. http://dx.doi.org/10.2307/3673305
WANG, G., TANDING, Y., YANG, G., XIAOXIN, Y., DAMBARU, B. K. (2011) A First-order Method to Identify Potentially Dangerous Glacial Lakes in a Region of the Southeastern Tibetan Plateau. Mountain Research and Development, vol. 31, no. 2, pp. 122–130. http://dx.doi.org/10.1659/MRD-JOURNAL-D-10-00059.1
WANG, S., ZHANG, M., PEPIN, N. C., LI, Z., SUN, M., HUANG, X., WANG, Q. (2014). Recent changes in freezing level heights in High Asia and their impact on glacier changes. Journal of Geophysical Research: Atmospheres, 119(4), 1753–1765. http://dx.doi.org/10.1002/2013jd020490
WATANABE, T., IVES, J. D., HAMMOND, J. E. (1994). Rapid growth of a glacial lake in Khumbu Himal, Himalaya: prospects for a catastrophic flood. Mountain Research and Development, 14(4), 329–340. http://dx.doi.org/10.2307/3673729
WATANABE, T., ROTHACHER, D. (1996). The 1994 Lugge Tsho glacial lake outburst flood, Bhutan Himalaya. Mountain Research and Development, 16(1), 77–81. http://dx.doi.org/10.2307/3673897
WORNI, R., STOFFEL, M., HUGGEL, C., VOLZ, C., CASTELLER, A., LUCKMAN, B. (2012). Analysis and dynamic modeling of a moraine failure and glacier lake outburst flood at Ventisquero Negro, Patagonian Andes (Argentina). Journal of Hydrology, 444–445, 134–145. http://dx.doi.org/10.1016/j.jhydrol.2012.04.013
YAMADA, T. (1998). Glacier lake and its outburst flood in the Nepal Himalaya. Tokyo, Japanese Society of Snow and Ice. Data Center for Glacier Research.
YAMADA, T., SHARMA, C.K. (1993). Glacier Lakes and Outburst Floods in the Nepal Himalaya. Snow and Glacier Hydrology (Proceedings of the Kathmandu Symposium, November 1992). IAHS Publ. no. 218, 319–330.
YEROKHIN, S. A. (2003). Study and monitoring of dam failure safety of high-mountain lakes in Kyrgyzstan. State Institute of Geology, Bishkek, 156 pp.
YESENOV, U. Y., DEGOVETS, A. S. (1979). Catastrophic mudflow on the Bol'shaya Almatinka River in 1977. Soviet Hydrology: Selected Papers, 18(2), 158–160.
ZHANG, X. S. (1992) Investigation of glacier bursts of the Yarkant River in Xinjiang, China. Annals of Glaciology, 16, 135–139.
ZHAO, Q., ZHANG, S., DING, Y. J., WANG, J., HAN, H., XU, J., ZHAO, C., GUO, W., SHANGGUAN, D. (2015). Modeling Hydrologic Response to Climate Change and Shrinking Glaciers in the Highly Glacierized Kunma Like River Catchment, Central Tian Shan. Journal of Hydrometeorology, 16(6), 2383–2402. http://dx.doi.org/10.1175/JHM-D-14-0231.1
ZHOU, S., KANG, S., GAO, T., ZHANG, G. (2010). Response of Zhadang Glacier runoff in Nam Co Basin, Tibet, to changes in air temperature and precipitation form. Chinese Science Bulletin, 55(20), 2103–2110. http://dx.doi.org/10.1007/s11434-010-3290-5

RESUMÉ

Časová analýza GLOF událostí ve vysokohorských oblastech Asie a zhodnocení jejich příčin

Práce se zabývá výskytem povodní způsobených vyprázdněním ledovcového jezera ve vysokohorských oblastech Asie a příčinami selhání hráze. V odborné literatuře byly vyhledány informace o tomto typu povodní v následujících horských masivech: Kavkaz, Pamír, Ťan Šan, Karákoram a Himálaj. Celkem bylo nalezeno 219 případů povodní z ledovcových jezer, z toho 145 případů u jezer hrazených ledovcem, ostatní hrazené morénou či nacházející se na morénovém valu. Co se časové distribuce týče, byla zjištěna období s nižším a vyšším výskytem průvalů ledovcových jezer, nejmarkantněji se to projevilo u jezer hrazených ledovcem. Drobné rozdíly mezi oběma skupinami jezer se vyskytly při analýze distribuce událostí v rámci roku. Většina povodní se vyskytla během ablační sezóny, ty způsobené vyprázdněním jezer hrazených ledovcem však byly zaznamenány spíše později (srpen–říjen) oproti povodním z morénových jezer (červen–srpen). Příčina vyprázdnění jezera byla zjištěna celkem pro 202 událostí, velká část z nich však nebyla dostatečně specifická. Mimo zvýšení úrovně hladiny a tím i zvýšení hydrostatického tlaku, jež vede k otevření podpovrchových odtokových kanálů, byla nejčastější příčinou ledová lavina zaznamenána pouze u případů z Himálaje. Další zjištěnou příčinou bylo tání pohřbeného ledu v hrázi.

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5.2 Lakes‘ development and related outburst susceptibility

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Development of proglacial lakes and evaluation of related outburst susceptibility at Adygine ice-debris complex, northern Tien Shan

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Abstract. The formation and development of glacial lakes in mountainous regions is one of the consequences of glacier recession. Such lakes may drain partially or completely when the stability of their dams are disturbed or as a consequence of impacts. We present a case study from the Central Asian mountain range of Tien Shan – a north-oriented tributary of the Adygine Valley, where the retreat of a polythermal glacier surrounded by permafrost has resulted in the formation of several generations of lakes. The aim of this study was to analyse the past development of different types of glacial lakes influenced by the same glacier, to project the site's future development, and to evaluate the outburst susceptibility of individual lakes with an outlook for expected future change. We addressed the problem using a combination of methods, namely bathymetric, geodetic and geophysical on-site surveys, satellite images and digital elevation model analysis, and modelling of glacier development. Based on this case of the glacial lakes being of varied age and type, we demonstrated the significance of glacier ice in lake development. Lake 3, which is in contact with the glacier terminus, has changed rapidly over the last decade, expanding both in area and depth, and increasing its volume by more than 13 times (7800 m^3 to 106 000 m^3). The hydrological connections and routing of glacier meltwater have proved to be an important factor as well, since most lakes in the region are drained by subsurface channels. As the site is at the boundary between continuous and discontinuous permafrost, the subsurface water flow is strongly governed by the distribution of non-frozen zones above, within or beneath the perennially frozen ground. In the evaluation of lake outburst susceptibility, we have highlighted the importance of field data, which can provide crucial information on lake stability. In our case, an understanding of the hydrological system at the site, and its regime, helped to categorise Lake 2 as low outburst susceptible, while Lakes 1 and 3 were labelled as lakes with medium outburst susceptibility. Further development of the site will be driven mainly by rising air temperatures and increasingly negative glacier mass balance. All three climate model scenarios predicted a significant glacier areal decrease by 2050, specifically leaving 73.2% (A1B), 62.3% (A2) and 55.6% (B1) of the extent of the glacier in 2012. The glacier retreat will be accompanied by changes in glacier runoff, with the first peak expected around 2020, and the formation of additional lakes.

1 Introduction

Glacial lakes often appear as gems gleaming amid harsh mountain environments. 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 'glacial lake outburst flood' (GLOF), first used for the Himalayan region, is ubiquitous. Mainly due to the development of satellite-based sensor technologies, the regions under the 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 2000 glacial lakes (>1000 m^2), almost 20% of which are potentially dangerous, and ~15–20 lakes are at risk of sudden drainage each year (Erokhin and Zaginaev, 2016). However, this dry and glacier-meltwater-dependent region of Central Asia has remained rather in the background of research 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. Lastly, there are lakes that are in a phase of enlargement. The expansion of glacial lakes broadly correlates with climatic warming trends and negative glacier mass balance patterns (Lei et al., 2012; Mergili et al., 2013; Zhao et al., 2015). In addition, a recent study has shown that the expansion of glacial lakes under a dry continental climate regime can be closely related to thermokarst processes in permafrost as well (Li et al., 2014). This means that lakes enlarge when in contact with a retreating glacier tongue by the melting of ice in the lake basin bed or around the sides, or by filling due to increased inflow from a glacier. Other factors that play a role in lake evolution include: the erosion of surface drainage channels, the formation, expansion or blockage of subsurface drainage channels, dam morphology changes, or slope movements adjacent to the lake.

Outbursts from mountain lakes occur almost annually in Kyrgyzstan (in recent years, these included: Lakes Merzbacher – 2017, 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 a hyperconcentrated flow or debris flow) will threaten areas in the lower parts of a valley. To reduce the risk, an assessment framework (see, e.g., the GAPHAZ Technical Guidance Document – GAPHAZ, 2017) can be applied, which includes 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 evaluation of the outburst susceptibility of selected lakes that had already been identified as potentially dangerous (Erokhin and Zaginaev, 2016). As the need for field data for such an assessment had been generally acknowledged, here, we present an assessment based on a combination of on-the-spot and remotely-obtained data.

Although the study site is not particularly suited for the purpose of only monitoring glacier retreat, we did find it most appropriate in terms of observing the site in a more holisticway. The glacier itself is considered to be representative of smaller glaciers, which are subject to relatively rapid retreat rates (Narama et al., 2010) and have generally shorter response times to changes in climatic conditions compared to larger ones (Wang et al., 2014). This allowed us to follow a larger development period at a smaller time scale. Also, the existence of glacial lakes documenting glacier shrinkage was convenient – being at various stages of development, they were seen as able to 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 populated areas, an outburst susceptibility assessment strategy, with an estimatation of future change, was considered to be desirable and of practical relevance.

Based on a detailed study of one particular site, we looked at the issue of glacier development and the associated, actively-evolving glacier forefield, and its consequences on downstream areas. We have used the term 'ice-debris complex' (as coined by Bolch et al., 2019) to describe the glacier, and the debris landforms with varying shares of buried ice in front of it. In order to address the problems associated with the rapid processes that operate in mountainous regions, we incorporated perspectives of future scenarios for the development of the site, and the related outburst susceptibility, into our study. Because our intention was to promote a more complex approach to the issue of susceptibility assessment, our study included three particular objectives: i) to analyse the past evolution of different types of glacial lakes influenced by the same glacier; ii) to provide an insight into the probable future evolution of the site; and iii) to evaluate the outburst susceptibility 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 present an analysis of the past development of the local setting, following both the glacier and the proglacial lakes; then, we report on the potential future conditions for the site; and, finally, we include evaluations of the lakes' outburst susceptibilities and estimations of how they will likely evolve under changing climatic conditions. In our opinion, to analyse glacier retreat and lake formation without touching on the topic of consequences (present and near future) would be as incomplete as evaluating outburst susceptibility without knowing the past evolution of the site.

2 Study area, data and methods

The research site is situated in northern Tien Shan, specifically Kyrgyz Ala-Too. The highest peak of this east–west-oriented range is Pik Semenova Tienshanskogo, at 4895 m a.s.l. Although the Kyrgyz Ala-Too is not as severely glaciated as the ranges in central Tien Shan, the currently glacier-free parts of the valleys do show signs of the former glacier extent. The main Ala Archa Valley currently has ~33 km^2 of its area covered by glaciers. It drains into the Chuy River, which is part of a large endorheic basin. The most common lake type of Kyrgyz Ala-Too are those formed in intramorainic depressions (Table 1). According to Erokhin (2012), 83.6% of potentially dangerous lakes are ice-cored moraine-dammed lakes, which include intramorainic and thermokarst lakes in the presence of buried ice.

Table 1. Representation of different lake types in Kyrgyz Ala-Too, northern Tien Shan, in 2017. Only lakes with a minimum area of 1500 m^2 were categorised. Based on manual mapping using satellite imagery in Google Earth and refined with the 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 %**

The development of glaciation in the Ala Archa Basin is monitored, in the long term, mainly due to its position in proximity to the Kyrgyz capital, Bishkek, and the popularity of the valley with tourists visiting the National Park. According to Aizen et al. (2006), there was a change in the glaciated area of the Ala Archa watershed by -0.12 km^2/yr (-0.29%/yr) between 1963 and 1981, and by -0.2 km^2/yr (-0.51%/yr) between 1981 and 2003, documenting an accelerated retreat by the end of the 20th century. The overall reduction in glaciated area over 60 years (1943–2003) was 15.22% (a loss of 6.52 km^2; Aizen et al., 2006), which is an above-average value; in Tien Shan, the glacier area shrank by 14.2% over the same interval (Aizen et al., 2007a). Farinotti et al. (2015) confirmed this trend, and Bolch (2015) pointed out that, although the number of glaciers increased due to the disintegration of several larger glaciers, the total glaciated 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 in most glaciers, which has also led to changes in the hydrological regime of the glacial streams (Aizen et al., 1996; Glazirin, 1996; Pieczonka et al., 2013). Sporadic permafrost occurs at elevations above 2700 m a.s.l., is discontinuous at 3200–3500 m a.s.l., and is continuous above ~3500 m a.s.l. (Gorbunov et al., 1996).

The Adygine watershed has an area of 39.6 km^2, and its glaciated area constitutes ~10% of the basin (3.9 km^2). The total elevation differs by up to 2370 m; the valley below the glacier has an average slope of 10.8°. The Adygine ice-debris complex (42°30'10''N, 74°26'20''E) closes this 8-km-long tributary valley with a northern orientation, and reaches an elevation of 3400–4200 m a.s.l. (Fig. 1), which means it lies mainly within the zone of continual permafrost (Gorbunov et al., 1996). The permafrost is known to reach depths of up to 100 m, and the active layer can be up to 4 m thick at similar altitudes in northern Tien Shan (Marchenko et al., 2007). The occurrence of ice-rich, perennially frozen ground thus controls the underground water flow at the site, allowing it only through non-frozen zones (mainly the active layer and taliks; Cheng and Jin, 2013). The large landforms of mixed ice-debris below the glacier were formed by long-term viscous creep (Bolch et al., 2019) and contain buried glacier remnants and ice-lenses. Several such ice-lenses were recently exposed due to surface erosion, and may have had an impact on lake evolution and stability (Karlsson et al., 2012).

The upper part of the complex involves a polythermal glacier (2.8 km²), followed by a three-level cascade of glacial lakes that have evolved as a consequence of glacier retreat over the past 50 years. The glacier terminus is currently situated at 3600 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 fine material, considerably lowering the albedo of its surface. The position of the equilibrium line altitude (ELA) on the northern slopes of Kyrgyz Ala-Too is at ~3900 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 ages, and have different positions in the connected hydrological system that drains meltwater down the valley. The most recently formed lakes are close to the glacier terminus, reacting sensitively to changes in glacier melt rates, 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) – that has 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 3450 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 being identifiable based on visual inspection only. The western part is composed of creeping ice-rich debris, with a typical relief of arcuate ridges and furrows in its terminal part, ice-lenses are present and occasionally exposed (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. Lying probably within the discontinuous permafrost zone, this landform has a prominently oversteepened front that is almost 700 m wide. In its western part, thermokarst processes are manifested in subsidence craters, cracks and sagging, and numerous thermokarst lakes and exposed ice-lenses (Fig. 1c) can be found here. The eastern part is considered to be of an older generation, with a rather flattened surface and stable lakes having lower turbidity.

Climatic conditions in the area are continental, characterised by relatively low precipitation and high annual and daily air temperature fluctuations. This is clearly evident at the Ala Archa station (2200 m a.s.l.), which reports a 22°C difference between the mean air temperature of the warmest and coldest months (13°C for July, -9°C for January), a mean annual air temperature (MAAT) of 3.3°C and precipitation totals of 450 mm. Some 1300 m higher, at the Adygine site, the MAAT is around -3.5°C (based on weather station data from 2008–2013) and annual precipitation is estimated to be 700–800 mm. On the northern slope of Kyrgyz Ala-Too, the winters are characterised by low precipitation totals, whereas the maximum precipitation is typically reached in late spring and the beginning of summer.

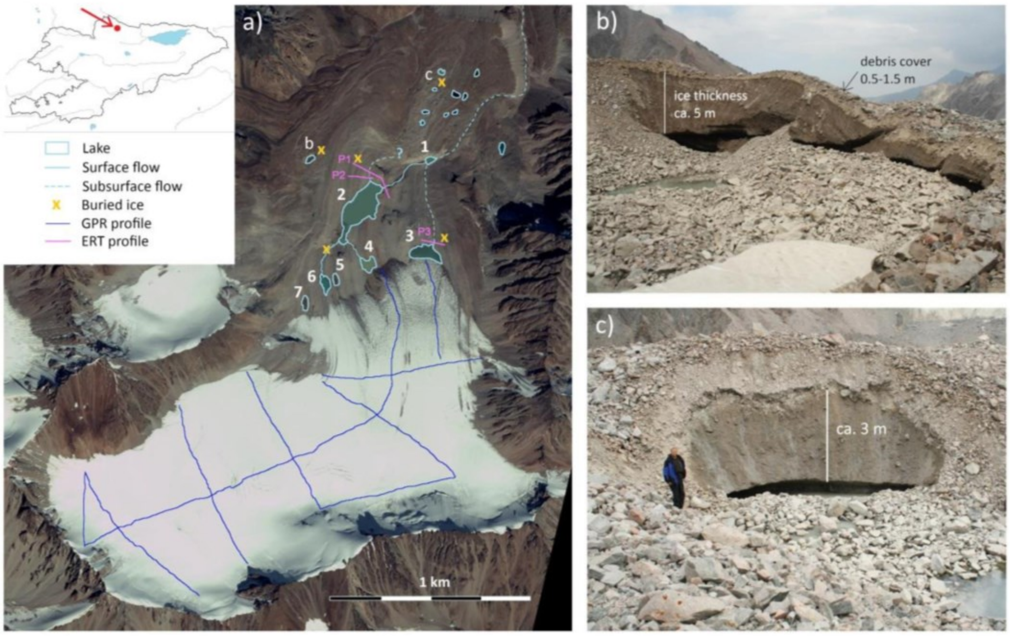

Figure 1. Overview of the Adygine ice-debris complex (a) and its recent dynamics; exposure of buried ice embedded within perennially frozen ground (b, c). 1–7 are the studied lakes; P1–P3 refers to geophysical profiles. Study site marked with an arrow on map of Kyrgyzstan (upper left corner). Positions of (b) and (c) are marked on Figure 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

2.1 Field mapping

A research station, built at the site in 2008 next to the Lake 2 surface outflow, provided an adequate location for field measurements to be taken from. Climatic data and ground temperatures were obtained from an automated weather station situated near the Adygine glacier terminus at 3560 m a.s.l. The hydrological regime of the lakes was 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 can result in the exposure of overdeepenings (potential future lakes). This time-series was 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 were measured by the same means as the glacier terminus. To capture changes in the lake basins and derive the volume of retained water, a repeated bathymetric survey was carried out at selected lakes (Sobr and Jansky, 2016). Depth was recorded in 2.5-m steps along 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 at the site (the lakes' surroundings and their dams) were surveyed by electrical resistivity tomography and spontaneous polarisation, in order to detect buried glacier remnants, frozen ground 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 Figure 1a.

The method of electrical resistivity tomography (ERT) consists of measuring the resistivity of the subsurface by means of a number of electrodes located along a profile. Interpretation of such measured data was performed using the software Res2Dinv (Loke and Barker, 1995), which provided 2D inversion of the measured resistance data and calculated the 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 model (GF Instruments, CR), with 48 electrodes at 4.4 m increments. The maximum range of the electrodes was 206.8 m, with a depth range of about 50 m.

The spontaneous polarisation (SP) method measures the natural electrical potential of the rock environment; in the case of lake dams, the filtering potential that arises from water filtration through the porous environment can be tested. Measurements were made with a GEOTOR I with offset compensation. Non-polarisable electrodes were used to minimise the effect of transient resistance between the measuring electrode and the geological environment. The measurements were carried out using potential variation, in which the electrical potential of a given point on the profile is measured relative to a fixed reference point. The measurement steps were around 10 m, and the location of every fifth point was recorded using GPS.

GPR was used to acquire information on the glacier thickness and subglacial topography. GPR data were collected along the central longitudinal profile, for five transverse and three short, connecting profiles. An unshielded 50 MHz rough terrain antenna and MALÅ CU-II control-unit were used for data collection. The signal acquisition time was set to 860 ns and the scan spacing was about 0.07 m. The collected data were processed and interpreted using Reflex-Win software version 4.5.

2.2 Modelling of glacier evolution

The glacier mass balance model – glacier evolution runoff model, or GERM – applied in this study has been 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, the future glacier mass balances of the Adygine glacier until 2050 were calculated with GERM, forced by empirically downscaled scenarios data for daily air temperature and precipitation, as described below. As the glacier mass balance was not measured at Adygine glacier, we used the mass balance data from the 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 were validated for both winter and summer. In addition, we were able to use discharge data from the Adygine River (from gauging near the confluence with the Ala Archa River) for 1960–1987 to validate the modelled discharge of GERM with observed data.

Despite the generally poor availability of data in Central Asia, daily air temperature means and precipitation totals were gathered from two stations close to the study area: Alplager (2130 m a.s.l., 1978–2010) and Baityk (1580 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 datasets 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). Although the SRES scenario data were replaced by the representative concentration pathways approach of the Intergovernmental Panel on Climate Change (IPCC, 2013), the uncertainty ranges of future large-scale temperature and precipitation changes provided by the two approaches are rather similar and smaller compared to the uncertainty range of climate data coming from further downscaling. We argue that, for the purpose of assessing future glacier extent, in particular for the future development of glacial lakes, these scenarios are sufficiently realistic.

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

2.3 Outburst susceptibility assessment

The presented lake outburst susceptibility assessment is qualitative, and was adapted to accommodate the distinctive regional features – dry climate, the common occurrence of buried ice and permafrost. We based the assessment procedure on knowledge we gained through fieldwork in the region, focused especially on the hydrological systems of proglacial lakes. The specificity of the region is the origin of the lakes – the most common type are lakes formed in intramorainic depressions (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 was 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 evaluations, which would include on-site investigation, need to be carried out.

In this paper, the overall susceptibility is introduced as being a combination of conditioning factors (i.e. a lake's inherent characteristics determining the lake's susceptibility to bursting and causing flooding), and the presence of possible triggers that would have the capacity to cause an outburst. To assess a lake's inherent susceptibility, the following parameters were selected and a simple qualitative rating suggested (Table 3):

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

Lake type is primarily connected to the stability of the dam. The material forming the dam (or the depression in which a lake exists) has varying characteristics and behaviours in relation to water. The rock-step type is considered to be the most stable, as water usually has only a minor effect on this; however, when covered with the loose morainic material left behind a retreating glacier, this part of the dam can be prone to erosion and piping, and will become less stable overall. The interaction of a lake with an ice-dam results in even higher risks – ice melting due to heat transfer from water and high hydrostatic pressure on the ice-dam leading to its uplift, causing lake drainage.

Ice contact is another parameter that influences lake stability. A lake in direct contact with a glacier is considered to be more dangerous than one at a distance, as it will be strongly affected by glacier behaviour. Its evolution and morphological changes can be dynamic as the glacier retreats, advances or disintegrates.

Drainage type is the parameter that represents a lake's hydrological regime, covering only two options – surface or subsurface drainage. It is the limited (or changing) capacity of a subsurface channel that 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.

The *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. Inherent susceptibility parameters and ratings of lakes. The scheme on the right shows possible combinations of parameter ratings (H - high, M - medium, L - low) and the resulting susceptibility. Note that this simplifying table serves as guidelines to susceptibility estimate and that even medium resulting 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

Parameter 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**

The determination of the degree to which a lake is susceptible to outburst is followed by an assessment of potential triggers. The means of determining whether a trigger has the potential to cause outburst in the specific lake or not may be distant (satellite imagery, DEM), as is the case for the first group of triggers in Table 4, or can be on-site, as is necessary for inner dam stability or subsurface channel functioning, for example.

The fall of mass into a lake is one of the most common triggers of lake outburst (Ding and Liu, 1992; Emmer and Cochachin, 2013; Falatkova, 2016), resulting in an impact wave that can overtop or destabilise the dam. Following a widely used procedure, the distance between a lake and a slope, and slope steepness are the parameters that determine 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.

The development and stability of a glacial lake can be affected by the melting of buried ice within a dam. 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 the dam's inner structure (Richardson and Reynolds, 2000). Besides that, dam stability is also a function of the overall frozen/unfrozen condition of the dam material.

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

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

Changes in lake outburst susceptibility result from a complex process associated with the interaction of geomorphological processes and the behaviour of the glacio-hydrological system. Development of the englacial and subsurface routing of meltwater is especially crucial, while being difficult to determine; this was beyond the scope of this paper. The presented estimate of susceptibility change is linked to climatic changes (rising MAAT) only, and involves impacts accompanying deglaciation that have been observed in other mountain ranges, such as glacier retreat, altered glacier melt rate, slope instability (debuttressing effect), permafrost degradation and buried ice melting (Allen et al., 2016; Frey et al., 2010; Haeberli et al., 2017). The general assumptions for future climatic conditions, glacier retreat and altered melt rates are supported by the glacier mass balance model (Section 3.3).

Table 4. Triggers with potential to cause lake 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 (rock-step, ice dam)	water getting to surface at downstream side of dam, moisture marks, sediment accumulation near a 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 has 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 has retreated ~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 ~3540 m a.s.l. (close to the current Lake 2 drainage channel); the narrower eastern tongue bypassed the outcrop, reaching 100 m lower, to 3450 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 (3640 m a.s.l.). In the latest period of observation (2007–2017), the terminus retreat rate had similar values to the earlier (1962–1988) phase, at 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 a^{-1}. The positional differences of retreat rate are linked to the variability of the glacier bed topography (rock outcrops) and to heat transfer with meltwater (contact with glacial lake).

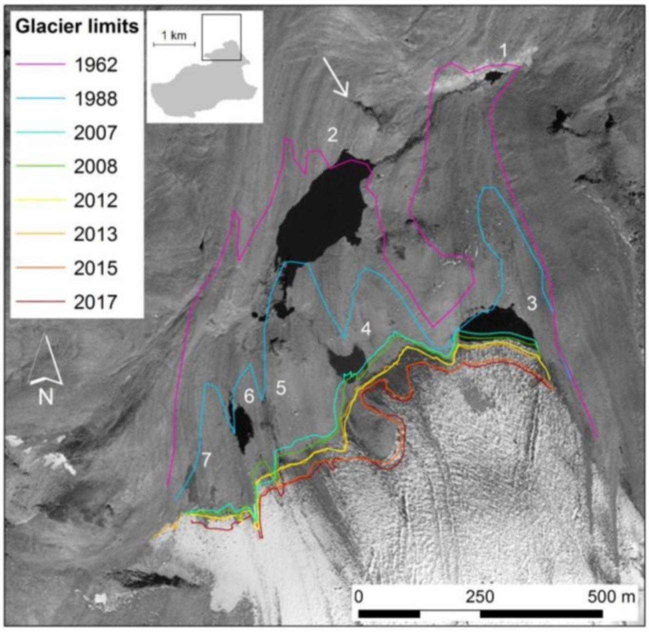

Figure 2. Retreat of the Adygine glacier tongue between 1962 and 2017. 1–7 are the proglacial lakes. Arrow marks the rock outcrop (riegel). Base image: Worldview-2 (2011).

3.2 Formation and development of the lakes

There are numerous lakes of varying size, type and stage of development in the study area. 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 from about a dozen thermokarst lakes, situated on a large morainic, perennially frozen platform at an elevation of ~3450 m. The largest lake (Lake 1; Janský et al., 2010) is situated directly below a rock riegel, in a depression covering ~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 unclear. 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 of between 10 000 and 15 000 m^3), caused by the limited capacity of the subsurface outflow channels. Besides the surface water intake, water is probably routed there by seepage from Lakes 2 and 3. There have been changes in the morphometry of the basin in recent years, as sliding of unconsolidated material from the basin walls has been observed. The part close to the lake drainage has changed recently, extending the basin in a north-easterly direction.

3.2.2 Lake 2

Lake 2, the largest lake at the study site, began forming in about 1960, after the western glacier tongue retreated behind a rock riegel at 3540 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. The 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's 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 also slightly decreased since the start 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 ice-rich frozen morainic material (Fig. 3). The overall thickness of ice and frozen ground exceeded the instrument depth range, being more than 40 m. The low resistivity values (<5000 Ωm) marked a thin, active layer at the surface, but also non-frozen material below the lake surface outflow and possibly the bedrock. Seepage was detected at the outflow (through the non-frozen ground above the bedrock) and at the western end of the dam, which can be linked to the active layer and non-frozen zones within the permafrost.

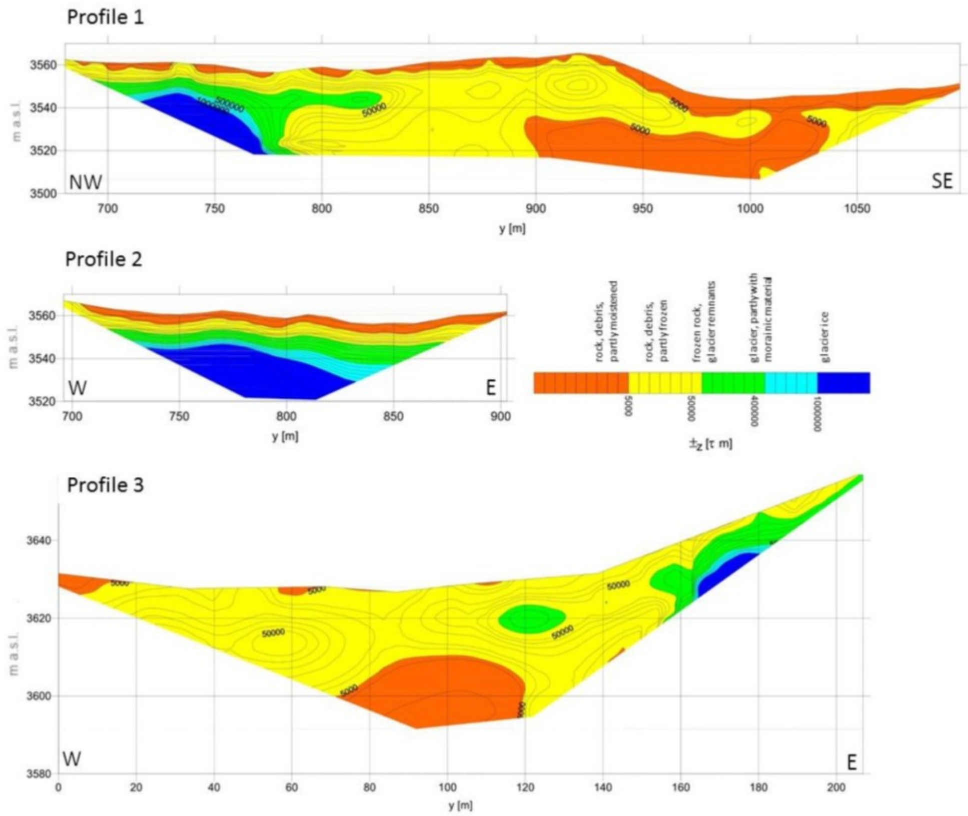

Figure 3. Electrical resistivity results along Profiles 1 and 2 for Lake 2, and Profile 3 for Lake 3. Resistivity values of materials depend on grain size, pore size, water content, salinity, temperature, and phase-resistivity increases due to the change from electrically conductive water to non-conductive ice (Kneisel et al., 2008). For location of the profiles (P1–P3), see Figure 1a.

3.2.3 Lake 3 and newly emerging lakes (4–7)

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 those lakes showed an increase in area. Lake 4 is an example of one that was enlarging slightly until it lost contact with the glacier in 2011–2012. Most lakes formed, enlarged or remained stable over a period of several years, and finally drained by subsurface routes (for detailed information on the lakes' morphometric changes, see the Supplement). An exception is Lake 6, which seemed stable over the monitoring period, until it suddenly drained in 2015, but was filled again the following year. The area exposed following the glacier retreat is underlain by buried ice-lenses (visual inspection, Fig. 1a) within frozen ground (documented by the measured resistivity, Fig. 3). A possible explanation for the lake's sudden drainage might be permafrost degradation caused by warming by the lake water (Shur and Jorgensen, 2007).

Although these proglacial lakes are highly unstable, the hazard they pose is negligible, as only a small amount of water (up to 4000 m^3 in Lakes 4 and 6) can be retained in such shallow depressions. Lake 3 is an exception; this easternmost lake is in direct contact with the glacier and grows 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 deep. Besides the climate-driven glacier retreat, the lake has been growing due to increased glacier melting, caused by heat transfer between the lake water and the ice. The lake expanded from 5710 m^2 in 2007 to 8830 m^2 in 2012. In the past few years, the growth has accelerated, leading to a lake area of 16 020 m^2 in 2017 (Fig. 4). Shortly after its formation, the lake had a surface outflow; however, the drainage soon switched to subsurface channels. Currently, during times of high lake-water levels, water overflows the dam. This minor surface drainage sinksbelow the surface after 5–6 metres, however. The geophysical survey (Fig. 3) revealed the dam structure, which is mainly composed of frozen debris with ice-lenses. In the eastern part, there is distinct glacier ice covered with ~10 m of ice-rich frozen material. About 15–20 m below the lake outflow, there is a zone of lower resistivity (several thousand Ωm), which could indicate either bedrock or a non-frozen zone (talik). If the latter is the case, the lake's further deepening, and expansion of the non-frozen zone beneath the basin, could result in a connection being formed with this talik. As the volume of the retained water increases rapidly (7800 m^3 in 2007, 29 300 m^3 in 2012 and 106 000 m^3 in 2017), this recently insignificant lake has become a focus of attention.

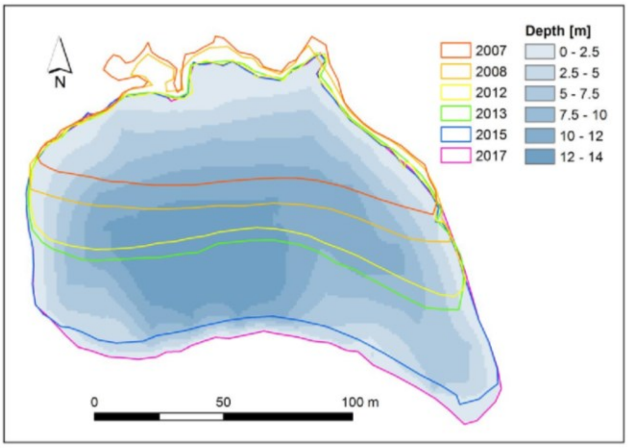

Figure 4. Bathymetric map and area changes of Lake 3 between 2007 and 2017. Lake depth data obtained during a field survey at the site carried out in 2017.

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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 metres of each other. Differences in lake water turbidity/colour suggest 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 were formed by thermokarst processes. As these lakes retain only small volumes of water, and no significant areal development has been observed, a closer investigation from the outburst susceptibility 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 develop further in the future, either by enlarging or draining through newly opened subsurface channels.

3.3 Expected future site conditions

According to all three modelled scenarios, an increase in air temperature in the area is highly likely in future decades, with an estimated higher temperature increase in spring and summer and a moderate temperature increase in autumn and winter. Scenarios of annual precipitation showed a reduction under all three scenarios. In particular, summer and autumn depicted a decrease in precipitation. Winter and spring displayed a minor reduction, or even a slight increase, in 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 were negative throughout the simulation period (Fig. 6a). Scenario A1B showed significantly less negative mass balances between 2015 and 2035 compared to the other scenarios. In this period, Scenario A2 was characterised by the highest decadal variation, with filtered values of between -1300 and -500 mm w.e. After 2030, all scenarios illustrated a significant negative trend, indicating that a tipping point for glacier shrinkage will be reached. Scenario A1B showed the steepest decline for this period, with a gradual recovery after 2045.

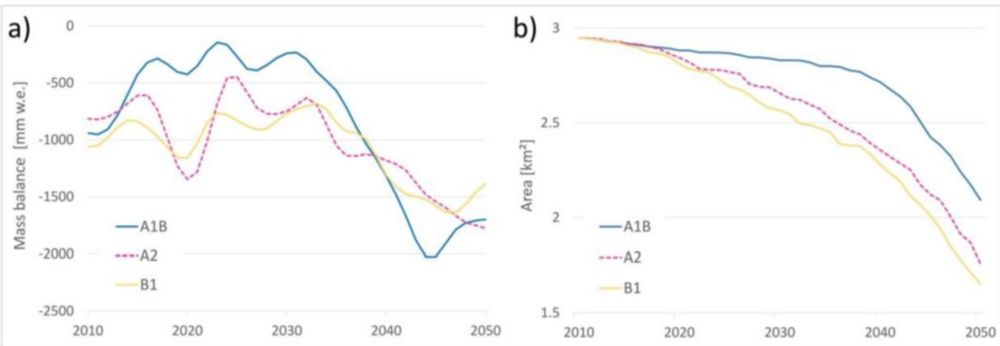

Figure 6. Glacier mass balance (a) and glacier areal evolution (b) modelled according to scenarios A1B, A2 and B1 for the period 2010–2050. Data smoothed using 10-year low-pass filter.

Due to the negative mass balances (Fig. 6a), all scenarios resulted in a reduction in glacier area. Under Scenario A1B, the glacier area was reduced only slightly in the first two decades (2015–2035), when compared to the other two scenarios (Fig. 6b). After 2035, Scenario A1B showed accelerated areal shrinkage (resulting from the accelerated mass loss). All three scenarios simulated a significant decrease in glacier area by 2050, specifically leaving only 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 will remain. 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, revealed several overdeepenings. Under Scenario A1B, three new lakes, with a total area of ~0.1 km², could emerge, while Scenario A2 presented the potential for seven new lakes to form, with a total area of 0.13 km². Scenario B1 exhibited the greatest modelled glacier retreat, and the exposed terrain had the potential for 11 new lakes, with a total area of 0.17 km².

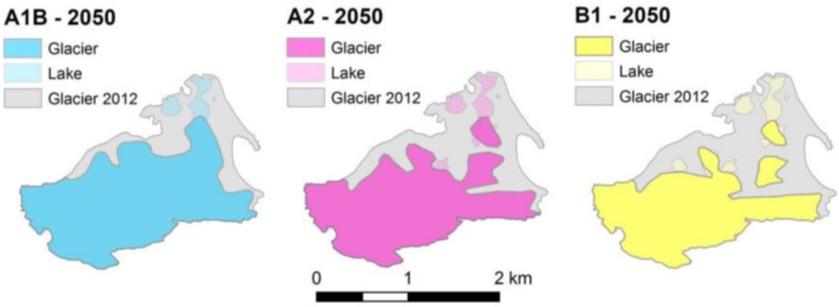

Figure 7. Modelled future glacier extent and potential areas for lake formation for 2050 under Scenarios A1B, A2 and B1.

As a result of greater negative glacier mass balances, total runoff from the Adygine glacier catchment is expected to rise. According to Scenarios A1B and A2, the total runoff will increase by 19.7% and 25%, respectively, between 2010 and 2050 (Fig. 8a). However, Scenario B1 showed a rather sensitive reaction of the glacier dynamics to climate change, with a clear peak in discharge values around 2020. This was 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. The most pronounced changes were seen at the beginning and end of an ablation period – specifically, in April, May and October. Slight increases in runoff are also expected in June and September. However, for the core period of the ablation season – July and August – no significant changes were modelled. Separating total runoff into its two main components – melt from snow and melt from ice – for both today and the future (Fig. 8b), highlighted a significant temporal shift for the snow component. The time of peak runoff from snow-melt, which was in mid-June for the first decade of 2010–2020, will change to mid-May in the decade 2040–2050. This will coincide with a slightly earlier start of ice-melt at the glacier, as the snow cover will have depleted earlier as well, thus exposing bare ice.

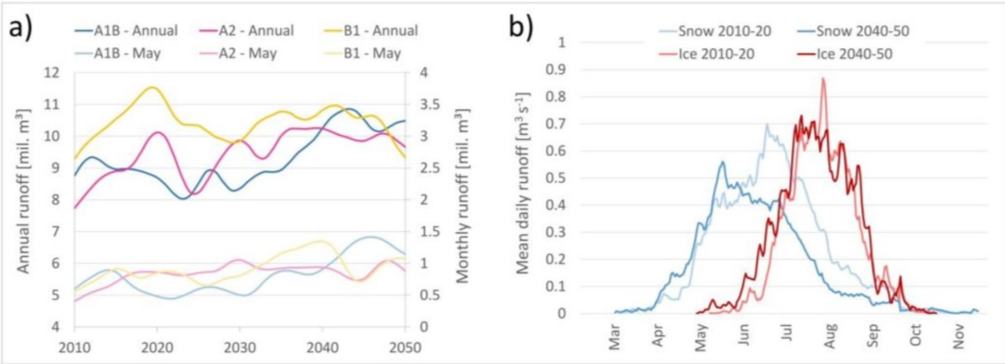

Figure 8. a) Accumulated annual runoff and monthly runoff for May between 2010 and 2050. Data smoothed with a 10-year low-pass filter. b) Mean daily runoff (averaged over all scenarios) from snow- and ice-melt for the periods of 2010–2020 and 2040–2050.

3.4 Outburst susceptibility assessment and its expected change

The inherent outburst susceptibility of the studied lakes, based on the lakes' characteristics, determined here by five parameters, is summarised in Table 5. Lakes 1 and 3 were marked as having medium susceptibility, whereas the largest lake of the site, Lake 2, was considered to be less inclined to burst and cause a flood. Several areas were determined as being 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 indicate 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 the ELA, the glacier is expected to shrink significantly in the coming decades. Sorg et al. (2014) confirmed that future susceptibility is likely to change due to glacier retreat and the accompanying consequences. For instance, the slopes may become more prone to destabilisation due to permafrost degradation, in particular those that are south-oriented (i.e. exposed to direct solar radiation).

Table 5. The resulting lakes' inherent susceptibility to outburst.

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

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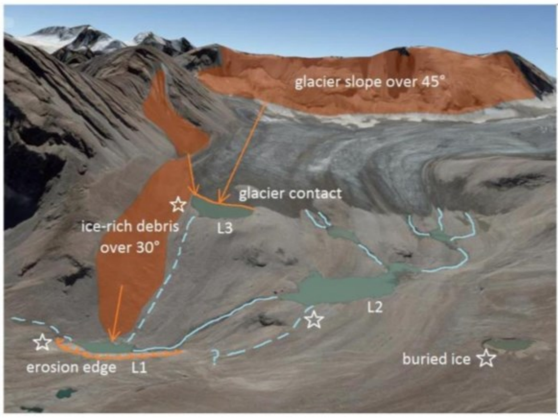

	Lake 1	Lake 2	Lake 3
Fall of mass into lake	ice avalanche	ice avalanche	Ice avalanche
	Landslide/ rockfall	landslide/ rockfall	Landslide/ rockfall
	calving	calving	calving
Inner dam stability	Buried ice melting	Buried ice melting	Buried ice melting
	piping	piping	Piping
	Earthquake	Earthquake	Earthquake
Hydrostatic pressure increase	Increased inflow	increased inflow	Increased inflow
	Subs. channel blockage	subs. channel blockage	Subs. channel blockage
	Upper lake outburst	upper lake outburst	upper lake outburst

Figure 9. Geomorphological conditions and potential triggers of lake outburst at Adygine ice-debris complex. View from the north (Google Earth, 2018).

Lake 1 has a medium susceptibility to cause flood, mainly due to its origin (intramorainic depression), subsurface drainage and the potential for enlargement. Outburst could 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 that leads through the moraine complex. Changes in the capacity and, above all, the blockage of such a channel would lead to the lake filling up and subsequent outburst. That could also be initiated by the outburst or overflow of Lake 3. Its outburst susceptibility is expected to increase in the future, due to the 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 the 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 the subsurface drainage network. A change in outburst susceptibility expected to be negligible in the future, as surface drainage would be able to deal with possible greater inflow, and the lake stability will not be threatened by unstable slopes.

Lake 3 has a medium susceptibility to cause flood, due to it being in contact with the glacier, having relatively unstable subsurface drainage and the potential to grow both in area and depth. Despite the contact with the glacier tongue, a dam overflow by impact wave from calving is not probable because the terminus is not crevassed and a calving front has not developed. However, the lake is within reach of an ice avalanche from the steep glacier slope, and also landslide from the 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. Outburst susceptibility is expected to increase in combination with increasing lake volume and hydrostatic pressure on the dam. Also, destabilised slopes in the lake's proximity will threaten its stability.

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

4 Discussion

4.1 Glacier thermal regime and impact of glacier contact on lake development

The GPR data show a two-layered structure of the glacier, with prevailing cold ice and a temperate layer along the glacier bed. The terminus is frozen to the ground, and cold ice dominates the glacier base 450 m up-valley. At mid-length, on a narrow glacier snout around 3700 m a.s.l., the ice thickness increases to more than 50 m, and temperate ice appears at the glacier base. Temperate ice spreads along the base over the next 1 km, and terminates below cold ice in the accumulation area. The cold–temperate transition along most of this section is located 30 to 50 m below the glacier's surface. The upper part of the glacier (above 3900 m a.s.l.) consists mainly of ice below the pressure-melting point, and is thus connected to the permafrost environment in bedrock.

The presence of both types of ice confirms the polythermal regime of the Adygine glacier that was originally assumed only for large valley glaciers in the Tien Shan region (Dyurgerov et al., 1995). The distribution of temperate ice is similar to that of the characteristic polythermal glaciers described by Copland and Sharp (2001) and Etzelmüller and Hagen (2005). Following the classification scheme by thermal structure proposed by Irvine-Fynn et al. (2011), the Adygine glacier corresponds to a thermal configuration with the presence of temperate ice near the bed in the ablation area. The smaller extent of warm ice in the accumulation zone implies a different temperature structure than that of the Tuyuksu glacier (the Zailiyskiy Alatau), which has a maximum thickness of warm ice in the accumulation zone, where the cold–temperate transition reaches the glacier surface (Nosenko et al., 2016).

Lake 2 is a typical example of a lake formed when a glacier retreats over a rock-step. The lake's 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 had a similar genesis, and is currently in the stage of growth. Thanks to intensive thermal exchange between lake water and ice, a 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 the 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 was not terminated or stabilised after loss of contact with the glacier. Its further evolution was linked to buried ice, its exposure and melting, and the degradation of permafrost. The depression formed in the 1960s and filled with water two decades later, yet even today, the lake is subject to changes in basin morphology and drainage. Such 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 to cause an outburst flood. In many cases, such lakes are even non-stationary, filling up only when a 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 a similar altitude to 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 the 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, causing dismay among the capital's inhabitants and material damage of ~US$100 000 (Zaginaev, 2013). It is the often-variable subsurface drainage, the common presence of buried glacier remnants and a steep valley full of loose sediment downstream that make these lakes potentially dangerous.

4.2 Model uncertainties and future conditions

Huss et al. (2014) described, in detail, the uncertainties of glacier and runoff modelling in general. In our case, GERM was calibrated using the summer and winter mass balance data of Golubina glacier from 1981 to 1994. Whereas simulated winter mass balances for Golubina showed no consistent trend of under- or overestimation, the modelled summer mass balance slightly overestimated the real conditions from 1981 to 1984, and generally underestimated it 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 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 a DEM from 1960, as well as the glacier area from this time, attenuated the accuracy of the calibration of the GERM simulations.

Both the 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) also predicted significant glacier degradation linked to the ELA shifting to higher altitudes because rising air temperatures will not be balanced out by sufficient increases in precipitation totals. Sorg et al. (2014) used the GERM for glacier and runoff modelling in Chon Kemin Valley (Zailiyskiy and Kungey Alatau, Tien Shan), where the glaciers are expected to have vanished 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 area, with 'warm' scenarios expecting peak runoff in the 2020s, and the main change in spring runoff (increase) being caused by higher winter precipitation and enhanced snow-melt in spring. A general prediction for runoff change in Tien Shan is a possible increase in the near future, followed by a steady decline until the end of the 21st century (Sorg et al., 2012). However, different regions in Tien Shan may exhibit varying responses and uncertainties connected to future runoff modelling. An advanced state of glacier degradation was 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 has been a significant decrease in August runoff, which is typical for the advanced stage of glacier runoff decline. Based on a comparison with expected future summer runoff from the Adygine glacier, we can estimate that Adygine is currently close to the stage of peak runoff.

Our approach to the identification of potential locations 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 a method to the whole region. Numerical models are a good substitute, and are widely used on large areas, from the Peruvian Andes (Colonia et al., 2017), to the Swiss Alps (Linsbauer et al., 2012) and the Himalayas–Karakoram (Linsbauer et al., 2016). The commonly used model is GlabTop, introduced by Linsbauer et al. (2009) and Paul and Linsbauer (2012), with an automated version (GlabTop2) introduced by Frey et al. (2014). The model calculates glacier ice thickness based on a DEM, and glacier outlines and branch lines; the automated version (GlabTop2) avoids manual delineation of the lines. A similar concept and results (Frey et al., 2014) were obtained from a model presented in Huss and Farinotti (2012). Potential sites with overdeepenings can also be identified, or confirmed, based on the three morphological criteria (glacier surface features) introduced by Frey et al. (2010). This is a more laborious approach, which includes manual analysis of the DEMs and high-resolution satellite/aerial imagery. If we apply these morphological criteria to the Adygine glacier, one location with a distinct narrowing, and also a change in slope, can be identified. This coincides with our results pointing to the presence of overdeepenings at the place of the current glacier tongue. Nevertheless, as Haeberli et al. (2016a) correctly pointed out, there are still significant limitations to understanding the principles of the depth erosion by glaciers, and thus to modelling of glacier-bed overdeepenings. Moreover, the estimation of debris cover volume left after glacier recession is a challenge, as excessive debris can smooth topographic irregularities and lead to the formation of outwash plains, and not lakes (Linsbauer et al., 2016). We are also aware of the limitations to our approach. 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 susceptibility assessment

The presented evaluation strategy combined knowledge from the field with the benefits of remotely-obtained data, providing a good insight into the interaction of the site's elements (a moraine, buried ice, perennially frozen ground, meltwater channels, a glacier, lakes) and a broader spatial and temporal perspective, thanks to the airborne imagery. Field investigations are necessary to produce a realistic assessment of lake outburst susceptibility. McKillop and Clague (2007) mentioned that knowledge of seepage and lake bathymetry and its changes is important information for understanding dam hydraulic conditions. Bolch et al. (2008) highlighted subsurface glacier meltwater routing, which we also believe to be extremely important, together with the buried ice distribution. At the Adygine ice-debris complex, the ERT technique helped us to discover an extensive ice-core near Lake 2, buried under a considerably thick (8–10 m) layer of mostly frozen debris. A similar structure was observed, for example, at the moraine dam of the 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. The resistivity values attributed to materials detected at our study s ite are consistent with the common resistivity value ranges of the different materials summarised in Kneisel et al. (2006). Frozen ground can exhibit a wide range of resistivity values, varying from 10^3 to 10^6 Ω m, depending on ice content, temperature (related to amount of unfrozen water) and content of impurities (Haeberli and Vonder Mühll, 1996). Glacier ice typically has relatively high resistivity values of well over 10^6 Ω m, which means it is practically non-conductive (Kneisel et al., 2003).

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; McKillop and Clague, 2007). Comprehensive hazard assessments elaborated as a basis for wide usage include, for example, the GAPHAZ Technical Guidance Document (GAPHAZ, 2017), the strategy prepared by ICIMOD, focused on the high-mountain region of the 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 trigger is often represented solely by a 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 a 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 did consider it to be one of the main factors that can increase the outburst susceptibility in future, as presented for the Alps by Haeberli et al. (2017). The presence of permafrost at the site is supported by other studies (e.g. Marchenko et al., 2007), the global permafrost model of Gruber et al. (2012) and the site's negative MAAT of recent years. Although the measurement period is rather short (2008–2013), the long-term MAAT of the site may, in fact, be even lower, given the low value (MAAT of -7.4°C) from another Kyrgyz station located at similar altitude (3614 m a.s.l.) in central Tien Shan (Bolch et al., 2019). An earthquake is definitely a phenomenon that can destabilise a structure via unconsolidated material, although McKillop and Clague (2007), for example, decided not to include it in their hazard assessment due to the 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 (~50 km from Adygine) in 1992. Also, increased inflow due to heavy rainfall has been 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 ice or snow avalanche-starting zones, but the rising MAAT could influence the glacier thermal regime and cause a shift in certain parts from cold- to warm-based, thereby lowering the friction between the glacier-bed and steep ice-masses (Huggel et al., 2008). Areas exposed after glacier recession often consist of unconsolidated material, which, if steep, may become a potential starting zone for landslides or debris flows (Haeberli et al., 2017). Already exposed steep slopes may become destabilised and fail due to the melting of interstitial ice or, where present, the degradation of mountain permafrost (Huggel et al., 2010). The formation of new lakes at higher elevations could increase the hazard of the site due to a possible cascade effect, where even a small volume release could trigger the outburst of a lower-lying lake. Such situations are expected to occur in the future, and the potentially arising challenges should be addressed in advance (Haeberli et al., 2016b).

5 Conclusions

This paper aimed to describe the impact of glacier retreat on its forefield, to highlight the differences in evolution and outburst susceptibility of proglacial lakes and to promote a more field-based, regionally-specified approach to hazard/susceptibility assessment.

The main results concerning the studied site are:

- The glacier is subject to relatively rapid recession, which is comparable to that of other glaciers in Tien Shan. Since 1962, it has retreated, on average, by 11.5 m a^{-1}. By 2050, it is expected to have shrunk to 55.6–73.2% of its 2012 extent.

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

- A three-level cascade of glacial lakes has formed in the glacier forefield, with the largest (Lake 2) retaining a volume of about 200 000 m^3. However, due to its stable surface drainage and lack of contact with the glacier tongue, its development terminated in the 1990s. Two other lakes have the potential to grow – Lake 1, thanks to filling and the melting of buried ice, and Lake 3 by further glacier retreat. The latter is currently the most dynamic lake; having formed in 2005, to date, it has expanded to a volume of 106 000 m^3 and a depth of >14 m.

- Lakes 1 and 3 were categorised as having medium outburst susceptibility, mainly due to their subsurface drainage and the presence of buried ice in their vicinities. Despite its large size, the outburst susceptibility of Lake 2 was assessed as low. The destabilisation of steep slopes, exposure and melting of buried ice and changes in glacier runoff may further deteriorate the stability of Lakes 1 and 3, leading to an increase in outburst susceptibility in the future.

A very similar development in other glacier complexes in this region is expected, with runoff peaks occurring at present or in the near future, accompanied by permafrost degradation, the gradual melting of buried ice and the formation of new lakes. This will result in new outburst hazards that could arise in a relatively short time. One of the main outputs of the study is an analysis of lake development dynamics, which indicate a lake's potential to grow and its related outburst susceptibility. As we have documented, lake size is not as important a factor, in terms of susceptibility, and we would rather emphasise the influence of buried ice in relation to the presence of perennially frozen ground and the unpredictability of subsurface drainage channels. In terms of future susceptibility change, we would like to draw attention to the influence of glacier runoff. The expected extension of the ablation season, and the related activity in glacial lakes, may lead to increased outburst susceptibility. The prolonged period of the melting season in relation to buried ice will result in a greater potential for the alteration of the hydrological network of drainage channels. Further research should be dedicated to the hydrological connections in proglacial morainic environments and the interactions of subsurface water flow with permafrost and buried ice, as there is still an insufficient understanding of these processes.

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

Author contributions. KF, MŠ and BJ carried out the fieldwork and processed the data obtained. KF designed the outburst susceptibility assessment, produced the figures and wrote the manuscript, with contributions from AN and WS for the modelling part. ZE provided results from the GPR survey; VB provided results from the ERT and SP surveys. AN and WS ran the glacier evolution model, HH supervised the EURAS-CLIMPACT project and initiated the concept of this joint paper. All authors contributed to improving drafts of the paper.

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References

Aizen, V. B., Aizen, E. M., Melack, J. M.: Precipitation, melt and runoff in the northern Tien Shan, J Hydrol, 186, 229-251, doi:10.1016/S0022-1694(96)03022-3, 1996.

Aizen, V. B., Kuzmichenok, V. A., Surazakov, A. B., Aizen, E. M.: Glacier changes in the central and northern Tien Shan during the last 140 years based on surface and remote-sensing data, Annals of Glaciology, 43, 202-213, doi: 10.3189/172756406781812465, 2006.

Aizen, V. B., Kuzmichenok, V. A., Surazakov, A. B., Aizen, E. M.: Glacier changes in the Tien Shan as determined from topographic and remotely sensed data, Global Planet Change, 56(3), 328-340, doi:10.1016/j.gloplacha.2006.07.016, 2007a.

Aizen, V. B., Aizen, E. M., Kuzmichonok, V. A.: Glaciers and hydrological changes in the Tien Shan: simulation and prediction, Environ Res Lett, 2(4), 045019, doi:10.1088/1748-9326/2/4/045019, 2007b.

Alean, J.: Ice avalanches: some empirical information about their formation and reach, J Glaciol, 31(109), 324–333, doi:10.3189/S0022143000006663, 1985.

Allen, S. K., Linsbauer, A., Randhawa, S. S., Huggel, C., Rana, P., Kumari, A.: Glacial lake outburst flood risk in Himachal Pradesh, India: an integrative and anticipatory approach considering current and future threats, Nat Hazards, 84(3), 1741-1763, doi:10.1007/s11069-016-2511-x, 2016.

Awal, R., Nakagawa, H., Fujita, M., Kawaike, K., Baba, Y., Zhang, H.: Study on Piping Failure of Natural Dam, Annuals of Disas. Prev. Res. Inst., Kyoto Univ., 54B, doi: 10.2208/jscejhe.67.i_157, 2011.

Benestad, R.E.: Empirical-statistical downscaling in climate modeling. Eos Trans Am Geophys Union, 85, 417-422, doi:10.1029/2004EO420002, 2004.

Benestad, R.E., Hanssen-Bauer, I., Førland, E.J.: An evaluation of statistical models for downscaling precipitation and their ability to capture long-term trends, Int J Climatol, 27, 649–665, doi: 10.1002/joc.1421, 2007.

Benn, D. I., Bolch, T., Hands, K., Gulley, J., Luckman, A., Nicholson, L. I., Quincey, D., Thompson, S., Toumi, R., Wiseman, S.: Response of debris-covered glaciers in the Mount Everest region to recent warming, and implications for outburst flood hazards, Earth Sci Rev, 114(1), 156-174, doi: 10.1016/j.earscirev.2012.03.008, 2012.

Bolch, T.: Glacier area and mass changes since 1964 in the Ala Archa Valley, Kyrgyz Ala-Too, northern Tien Shan, Lëd i Sneg, 55(1), 28-39, doi: 10.15356/2076-6734-2015-1-28-39, 2015.

Bolch, T., Buchroithner, M. F., Peters, J., Baessler, M., Bajracharya, S.: Identification of glacier motion and potentially dangerous glacial lakes in the Mt. Everest region/Nepal using spaceborne imagery, Nat Hazards Earth Sys, 8(6), 1329-1340, doi: 10.5194/nhess-8-1329-2008, 2008.

Bolch, T., Peters, J., Yegorov, A., Pradhan, B., Buchroithner, M., Blagoveshchensky, V.: Identification of potentially dangerous glacial lakes in the northern Tien Shan, Nat Hazards, 59(3), 1691-1714, doi: 10.1007/s11069-011-9860-2, 2011.

Bolch, T., Rohrbach, N., Kutuzov, S., Robson, B. A., Osmonov, A.: Occurrence, evolution and ice content of ice-debris complexes in the Ak-Shiirak, Central Tien Shan revealed by geophysical and remotely-sensed investigations, Earth Surf Proc Land, 44, 129-143, doi:10.1002/esp.4487, 2019.

Chen, Y., Xu, C., Chen, Y., Li, W., Liu, J.: Response of glacial-lake outburst floods to climate change in the Yarkant River basin on northern slope of Karakoram Mountains, China, Quatern Int, 226(1), 75-81, doi: 10.1016/j.quaint.2010.01.003, 2010.

Cheng, G., Jin, H.: Permafrost and groundwater on the Qinghai-Tibet Plateau and in northeast China, Hydrogeol J, 21(1), 5-23, doi: 10.1007/s10040-012-0927-2, 2013.

Clague, J. J., Evans, S. G.: A review of catastrophic drainage of moraine-dammed lakes in British Columbia, Quatern Sci Rev, 19(17), 1763-1783, doi: 10.1016/S0277-3791(00)00090-1, 2000.

Colonia, D., Torres, J., Haeberli, W., Schauwecker, S., Braendle, E., Giraldez, C., Cochachin, A.: Compiling an Inventory of Glacier-Bed Overdeepenings and Potential New Lakes in De-Glaciating Areas of the Peruvian Andes: Approach, First Results, and Perspectives for Adaptation to Climate Change, Water, 9(5), 336-354, doi: 10.3390/w9050336, 2017.

Copland, L., and M. Sharp.: Mapping thermal and hydrological conditions beneath a polythermal glacier with radio-echo sounding, J Glaciol, 47, 232–242, doi:10.3189/172756501781832377, 2001.

Ding, Y., Liu, J.: Glacier lake outburst flood disasters in China, Ann Glaciol, 16, 180-184, 1992.

Dussaillant, A., Benito, G., Buytaert, W., Carling, P., Meier, C., Espinoza, F.: Repeated glacial-lake outburst floods in Patagonia: an increasing hazard?, Nat Hazards, 54(2), 469-481, doi: 10.1007/s11069-009-9479-8, 2010.

Dyurgerov, M. B., Liu, C., Xie, Z.: Oledenenie Tyan'-Shanya [Tien Shan glaciers]. VINITI, Moscow, 1995

Emmer, A., Cochachin, A.: The causes and mechanisms of moraine-dammed lake failures in the Cordillera Blanca, North American Cordillera, and Himalayas, AUC Geographica, 48(2), 5-15, doi: 10.14712/23361980.2014.23, 2013.

Emmer, A., Vilímek, V.: Lake and breach hazard assessment for moraine-dammed lakes: an example from the Cordillera Blanca (Peru), Nat Hazard Earth Sys, 13(6), 1551-1565, doi: 10.5194/nhess-13-1551-2013, 2013.

Engel, Z., Sobr, M., Erokhin, S.A.: Changes of Petrov glacier and its proglacial lake in the Akshiirak massif, central Tien Shan, since 1977, J Glaciol, 58(208), 388-398, doi: 10.3189/2012JoG11J085, 2012.

Erokhin, S. A.: Monitoring of hazardous lakes in Kyrgyzstan. PhD thesis, 220 pp. Kyrgyz National Academy of Sciences, Institute of Water Problems and Hydropower, 2012. (in Russian)

Erokhin S. A., Zaginaev V. V.: Forecasting of hazards from mountain lakes based on the catalogue. Report book of Kyrgyz Ministry of Emergency Situations, Part III, 542-555, 2016. (in Russian)

Erokhin, S.A., Zaginaev, V.V., Meleshko, A.A., Ruiz-Villanueva, V., Petrakov, D.A., Chernomorets, S. S, Viskhadzhieva, K.S., Tutubalina, O.V., Stoffel, M.: Debris flows triggered from non-stationary glacier lake outbursts: the case of the Teztor Lake complex (Northern Tian Shan, Kyrgyzstan), Landslides, 1-16, doi: 10.1007/s10346-017-0862-3, 2017.

Etzelmüller, B., Hagen, J. O.: Glacier-permafrost interaction in Arctic and alpine mountain environments with examples from southern Norway and Svalbard, in: Cryospheric Systems: Glaciers and Permafrost, C. Harris and J. B. Murton (eds.), Geol Soc Spec Publ, 242,11–27, doi: 10.1144/gsl.sp.2005.242.01.02, 2005.

Falatkova, K.: Temporal analysis of GLOFs in high-mountain regions of Asia and assessment of their causes, AUC Geographica, 51(2), 145-154, doi: 10.14712/23361980.2016.12, 2016.

Falatkova, K., Sobr, M., Kocum, J., Jansky, B.: Hydrological regime of Adygine Lake, Tien Shan, Kyrgyzstan, Geografie, 119(4), 320-341, 2014.

Farinotti, D., Longuevergne, L., Moholdt, G., Duethmann, D., Mölg, T., Bolch, T., Vorogushin, S., Güntner, A.: Substantial glacier mass loss in the Tien Shan over the past 50 years, Nat Geosci, 8(9), 716-722, doi: 10.1038/ngeo2513, 2015.

Farr, T.G., Rosen, P.A., Caro, E., Crippen, R., Duren, R., Hensley, S., Kobrick, M., Paller, M., Rodriguez, E., Roth, L., Seal, D., Shaffer, S., Shimada, J., Umland, J., Werner, M., Oskin, M., Burbank, D., Alsdorf, D.: The Shuttle Radar Topography Mission, Rev Geophys, 45(2), 583-585, doi: 10.1029/2005RG000183, 2007.

Frey, H., Haeberli, W., Linsbauer, A., Huggel, C., Paul, F.: A multi-level strategy for anticipating future glacier lake formation and associated hazard potentials, Nat Hazards Earth Sys, 10(2), 339-352, doi: 10.5194/nhess-10-339-2010, 2010.

Frey, H., Machguth, H., Huss, M., Huggel, C., Bajracharya, S., Bolch, T., Kulkarni, A., Linsbauer, A., Salzmann, N., Stoffel, M.: Estimating the volume of glaciers in the Himalayan–Karakoram region using different methods, Cryosphere, 8(6), 2313–2333, doi: 10.5194/tc-8-2313-2014, 2014.

Fischer, L., Purves, R. S., Huggel, C., Noetzli, J., Haeberli, W.: On the influence of topographic, geological and cryospheric factors on rock avalanches and rockfalls in high-mountain areas, Nat Hazards Earth Sys, 12(1), 241-254, doi: 10.5194/nhess-12-241-2012, 2012.

Fujita, K., Sakai, A., Nuimura, T., Yamaguchi, S., Sharma, R. R.: Recent changes in Imja Glacial Lake and its damming moraine in the Nepal Himalaya revealed by in situ surveys and multi-temporal ASTER imagery, Environ Res Lett, 4(4), 1-7, 2009.

GAPHAZ 2017: Assessment of Glacier and Permafrost Hazards in Mountain Regions – Technical Guidance Document. Prepared by Allen, S., Frey, H., Huggel, C. et al. Standing Group on Glacier and Permafrost Hazards in Mountains (GAPHAZ) of the International Association of Cryospheric Sciences (IACS) and the International Permafrost Association (IPA). Zurich, Switzerland / Lima, Peru, 72 pp, 2017.

Glazirin, G. E.: The reaction of glaciers in west Tien Shan to climate changes, Zeitschrift fur Gletscherkunde und Glacialgeologie, 32, 33-39, 1996.

Google Earth Pro 7.3.2.5491. (August 30, 2014): Adygine, Kyrgyzstan. 42°30'11'' N 74°26'20'' E. DigitalGlobe 2018, http://www.earth.google.com [January 10, 2018].

Gorbunov, A. P.: Monitoring the evolution of permafrost in the Tien Shan, Permafrost Periglac, 7(3), 297-298, doi: 10.1002/(SICI)1099-1530(199609)7:3%3C297::AID-PPP223%3E3.0.CO;2-C, 1996.

Gruber, S.: Derivation and analysis of a high-resolution estimate of global permafrost zonation, Cryosphere, 6(1), 221-233, doi: 10.5194/tc-6-221-2012, 2012.

Haeberli, W., Buetler, M., Huggel, C., Friedli, T. L., Schaub, Y., Schleiss, A. J.: New lakes in deglaciating high-mountain regions–opportunities and risks, Climatic change, 139(2), 201-214, doi: 10.1007/s10584-016-1771-5, 2016b.

Haeberli, W., Linsbauer, A., Cochachin, A., Salazar, C., Fischer, U. H.: On the morphological characteristics of overdeepenings in high-mountain glacier beds, Earth Surf Proc Land, 41(13), 1980-1990, doi: 10.1002/esp.3966, 2016a.

Haeberli, W., Schaub, Y., Huggel, C.: Increasing risks related to landslides from degrading permafrost into new lakes in de-glaciating mountain ranges, Geomorphology, 293, 405-417, doi: 10.1016/j.geomorph.2016.02.009, 2017.

Haemmig, C., Huss, M., Keusen, H., Hess, J., Wegmüller, U., Ao, Z., Kulubayi, W.: Hazard assessment of glacial lake outburst floods from Kyagar glacier, Karakoram mountains, China, Ann Glaciol, 55(66), 34-44, doi: 10.3189/2014AoG66A001, 2014.

Hock, R.: A distributed temperature-index ice- and snowmelt model including potential direct solar radiation, J Glaciol, 45, 101–111, doi: 10.3189/S0022143000003087, 1999.

Huggel, C., Haeberli, W., Kääb, A., Bieri, D., Richardson, S.: An assessment procedure for glacial hazards in the Swiss Alps, Can Geotech J, 41(6), 1068-1083, doi: 10.1139/t04-053, 2004.

Huggel, C., Caplan-Auerbach, J., Gruber, S., Molnia, B., Wessels, R.: The 2005 Mt. Steller, Alaska, rock-ice avalanche: A large slope failure in cold permafrost, In Proceedings of the 9th International Conference on Permafrost, 29, 747-752, 2008.

Huggel, C., Fischer, L., Schneider, D., Haeberli, W.: Research advances on climate-induced slope instability in glacier and permafrost high-mountain environments, Geographica Helvetica, 65(2), 146-156, doi: 10.5194/gh-65-146-2010, 2010.

Huss, M., Farinotti, D., Bauder, A., Funk, M.: Modelling runoff from highly glacierized alpine drainage basins in a changing climate, Hydrol Process, 22(19), 3888-3902, doi: 10.1002/hyp.7055, 2008.

Huss, M., Farinotti, D.: Distributed ice thickness and volume of all glaciers around the globe, J Geophys Res-Earth, 117(F4), doi: 10.1029/2012jf002523, 2012.

Huss, M., Zemp, M., Joerg, P.C., Salzmann, N.: High uncertainty in 21st century runoff projections from glacierized basins, J. Hydrol. 510, 35–48, doi: 10.1016/j.jhydrol.2013.12.017, 2014.

Huss, M., Fischer, M.: Sensitivity of very small glaciers in the Swiss Alps to future climate change, Frontiers in Earth Science, 4:34, 1-17, doi: 10.3389/feart.2016.00034, 2016.

ICIMOD 2011: Glacial lakes and glacial lake outburst floods in Nepal. Kathmandu: ICIMOD, 109 pp, 2011.

IPCC 2007: Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change [Solomon, S., D. Qin, M. Manning, Z. Chen, M. Marquis, K.B. Averyt, M. Tignor, H.L. Miller (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, 996 pp, 2007.

IPCC 2013: Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change [Stocker, T.F., D. Qin, G.-K. Plattner, M. Tignor, S.K. Allen, J. Boschung, A. Nauels, Y. Xia, V. Bex, P.M. Midgley (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, 1535 pp, 2013.

Irvine-Fynn, T. D. L., Hodson, A. J., Moorman, B. J., Vatne, G., Hubbard, A. L.: Polythermal glacier hydrology: a review, Rev Geophys, 49, 1–37, doi: 10.1029/2010RG000350.1.INTRODUCTION, 2011.

Janský, B., Šobr, M., Engel, Z.: Outburst flood hazard: case studies from the Tien-Shan Mountains, Kyrgyzstan, Limnologica, 40(4), 358-364, doi: 10.1016/j.limno.2009.11.013, 2010.

Kalnay, E., Kanamitsu, M., Kistler, R., Collins, W., Deaven, D., Gandin, L., Iredell, M., Saha, S., White, G., Woollen, J., Zhu, Y., Leetmaa, A., Reynolds, R., Chelliah, M., Ebisuzaki, W., Higgins, W., Janowiak, J., Mo, K.C., Ropelewski, C., Wang, J., Jenne, R., Joseph, D.: The NCEP/NCAR 40-Year Reanalysis Project. Bull Am Meteorol Soc, 77, 437–471, doi: 10.1175/1520-0477(1996)077%3C0437:TNYRP%3E2.0.CO;2, 1996.

Karlsson, J. M., Lyon, S. W., Destouni, G.: Thermokarst lake, hydrological flow and water balance indicators of permafrost change in Western Siberia, J Hydrol, 464, 459-466, doi: 10.1016/j.jhydrol.2012.07.037, 2012.

Kneisel, C.: Permafrost in recently deglaciated glacier forefields - measurements and observations in the eastern Swiss Alps and northern Sweden, Z Geomorphol, 47, 289–305, 2003.

Kneisel, C.: Assessment of subsurface lithology in mountain environments using 2D resistivity imaging, Geomorphology, 80(1-2), 32-44, doi: 10.1016/j.geomorph.2005.09.012, 2006.

Kneisel, C., Hauck, C., Fortier, R., Moorman, B.: Advances in geophysical methods for permafrost investigations, Permafrost Periglac, 19(2), 157-178, doi: 10.1002/ppp.616, 2008.

Lei, Y., Yao, T., Yi, C., Wang, W., Sheng, Y., Li, J., Joswiak, D.: Glacier mass loss induced the rapid growth of Linggo Co on the central Tibetan Plateau, J Glaciol, 58(207), 177-184, doi: 10.3189/2012JoG11J025, 2012.

Li, S., Zhan, H., Lai, Y., Sun, Z., Pei, W.: The coupled moisture-heat process of permafrost around a thermokarst pond in Qinghai-Tibet Plateau under global warming, J Geophys Res-Earth, 119(4), 836-853, doi: 10.1002/2013JF002930, 2014.

Linsbauer, A., Paul, F., Hoelzle, M., Frey, H., Haeberli, W.: The Swiss Alps without glaciers–a GIS-based modelling approach for reconstruction of glacier beds, Proceedings of Geomorphometry, 31, 243-247, 2009.

Linsbauer, A., Paul, F., Haeberli, W.: Modeling glacier thickness distribution and bed topography over entire mountain ranges with GlabTop: Application of a fast and robust approach, J Geophys Res-Earth, 117(F3), doi: 10.1029/2011jf002313, 2012.

Linsbauer, A., Frey, H., Haeberli, W., Machguth, H., Azam, M. F., Allen, S.: Modelling glacier-bed overdeepenings and possible future lakes for the glaciers in the Himalaya—Karakoram region, Ann Glaciol, 57(71), 119-130, doi: 10.3189/2016aog71a627, 2016.

Liu, J. J., Cheng, Z. L., Su, P. C.: The relationship between air temperature fluctuation and Glacial Lake Outburst Floods in Tibet, China, Quatern Int, 321, 78-87, doi: 10.1016/j.quaint.2013.11.023, 2014.

Lliboutry, L., Arnao, B. M., Pautre, A., Schneider, B.: Glaciological problems set by the control of dangerous lakes in Cordillera Blanca, Peru. I. Historical failures of morainic dams, their causes and prevention, J Glaciol, 18(79), 239-254, doi: 10.3189/S0022143000029610, 1977.

Loke, M. H., Barker, R. D.: Least-squares deconvolution of apparent resistivity pseudosections, Geophysics, 60(6), 1682-1690, doi: 10.1190/1.1443900, 1995.

Marchenko, S. S., Gorbunov, A. P., Romanovsky, V. E.: Permafrost warming in the Tien Shan mountains, central Asia, Global Planet Change, 56(3), 311-327, doi: 10.1016/j.gloplacha.2006.07.023, 2007.

McKillop, R. J., Clague, J. J.: A procedure for making objective preliminary assessments of outburst flood hazard from moraine-dammed lakes in southwestern British Columbia, Nat Hazards, 41(1), 131-157, doi: 10.1007/s11069-006-9028-7, 2007.

Mergili, M., Schneider, J. F.: Regional-scale analysis of lake outburst hazards in the southwestern Pamir, Tajikistan, based on remote sensing and GIS, Nat Hazards Earth Sys, 11(5), 1447-1462, 2011.

Mergili, M., Müller, J. P., Schneider, J. F.: Spatio-temporal development of high-mountain lakes in the headwaters of the Amu Darya River (Central Asia), Global Planet Change, 107, 13-24, doi: 10.5194/nhess-11-1447-2011, 2013.

Narama, C., Kääb, A., Duishonakunov, M., Abdrakhmatov, K.: Spatial variability of recent glacier area changes in the Tien Shan Mountains, Central Asia, using Corona (~ 1970), Landsat (~ 2000), and ALOS (~ 2007) satellite data, Global Planet Change, 71(1), 42-54, doi: 10.1016/j.gloplacha.2009.08.002, 2010.

Narama, C., Daiyrov, M., Tadono, T., Yamamoto, M., Kääb, A., Morita, R., Ukita, J.: Seasonal drainage of supraglacial lakes on debris-covered glaciers in the Tien Shan Mountains, Central Asia, Geomorphology, 286, 133-142, doi: 10.1016/j.geomorph.2017.03.002, 2017.

Nash, J. E., Sutcliffe, J. V.: River flow forecasting through conceptual models part I—A discussion of principles, J Hydrol, 10(3), 282-290, doi: 10.1016/0022-1694(70)90255-6, 1970.

Noetzli, J., Hoelzle, M., Haeberli, W.: Mountain permafrost and recent Alpine rock-fall events: a GIS-based approach to determine critical factors, Proceedings of the 8th International Conference on Permafrost, 2, 827-832, 2003.

Nosenko, G. A., Lavrentiev, I. I., Glazovskii, A.F., Kasatkin, N. E., Kokarev, A. L.: The polythermal structure of Central Tuyuksu glacier, Earth's Cryosphere (Kriosfera Zemli), XX (4), 93–102, doi: 10.21782/kz1560-7496-2016-4(105-115), 2016.

Paul, F., Linsbauer, A.: Modeling of glacier bed topography from glacier outlines, central branch lines, and a DEM, Int J Geogr Inf Sci, 26(7), 1173-1190, doi: 10.1080/13658816.2011.627859, 2012.

Pieczonka, T., Bolch, T., Junfeng, W., Shiyin, L.: Heterogeneous mass loss of glaciers in the Aksu-Tarim Catchment (Central Tien Shan) revealed by 1976 KH-9 Hexagon and 2009 SPOT-5 stereo imagery, Remote Sens Environ, 130, 233-244, doi: 10.1016/j.rse.2012.11.020, 2013.

Richardson, S. D., Reynolds, J.M.: Degradation of ice-cored moraine dams: implications for hazard development, Debris-Covered Glaciers, Proceedings of a workshop, Seattle, Washington, USA, Sept 2000.

Roeckner, E., Bauml, G., Bonaventura, L., Brokopf, R., Esch, M., Giorgetta, M., Hagemann, S., Kirchner, I., Kornblueh, L., Manzini, E., Rhodin, A., Schlese, U., Schulzweida, U., Tompkins, A.: The atmospheric general circulation model ECHAM5. Part I: Model description, Rep. 349, Max Planck Institute for Meteorology, 127 pp, 2003.

Shatravin, V. I.: Reconstruction of the Pleistocene and Holocene glaciations of the Tian-Shan and Pamir: new results. International Workshop, July 22–23, Institute of Soil Science and Soil Geography, University of Bayreuth, Germany, 2000.

Shrestha, A. B., Eriksson, M., Mool, P., Ghimire, P., Mishra, B., Khanal, N. R.: Glacial lake outburst flood risk assessment of Sun Koshi basin, Nepal, Geomat Nat Haz Risk, 1(2), 157-169, doi: 10.1080/19475701003668968, 2010.

Shur, Y. L., Jorgenson, M. T.: Patterns of permafrost formation and degradation in relation to climate and ecosystems, Permafrost and Periglacial Processes, 18(1), 7-19, doi: 10.1002/ppp.582, 2007.

Sobr, M., Jansky, B.: The morphometric parameters of glacial lakes in the Bohemian Forest, Silva Gabreta, 22, 31-61, 2016.

Sorg, A., Bolch, T., Stoffel, M., Solomina, O., Beniston, M.: Climate change impacts on glaciers and runoff in Tien Shan (Central Asia), Nat Clim Change, 2(10), 725-731, 2012.

Sorg, A., Huss, M., Rohrer, M., Stoffel, M.: The days of plenty might soon be over in glacierized Central Asian catchments, Environ Res Lett, 9(10), 104018, doi: 10.1038/nclimate1592, 2014.

Thompson, S., Kulessa, B., Luckman, A.: Integrated electrical resistivity tomography (ERT) and self-potential (SP) techniques for assessing hydrological processes within glacial lake moraine dams, J Glaciol, 58(211), 849-858, doi: 10.3189/2012jog11j235, 2012.

Wang, X., Siegert, F., Zhou, A. G., Franke, J.: Glacier and glacial lake changes and their relationship in the context of climate change, Central Tibetan Plateau 1972–2010, Global Planet Change, 111, 246-257, doi:10.1016/j.gloplacha.2013.09.011, 2013.

Wang, L., Li, Z., Wang, F., Edwards, R.: Glacier shrinkage in the Ebinur lake basin, Tien Shan, China, during the past 40 years, J Glaciol, 60(220), 245-254, doi: 10.3189/2014JoG13J023, 2014.

WGMS: Fluctuations of Glaciers Database, World Glacier Monitoring Service, Zurich, Switzerland. Online access: http://dx.doi.org/10.5904/wgms-fog-2017-10, 2017.

Xu, D.: Characteristics of debris flow caused by outburst of glacial lake in Boqu river, Xizang, China, 1981, GeoJournal, 17(4), 569-580, doi: 10.1007/BF00209443, 1988.

Yamada, T.: Glacier lake and its outburst flood in the Nepal Himalaya, Data Center for Glacier Research, Japanese Society of Snow and Ice, Tokyo, 1998.

Zaginaev, V.: The Teztor lake outburst event (northern Tien-Shan). Monitoring, forecasting of hazardous processes and phenomena in the territory of the Kyrgyz Republic, Ministry of Emergency Situations of the Kyrgyz Republic, 10th ed., 563-570, 2013.

Zaginaev, V., Ballesteros-Cánovas, J. A., Erokhin, S., Matov, E., Petrakov, D., Stoffel, M.: Reconstruction of glacial lake outburst floods in northern Tien Shan: Implications for hazard assessment, Geomorphology, 269, 75-84, doi:10.1016/j.geomorph.2016.06.028, 2016.

Zhang, G., Yao, T., Piao, S., Bolch, T., Xie, H., Chen, D., Gao, Y., O'Reilly, C.M., Shum, C.K., Yang, K., Yi, S.: Extensive and drastically different alpine lake changes on Asia's high plateaus during the past four decades, Geophys Res Lett, 44(1), 252-260, doi: 10.1002/2016GL072033, 2017.

Zhao, Q., Zhang, S., Ding, Y. J., Wang, J., Han, H., Xu, J., Zhao, C., Guo, W., Shangguan, D.: Modeling hydrologic response to climate change and shrinking glaciers in the highly Glacierized Kunma Like River catchment, Central Tian Shan, J Hydrometeorol, 16(6), 2383-2402, doi: 10.1175/JHM-D-14-0231.1, 2015.

5.3 Hydrological regime of a proglacial lake

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Falátková, K., Šobr, M., Kocum, J., Janský, B. (2014): Hydrological Regime of Lake Adygine, Tien Shan, Kyrgyzstan. Geografie, 119(4), 320-341.

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HYDROLOGICAL REGIME OF ADYGINE LAKE, TIEN SHAN, KYRGYZSTAN

FALÁTKOVÁ, K., ŠOBR, M., KOCUM, J., JANSKÝ, B. (2014): Hydrological regime of Adygine lake, Tien Shan, Kyrgyzstan. Geografie, 119, No. 4, pp. 320–341. – This article provides a detailed analysis of the hydrological regime of the Adygine glacial lake in Tien Shan, Kyrgyzstan, and of the specific factors which affect it. Glaciers of central Tien Shan are considered to be very sensitive indicators of climate change. The studied lake belongs to a system of relatively recently formed lakes situated near the front of the retreating glacier and is numbered amongst the potentially dangerous. The lake's water level and its dependence on the development of climatic conditions in the area were monitored in detail in 2007–2012. A substantial part of this paper is the evaluation of the inflow and outflow balance of the lake's basin. The results confirmed that the hydrological regime is glacial and exhibits its typical characteristics, such as a seasonal evolution of runoff delay or significant diurnal fluctuation of the lake's water level. During the monitored period, no major changes in annual lake level fluctuation were recorded.
KEY WORDS: hydrological regime – glacial lake – glacier melt – water level fluctuation – surface outflow – hydrological balance.

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1. Introduction

Climatic changes and their manifestation currently represent one of the main topics of hydrology (Aizen et al. 1997; Dvořák, Hladný, Kašpárek 1997; Nijssen et al. 2001; Middelkoop et al. 2001; Bates et al., eds. 2008). Glaciers in high mountain regions of the world are considered very sensitive indicators of climatic change (Beniston, Diaz, Bradley 1997; Li et al. 2007; Hagg et al. 2007; Zhou, Jing, Zhao et al. 2010; Janský et al. 2011). These areas include also the northern part of the Tien Shan mountain range (Gao et al. 2010). Development of glacial lakes within Kyrgyz range is monitored in detail since 2004 (Černý et al. 2010). The rapid retreat of glaciers causes extreme increase in meltwater flows in the summer periods, which leads to overfilling of lake basins and destabilization of their moraine dams (Ageta et al. 2000). These processes result in frequent outbursts of glacial lakes (Clague, Evans 2000; Kattelmann 2003). The studied lake is a part of a lake system and is a very suitable research subject. At this site there is a risk of outbursts and subsequent development of floods or mudflows threatening Ala Archa Valley, which is very popular with tourists and is directed towards Bishkek, the capital of Kyrgyzstan.

Adygine Lake is a relatively young glacial lake (began to develop in 1960's) dammed by a rock step which is covered by moraine material with buried ice (Černý et al. 2010). This type of dam is very frequent in high mountain areas (Janský, Šobr, Yerokhin 2006) and is relatively vulnerable in terms of its stability and given its subsurface outflow paths. The danger then lies in the fact that Adygine Lake is part of an entire system of dynamically forming and developing lake basins.

The aim of this paper is to evaluate the hydrological regime of the largest lake of the site – Adygine Lake – based on climatic and hydrological measurements. The results presented here could be used in the context of further procedures associated with modelling of possible lake outburst scenarios, or with installation of a warning system for Bishkek. The results of the work are compared with relevant literature focusing on similar topics.

2. Glacial Hydrological Regime

Lakes with glacier meltwater inflow show a similar annual and diurnal cycle of water level fluctuations as glacial streams. Changes in water volume of water in the lake are caused by changes in the hydrological balance, especially the inflow. The highest water levels correlate with periods of highest water inflow, which usually occurs in July and August. The hydrological regime of the lake inflow is influenced by climatic conditions in the lake catchment. Thanks to the retention capacity of the lake, however, changes in the water level are considerably smoother than in rivers (Singh, Singh 2001).

Inflow into a lake is composed of four main sources, the importance and influence of them changes throughout the year. These are the base flow, snowmelt, glacier melt and atmospheric precipitation (Singh, Singh 2001). Base flow has multiple sources, which are described in detail by Röthlisberger and Lang (1987). One of them is the runoff resulting from melting on the glacier base. Melting of ice occurs due to friction and energy exchange between the glacier and its bedrock. Other sources can include gradually released water retained in glacial cavities, meltwater leaking through the snow or firn or groundwater. Base flow is a relatively constant source of water for the lake throughout the year. It is virtually the only source in winter and its share of the total inflow decreases distinctly in summer.

Another source is water from melting snow. In the studied area, snow begins to melt in April-May and continues until June, sometimes July. Due to the high albedo of a new snow, melting is not as intensive as in the case of glacier ice and the retention capacity of the snow cover slows the meltwater runoff even further (Singh, Singh 2001). This results in a relatively balanced inflow from snow melt with rather smaller diurnal fluctuations, which reaches the lake with a certain delay.

Inflow from glacier surface melting is the main source of water during the summer period. Glacial ice has lower albedo than snow, therefore it melts faster and, moreover, it does not have the same retention capacity as snow. However, meltwater may be retained in cracks and cavities in the glacier (Röthlisberger, Lang 1987). Therefore, water flows on the glacier surface as well as through moulins and channels inside the glacier (R-channels; Röthlisberger 1972) and then through channels incised into the bedrock or sediment (N-channels; Nye 1973). As summer season progresses, this drainage system is so developed that the meltwater from the glacier reaches the lake in a relatively short time. Thanks to the rapid response of the glacier to increased solar radiation and air temperature, a significant diurnal regime is developed for lakes and streams supplied from the glacier (Singh, Singh 2001). The water level maximum occurs several hours after sun culmination, then the inflow slowly declines to the minimum in the morning followed by a rapid increase (Röthlisberger, Lang 1987). The time lag therefore ranges from dozens of minutes to several hours (depending on the lake distance from the glacier front). However, shortening of this time during the ablation season (due to increasingly efficient system of drainage channels) is particularly important (Singh, Singh 2001).

The fourth source is water from precipitation that occurs mostly during summer at the studied site. Although the direct effect of precipitation on lake water level is usually not very important, its indirect effect is rather significant (Collins 1998). Warm liquid precipitation, especially in late summer, falling on the glacier surface accelerates its melting, and thus may cause a significant increase in runoff from the glacier while solid precipitation (snow) has a dampening effect, as it almost completely stops surface glacier melting and the daily glacial regime is replaced by a low base flow even for several days after the precipitation event (Hubbard, Glasser 2005).

Changes in the runoff volume from a glacier can be monitored at several time scales. Short-term daily and annual variability is caused by the variability of meteorological conditions. However, changes in runoff reflecting changes in mass balance of the glacier can be seen in the long term. Gradual air warming first leads to an increased runoff from the glacier, but later the runoff decreases due to a reduction of the glacier area (Benn, Evans 2010).

3. Study Area

The Adygine area is located in the higher eastern part of the Kyrgyz range which is also called Northern Tien Shan (Adyshev, ed. 1987, Koppes et al. 2008). The Adygine glacier complex (3,400–4,200 m a.s.l.) is located on the northern side of the Kyrgyz ridge about 40 km south of Bishkek. The study area is at the end of a tributary valley that joins the main Ala Archa valley from the west. Several genetic types of glacial lakes could be found in this area (Fig. 1): morainic-glacier lakes, thermokarst lakes or lakes dammed by a rock step. The Adygine Glacier as well as the majority of glaciers in Central Asia has been undergoing significant degradation since mid-1920s, which has even intensified in recent years (Dyurgerov 2003, Aizen et al. 2006, Unger-Shayesteh et al. 2013).

The retreat of the glacier front has resulted in formation of five small lakes in the adjacent depressions dammed by a hummocky moraine. Despite the low volume of retained water, these lakes are considered potentially dangerous due to their rapid development. The first signs of the lakes were reported in 2004 and a year after the largest of them (Lake 3) had a surface area of 5,700 m^2 and a depth of 3.8 m (Černý et al. 2010). In 2012, the maximum depth of this lake was already 10.3 m and the surface area increased to 8,800 m^2. The largest lake of this area, also called Upper Adygine Lake, is located lower and is dammed by a rock step (riegel) and partly also by a moraine with buried ice. There are two drainage paths from the lake: one on the surface over a considerably weathered rock outcrop and the second through subsurface channels. The entire complex is complemented by the Lower Adygine Lake supplied by the surface flow from the higher located Upper Adygine Lake. It is a thermokarst lake without surface outflow with significant fluctuation in water level and volume of retained water during the day (between 10,000 and 15,000 m^3), which is probably caused by the limited capacity of subsurface drainage routes (Janský, Šobr, Engel 2010).

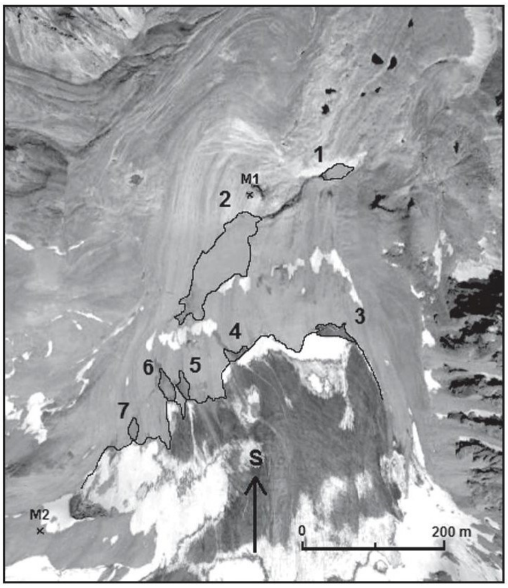

Fig. 1 – Satellite image of study area Adygine, 2006. 1 – Lower Adygine, 2 – Upper Adygine, 3–7 – present ice-contact lakes, M1, M2: meteorological stations. Source: Google Earth.

The present study is focused mainly on the largest lake, which is further on refered to as “Adygine Lake”.

The climate in the study area is continental. In winter it is affected mainly by the Siberian anticyclone that blocks the progress of air masses from the west and thus reduces the amount of winter precipitation in the region (Aizen, Aizen 1994). This is seen especially in the Ala Archa Valley. Total precipitation in winter is therefore very low (10–20 mm per month, Hagg et al. 2007).

Aizen, Aizen, Melack (1995) describe the distribution of precipitation in the mountains of Tien Shan during the year. The occurrence of one precipitation maximum in May–July is typical for mountainous regions of the northern hemisphere. At altitudes, where the Adygine area is situated, i.e. above 3,400 m a.s.l., summer precipitation represents according to Aizen, Aizen, Melack (1996) up to 72% of the annual amount, about 65% of this precipitation falls as snow. The precipitation minimum occurs in December and January when only 2–5% of the annual precipitation falls on average (Aizen, Aizen, Melack 1996).

The annual variation in temperature in the lower part of the Ala Archa Valley (2,200 m a.s.l.) is shown in Figure 2. The temperatures drop to the lowest point in January (average temperature of –9 °C in 2002–2010, minimum temperature goes down to –25 °C) and the air temperatures culminate during July and August with average temperature of 13 °C (Černý et al. 2010). About 1,500 metres higher in the Adygine area, the temperature cycle during the year

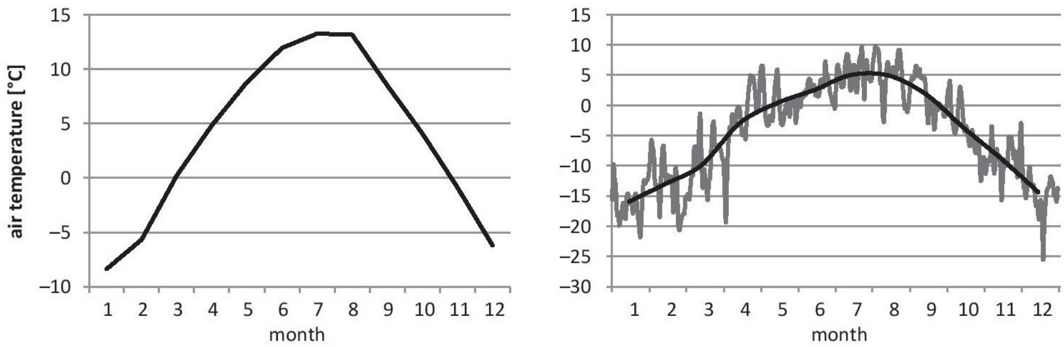

Fig. 2 – Mean monthly air temperature (left), meteorological station Ala Archa, 2,200 m a.s.l., 2002–2008 and mean monthly and daily air temperature (right), heat sensor at Adygine, 3,653 m a.s.l., 2011. Source: Černý et al. 2010.

is similar, however, the air temperatures are lower and the annual amplitude is smaller. The average temperature of the warmest month (August) is around 5 °C, in January the average temperature is mostly around –10 to –15 °C (Černý et al. 2010).

4. Material and Methods

4.1. Data

Background for research in Adygine area is provided by a research station, which was built near the outlet from Adygine Lake in 2008. Meteorological data are provided by two automatic meteorological stations, one of them is located at an altitude of 3,550 m ("lower") and the other one ("upper") is located at an altitude of 3,800 m a.s.l. To complement the temperature range from weather stations, data from a sensor located on the right side of the glacier basin at an altitude of 3,653 m a.s.l. were also used. Data from the national meterological station (2,200 m a.s.l.) located near the Ala Archa National Park were available for the description of atmospheric conditions in the lower part of the Ala Archa Valley.

Hydrological data are provided by a pressure sensor (water level indicator) which was installed in Adygine Lake in August 2007. From August 19, 2007 until July 24, 2013, only two outages were reported. Unfortunately they were both relatively long – the series is missing from July 31 till August 22, 2008 and from August 1 till August 17, 2011. A detailed bathymetric map based on measurements in 2008 (Černý et al. 2010) has been used to calculate the volume of the lake at various water levels.

4.2. Methods

Despite annual maintenance, continuous data series are not available from any of the meteorological stations, which is probably caused by low resistance

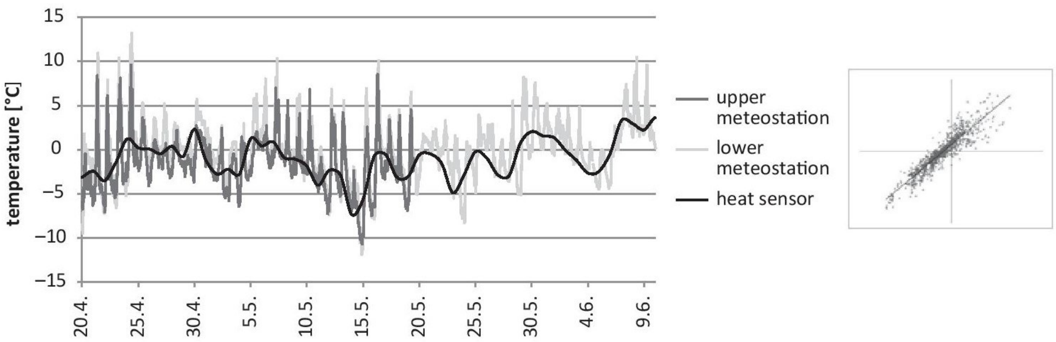

Fig. 3 – Air temperature fluctuations measured at upper meteorological station (left), lower meteorological station and mean daily air temperature measured by a heat sensor, 2010 and correlation graph of air temperature (right) from upper (axis x) and lower (axis y) meteorological station.

to weather conditions (strong wind, low temperatures) and mainly to static electricity effect during storms. Series from various sources were used for individual years due to lack of continuous data from a single station. The maximum altitude difference between the two meteorological stations is about 250 m, therefore the measured temperatures differ, however, their time development is analogous (Fig. 3) – the correlation coefficient of the two monthly temperature series from the stations is 0.92.

On July 26, 2008 at 7:30 a.m., the lake level was measured by total geodetic station at 3,542.966 m a.s.l. and this corresponded to 5,966 m at the level indicator. Therefore, a value of 3,537 m was added to all values measured by the level indicator to obtain information about the altitude of water levels.

The methodology used for the bathymetry measurements is described by Česák and Šobr (2005). This methodology was modified for use in high mountain areas and for research of glacial lakes.

Discharge at the outflow from Adygine Lake was measured using a hydrometric propeller OTT C2 and in accordance with ČSN ISO 748 on 19 verticals at an interval of 0.5 m (Fig. 4). The only exception was the measurement at the highest water level (0.67 m), where the width of the outflow was divided into 25 verticals at 0.5 m. The measurements were taken from August 3 to August 5, 2012 at the same location at different water levels. During the measurement, the value on the water-level measuring pole located near the outlet was recorded, so that the corresponding altitude of the lake level could be assigned to the measured depths of the profiles. Therefore, it was necessary to identify the relationship between the values on the water-level measuring pole and on the level indicator: 0.47 m on the pole corresponded to 6.035 m on the level indicator.

When measuring the discharge, the maximum water level of the lake reached 3,543.236 m a.s.l., however, the highest level recorded by the level indicator was 104 mm higher (3,543.340 m on July 6, 2008 at 3:30 p.m.). When creating the rating curve, it was therefore necessary to extrapolate flow rates for the highest water levels (above 3,543.236 m a.s.l.).

The annual cycle of lake level fluctuations was divided into three parts. Water level decline (phase I) was determined from the moment, when the water

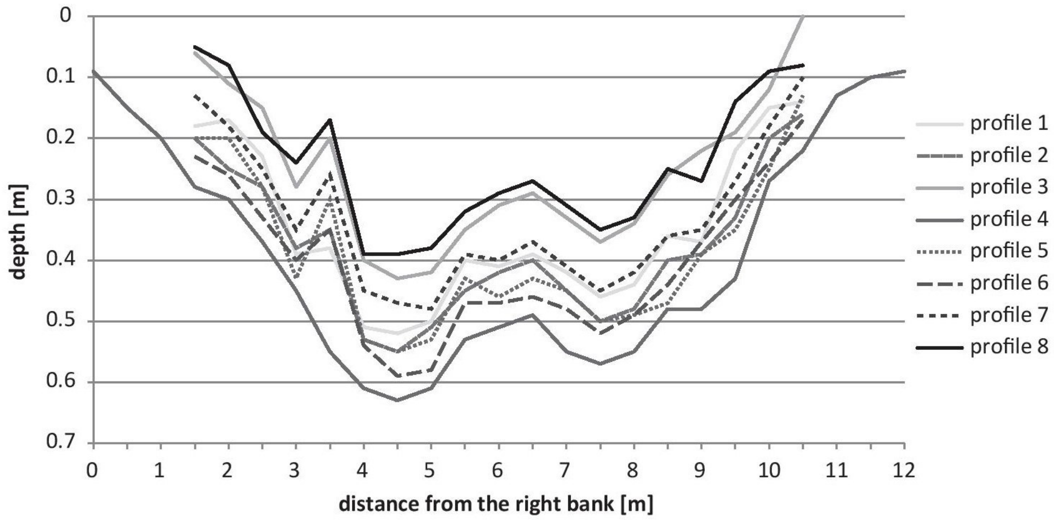

Fig. 4 – Cross section profiles of lake outflow at different water stages. Measurement time: profile 1 – 3.8. 15:30–16:00; profile 2 – 3.8. 17:45–18:05; profile 3 – 4.8. 8:00–8:15; profile 4 – 4.8. 17:20–17:40; profile 5 – 4.8. 20:10–20:35; profile 6 – 5.8. 10:10–10:30; profile 7 – 5.8. 13:50–14:10; profile 8 – 5.8. 15:00–15:15.

level was only declining to the minimum of the given cycle. This phase was additionally divided into two parts – in the first part, the level was declining rapidly, in the second one the decline was significantly slower. The boundary between the two parts was determined visually from the graph of the annual variation to November 6 (this day is included in the first part of the decline). When the minimum water level is reached (stage 1), the filling phase begins (phase II), which continues until the moment when the water level first begins to decline after the restoration of the surface outflow from the lake (stage 2). After the lake basin is filled with meltwater from snow and the glacier, the stabilization phase occurs (phase III) during which the water level fluctuates around an altitude of approximately 3,543 m and shows a daily amplitude of as much as 0.34 m.

There are multiple inflows to the lake, but they are very difficult to measure and a significant part of water also flows into the lake through subsurface channels. The inflow was therefore derived from the change in water level, volume of surface outflow and an approximate calculation of the volume of subsurface outflow. The arithmetic mean of the values of the decline rate in the first phase of all monitored years was calculated to determine the approximate subsurface outflow in the stabilization phase. During filling of the lake, the arithmetic mean of the second part of the decline was used up to the altitude of 3,541 m a.s.l. and from this level the same value as during stabilization was used.

5. Results

5.1. Lake Level Fluctuations

The annual cycle of lake level fluctuations shows a considerable regularity within a monitored period of five consecutive cycles as demonstrated in Figure 5. Each curve can be divided into three parts: I. level decline, II. filling, and III. stabilization.

The longest lasting is the first part (about 7 months) during which the lake water level declines due to outflow through subsurface channels. There is a significant drop in temperature in September and thus the volume of meltwater flowing into the lake from the glacier decreases. Around mid-September, the lake level declines to the point, from which the lake has no surface outflow. This occurs at a lake level altitude of 3,542.606 m. From this level, the lake is drained only through subsurface channels, the capacity of which remained relatively unchanged throughout the monitored period. The convex shape of the curves shows that some outflow routes are uncovered with the water level decline and therefore the rate of water level decline decreases (Table 1).

The second phase – filling of the lake – begins after the lowest water level is reached. The minimum water level varies from year to year – the largest difference between the measured lowest water levels in individual years was 288 mm (2008: 3,540.025 m a.s.l.; 2010: 3,540.313 m a.s.l.). Reaching of the minimum water level is affected by the capacity of drainage channels, available water input from snowmelt in spring, the volume of base flow, but also by relatively high air temperatures in September (or October) that could delay the onset of the water level decline and thereby shorten this period. The beginning, course and time of filling vary considerably between the monitored years (Table 2). The earliest beginning of the filling occurred in 2012 (April 17) and the latest filling began May 3, 2009. The entire filling process depends only on the course of air temperature and radiation, which affects the intensity of snow melting.

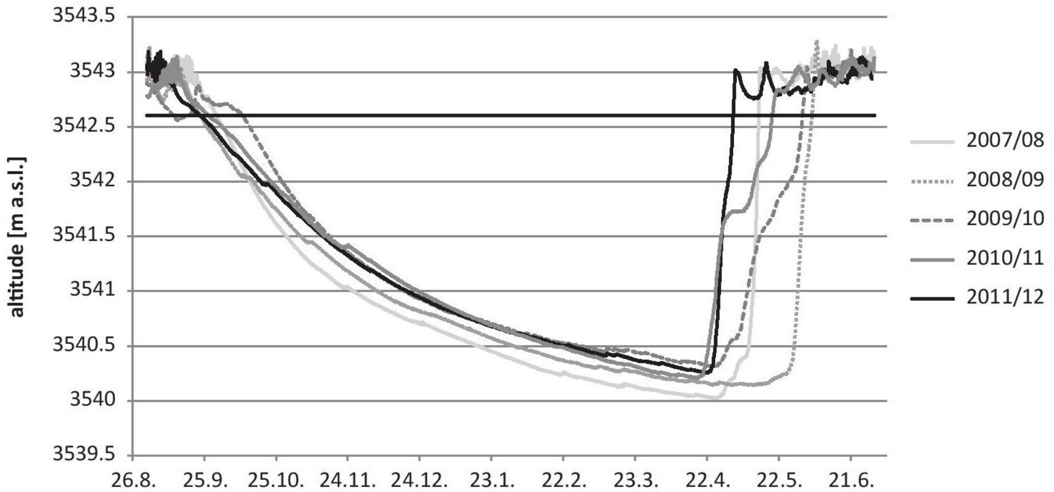

Fig. 5 – Annual course of water level fluctuation at lake Adygine. Surface outflow altitude marked with a solid black line.

Tab. 1 – Parameters of lake water level decline

Year	1st part of decline (until 6th Nov)			2nd part of decline (since 7th Nov)		
	Volume (m^3) / water level (cm) change	Time (days)	Velocity (l/s) / (cm/day)	Volume (m^3) / water level (cm) change	Time (days)	Velocity (l/s) / (cm/day)
2007/08	31,457 / 131.6	39	*9.34 / 3.4*	26,854 / 126.5	172	*1.81 / 0.74*
2008/09	27,691 / 115.0	45	*7.12 / 2.6*	28,230 / 131.2	179	*1.83 / 0.73*
2009/10	21,958 / 90.2	28	*9.07 / 3.2*	30,520 / 139.1	172	*2.05 / 0.81*
2010/11	23,659 / 97.5	46	*5.95 / 2.1*	30,007 / 137.6	169	*2.06 / 0.81*
2011/12	23,054 / 94.9	42	*6.35 / 2.3*	31,529 / 144.7	163	*2.23 / 0.89*

Tab. 2 – Parameters of lake filling

Year	Beginning	Stage 1 (m a.s.l.)	End	Stage 2 (m a.s.l.)	Time (days)	Volume 1 (m^3)	Volume 2 (m^3)	Volume change (m^3)	Filling velocity (l/s)
2008	26.4.	3,540.025	14.5.	3,543.041	*18*	137,698	207,076	*69,378*	*44.6*
2009	3.5.	3,540.144	8.6.	3,543.292	*36*	140,088	213,902	*73,814*	*23.7*
2010	26.4.	3,540.313	9.6.	3,542.986	*44*	143,531	205,791	*62,260*	*16.4*
2011	23.4.	3,540.255	5.5.	3,543.020	*12*	142,343	206,682	*64,339*	*62.0*
2012	17.4.	3,540.210	21.5.	3,542.839	*34*	141,426	201,969	*60,543*	*20.6*

The third phase of the cycle of the annual evolution of lake water level is stabilization and daily fluctuations. The longest period of stabilization was recorded in 2011, when the lake was filled quite early due to increased temperatures in late April and the phase lasted for 133 days. In contrast, the shortest time of summer water level fluctuation was in 2010, when the water level began to fluctuate on June 10 and the water level declined already on September 12 due to rather low temperatures in early September. Overall, this phase lasted only 93 days, which is 70% of the longest period.

The diurnal cycle of lake level fluctuation depends on the diurnal temperature variability that affects the rate of snow and glacier melting. The highest water level is usually recorded several hours after the sun culmination, i.e. around 3–5 p.m. After that, the water level gradually declines and reaches its minimum in the morning, around 8–10 a.m., which is related to the delayed inflow into the lake.

A very similar pattern of daily water level fluctuations in the ablation season (Fig. 6) was identified in the monitored period. The amplitude of the water level during the day ranges within several tens of millimetres in first weeks. The daily water level fluctuations increase gradually with increasing air temperature and radiation intensity, while maximum values were identified in August, usually in its second half. A maximum value of about 300 mm was identified in each of the monitored period; however, in 2009 the water level fluctuation did not exceed 205 mm. The daily fluctuation declines relatively quickly until it stops completely and is replaced again by a phase of decline.

An interesting phenomenon typical for glacial hydrological regime is a runoff delay, which is variable throughout the ablation season. A considerable reduction of the runoff delay is observed at the beginning of the ablation

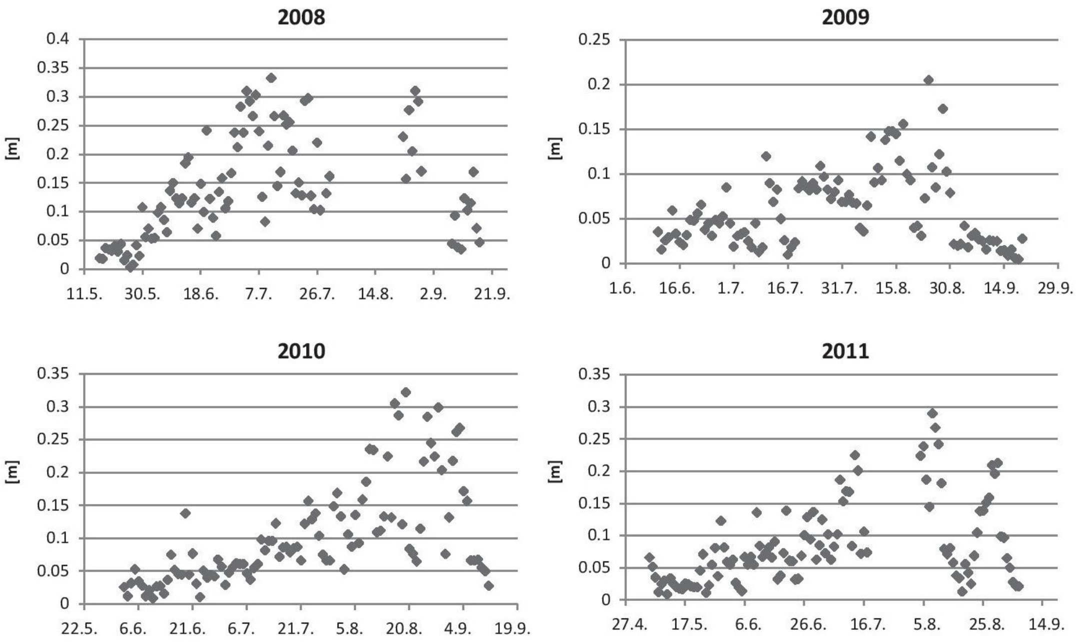

Fig. 6 – Water level fluctuation in ablation seasons of 2008–2011

season. It is attributed to the gradual melting of the snow cover, which slows down the runoff. At the peak of summer, the glacier responds to the increase in air temperature more quickly. The delay period thus decreases which is enhanced by a drainage system created and deepened during the ablation period and it extends up the glacier so the transport of meltwater becomes more effective (Hubbard, Glasser 2005). Beitlerová (2010) monitored the run-off delay for the ablation season of 2008 (Fig. 7). The first 10 days were of the largest delay, specifically 10–12 hours, followed by a period with a delay of 4–9 hours and at the peak of summer with a fully evolved diurnal cycle the delay reached only around 4 hours and even significantly lower values (1 hour) occurred. In late August, delays of 2–6 hours were detected (Beitlerová 2010).

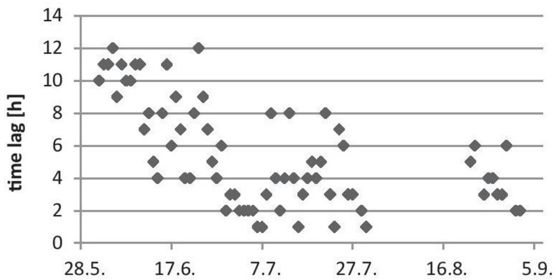

Fig. 7 – Runoff delay development in ablation season of 2008 (Beitlerová 2010)

5.2. Surface Outflow

The surface outflow from Adygine Lake runs over the lowest point of the dam starting from the water level reaching the altitude of 3,542.606 m. The discharge derived from the rating curve depends mainly on the air temperature at the given location. Due to rapid response of the glacier to increased solar

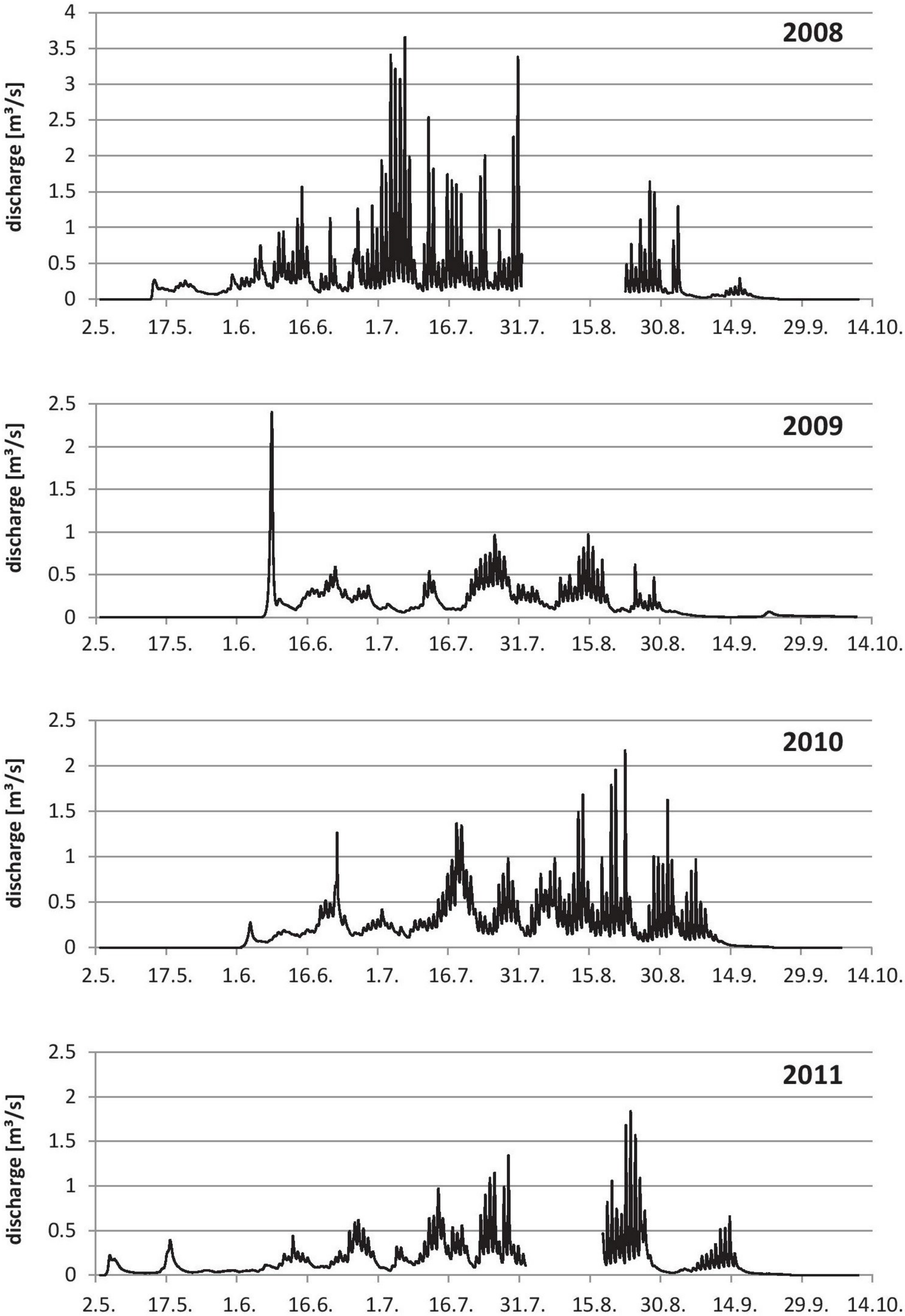

Fig. 8 – Discharge at lake Adygine outflow during ablation seasons of 2008–2011

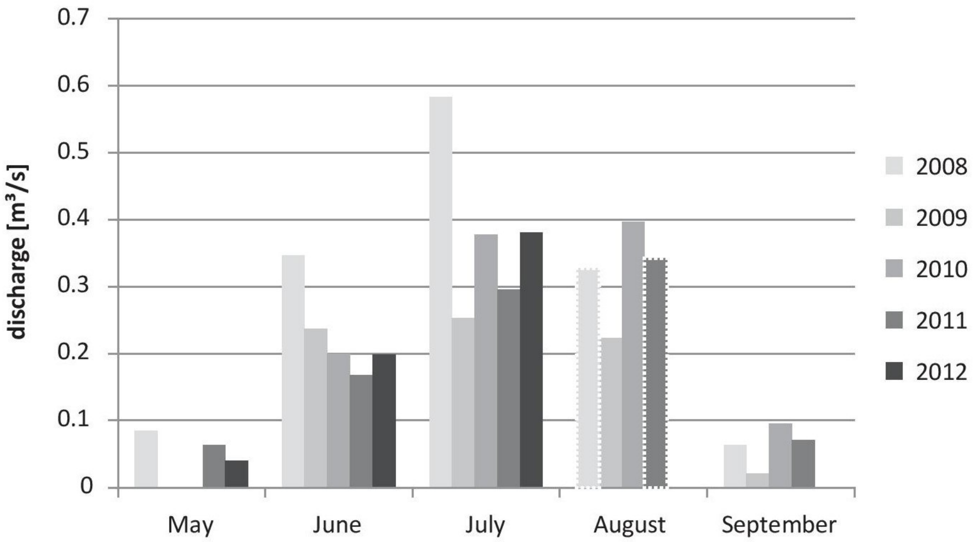

Fig. 9 – Mean monthly discharge at lake Adygine outflow, 2008–2012; white dashed outline noting incomplete data

radiation and air temperature, the lake shows a significant diurnal discharge regime at the lake outflow.

It is apparent from Figure 8 that the course of the ablation season significantly varies between the monitored years. The year 2008 was above-average in terms of water discharge at the outflow. Surface outflow was recorded relatively early and the high melting intensity of glacier ice showed in early July due to high air temperatures. A significant daily regime with maximum flow rates of over 3 $m^3.s^{-1}$ evolved and lasted with short breaks at least throughout July. In contrast, rather below-average flow rates were recorded in the following year. There was a very rapid increase in water level and thus also the discharge right at the beginning of the ablation season. It is likely that it was caused by intense snow melting over a large area. However, the discharge decreased and during the summer it did not exceed 1 $m^3.s^{-1}$. The diurnal regime evolved only in a limited extent. In the following two years (2010, 2011), summer flow rates were without any abnormalities. As the ablation season progressed, the diurnal regime evolved and flow rates reached the highest values in the second half of August.

The year of 2008 has high values of average flow rates for individual months (Fig. 9). Average monthly flow rates for 2009 are rather low and the similarity of values for June, July and August is unusual.

5.3. Balance of Lake Inflow and Outflow

The cumulative volume of surface outflow from Adygine Lake for individual monitored years 2008–2012 is shown in Figure 10. Differences in the initiation date, the course of outflow accumulation and its total volume for the entire ablation season are clearly noticeable.

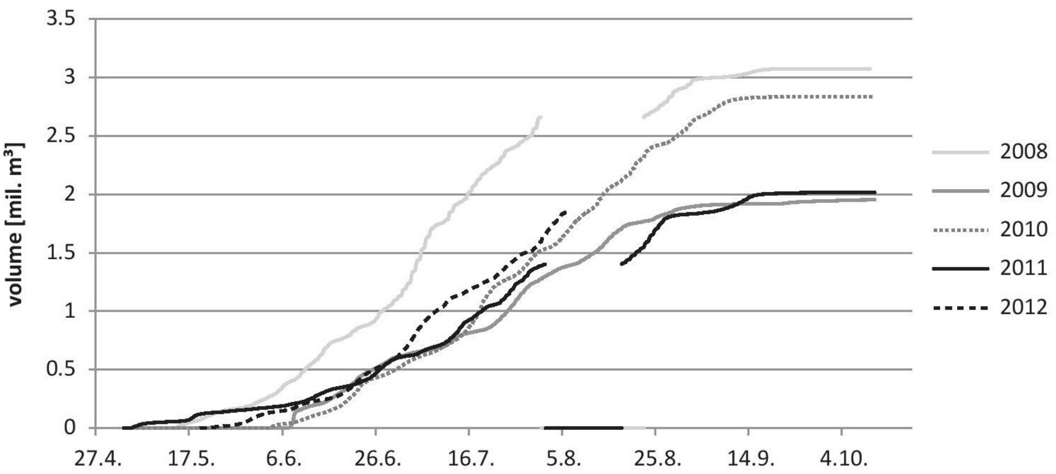

Fig. 10 – Cumulative volume of surface runoff for years 2008–2012

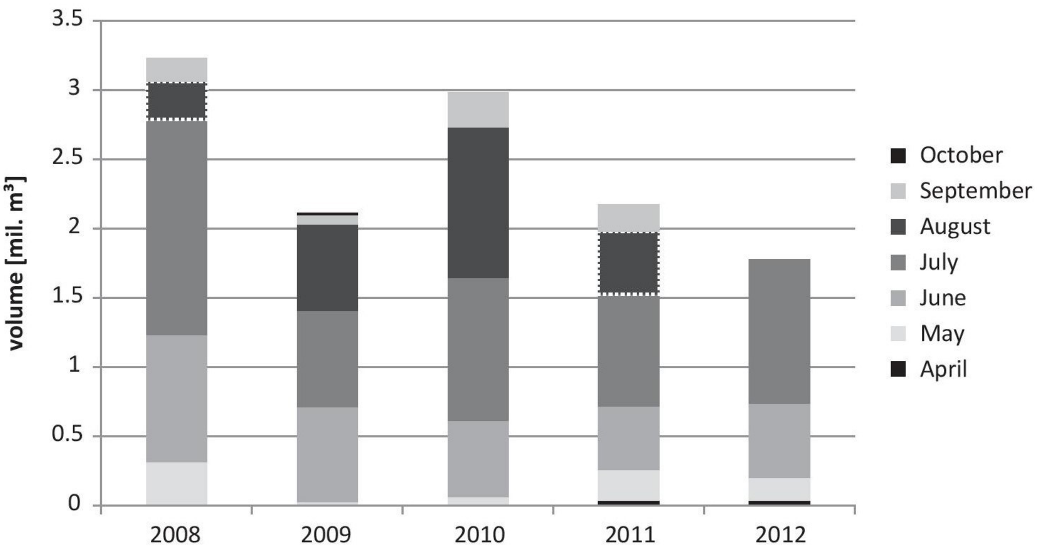

Fig. 11 – Volume of lake inflow for individual months of 2008–2012; white dashed outline noting incomplete data

Despite the different date of outflow initiation in individual years, the cumulative volume was around 500,000 m^3 on June 26 for most of the years. The only exception was the year 2008 when at this time the accumulated volume was nearly double. The total outflow volumes for individual years are significantly different. The most water drained from the lake in 2008 – it was 3.1 million m^3 for the measured period. Due to a lack of data (see 4.1. Data), however, this value does not correspond to the total volume for the entire ablation season. The actual volume of outflow water drained from the lake is estimated at 3.5–3.8 million m^3. The least amount of water drained from the lake through surface channel in 2009, specifically 1.9 million m^3.

For a complete description of the hydrological regime of Adygine Lake, it is necessary to add information on the inflow. The different height of the bars in the graph (Fig. 11) points to seasonal climate variability in the monitored area. The graph shows the distribution of inflowing water in individual months – April to October. Water started to flow into the lake (in a recordable amount) already in April during the last two monitored years (2011, 2012). In May, a significant inflow of water was detected in 2008 (nearly 309,000 m^3), however, in 2009 and 2010 the inflow was rather low for this month (27,000 and 59,000 m^3, respectively). There are no such major differences for June inflow volumes of individual years. June 2008 dominates with nearly 920,000 m^3 and June of the otherwise cold year 2009 was also relatively rich in water. July and August are most significant in terms of inflow. During these months >2 million m^3 of water flows into the lake, which was the case in 2010 and probably also in 2008. Air temperatures in September often drop below 0 °C and melting is thus limited reducing the inflow significantly. Relatively high temperatures in the first half of September 2010 caused that about 254,000 m^3 of water flowed into the lake in this month. In 2009, it was only 72,000 m^3, however, the air temperature did not drop in the second half of September and therefore inflow into the lake was recorded even in early October, which is rather unusual.

6. Discussion

6.1. Uncertainties of Measurements and Data Processing

It is very difficult to measure precipitation under the given conditions (see 4.2. Methods) at the studied area. Particularly solid precipitation is difficult to capture, as it is blown out of the rain gauge, which is not heated. There is not enough sunshine for a solar panel, which could provide heating of the rain gauge. The rain gauge in the upper meteorological station captures about 300–500 mm of precipitation per year. Given the altitude and the northern aspect of the site, the estimated values are around 1,200 mm (Černý et al. 2010). Aizen et al. (1996) mention, that there is a single precipitation maximum in the mountainous areas of the Kyrgyz range, specifically in May – July. At altitudes where the Adygine complex is situated, i.e. above 3,400 m a.s.l., precipitation in this time period represents up to 72 % of the annual precipitation and about 65% of the precipitation falls down in the form of snow (Aizen, Aizen, Melack 1996). Figure 12 shows daily precipitation for three months of 2012 with the highest precipitation – May: 58.4 mm, June: 114 mm, July: 79.9 mm (total of 252.3 mm). It is evident that the measured values do not correspond with the expected precipitation in this area and substantial part of precipitation is not detected by the measuring device. The meteorological stations also have incomplete data series due to technical problems. Therefore, the precipitation data could be used only in a limited way.

Other inaccuracies can occur during processing of the measured data. Depth of a certain spot was adjusted within two profiles when calculating the flow rates. After this adjustment, a rating curve with a higher reliability value was

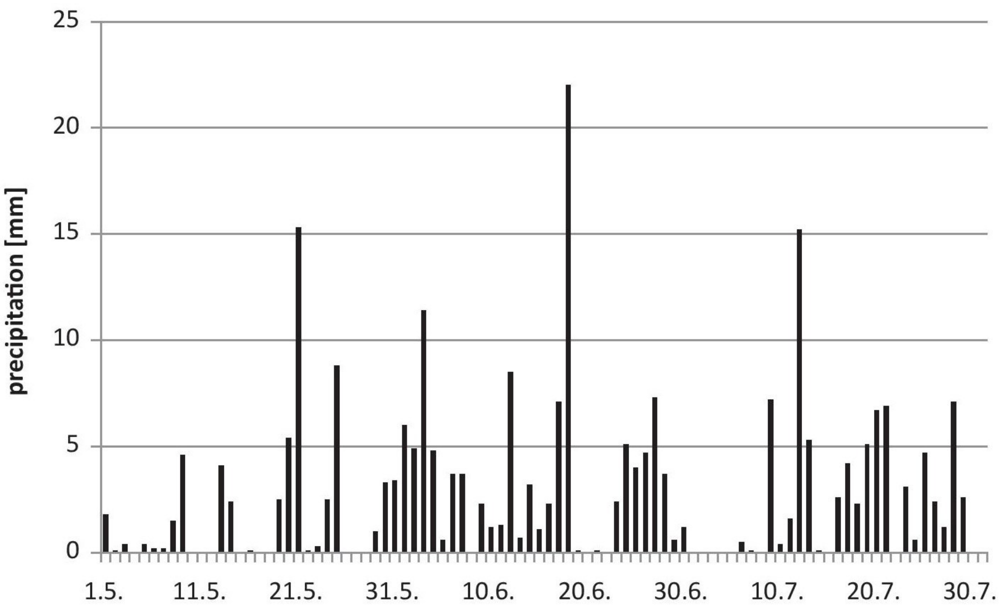

Fig. 12 – Daily amount of precipitation for May–June period of 2012, upper meteorological station

fitted through the points (x – discharge, y – water level). Flow rates for water levels of the entire monitored period were calculated using the equation of this curve. Generalization is necessary here, but the question is whether the number of measurements (eight) is sufficient. Except for one case, all points were located close to the curve. The reliability value of 0.9914 is satisfactory in this case. Nevertheless, it is necessary to mention that greater errors could probably occur here as the values of calculated flow rates are used for further calculations (e.g. outflow volume).

The calculation of the inflow to the lake is indirect. It was estimated on the basis of a change in the volume of the lake, surface and subsurface outflow. Uncertainties associated with the calculation of the volume of the lake and surface outflow have been outlined above. The exact volume of subsurface runoff is not known, it is estimated according to the water level decline without surface outflow. However, the lake is constantly supplied from the base flow from the glacier and therefore the subsurface runoff from the lake will be actually slightly higher. In addition, the volume of runoff through subsurface channels varies during the season and even between individual years as some of the channels could be closed and new ones opened. Complicated system of subsurface channels and cracks was confirmed by geophysical measurements in 2008 (Černý et al. 2010). This phenomenon was observed by the research team at similar lake sites in Kyrgyzstan as well. The arithmetic mean of all monitored years was used to approximate the calculated values to the actual volume of the inflow.

6.2. Glacial hydrological regime

In this study, the hydrological regime of Adygine Lake situated near the glacier front at an altitude of 3,643 m was described. Individual characteristics of the outflow exhibit elements typical for a glacial hydrological regime, which is described in detail in studies by Jobard and Dzikowski (2006), Singh et al. (2006) or Han et al. (2013).

The flow rate in the ablation season is considerably variable and depends mainly on air temperature and radiation. Starting with rather lower values, discharge at the outflow from Lake Adygine peaks in mid-summer (mostly in August). Similar discharge evolution during ablation season was observed also by Fountain (1985) in North Cascade Mts, USA, or by Jobard and Dzikowski (2006) in French Alps.

Swift et al. (2005) point out the fact that in spring a large part of short-wave radiation is reflected due to high albedo of snow and therefore melting occurs after an increase in air temperature. Therefore, the influence of radiation is limited here. On the other hand, relatively lower albedo of exposed glacier increases the importance of radiation, which accelerates melting (the ice-albedo positive feedback).

The influence of precipitation on flow rate is probably more complex. Some studies have considered it to be minimal, e.g. Han et al. (2013) did not find any significant dependence of flow rate on precipitation. In contrast, Fujita (2008) describes summer precipitation that can accelerate melting and increase flow rate in a single occurrence. Unfortunately, we don't have data on state of precipitation that fall on the study area. However, as maximum precipitation occurs in summer, it probably has certain influence on the melting process.

The onset and end of the ablation season depends on the latitude, but also on basin aspect and climatic conditions (air temperature, precipitation) of a study site. While at the study area and elsewhere in the northern Tien Shan (Han et al. 2013) or the Swiss Alps (Swift et al. 2005) the temperatures drop below 0 °C already during September, in the Himalayas at 31° N the ablation season usually does not end before the turn of October and November (Singh et al. 2006). The complex Adygine is north-oriented and a significant reduction in solar energy input in September causes rather rapid decline of lake water level.

Well-pronounced daily water level fluctuation was noticed in July and August at the study site. By this time most snow is already melted and the main source of meltwater is glacier ice. The proportion of meltwater from snow, firn and glacier ice in discharge is described by Jansson, Hock, Schneider (2003), who has observed changes in daily regime as well. A study by Hock (1998) also confirms that a more distinct daily regime develops after melting of most snow and the flow rate increases after uncovering of larger area of the glacier. The development of discharge daily amplitude is similarly described by others as well (Jobard, Dzikowski 2006; Singh et al. 2006; Han et al. 2013). As the melting season at Adygine progresses, there are greater differences between the daily minimum and maximum discharge at lake's outflow (Fig. 13). Very similar evolution of daily discharge amplitude (Fig. 14) was monitored in the Alps on a stream draining Gornergletscher (Elliston 1973).

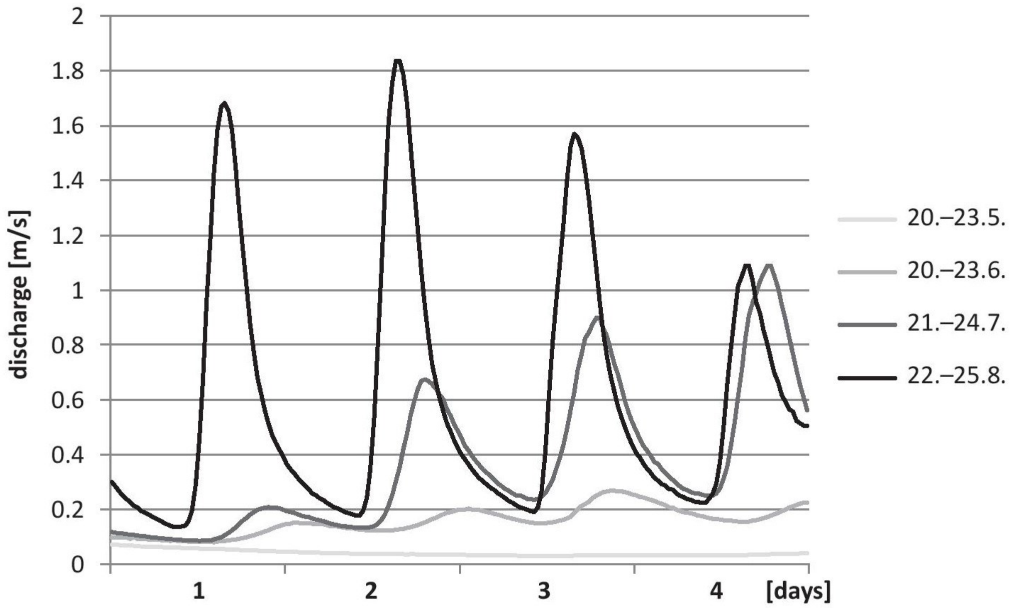

Fig. 13 – Daily discharge amplitude evolution at lake Adygine outflow

Han et al. (2013) focus on the differences between discharges measured during the day and night. It is affected by cloud cover during the day – "a clear-weather day" results in larger differences than an overcast day.

Runoff delay is another feature of glacial hydrological regime observed at Lake Adygine. Richards et al. (1996) also examined the development of runoff delay in the ablation season finding similar tendency to time lag decline as in this study. According to Jobard and Dzikowski (2006) snow cover significantly slows down the runoff, however, when it's melted, there is a rapid decrease in the lag time mainly due to gradually developing effective network of drainage channels (Hubbard, Glasser 2005; Singh et al. 2006). That is why the diurnal maximum discharge value occurs earlier in peak summer than at the beginning of the ablation season (Fig. 13, Fig. 14). Data presented by Jobard and Dzikowski (2006) contain similar values of runoff delay to those noticed at Lake Adygine: in early June values of 8–10 hours, continued by time lag decline till 2–4 hours in mid-July and then time lag stabilization for the rest of the season. Han et al. (2013) also mention in this context the influence of "consecutive clear-weather days" which may further reduce the delay. In the middle of an ablation season, radiation plays a major role – its maximum occurs even several hours before the maximum air temperature is reached and therefore the runoff delay gets shorter.

7. Conclusions

The hydrological regime of Adygine Lake shows features typical of the regime of glacial lakes. Changes in the water level have annual periodicity and a daily regime develops during the ablation season. Water level fluctuations during the year can be divided into three phases: decline, filling and fluctuations. The longest-lasting phase is the first mentioned one, when the water level usually declines from mid-September with a slowing tendency until April/May. The filling phase is highly variable in terms of timing and duration. The shortest recorded time was 12 days (2011), the longest up to 44 days (2010). It is followed by a phase of fluctuations during which a daily regime develops. The fluctuation of the lake level during the day is affected by air temperature and solar radiation intensity and as the ablation season progresses, the fluctuation increases due to more intensive melting on the glacier surface. The maximum usually occurs around mid-August. This is associated with shortening of the runoff delay during the summer due to developing drainage system of the glacier.

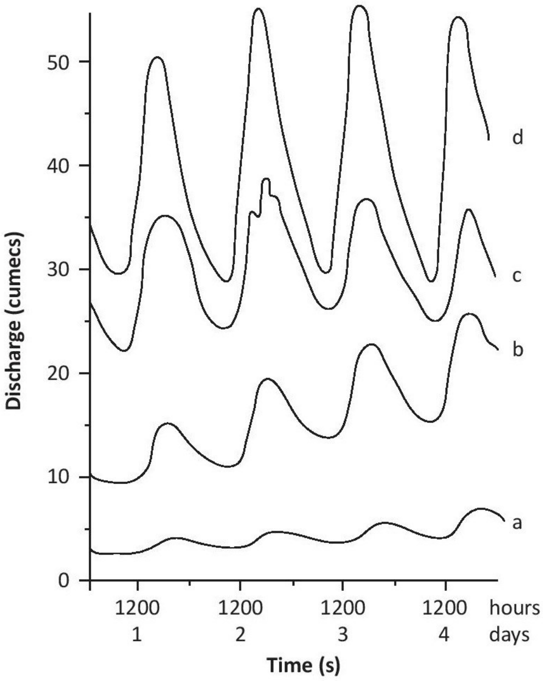

Fig. 14 – Course of daily discharge amplitude at a stream draining Gornergletscher glacier, Switzerland. Monitored periods in 1959: a) 17th–20th May b) 14th–17th June c) 23rd–26th June d) 19th–22nd July. Source: Elliston 1973.

Adygine Lake has a relatively regular cycle of fluctuations. No major changes in the hydrological regime were recorded in the monitored period. Its changes depend on meteorological conditions in individual years.

Subsurface drainage channels divert water from the lake through the dam and thereby supplement the surface outflow. Their capacity varies between years. However, not even here a tendency towards increase or decrease in the drainage capacity was observed. The slowing decline of the lake water level during winter indicates the approximate distribution of the drainage channels. However, their precise distribution and reasons for changes in their capacity should be subject to further research. Subsurface outflow may lower the dam stability and finding out more information about the functioning and development of these drainage channels could help better identify the risk of an outburst of Adygine Lake.

References:

ADYSHEV, M.M. ed. (1987): Atlas Kirgizskoj CCP. GUGK SSSR, Moscow, 157 pp.

AGETA, Y., IWATA, S., YABUKI, H., NAITO, N., SAKAI, A., NARAMA, C., KARMA, T. (2000): Expansion of glacier lakes in recent decades in the Bhutan Himalayas. IAHS Publication, pp. 165–176.

AIZEN, V.B., AIZEN, E.M. (1994): Features of regime and mass exchange of some glaciers on central Asia periphery. Bulletin of Glacier Research, 12, pp. 9–24.
AIZEN, V.B., AIZEN, E.M., MELACK, J. (1995): Climate, snow cover, glaciers, and runoff in the Tien-Shan, Central Asia. Water Resources Bulletin, 31, No. 6, pp. 1113–1129.
AIZEN, V.B., AIZEN, E.M., MELACK, J.M. (1996): Precipitation, melt and runoff in the northern Tien Shan. Journal of Hydrology, 186, pp. 229–251.
AIZEN, V.B., AIZEN, E.M., MELACK, J.M., DOZIER, J. (1997): Climate and hydrologic changes in the Tien Shan, central Asia. Journal of Climate, 10, pp. 1393–1404.
AIZEN, V.B., KUZMICHENOK, V.A., SURAZAKOV, A.B., AIZEN, E.M. (2006): Glacier changes in the central and northern Tien Shan during the last 140 years based on surface and remote-sensing data. Annals of Glaciology, 43, pp. 202–213.
BATES, B.C., KUNDZEWICZ, Z.W., WU, S., PALUTIKOF, J.P., eds. (2008): Climate Change and Water. Technical Paper of the Intergovernmental Panel on Climate Change, IPCC Secretariat, Geneva, 210 pp.
BEITLEROVÁ, H. (2010): Ledovcový hydrologický režim, případová studie jezera Adygine (Ťan Šan, Kyrgyzstán). Bc. Thesis. Univerzita Karlova v Praze, Přírodovědecká fakulta, katedra fyzické geografie a geoekologie, Praha, 75 pp.
BENISTON, M., DIAZ, H.F., BRADLEY, R.S. (1997): Climatic change at high elevation sites: an overview. Climatic Change, 36, No. 3–4, pp. 233–251.
BENN, D.I., EVANS, D.J.A. (2010): Glaciers and Glaciation. 2nd ed., Hodder Education, London, 816 pp.
CLAGUE, J.J., EVANS, S.G. (2000): A review of catastrophic drainage of moraine-dammed lakes in British Columbia. Quaternary Science Reviews,19, No. 17, pp. 1763–1783.
COLLINS, D.N. (1998): Rainfall-inducted high-magnitude runoff events in highly-glacialized Alpine basins. In: Hydrology, Water Resources and Ecology in Headwaters. Proceedings of the HeadWater'98 Conference, Merano, Italy, 248, pp. 69–78.
ČERNÝ, M., ŠOBR, M., ENGEL, Z., KOCUM, J., HRDINKA, T., JANSKÝ, B., HŮLKA, L., ŽÁČEK, M., JEROCHIN, S. A., ERMANGAEV, B., ZAGINAEV, V.V., KUZNECOV, B.A., ZAJAGIN, D. (2010): Analýza rizik a omezení důsledků protržení hrází vysokohorských jezer. Závěrečná zpráva projektu rozvojové spolupráce mezi Českou a Kyrgyzskou republikou RP9/2007, GEOMIN družstvo, Jihlava, 2 parts, 20 suppl., 242 pp.
ČESÁK, J., ŠOBR, M. (2005): Metody batymetrického mapování českých jezer. Geografie, 109, No. 3, pp. 141–151.
DVOŘÁK, V., HLADNÝ, J., KAŠPÁREK, L. (1997): Climate change hydrology and water resources impact and adaptation for selected river basins in the Czech Republic. Climatic Change, 36, No. 1–2, pp. 93–106.
DYURGEROV, M. (2003): Mountain and subpolar glaciers show an increase in sensitivity to climate warming and intensification of the water cycle. Journal of Hydrology, 282, pp. 164–176.
ELLISTON, G.R. (1973): Water movement through the Gornergletscher. IASH Publ, 95, pp. 79–84.
FOUNTAIN, A.G. (1985): The effect of glaciers on streamflow variations. Water Resources Research, 21, No. 4, pp. 579–586.
FUJITA, K. (2008): Influence of precipitation seasonality on glacier mass balance and its sensitivity to climate change. Annals of Glaciology, 48, pp. 88–92.
GAO, X., YE, B., ZHANG, S., QIAO, C., ZHANG, X. (2010): Glacier runoff variation and its influence on river runoff during 1961–2006 in the Tarim River basin, China. Science China Earth Sciences, 53, No. 6, pp. 880–891.
HAGG, W., BRAUN, L.N., KUHN, M., NESGAARD, T.I. (2007): Modelling of hydrological response to climate change in glacierized Central Asian catchments. Journal of Hydrology, 332, pp. 40–53.
HAN, T., GAO, M., YE, B., JIAO, K. (2013): Characteristics of melt water discharge in the Glacier No. 1 basin, headwater of Urumqi River. Journal of Hydrology, 489, pp. 180–188.
HOCK, R. (1998): Modelling of glacier melt and discharge. PhD dissertation, Department of Geography. ETH, Zürich, Switzerland. Zürcher Geographische Schriften, 70, 126 pp.

HUBBARD, B., GLASSER, N. (2005): Field Techniques in Glaciology and Glacial Geomorphology. John Wiley and Sons Ltd., Chichester, England, 400 pp.
JANSKÝ, B. ŠOBR, M., YEROKHIN, S. (2006): Typology of high mountain lakes of Kyrgyzstan with regard to the risk of their rupture. Limnological Review, 6, pp. 135–140.
JANSKÝ, B., ŠOBR, M., ENGEL, Z. (2010): Outburst flood hazard: Case studies from the Tien-Shan Mountains, Kyrgyzstan. Limnologica, 40, pp. 358–364.
JANSKÝ, B., ENGEL, Z., KOCUM, J., ŠEFRNA, L., ČESÁK, J. (2011): The Amazon River headstream area in the Cordillera Chila, Peru: hydrographical, hydrological and glaciological conditions. Hydrological Sciences Journal, 56, No. 1, pp. 138–151.
JANSSON, P., HOCK, R., SCHNEIDER, T. (2003): The concept of glacier storage: a review. Journal of Hydrology, 282, pp. 116–129.
JOBARD, S., DZIKOWSKI, M. (2006): Evolution of glacial flow and drainage during the ablation season. Journal of Hydrology, 330, pp. 663–671.
KATTELMANN, R. (2003): Glacial lake outburst floods in the Nepal Himalaya: a manageable hazard? Natural Hazards, 28, No. 1, pp. 145–154.
KOPPES, M., GILLESPIE, A.R., BURKE, R.M., THOMPSON, S.C., STONE, J. (2008): Late Quaternary glaciation in the Kyrgyz Tien Shan. Quaternary Science Reviews, 27, pp. 846–866.
LI, Z. Q., SHEN, Y.P., WANG, F.T., LI, H.L., DONG, Z.W., WANG, W.B., WANG, L. (2007). Response of glacier melting to climate change-take Urumqi Glacier No. 1 as an example. Journal of Glaciology and Geocryology, 29, pp. 333–341.
MIDDELKOOP, H., DAAMEN, K., GELLENS, D., GRABS, W., KWADIJK, J.C.J., LANG, H., PARMET, B.W.A.H., SCHÄDLER, B., SCHULLA, J., WILKE, K. (2001): Impact of climate change on hydrological regimes and water resources management in the Rhine basin. Climatic change, 49, No. 1–2, pp. 105–128.
NIJSSEN, B., O›DONNELL, G.M., HAMLET, A.F., LETTENMAIER, D.P. (2001): Hydrologic sensitivity of global rivers to climate change. Climatic change, 50, No. 1–2, pp. 143–175.
NYE, J.F. (1973): Water at the bed of a glacier. Symposium on the Hydrology of Glaciers, Cambridge, September 1969, pp. 189–194 (Publication No. 95 de l'Association Internationale d'Hydrologie Scientifique).
RICHARDS, K.S., SHARP, M., ARNOLD, N., GURNELL, A.M., CLARK, M., TRANTER, M., NIENOW, P., BROWN, G., WILLIS, I., LAWSON, W. (1996): An integrated approach to modelling hydrology and water quality in glacierized catchments. Hydrological Processes, 10, pp. 475–660.
RÖTHLISBERGER, H. (1972): Water pressure in intra- and subglacial channels. Journal of Glaciology, 11, No. 62, pp. 177–203.
RÖTHLISBERGER, H., LANG, H. (1987): Glacial Hydrology. In: Gurnell, A.M., Clark, M.J. (eds.): Glacio-fluvial Sediment Transfer. Wiley, New York, pp. 207–284.
SINGH, P., HARITASHYA, U. K., KUMAR, N., SINGH, Y. (2006): Hydrological characteristics of the Gangotri Glacier, central Himalayas, India. Journal of Hydrology, 327, pp. 55–67.
SINGH, P., SINGH, V.P. (2001): Snow and Glacier Hydrology. Kluwer Academic Publishers, Dordrecht, The Netherlands, 742 pp.
SWIFT, D.A., NIENOW, P.W., HOEY, T.B., MAIR, D.W.F. (2005): Seasonal evolution of runoff from Haut Glacier d'Arolla, Switzerland and implications for glacial geomorphic processes. Journal of Hydrology, 309, pp. 133–148.
UNGER-SHAYESTEH, K., VOROGUSHYN, S., FARINOTTI, D., GAFUROV, A., DUETHMANN, D., MANDYCHEV, A., MERZ, B. (2013): What do we know about past changes in the water cycle of Central Asian headwaters? A review. Global and Planetary Change, 110, pp. 4–25.
ZHOU, Z., JING, Z., ZHAO, S., et al. (2010): Response of Glacier moving speed to climate change – take Urümqi Glacier No. 1 as an example. Journal of Acta Geoscientica Sinica, 31, No. 2, pp. 2237–2244.

Shrnutí

HYDROLOGICKÝ REŽIM JEZERA ADYGINE, TIEN SHAN, KYRGYZSTÁN

Tání ledovců ve vysokohorských oblastech a s ním související vznik a vývoj jezer je předmětem zájmu mnoha výzkumných týmů působících po celém světě. Ledovcová jezera mohou být velmi nestabilní a ohrožovat tak údolí pod nimi. Zkoumaná lokalita se nachází v Kyrgyzském hřbetu pohoří Tien Shan, asi 40 km jižně od hlavního města Kyrgyzstánu, Biškek. Článek se zabývá rozborem hydrologického režimu největšího jezera lokality, jezera Adygine Horní.

Díky získaným datům z hladinoměru, meteorologických stanic a měření průtoků na odtoku z jezera bylo vyhodnoceno kolísání hladiny jezera, vývoj průtoku v ablační sezóně a objem přiteklé a odteklé vody. Roční cyklus kolísání byl rozdělen do tří fází – pokles hladiny, plnění a ustálení. Během nejdelší fáze poklesu hladiny byl odtok uskutečňován pouze podzemními odtokovými kanály. Díky podrobné batymetrické mapě jezera tak byla vypočtena změna objemu jezera v této fázi, a tak i přibližná kapacita odtokových kanálů. Rychlost poklesu se v průběhu této fáze snižovala, což naznačuje rozmístění těchto kanálů zejména v horní části jezerní pánve. Fáze plnění představuje poměrně variabilní část celého cyklu. V závislosti na intenzitě slunečního záření a na vzestupu teploty vzduchu se začalo jezero plnit tavnou vodou ze sněhové pokrývky nejdříve 17. 4. (v roce 2012), nejpozději pak 3. 5. (v roce 2009). Nejdéle trvala tato fáze 44 dní (rok 2010, 26. 4.–9. 6.), nejkratší dobu pak v roce 2011, a to pouhých 12 dní (23. 4.–5. 5.). Následuje fáze ustálení, kdy se úroveň hladiny jezera pohybuje kolem nadmořské výšky 3 643 m a je obnoven povrchový odtok. V rámci této fáze byly sledovány dva jevy, a to denní rozkolísanost hladiny a zpoždění odtoku.

Rozkolísanost hladiny během dne je zpočátku nevýrazná, příčinou je tlumivý efekt sněhu, který tavnou vodu částečně zadržuje. Po roztátí sněhové pokrývky je jezero zásobeno vodou z tajícího ledovce a kolísání hladiny se zvětšuje. Největší rozkolísanost byla zaznamenána zpravidla ve druhé polovině srpna, kdy bylo tání díky vysokým teplotám vzduchu a radiaci nejintenzivnější. S příchodem září se však teploty poměrně rychle snižují a nezřídka se dostávají i pod 0 °C, proto je i denní kolísání hladiny značně shlazeno. Na vývoj zpoždění odtoku má značný vliv opět sněhová pokrývka, která zpočátku odtok značně zpomaluje. Po jejím roztátí se však na povrchu ledovce, uvnitř i pod ním začne vyvíjet drenážní systém odvádějící tavnou vodu z ledovce. Zpoždění odtoku se tak nadále snižuje, nejnižších hodnot pak dosahuje v srpnu.

Průtoky na odtoku z jezera byly vypočteny pomocí konsumpční křivky z údajů o úrovni hladiny. Z nich byl dále vypočten objem vody odteklé povrchovou cestou. Nejvyšší průtoky i celkový objem odteklé vody za ablační sezónu byl zaznamenán v roce 2008, následující rok byl naopak málo vodný. Tomu odpovídaly i vypočtené průměrné měsíční průtoky. Přítoků do jezera je více a jejich změření nebylo možné, proto bylo množství vody odhadnuto ze změny úrovně hladiny, povrchového a podzemního odtoku z jezera. Nejvíce vody přitéká do jezera v červenci a srpnu, celkem za tyto dva měsíce může přitéct i přes 2 mil. m^3 vody (rok 2010, nejspíše i 2008).

Průběh kolísání hladiny se mezi jednotlivými roky ve sledovaném období lišil, nebyly však zaznamenány výrazné změny. Poměrně proměnlivá je kapacita podzemních odtokových kanálů. Příčin může být více, opět však nebyl zjištěn stálý posun ať již k zvětšení či zmenšení kapacity těchto odtokových cest. Jejich rozmístění a proměnlivá kapacita by měly být podrobeny dalšímu zkoumání.

Obr. 1 – Satelitní snímek zájmové oblasti Adygine, rok 2006. 1 – Adygine Dolní, 2 – Adygine Horní, 3–7 – nová jezera u čela ledovce, M1, M2 – meteostanice. Zdroj: Google Earth.

Obr. 2 – Průměrné měsíční teploty vzduchu (vlevo), meteostanice Ala Archa, 2 200 m n. m., 2002–2008 a průměrné denní a měsíční teploty vzduchu v lokalitě Adygine (vpravo), 3 653 m n. m., 2011.

Obr. 3 – Průběh teplot naměřených na horní meteostanici, dolní meteostanici a denní průměry teplot naměřené teplotním čidlem (vlevo) a graf korelace hodnot teploty vzduchu (vpravo) z horní (osa x) a dolní (osa y) meteostanice.

Obr. 4 – Příčné profily výtoku z jezera při různých vodních stavech. Doba měření: profil 1 – 3. 8. 15:30–16:00; profil 2 – 3. 8. 17:45–18:05; profil 3 – 4. 8. 8:00–8:15; profil 4 – 4. 8. 17:20–17:40; profil 5 – 4. 8. 20:10–20:35; profil 6 – 5. 8. 10:10–10:30; profil 7 – 5. 8. 13:50–14:10; profil 8 – 5. 8. 15:00–15:15.
Obr. 5 – Roční průběh kolísání hladiny jezera Adygine. Nadmořská výška povrchového odtoku vyznačena plnou černou čarou.
Obr. 6 – Rozkolísanost hladiny v ablační sezóně let 2008–2011.
Obr. 7 – Vývoj zpoždění odtoku v ablační sezóně roku 2008 (podle Beitlerové 2010).
Obr. 8 – Průtok na odtoku z jezera Adygine v průběhu ablační sezóny, 2008–2011.
Obr. 9 – Průměrné měsíční průtoky. Sloupce za srpen 2008 a 2011 jsou ohraničeny čárkovaně kvůli neúplnosti dat.
Obr. 10 – Kumulativní objem povrchového odtoku za roky 2008–2012.
Obr. 11 – Objem přítoku do jezera za jednotlivé měsíce v letech 2008–2012.
Obr. 12 – Denní úhrny srážek za období květen–červenec roku 2012, horní meteostanice.
Obr. 13 – Vývoj denní amplitudy průtoku na odtoku z jezera Adygine.
Obr. 14 – Vývoj denní amplitudy průtoku na toku odvodňující ledovec Gornergletscher, Švýcarsko. Sledovaná období roku 1959: a) 17.–20. 5. b) 14.–17. 6. c) 23.–26. 6. d) 19.–22. 7. Zdroj: Elliston (1973).

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5.4 Glacial meltwater flow through proglacial area

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Hydrological and isotopic characterization of proglacial lakes and their connectivity, Adygine glacier-moraine complex, northern Tien Shan

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Abstract. This study characterized the glacial meltwater running through a proglacial area with focus on proglacial lakes, their hydrological regime, and a morainic landform with small tarns. The studied proglacial lakes showed a typical glacial hydrological regime with a distinct development throughout an ablation season. The annual course of water level fluctuation revealed that there are major differences among the lakes' subsurface drainage channels capacity. Glacial meltwater flows through the lakes and further downstream through the morainic landform rather fast, moving at 0.085 m s^{-1}. However, based on the low dye recovery in the stream (0.03%), only a small portion of water was routed efficiently, large part was delayed in the porous environment. The complexity of the site's drainage system was supported by varying share of hydrological balance components in the tarns situated on the moraine.

Keywords: Glacial lakes; Tracer test; Stable water isotopes; Alpine hydrology; Tien Shan

1 Introduction

The retreat of mountain glaciers (Bolch, 2007; Schiefer et al., 2007; Paul et al., 2007; Kutuzov and Shahgedanova, 2009; Bolch et al., 2012; Rabatel et al., 2013; Paul and Mölg, 2014) has raised concerns, especially in regions where the population greatly depends on glacier meltwater (Vergara et al., 2007; Akhtar et al., 2008; Sorg et al., 2012). In addition to glacier melt rates, the functioning of proglacial systems has a significant influence on the overall hydrological regime of basins (Moorman and Michel, 2000). This is particularly true for the dry regions in central Asia, where glacier meltwater contributes up to 40–70% of the total amount of summer runoff (Aizen et al., 1996). A considerable amount of attention has focused on glaciers and glacier streams (Shiyin et al., 2003; Bliss et al., 2014; Petrakov et al., 2016; Shahgedanova et al., 2018), but there has been less focus on the transition areas between glaciers and glacier streams. In most instances, a stream emerges directly from a glacier terminus; sometimes, this part is constituted by a proglacial lake, but there are many cases where the meltwater is routed underground through morainic sediment before forming a stream.

Groundwater sourced from proglacial landforms, such as rock glaciers or ice-cored moraines, has been shown to be important both in the timing and quantity of alpine watershed discharge (Nienow et al., 1998; Winkler et al., 2016; Jones et al., 2018). Studies conducted in the Andes indicated that these landforms store significant amounts of water, making them important water reservoirs, especially in times of drought (Croce and Milana, 2002; Brenning, 2005). However, they represent a complicated system where layers of varying permeability occur (fine and coarse sediments, boulders, buried ice, etc.); thus, several drainage systems (slow/delayed and fast flow component) may be present (Winkler et al., 2016). Although it is clear that proglacial lakes and landforms play an important role in transmitting and temporarily storing water, their internal hydrological processes and pathways are still not well understood.

Glacial streams are characterized by a glacial regime that has a typical daily and seasonal variation of water discharge (Singh et al., 2005; Swift et al., 2005). Hydrology of proglacial environments has been studied using various methods, including geophysical soundings of moraines (Langston et al., 2011), an analysis of glacier runoff hydrographs (Swift et al., 2005), tracer tests (Buchli et al., 2013), the isotopic composition of water sources (Turner et al., 2010; Penna et al., 2014), or by calculating the energy balance of lake water (Langston et al., 2013).

Tracer tests, specifically the dye tracing technique, has commonly been applied in many fields, for example, in studying groundwater flow and karstic or glacier channel systems (Nienow, 2011). The fluorescent dyes used in glacier environment include rhodamine, uranine (Na-fluorescein), or Optical brighteners. The use of dye-tracing techniques for examining the internal hydrology of glaciers became common in the 1970s (Burkimsher, 1983). However, using this technique in a proglacial area can be problematic; there is a risk of the adsorption of the tracer into sediment and a very long residence time in morainic sediment.

Besides the dye-tracing method, the stable isotopes of water ($\delta^{18}O$ and $\delta^{2}H$) can be used as natural tracers in order to study hydrological processes and reveal changes in the lake water balance (Isokangas et al., 2015; Wu et al., 2015; Cui et al., 2016; Yang et al., 2016; Kang et al., 2017). The isotopic ratio of ^{18}O to ^{16}O and ^{2}H to ^{1}H, in addition to the relationship of $\delta^{18}O$ and $\delta^{2}H$ in a water sample, differs according to altitude, latitude, distance from ocean, and influence of evaporation (Jouzel et al., 2000). A region's hydrological specificity is characterized by a local meteoric water line (LMWL) - for precipitation, and local evaporation line (LEL) - for water undergoing evaporation (i.e., surface water). The position of a sample given by $\delta^{18}O$ and $\delta^{2}H$ relative to the LMWL, LEL, or to other samples can reveal useful information about the origin of water, helping determine to which degree evaporation influences a lake's water balance or whether recharge from rain, melting snow, or ice outweighs the influence of evaporation (Stewart, 1975).

In the current study, we present research on the hydrological regime of proglacial lakes, their connectivity, and the characteristics of subsurface drainage systems. The objectives of the present study are the following: 1) to analyze the daily and annual hydrological regime of the main proglacial lakes fed by glacial meltwater based on lake water fluctuation; 2) to provide information on a groundwater flow system in a moraine complex using a dye tracing technique; and 3) to examine the linkage of tarns to glacial meltwater based on the isotopic composition of water. These findings can contribute to a better understanding of groundwater flow through proglacial landforms and to their more accurate representation in the hydrological modeling of glacier runoff.

2 Materials and Methods

2.1 Study site

The study area lies within the Adygine Valley, northern Tien Shan, Kyrgyzstan. The Adygine glacier-moraine complex (42°30’10’’ N, 74°26’20’’ E) closes this 8-km-long tributary valley, has a northern aspect and reaches an elevation of 3400–4200 m a.s.l. It belongs to the Ala Archa watershed, which drains to the endorheic Chu basin stretching across Kyrgyzstan and Kazakhstan. The upper part of the Adygine Valley contains large amounts of loose debris and is dominated by several generations of moraines that are up to 3.5 km away from the current glacier terminus (3600 m a.s.l.). According to Gorbunov et al. (1996), the site lies at the boundary (3500 m a.s.l.) between zones of continuous and discontinuous permafrost, and the occurrence of perennially frozen ground at the site is confirmed by the global permafrost model (Gruber et al., 2012).

The site includes a relatively small glacier (2.8 km²), a three-level cascade of proglacial lakes that have evolved because of glacier retreat over the past 60 years (Falatkova et al., 2019), and a moraine complex. The latter (3450 m a.s.l.) consists of several parts that have varying origin (Shatravin, 2000). Distinctive geomorphologic features like topographic highs and lows, exposed shallow buried ice, indicate that this moraine complex contains ice blocks (glacier remnants). The western part shows surface features signalling viscous creep of ice-rich debris. The central landform of the moraine complex can be categorized as a rock glacier with glacial origins (Knight et al., 2019), its steep frontal slope is almost 700 m wide. It developed from a stagnant glacier tongue buried under a thick debris cover, which has lost contact with the glacier itself. Debris at the surface consists of predominantly large angular boulders of granite and granodiorite, but there are several older parts with coarse sand and gravel, pointing at previous changes in the hydrological conditions. Currently, there are no surface channels, so all snow and glacier meltwater pass underground through the moraine complex to reach the stream. Estimated major routes of meltwater below the surface (Fig. 1) are based on results of geophysical sounding near lake dams and on visual inspection (Falatkova et al., 2019).

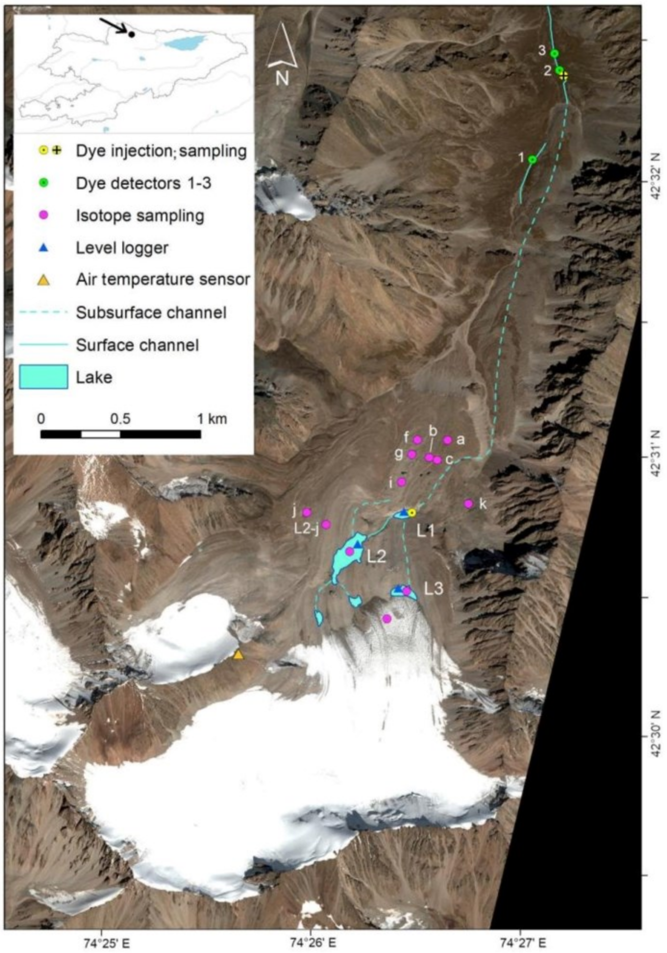

Figure 1. Map of Adygine glacier-moraine complex with location of installed measuring devices, dye tracer experiment, and isotope sampling sites. The main proglacial lakes are marked as L1 (Lake 1), L2 (Lake 2), L3 (Lake 3), the tarns are marked with letters a-k. Study site marked with an arrow on map of Kyrgyzstan (upper left corner).

The main lakes of the site (Lakes 1–3; Table 1) are of varying ages. The most recently formed lakes are close to the glacier terminus and sensitively react to changes in glacier melt rates, redistributing the water further downstream, either by surface or subsurface channels. The most prominent of these lakes is Lake 3, which is supplied with meltwater directly from the glacier by englacial and/or subglacial channels. The eastern part of the glacier, which is most likely forming the drainage basin of Lake 3, is shaded by a steep ridge and receives direct radiation only later in the afternoon. The middle level is represented by the largest lake of the site, Lake 2, which has a fairly stable (consistent over the monitored period) annual hydrological regime (Falatkova et al., 2014). The central and western parts of the glacier form Lake 2's watershed, so most glacier and snow meltwater is transferred through this lake. It supplies water to the lowest part of the cascade (Fig. 2), Lake 1. Not being permanent, Lake 1 is an intramorainic depression that is filled with water only during an ablation season when the rate of incoming meltwater is higher than the capacity of its subsurface drainage. There are also several tarns (named with letters a–k) situated on a moraine complex at the similar level as Lake 1. Except for one (tarn a), all of these tarns show a certain level of turbidity (similarly to the main lakes) due to suspended sediment ('glacial flour'). The term "tarn" is used in this paper as an equivalent to a small pond on a proglacial moraine that has no surface inflow or outflow. The origin of the tarn depressions can be linked to the melting of buried ice blocks. These tarns have an unknown connection to the groundwater system draining the site. The meltwater is transported from the glacier through to the main lakes (Lakes 1–3), and then, it is routed underground through the moraine complex, finally emerging as a stream at an elevation of 2900 m a.s.l., that is, about 3.6 km from the glacier terminus (average slope of 11.3°).

Table 1. Morphological parameters and connectivity characterization of the three main lakes of the site. Data based on field measurement in July 2017.

Lake No.	Elevation [m a.s.l.]	Max area [m^2]	Max depth [m]	Max volume [m^3]	Prevailing inflow type	Prevailing drainage type
1	3450	3400	4.4	13,300	surface	subsurface
2	3540	31,900	21.3	194,000	subsurface	surface
3	3580	16,020	14	106,000	subsurface	subsurface

Lake connection	Distance [m]	Elevation difference [m]	Slope	Hydrological connection
Lake 2 → Lake 1	280	90	18.8°	surface
Lake 3 → Lake 1	490	130	15.4°	subsurface
Lake3 → Lake 2	270	40	8.5°	none

The climatic conditions in the area are continental and are characterized by relatively low precipitation and high annual and daily air temperature amplitude. At the Adygine glacier, the mean air temperature for January drops to -22 °C and reaches 5 °C in July; mean annual precipitation is estimated to be 700–800 mm. On the northern slope of Kyrgyz range, winters are characterized by low precipitation totals, whereas the maximum precipitation amounts are typically reached in late spring and the beginning of summer. Although the region is characterized by an overall low humidity, the cold high-altitude areas have medium to high relative air humidity. The ablation season usually spans from June to mid-September.

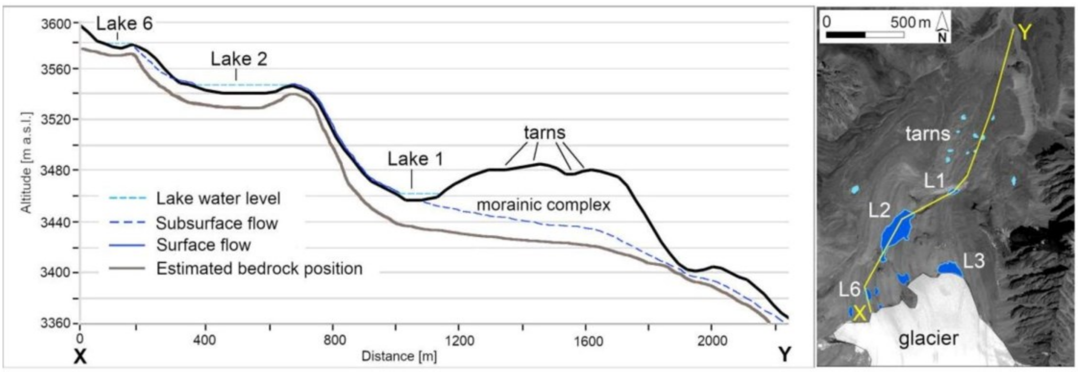

Figure 2. A profile (XY) from the glacier terminus through the 3-level cascade of proglacial lakes at Adygine glacier-moraine complex. The profile supplemented with hydrological connections between lakes, drainage of Lake 1, and estimated position of the bedrock.

2.2 Lake water level measurements

To analyze the daily and annual hydrological regime of the lakes supplied by glacier meltwater, level loggers (water pressure sensors manufactured by Solinst, Canada) were installed in the three main lakes of the site (Figure 1). The data from Lake 2 are available for the period of 2007–2017, the water level fluctuation data of Lake 1 and Lake 3 are available for 2012–2015. The measured pressure is compensated with air pressure data from a barologger (Solinst, Canada) installed at Lake 2. The sensors recorded the water level in 0.5 h intervals with an accuracy ± 0.006 m. The lake water level fluctuation data were processed with R software and supplemented with air temperature data from a sensor with radiation shield installed on site (3900 m a.s.l., see Figure 1) at 2 m above ground.

2.3 Tracer test

To examine the passage of meltwater into a stream, a dye-tracing method was used. Sodium fluorescein (uranine) was selected as it is detectable even in very small concentrations (up to 10^{-12} g ml^{-1}), it is nontoxic, and it is readily soluble in water. Uranine also sorbs less to organic matter and clay minerals than most of other dyes (Käss, 1998). The limit of detection when using a fluorimeter is 10^{-12} g ml^{-1}.

Conducting the tracing test, the water passage between Lake 1 (the furthermost glacier meltwater on surface) and a stream resurgence about 3700 m away from the glacier at an altitude of 2900 m a.s.l. was examined. The total straight-line distance between these two spots is 3100 m. A sampling strategy was set up by estimating the minimum and maximum dye tracer travel time (4.4–139 h), which was based on possible minimum and maximum flow velocities (Krainer and Mostler, 2002; Pauritsch et al., 2017) and travel distances (straight-line/sinuous path).

To detect uranine in water, three passive samplers were installed with granulated active charcoal at different stream reaches, and water samples were collected for subsequent laboratory analysis using a fluorimeter. Because uranine is photosensitive, the injection time was set for evening. Before the tracer test, blank samples were collected to establish background concentrations.

On June 22, 2017 at 18:30 local time (GMT +5 h), 3000 g of powder uranine mixed with water was injected into Lake 1 near the expected position of the drainage channels (i.e., at the eastern part of the lake basin). By that time, the lake's water level had reached its daily maximum, and although hardly perceptible, it was possible to observe water flowing toward the presumed drainage spot. At 22:00, the stream water was examined using a rectangular glass container put under a blue LED diode light with a wavelength of 470 nm. This visual testing was carried out at one-hour intervals until the dye was detected and the first sample was collected. The samples were collected continually at one-hour and later at two-hour intervals over 32 hours using 10 ml dark glass bottles. The bottles were stored in a black container to prevent the uranine in the samples from decomposing because of exposure to light. The passive samplers were picked up after the experiment to confirm uranine arrival. The charcoal from the passive samplers was rinsed and leached in a 1:1 mixture of 15% KOH and 96% medicinal grade ethanol solution. The resulting solution was analyzed in a fluorimeter.The samples were analyzed using a fluorimeter (LS55, Perkin Elmer, United States) with an excitation wavelength of 492 nm; the emission peak was observed at the intensity of 512 nm for uranine (Käss, 1998) and its height was compared with standard concentrations of 10^{-10} g ml^{-1}, 10^{-11} g ml^{-1}, 10^{-12} g ml^{-1}. For creating the standards, water with similar mineralization as the stream water (low mineralized groundwater) was used.

2.4 O and H stable isotopes

The tarns' connectivity to glacial meltwater (or hydrological isolation) has not been studied yet. Based on water color and turbidity it was estimated that these tarns have varying portion of glacial meltwater recharge, thus a water isotope analysis was used to resolve the tarns' prevailing water source. The lake and tarn water is considered to be a mixture of glacier and snow melt (generally more depleted in ^{2}H and ^{18}O; Clark and Fritz, 1997) and rain (generally more enriched in ^{2}H and ^{18}O); there also might be a certain influence of evaporation. The term 'glacial meltwater' refers to combined portions of melting (older) ice and recent snow, we do not differentiate between the two sources. The main lakes' (Lakes 1–3) balance is obviously governed by glacial meltwater, however, the tarns have unknown portion of meltwater recharge. In order to estimate the portion of glacial meltwater, a precipitation sample of rain (most enriched in ^{2}H and ^{18}O; Bishkek 1) and snow (most depleted in ^{2}H and ^{18}O; Bishkek 2) from the city of Bishkek (~40 km away) were selected to serve as limit values (Table 2). As having a strict threshold (at 50%) is inappropriate given that the range is based on two single values, three categories were defined; two with a prevailing source and one transitional category (50% ± ~10%).

On the 25 July 2017, ten water samples were collected from eight tarns, Lake 2, Lake 3, from a spring situated west of Lake 2 (L2-j), and a sample from the snow covering the lower part of the glacier. The lake/tarn water samples were collected from about 40 cm below the water's surface to reduce the effect of evaporation on isotopic composition. Freely available data from the Global network of isotopes in precipitation (GNIP) database was used to compare our results with other climatic and hydrologic data. A δ^{18}O and δ^{2}H analyses were done by the Isotope Laboratory of the Czech Academy of Sciences, Ceske Budejovice, Czechia (Table 2). The measured values were corrected by linear regression between the Vienna Standard Mean Ocean Water (VSMOW) and Greenland Ice Sheet Precipitation standards. Here, isotopic composition is expressed as δ-values, representing deviations in per mil (‰) from the VSMOW according to $\delta_{sample} = [(R_{sample}/R_{vsmow}) - 1] * 10^{3}$, where R is a ratio ^{18}O/^{16}O or ^{2}H/^{1}H. The standard deviation of the values is 1.1‰ for deuterium and 0.26‰ for ^{18}O.

Table 2. Stable water isotopes in samples from the study site Adygine and the share of meltwater component based on precipitation samples from Bishkek (rain: Bishkek 1; snow: Bishkek 2). Data source: IAEA/WMO, 2018.

Site	**δ^2H (σ ±1.1‰)**	**Meltwater component (%)**	**δ^{18}O (σ ±0.26‰)**	**Meltwater component (%)**
a	-36.075	29.88	-6.253	24.56
i	-50.431	41.80	-8.676	39.25
g	-54.949	45.56	-9.05	41.52
f	-60.948	50.54	-9.725	45.61
k	-68.768	57.03	-10.943	52.99
b	-75.368	62.51	-11.572	56.80
L3	-76.883	63.77	-12.359	61.57
c	-78.677	65.26	-11.882	58.68
j	-80.779	67.01	-12.334	61.42
L2	-81.444	67.56	-12.755	63.97
L2-j	-81.743	67.81	-12.732	63.83
snow	-99.968	-	-14.921	-
Bishkek 1	-0.1	0	-2.2	0
Bishkek 2	-120.5	100	-18.7	100

3 Results and discussion

3.1 Hydrological regime of the main lakes

3.1.1 Daily water level fluctuations and causal factors

Although the lakes have the same source for their water supply (glacier Adygine) and react very similarly to changes in the glacier melt rate, there is a difference in timing because these lakes are at different stages of water passage - their distance from the glacier are 600 m, 240 m, 0 m for Lakes 1–3, respectively (Table 1). Figure 3 shows the daily fluctuation of the lakes and the course of the air temperature at the site (3900 m a.s.l.). When the air temperature decreases and the daily peak is partially suppressed, the water level reacts accordingly (for example, on the 31 August, Fig. 3b). In the case of Lakes 2 and 3, the peak becomes less pronounced, whereas in the case of Lake 1, the water level starts to drop more dramatically.

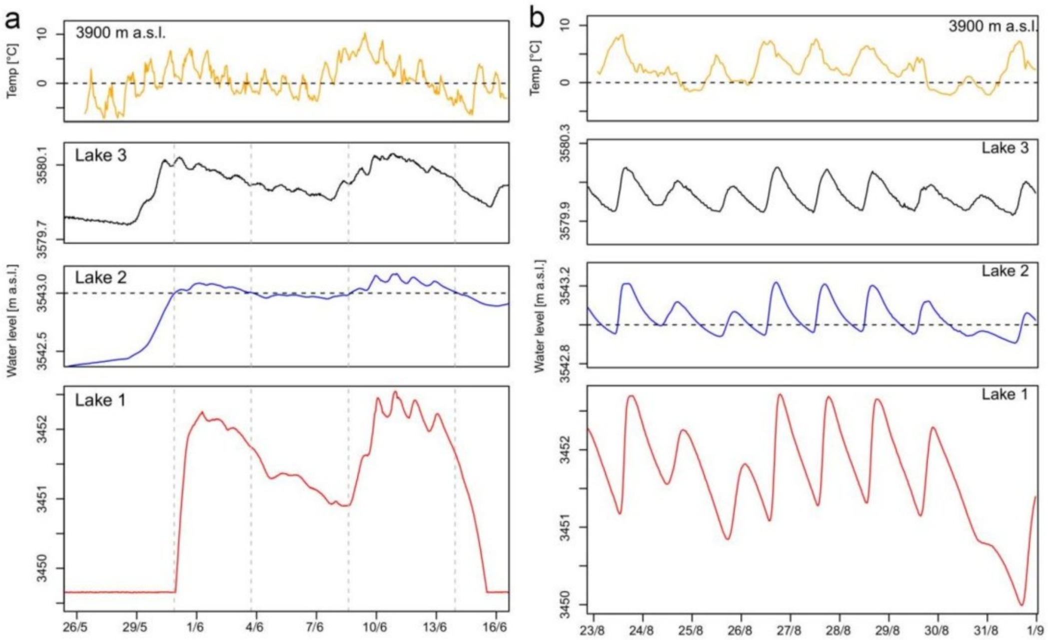

Figure 3. Water level fluctuation of Lake 1, Lake 2, Lake 3, and air temperature in ablation season of 2013. (a) Beginning of the ablation season. A horizontal dashed line corresponds to the overflow level of Lake 2, which enables filling of Lake 1. Vertical dashed lines and arrows signify time when this threshold is reached. (b) Daily water level fluctuation of the main lakes shows lakes' response to daily course of air temperature at the site (late August 2013).

The beginning of an ablation season (Fig. 3a) shows the influence that Lake 2's water supply has on Lake 1. The water level of the latter starts to increase only after the level of Lake 2 reaches a certain threshold, that is, high capacity overflow, which occurs when the surface inflow to the lake is sufficiently large. Based on our rating curve, which was established in 2012 (Falatkova et al., 2014), this threshold is around 0.3 $m^3 s^{-1}$ and corresponds to the lake water level at ca 3543.05 m a.s.l. The subsurface water supply from Lake 3 is likely not enough to fill Lake 1 because the former reaches its maximum water level without a response from the latter (Fig. 3a). A similar pattern is repeated over an ablation season when Lake 1 is drained, and its level rises again only after the surface inflow from Lake 2 is sufficiently high.

The daily lake water level changes vary for individual lakes. Lake 3 exhibits the least pronounced fluctuation, 0.15–0.2 m in the peak ablation season, whereas Lake 2 has a daily water level amplitude of 0.25–0.35 m. Taking into account the lake's areal extent, the volume of Lake 2 fluctuates by ~10,000 m^3 per day compared with Lake 3, which fluctuates only by ~3200 m^3 per day. At over 1.5 m, Lake 1 shows the largest daily amplitude of water level fluctuation, changing its volume by ~5100 m^3 per day. Considerable changes in lake level are possible because of the steep slopes of the basin and the high drainage channel capacity. The daily volume change of the stored water in Lake 1 constitutes up to 38% of the lake's volume, but in the case of Lake 2 and Lake 3, it is only 5% and 3%, respectively.

Lake level fluctuation evolves during an ablation season; there is a gradual increase of its daily amplitude from June to August, and then, it drops again during September (Figure 4). This general pattern is caused by a changing share of snow and glacier melting, by rising air temperature, its daily amplitude, and solar radiation pattern. In June, the mean daily water level fluctuation amplitude of Lake 2 is 0.07 m, it rises to 0.11 m in July, and it finally reaches 0.18 m in August (Figure 4a). The amplitude is still high at the beginning of September, but toward the end of the month, it drops to only a few centimeters (the mean amplitude is 0.07 m). The high values recorded at the beginning of June correspond to the filling phase when the lake's water level rises sharply; thus, the minimum value occurs after midnight and the maximum at the end of the same day. A very similar pattern was found at Lake 3 (Figure 4b) although it was based on less data compared with Lake 2. Peak daily fluctuation is reached in August as well; a few extraordinarily high amplitudes were recorded at the end of June and can be linked to the sudden release of snow meltwater from an englacial cavity.

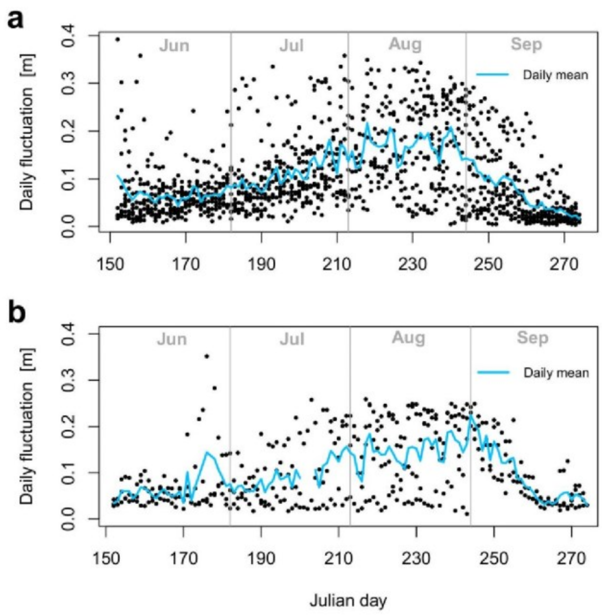

Figure 4. Daily water level fluctuation amplitude during ablation seasons 2007–2017 for Lake 2 (a) and 2012–2015 for Lake 3 (b). The line represents mean daily fluctuation amplitude.

Figure 5 presents the values recorded in July and August between August 2012 and July 2015. As expected, Lake 1 shows a lag behind the Lake 2 when it comes to the timing of the daily lowest and peak water level (2.4 h and 2 h, respectively). Because Lake 3 is the closest to the glacier, a typical glacial regime and the daily peak being reached first here was expected. There is a similar temporal pattern regarding the maximum and minimum water level as at Lakes 1 and 2; however, there are also many cases that do not occur during the expected daytime period. This is obvious especially for the maximum daily water levels—besides the late afternoon cases, there are many that occur between midnight and 8:00. Accordingly, the periods for the minimum daily water level were, besides the usual times, recorded in the late afternoon and evening. These irregularities (~15% of all cases) may be linked to the englacial and subglacial system of cavities and diurnal changes in water pressures (Röthlisberger, 1972; Gordon et al., 1998).

For all three lakes, there were outlier cases where the maximum water level was recorded at midnight (occurring in 9.8%, 8.2%, 6% of days for Lakes 1–3, respectively). These represent days with an overall water level decline and diminished or absent daily fluctuation of water level. The timing of the daily maximum and minimum water level evolves throughout an ablation season, similarly to the daily water level fluctuation amplitude (Fig. 4). Generally, both the daily minimum and maximum water levels tend to occur earlier during the day throughout the month of July. This is caused by increasing efficiency of the supraglacial and englacial drainage channels transporting meltwater to the proglacial area; therefore, the time it takes the meltwater to reach a lake is shortened. However, during the second half of the observed period (in August), the occurrence of the lowest water level slightly shifts to later hours, while the timing of the daily maximum remains the same. Once an efficient mode of meltwater transport has been reached, it does not change much; however, the solar radiation keeps changing. The path of the sun in a day becomes shorter and the peak temperature timing stays the same, but because of a later sunrise, the daily water level minimum occurs later.

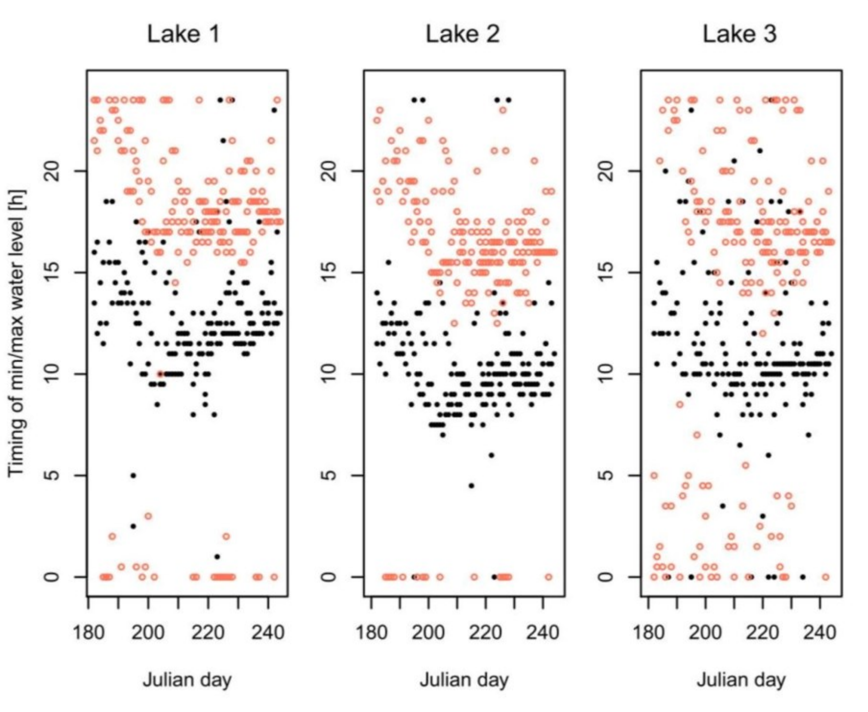

Figure 5. Timing of daily maximum and minimum water level for Lake 1, Lake 2, and Lake 3. The data are from the peak ablation periods (months of July and August) in 2013–2015. The maximum marked with an empty circle, minimum with a black full point.

The water level fluctuation of the main lakes that are fed by glacier meltwater shows the typical characteristics of a glacial regime both on a daily and seasonal scale. However, a slight difference was found in the daily water fluctuation curve of Lake 3. The pressure sensor recorded that the water level gradually rises to reach its daily peak (the rising limb is smooth), whereas the decline in the water level shows a step-like pattern (Figure 6). In days with low lake inflow (i.e., slow water level rise), there is also a step-like pattern for the water level as it rises. This means that the high flow rate of water inflow is gradual, but when the inflow discharge is under a certain threshold, it is delivered to the lake unevenly. A possible explanation for this behavior can be merging waves (or roll waves, described in Fowler, 2011); the flow velocity in the englacial channels increases with the volume of the transported water. The more water there is, the faster it flows, moving the water forward where it merges with another wave. Therefore, there is a section with a high amount of water followed by section with less water, which is a pattern that repeats somewhat periodically. The channel surface would have to be free of obstacles for this phenomenon to develop. Fountain and Walder (1998) describe the fast (arborescent) and slow spatially widespread (nonarborescent) channel network delivering meltwater from the glacier's surface to its proglacial area. The latter is often poorly interconnected and contains cavities. Provided that the englacial conduits route water from the ablation area, the water flow will be pressurized only in times of daily peak discharge; otherwise, it is an open-channel flow (Fountain and Walder, 1998). This supports the explanation of the difference between the falling and rising limb of the daily water level fluctuation curve.

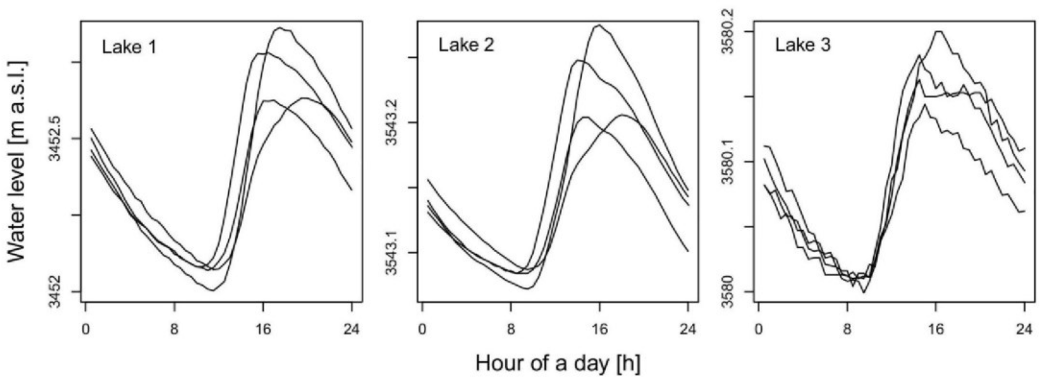

Figure 6. An example of a daily water level fluctuation (1–4 August 2013) at the main proglacial lakes at the Adygine complex.

3.1.2 Lake annual water level fluctuations

The basin of Lake 1 is empty for most of the year and is filled when the water supply is large enough to balance the capacity of the subsurface channel drainage. Although short-term emptying of the lake (taking a few days) was recorded also during an ablation season due to cold spells, the lake is usually drained during first half of September. As the capacity of the subsurface drainage is large, once the surface inflow from Lake 2 is subdued, the lake is drained within 1–2 days. The basin remains empty throughout the cold season and is filled again (within a few tens of hours) in late May or early June when the inflow is restored.

The pattern of annual water level fluctuation of Lake 2 has proven to be rather unchanged over the monitored period (Figure 7a). During ablation season, the water level fluctuates above 3543 m a.s.l. and then drops throughout the cold season until late April/May when the minimum water level is reached. Decreasing by ~3 m (to 3540–3540.3 m a.s.l.), the lake's volume is reduced to ~140,000 m^3, that is, 72% of the maximum volume. The decrease is not linear; first, the level drops at a faster rate (0.03–0.05 m per day), but toward the end of a cold season, it drops very slowly (≤0.01 m per day; Figure A1). This reveals the approximate depth of drainage channels or seepage routes, which seem to be situated in the upper part of the basin (up to 3 m below overflow point). Over the monitored period (2007–2017), the rate of water level decline at the beginning of the cold season decreased from ~0.05 m to less than ~0.03 m (Figure A1). This could have been caused by gradual siltation of the drainage routes. The phase when the lake is being filled (water level rise, Fig. 5a) is short (13–48 days) compared with the period of water level decline (Fig. 7a) which takes 200–225 days.

During an ablation season, the water level of Lake 3 fluctuates at the altitude of 3580 m a.s.l. Starting in September, it drops by ~0.5 m to reach its annual minimum level in early November (Figure 7b). Unlike the water level of Lake 2, the water level of Lake 3 rises during most of the cold season. This gradual increase in lake level (0.2 m in 6 months) lasts until May, when higher air temperatures cause a shift in the lake's hydrological regime to an ablation fluctuation mode. This may indicate that the drainage routes are probably very close to the surface (~1 m) and that the lower part of the basin is formed by impermeable material (ice). Around mid-November, the water flowing near the surface freezes and blocks these routes. Thus, the outflow is stopped, and because of a continuous small basal water supply from the glacier, the lake level rises (Figure 7b).

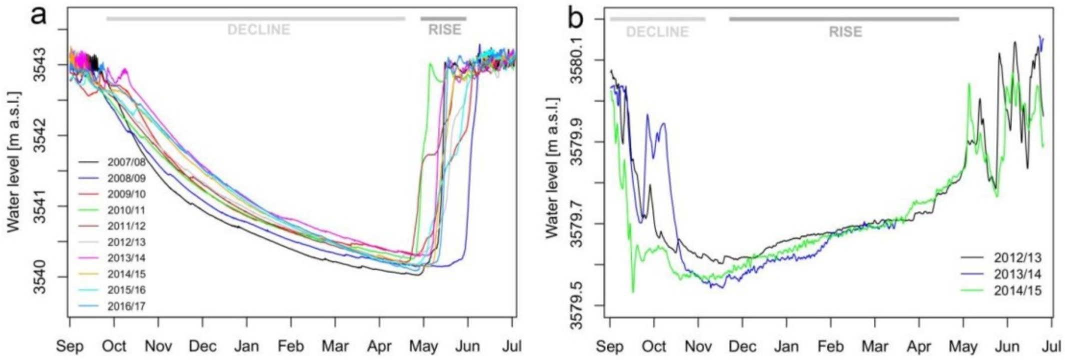

Figure 7. Water level fluctuation of Lake 2 (a) and Lake 3 (b) during the cold part of a year (September-June). Horizontal lines at the top represent mean period of lake water level decline and rise. Lake 1 is not included - as a temporary lake it is empty during the cold part of a year.

The 10-year data series of Lake 2's water level fluctuation allows for a comparison of the monthly water level ranges (Figure 8). The sensor recorded the unusually warm ablation seasons of 2007 and 2008, which were reflected in a distinct glacier tongue retreat and enlargement of some of the proglacial lakes (Falatkova et al., 2019). For the months of June and July, a statistically significant ($p<0.1$) increase of mean water level was recorded, whereas August and September did not show any particular trend. However, the minimum recorded September water levels for the period of 2007–2017 showed a statistically significant ($p<0.1$) rising trend. Water levels well above the surface drainage limit (3542.61 m a.s.l., marked by a solid line in Fig. 8) occurred more often in the last 5 years, which could be because of the expected trend of the ablation season's prolongation (Sorg et al., 2012). Water levels recorded in September have larger variance than other months because lake level often drops below the surface drainage limit toward the end of the month, but high water levels occur as well, because the ablation season may continue till mid-September. Therefore, September is considered a transition period between the ablation season's water level fluctuation and cold season's lake level decline.

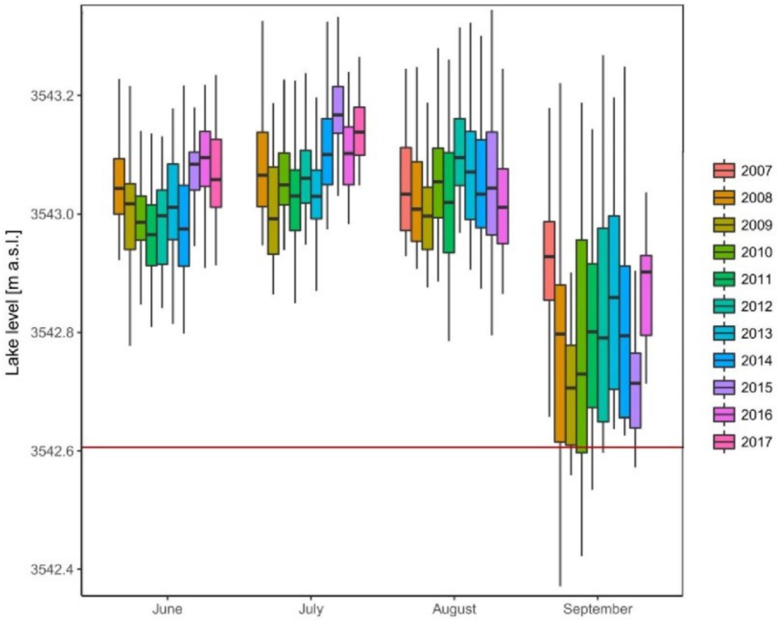

Figure 8. Monthly water level fluctuation ranges of Lake 2 in ablation seasons 2007-2017. The altitude that marks surface drainage channel (3542.61 m a.s.l.) is represented by a horizontal line.

3.2 Water passage through the moraine to the stream

After the uranine injection into the lake, the dye moved slowly toward the outflow area (Figure 9a), and within 10 minutes, all the visibly bright green-colored water infiltrated underground. At the time of injection, the water inflow from Lake 2 was already declining (daily max: 1.5 m^3 s^{-1} at 13:00, 0.9 m^3 s^{-1} at 18:30, 0.7 m^3 s^{-1} at 21:30) yet still rather high, which facilitated the fast water transport and limited dye dispersion. The estimated concentration of uranine in the lake was 10^{-4} g ml^{-1}.

The first observation of uranine in the stream was at 5 a.m., which is 10.5 h after the injection (Table 3). This first collected sample showed the highest uranine concentration: $\sim 5 \cdot 10^{-11}$ g ml^{-1}. Except for the first two samples (5 a.m. and 6 a.m.), the fluorimeter output curve did not show a distinct peak over 512 nm (which equals the uranine emission peak). Although not marked with a peak, the significant irregularity between the rising and falling limb of the output curve indicates the presence of uranine (compared with the output curve of the stream water collected before dye injection). As shown in Figure 9b, a so-called 'tail' is observed for the following 30 hours. These samples were categorized as being at the detection limit (10^{-12} g ml^{-1}), that is, uranine was possibly present, but it could not be proved because the concentration was very low.

A breakthrough curve can provide information about tracer dispersion and the drainage channel network. Our breakthrough curve (Figure 9b) is characterized by a prominent peak and a rather gradual decline of the falling limb. The overall tracer recovery was extremely small at about only 0.03% of injected uranine, which reflects the drainage system's complexity. The maximum water flow velocity through the system (based on straight-line distance between injection and sampling site) is 0.085 m s^{-1}. This relatively high flow velocity indicates that water is directed through morainic material with washed-out routes among debris, and the low recovery indicates retention of water and dye in intergranular pores. A lack of turbidity at stream resurgence confirms that fines are captured in the system because of low velocity and thus a low sediment-carrying capacity of the water. The flow path has considerable dimensions (volume of mobile water >80,000 m^3, mean water-filled cross section >25 m^2). The volume of the relatively fast-flowing portion of water between the injection point and the resurgence (0.03% of total flow) is, however, small (only tens of m^3). Therefore, most of the water is captured by the slow water paths in intergranular pores, where the residence time exceeds tens of hours (probably weeks, or even months).

Proglacial depositional landforms (moraine, rock glacier, sandur, etc.) are known for their capacity to store water (Krainer and Mostler, 2002; Brenning, 2005; Rangecroft et al., 2015) and represent an important water source. There is an undefined volume of water within the unfrozen saturated layer and numerous connecting channels (Langston et al., 2011). This saturated layer's thickness depends largely on permafrost (and taliks) occurrence, ice content, and overall structure of a moraine; values in range of several meters were observed (e.g., 2–5 m in McClymonth et al., 2011 or Lehmann-Horn et al., 2011). This volume is sufficient to store water for long periods of time, thus diluting the tracer below detection limits.

The flow velocities calculated from the measured travel times and straight-line distances of most of the reported tracer tests in a similar environment (i.e., morainic material with buried ice) were between 0.0002 and 0.09 m s^{-1} (Tenthorey, 1992; Krainer and Mostler, 2002; Caballero, 2002; Buchli et al., 2013). This variety indicates that internal drainage systems may be developed at varying levels of efficiency, with perennially frozen ground, buried ice, and bedrock topography playing an important role as impermeable materials. Buchli et al. (2013) highlights the importance of the 0°C isotherm position, which can alter the routing of the water flow and therefore the residence time (using the term "suprapermafrost flow"). Moreover, numerous past studies have differentiated between fast (quick flow response, supraflow) and slow flow (base flow, intraflow) within a moraine.

Table 3. Parameters related to the dye tracing experiment in the Adygine Valley.

Injection site – A	Lake 1 (3450 m a.s.l.)
Sampling site – B	Stream Adygine (2900 m a.s.l.)
Straight-line distance A-B	3200 m
Average slope between A and B	9.9°
Tracer	Sodium Fluorescein (uranine)
Tracer amount	3000 g
Time of injection	18:30 (7/22 2018)
Time of first tracer arrival	5:00 (7/23 2018) (10.5 hours)
Mean residence time	> 11 hours
Water flow velocity (based on straight-line distance)	0.085 m s^{-1}
Mean discharge at A	0.6 m^3
Mean discharge at B	~2 m^3 s^{-1}
Volume of mobile water between A and B	2 m^3 s^{-1}*(11 hours *3600)= >80,000 m^3
Mean water-filled cross-section	80,000 m^3/3200 m= >25 m^2
Dye recovery	0.03 %
pH at A	7.22
pH at B	7.13
Conductivity at A	19.34 µS cm^{-1}
Conductivity at B	36.9 µS cm^{-1}

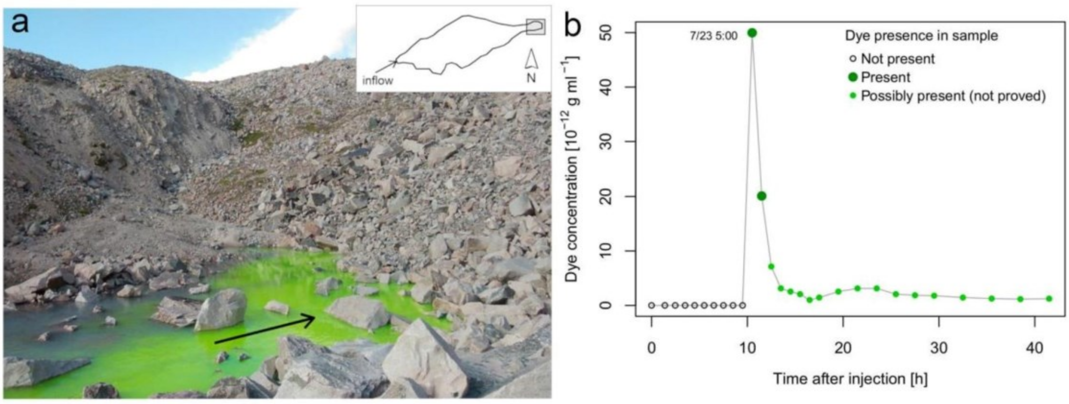

Figure 9. a) Eastern part of Lake 1 and location of subsurface drainage. The flow direction from injection to drainage spot marked with an arrow. The scheme in the upper right corner shows the shoreline of Lake 1 and the part captured in the photograph. b) A breakthrough curve for uranine experiment carried out at the Adygine complex on 22-24 July 2017. The samples were collected in hourly and later in two-hour steps.

3.3 Source of water in lakes and tarns based on stable isotopes

Based on the water isotopic composition, three categories of lakes in terms of their link to the glacial meltwater were earmarked: (i) lakes with significant glacial meltwater recharge, (ii) lakes with limited glacial meltwater recharge, and (iii) lakes with little or no glacial meltwater recharge (Figure 10a). Lakes 2 and 3 constitute the first group, along with four tarns (b, c, j, and k), which showed surprisingly similar water isotopic compositions to the main lakes. These lakes' main feature is continual meltwater through-flow and, thus, the diminished influence of liquid precipitation and evaporation. The basins lie at the local water table. The second group of lakes is characterized by waters noticeably more enriched in heavy isotopes. The subsurface glacial meltwater recharge is present; however, it is less pronounced, and the importance of liquid precipitation is enhanced compared with the first group. This transitional category of lakes with both meltwater and liquid precipitation influence include tarns f, g, and i. The third category – lakes with little or no glacial meltwater recharge, is represented by the tarn a. This tarn's water is substantially ^{2}H- and ^{18}O-enriched compared with the rest of the samples. It is very likely a perched lake (its basin lies above the local water table) and its water balance is governed predominantly by incoming liquid precipitation. The results correspond to the observed water turbidity in the tarns, as tarn a was highlighted having little turbidity compared with the others.

Lake hydrological balance and share of different water sources have been studied thanks to isotopic water composition successfully, for example, in northern Canada (Wolfe et al., 2007; Yi et al., 2008; Brock et al., 2009). Turner et al. (2010) distinguish four lake types on the basis of lake water isotopes – these are: rainfall-dominated, snowmelt-dominated, groundwater-influenced, and evaporation-dominated. Our simple analysis includes only comparison among the lakes at one particular site and its ambition was to determine approximate share of glacial meltwater recharge on lake's balance. Comparing our categories of lakes with those by Turner et al. (2010), the first (i) and third (iii) one could be assigned to the snowmelt- and rainfall-dominated type, respectively. The second category would be a transitional one between the two.

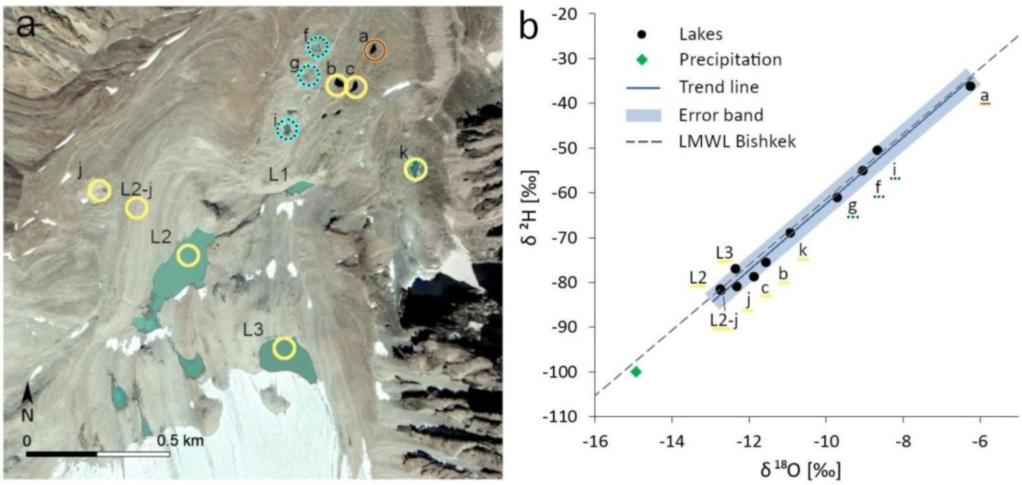

Figure 10. Isotopic composition of water in proglacial lakes and tarns at Adygine complex. a) The water bodies were categorized according to their isotopic composition into three groups: significant glacial meltwater recharge (yellow-blank), limited glacial meltwater recharge (blue-dotted), and little or no glacial meltwater recharge (orange-stripe). b) Isotopic composition of precipitation (snow), individual lake samples, their trend line with error band, and the position of local meteoric water line for Bishkek (data source: IAEA/WMO, 2018).

The isotopic composition of water in the lakes and tarns spans a wide range in the $\delta^{18}O$ (-12.76‰ to -6.25‰) and $\delta^{2}H$ (-81.74‰ to -36.08‰) values, which reflects the diverse water balance conditions of the observed lakes. The most depleted $\delta^{18}O$ and $\delta^{2}H$ values were found in Lake 2's water sample, the most enriched at tarn a (Figure 10). All sampled lakes and tarns are located at limited area, in a horizontal distance of a few hundered meters and elevation difference of up to 130 m from one another. Therefore, the difference (enrichement/depletion in ^{2}H and ^{18}O) is very likely not due to isotopic composition of precipitation falling at lower and higher parts of the site.

Lake samples are generally more enriched in heavy isotopes later in an ablation season compared with its beginning, because of the isotopic fractionation of water during evaporation (Turner et al., 2010), but also due to admixing of summer liquid precipitation. At this altitude (3500 m a.s.l.), snow around lakes melts mainly during May and June, and since then, the dominant components of water balance is glacial meltwater, precipitation (rain), and possibly also evaporation. Due to the timing of sample collection during the peak of ablation season, more prominent differences among the tarns were found. Some stayed close to glacial meltwater isotopic composition, but those without significant recharge of meltwater became more enriched in heavy isotopes.

D-excess of the samples ranges from 13.92‰ to 20.34‰ (d=10‰ for the global meteoric water line) and is calculated from the following equation: $d = \delta^{2}H - 8\,\delta^{18}O$ (Dansgaard, 1964). The higher D-excess points to enhanced moisture recycling and/or a lower relative humidity of vapor above the ocean (Cui et al., 2016). For central Asia, the addition of re-evaporated moisture from continental basins to the water vapor traveling inland is significant, as the residence time of this moisture over the continents is often large (Numaguti, 1999).

There is a close linear relationship between the $\delta^{18}O$ and $\delta^{2}H$ values measured in the lake water samples, and the trend line is given by the following equation: $\delta^{2}H = 7.27\,\delta^{18}O + 10.39$ (n=11, R^2=0.989). The results were supplemented with a LMWL based on precipitation data from a nearby city of Bishkek. The mild slope of the trend line for Adygine (a=7.3) corresponds with that of LMWL for Bishkek (Fig. 10b). Evaporation effect on hydrological balance of lakes and tarns was not proved at any sample as all points are situated close to the line. However, the very small number of samples means that the line's slope may be biased. Since stable water isotopes data available from the sites closest to Adygine (Wulumuqui, China, and sites in Tajikistan) exhibit higher LMWL slope (8.4 and 9.2, respectively; IAEA/WMO, 2018; IAEA, 2018; Liu et al., 2015), it is possible that the slope of the Adygine LMWL can be higher. As a result, the evaporation effect on lake water then might be visible for the tarn a (and maybe also tarns f, g, and i), which would lie below the meteoric water line. The estimation of possible evaporation lines' position (and the effect of seasonally varying precipitation) is well presented in Benettin et al. (2018).

4 Conclusions

The study presents an analysis of hydrological regime and connectivity of proglacial lakes and provides the characteristics of the water flow path through the morainic complex to a stream. The daily lake water level fluctuation revealed dependence of Lake 1 on surface water supply from Lake 2. Hydrological regime of these two lakes showed a fairly similar development throughout an ablation season, only the daily course of fluctuation of Lake 1 lagged by ~2 hours behind Lake 2. Lake's direct contact with glacier terminus, in case of Lake 3, proved to be causing irregularities in the lake's hydrological regime as the glacier meltwater is delivered to the lake solely by englacial and/or subglacial channels. Subsurface drainage of the studied lakes is dependant on the extent of impermeable layers forming the basin and distribution of the drainage channels. The field observations and annual course of lake water level fluctuations revealed that Lake 1 has high capacity of drainage channel that is situated at the bottom, whereas basins of Lake 2 and 3 are formed mostly by impermeable (very likely frozen, ice-rich) material and drainage is allowed only through positions up to a few meters below surface. The glacial meltwater is routed from the glacier throught the lakes and further downstream by a subsurface system in the moraine complex. The results of a dye tracer test show a high water flow velocity (0.085 m s^{-1}) that corresponds to the flow in eroded paths in gravely/bouldery material. However, the low dye recovery indicates that only a small portion of water is routed efficiently to the stream; most water is delayed in the system so the moraine complex is expected to retain considerable volume of water. The complexity of the site's drainage system is supported by varying share of the hydrological balance components in the tarns situated on the moraine. Spatial distribution of permeable and impermeable positions within the morainic landform determine the hydrological connections and drainage paths and causes varying degree of connection to subsurface water routes of individual tarns. Long-term monitoring of these tarns could provide information on the temperature-related degradation and internal changes of the morainic landform.

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References

Aizen, V. B., Aizen, E. M., Melack, J. M., 1996. Precipitation, melt and runoff in the northern Tien Shan. J Hydrol, 186, 229-251. Doi: 10.1016/s0022-1694(96)03022-3

Akhtar, M., Ahmad, N., Booij, M. J., 2008. The impact of climate change on the water resources of Hindukush–Karakorum–Himalaya region under different glacier coverage scenarios. J Hydrol, 355(1-4), 148-163. Doi: 10.1016/j.jhydrol.2008.03.015

Benettin, P., Volkmann, T. H., Freyberg, J. V., Frentress, J., Penna, D., Dawson, T. E., Kirchner, J. W., 2018. Effects of climatic seasonality on the isotopic composition of evaporating soil waters. Hydrol Earth Sys Sc, 22(5), 2881-2890. Doi: 10.5194/hess-2018-40

Bliss, A., Hock, R., Radić, V., 2014. Global response of glacier runoff to twenty-first century climate change. J Geophys Res-Earth, 119(4), 717-730. Doi: 10.1002/2013jf002931

Bolch, T., 2007. Climate change and glacier retreat in northern Tien Shan (Kazakhstan/Kyrgyzstan) using remote sensing data. Global Planet Change, 56(1-2), 1-12. Doi: 10.1016/j.gloplacha.2006.07.009

Bolch, T., Kulkarni, A., Kääb, A., Huggel, C., Paul, F., Cogley, J. G., Frey, H., Kargel, J. S., Fujita, K., Scheel, M., Bajracharya, S., Stoffel, M., 2012. The state and fate of Himalayan glaciers. Science, 336(6079), 310-314. Doi: 10.1126/science.1215828

Brenning, A., 2005. Geomorphological, hydrological and climatic significance of rock glaciers in the Andes of Central Chile (33–35 S). Permafrost Periglac, 16(3), 231-240. Doi: 10.1002/ppp.528

Brock, B. E., Yi, Y., Clogg-Wright, K.P., Edwards, T. W. D., Wolfe, B. B., 2009. Multi-year landscape-scale assessment of lakewater balances in the Slave River Delta, NWT, using water isotope tracers. J Hydrol, 379, 81–91. Doi: 10.1016/j.jhydrol.2009.09.046

Buchli, T., Merz, K., Zhou, X., Kinzelbach, W., Springman, S. M., 2013. Characterization and monitoring of the Furggwanghorn Rock Glacier, Turtmann Valley, Switzerland: results from 2010 to 2012. Vadose Zone J, 12(1), 1-15. Doi: 10.2136/vzj2012.0067

Burkimsher, M., 1983. Investigations of glacier hydrological systems using dye tracer techniques: Observations at Pasterzengletscher, Austria. J Glaciol, 29 (103), 403-416. Doi: 10.3189/s002214300003032x

Caballero, Y., V. Jomelli, P. Chevallier, P. Ribstein, 2002. Hydrological characteristics of slope deposits in high tropical mountains (Cordillera Real, Bolivia). Catena, 47, 101–116. Doi: 10.1016/s0341-8162(01)00179-5

Clark, I. D., Fritz, P., 2013. Environmental isotopes in hydrogeology. CRC press, 352 pp.

Croce, F. A., Milana, J. P., 2002. Internal structure and behaviour of a rock glacier in the Arid Andes of Argentina. Permafrost Periglac, 13(4), 289-299. Doi: 10.1002/ppp.431

Cui, J., Tian, L., Biggs, T. W., Wen, R., 2016. Deuterium-excess determination of evaporation to inflow ratios of an alpine lake: Implications for water balance and modeling. Hydrol Process, 31, 1034–1046. Doi: 10.1002/hyp.11085

Dansgaard, W., 1964. Stable isotopes in precipitation. Tellus, 16, 436-468. Doi: 10.1111/j.2153-3490.1964.tb00181.x

Falatkova, K., Sobr, M., Kocum, J., Jansky, B., 2014. Hydrological regime of Adygine Lake, Tien Shan, Kyrgyzstan. Geografie, 119(4), 320-341.

Falatkova, K., Šobr, M., Neuroiter, A., Schöner, W., Janský, B., Häusler, H., Engel, Z., Beneš, V., 2019. Development of proglacial lakes and evaluation of related outburst susceptibility at Adygine ice-debris complex, northern Tien Shan. Earth Surf Dynam.

Fountain, A. G., Walder, J. S., 1998. Water flow through temperate glaciers. Rev Geophys, 36(3), 299-328. Doi: 10.1029/97rg03579

Fowler, A., 2011. Mathematical geoscience (Vol. 36). Springer Science & Business Media, London, 883 pp. Doi: 10.1007/987-0-85729-721-1

Gorbunov, A. P., 1996. Monitoring the evolution of permafrost in the Tien Shan, Permafrost Periglac, 7(3), 297-298. Doi: 10.1002/(SICI)1099-1530(199609)7:3%3C297::AID-PPP223%3E3.0.CO;2-C

Gordon, S., Sharp, M., Hubbard, B., Smart, C., Ketterling, B., Willis, I., 1998. Seasonal reorganization of subglacial drainage inferred from measurements in boreholes. Hydrol Process, 12 (1), 105–133. Doi: 10.1002/(sici)1099-1085(199801)12:1<105::aid-hyp566>3.0.co;2-#

Gruber, S., 2012. Derivation and analysis of a high-resolution estimate of global permafrost zonation. Cryosphere, 6(1), 221-233. Doi: 10.5194/tc-6-221-2012

IAEA, 2018. Global Network of Isotopes in Rivers. The GNIR Database. Accessible at: https://nucleus.iaea.org/wiser

IAEA/WMO, 2018. Global Network of Isotopes in Precipitation. The GNIP Database. Accessible at: https://nucleus.iaea.org/wiser

Isokangas, E., Rozanski, K., Rossi, P. M., Ronkanen, A. K., Kløve, B., 2015. Quantifying groundwater dependence of a sub-polar lake cluster in Finland using an isotope mass balance approach. Hydrol Earth Sys Sc, 19, 1247–1262. Doi: 10.5194/hess-19-1247-2015

Jones, D. B., Harrison, S., Anderson, K., Betts, R. A., 2018. Mountain rock glaciers contain globally significant water stores. Sci Rep-UK, 8(1), 123-142. Doi: 10.1038/s41598-018-21244-w

Jouzel, J., Hoffmann, G., Koster, R. D., Masson, V., 2000. Water isotopes in precipitation: data/model comparison for present-day and past climates. Quaternary Sci Rev, 19(1-5), 363-379. Doi: 10.1016/s0277-3791(99)00069-4

Kang, S., Yi, Y., Xu, Y., Xu, B., Zhang, Y., 2017. Water isotope framework for lake water balance monitoring and modelling in the nam co basin, Tibetan plateau. J Hydrol: Regional Studies, 12, 289-302. Doi: 10.1016/j.ejrh.2017.05.007

Käss, W., 1998. Tracing Technique in Geohydrology. Balkema, Rotterdam, 581 pp.

Knight, J., Harrison, S., Jones, D. B., 2018. Rock glaciers and the geomorphological evolution of deglacierizing mountains. Geomorphology, 324, 14-24. Doi: 10.1016/j.geomorph.2018.09.020

Krainer, K., W. Mostler, 2002. Hydrology of active rock glaciers: Examples from the Austrian Alps. Arct Antarct Alp Res, 34, 142–149. Doi: 10.1080/15230430.2002.12003478

Kutuzov, S., Shahgedanova, M., 2009. Glacier retreat and climatic variability in the eastern Terskey–Alatoo, inner Tien Shan between the middle of the 19th century and beginning of the 21st century. Global Planet Change, 69(1-2), 59-70. Doi: 10.1016/j.gloplacha.2009.07.001

Langston, G., Bentley, L. R., Hayashi, M., McClymont, A., Pidlisecky, A., 2011. Internal structure and hydrological functions of an alpine proglacial moraine. Hydrol Process, 25(19), 2967-2982. Doi: 10.1002/hyp.8144

Langston, G., Hayashi, M., Roy, J. W., 2013. Quantifying groundwater-surface water interactions in a proglacial moraine using heat and solute tracers. Water Resour Res, 49(9), 5411-5426. Doi: 10.1002/wrcr.20372

Lehmann-Horn, J. A., Walbrecker, J. O., Hertrich, M., Langston, G., McClymont, A. F., Green, A. G., 2011. Imaging groundwater beneath a rugged proglacial moraine. Geophysics, 76(5), B165-B172. Doi: 10.1190/geo2011-0095.1

Liu, Q., Tian, L. D., Wang, J. L., Wen, R., Weng, Y. B., Shen, Y. P., Vladislav, M., Kanaev, E., 2015. A study of longitudinal and altitudinal variations in surface water stable isotopes in West Pamir, Tajikistan. Atmos Res, 153, 10-18. Doi: 10.1016/j.atmosres.2014.07.029

McClymont, A. F., Roy, J. W., Hayashi, M., Bentley, L. R., Maurer, H., Langston, G., 2011. Investigating groundwater flow paths within proglacial moraine using multiple geophysical methods. J Hydrol, 399(1-2), 57-69. Doi: 10.1016/j.jhydrol.2010.12.036

Moorman, B. J., Michel, F. A., 2000. Glacial hydrological system characterization using ground-penetrating radar. Hydrol Process, 14(15), 2645-2667. Doi: 10.1002/1099-1085(20001030)14:15<2645::aid-hyp84>3.3.co;2-u

Nienow, P., 2011. Dye Tracer Investigations of Glacier Hydrology. In: Singh, V. P., Singh, P., Haritashya, U. K. (eds). Encyclopedia of Snow, Ice and Glaciers. Encyclopedia of Earth Sciences Series. Springer, Dordrecht. Doi: 10.1007/978-90-481-2642-2

Nienow, P., Sharp, M., Willis, I., 1998. Seasonal changes in the morphology of the subglacial drainage system, Haut Glacier d'Arolla, Switzerland. Earth Surface Processes and Landforms: The Journal of the British Geomorphological Group, 23(9), 825-843. Doi: 10.1002/(sici)1096-9837(199809)23:9<825::aid-esp893>3.0.co;2-2

Numaguti, A., 1999. Origin and recycling processes of precipitating water over the Eurasian continent: experiments using an atmospheric general circulation model. J Geophys Res, 104, 1957–1972. Doi: 10.1029/1998jd200026

Paul, F., Kääb, A., Haeberli, W., 2007. Recent glacier changes in the Alps observed by satellite: Consequences for future monitoring strategies. Global Planet Change, 56(1-2), 111-122. Doi: 10.1016/j.gloplacha.2006.07.007

Paul, F., Mölg, N., 2014. Hasty retreat of glaciers in northern Patagonia from 1985 to 2011. J Glaciol, 60(224), 1033-1043. Doi: 10.3189/2014jog14j104

Pauritsch, M., Wagner, T., Winkler, G., Birk, S., 2017. Investigating groundwater flow components in an Alpine relict rock glacier (Austria) using a numerical model. Hydrogeol J, 25(2), 371-383. Doi: 10.1007/s10040-016-1484-x

Penna, D., Engel, M., Mao, L., Dell'Agnese, A., Bertoldi, G., Comiti, F., 2014. Tracer-based analysis of spatial and temporal variations of water sources in a glacierized catchment. Hydrol Earth Sys Sc, 18(12), 5271-5288. Doi: 10.5194/hess-18-5271-2014

Petrakov, D., Shpuntova, A., Aleinikov, A., Kääb, A., Kutuzov, S., Lavrentiev, I., Stoffel, M., Tutubalina, O., Usubaliev, R., 2016. Accelerated glacier shrinkage in the Ak-Shyirak massif, Inner Tien Shan, during 2003–2013. Sci Total Environ, 562, 364-378. Doi: 10.1016/j.scitotenv.2016.03.162

Rabatel, A., Francou, B., Soruco, A., Gomez, J., Cáceres, B., Ceballos, J. L., Basantes, R., Vuille, M., Sicart, J.-E., Huggel, C. et al. (2013). Current state of glaciers in the tropical Andes: a multi-century perspective on glacier evolution and climate change. Cryosphere, 7(1), 81-102. Doi: 10.5194/tc-7-81-2013

Rangecroft, S., Harrison, S., Anderson, K., 2015. Rock glaciers as water stores in the Bolivian Andes: an assessment of their hydrological importance. Arct Antarct Alp Res, 47(1), 89-98. Doi: 10.1657/aaar0014-029

Rothlisberger, H., 1972. Water pressure in intra- and subglacial channels. J Glaciol, 11(62), 177-204. Doi: 10.3189/s0022143000022188

Schiefer, E., Menounos, B., Wheate, R., 2007. Recent volume loss of British Columbian glaciers, Canada. Geophys Res Lett, 34(16), L16503. Doi:10.1029/2007GL030780.

Shahgedanova, M., Afzal, M., Severskiy, I., Usmanova, Z., Saidaliyeva, Z., Kapitsa, V., Kasatkin, N., Dolgikh, S., 2018. Changes in the mountain river discharge in the northern Tien Shan since the mid-20th Century: Results from the analysis of a homogeneous daily streamflow data set from seven catchments. J Hydrol, 564, 1133-1152. Doi: 10.1016/j.jhydrol.2018.08.001

Shatravin, V. I., 2000. Reconstruction of the Pleistocene and Holocene glaciations of the Tian-Shan and Pamir: new results. International Workshop, July 22–23, Institute of Soil Science and Soil Geography, University of Bayreuth, Germany.

Shiyin, L., Wenxin, S., Yongping, S., Gang, L., 2003. Glacier changes since the Little Ice Age maximum in the western Qilian Shan, northwest China, and consequences of glacier runoff for water supply. J Glaciol, 49(164), 117-124. Doi: 10.3189/172756503781830926

Singh, P., Haritashya, U. K., Ramasastri, K. S., Kumar, N., 2005. Diurnal variations in discharge and suspended sediment concentration, including runoff-delaying characteristics, of the Gangotri Glacier in the Garhwal Himalayas. Hydrol Process: An International Journal, 19(7), 1445-1457. Doi: 10.1002/hyp.5583

Sorg, A., Bolch, T., Stoffel, M., Solomina, O., Beniston, M., 2012. Climate change impacts on glaciers and runoff in Tien Shan (Central Asia). Nat Clim Change, 2(10), 725-731. Doi: 10.1038/NCLIMATE1592

Stewart, M. K., 1975. Stable isotope fractionation due to evaporation and isotopic exchange of falling waterdrops: Applications to atmospheric processes and evaporation of lakes. J Geophys Res, 80(9), 1133-1146. Doi: 10.1029/jc080i009p01133

Swift, D. A., Nienow, P. W., Hoey, T. B., Mair, D. W., 2005. Seasonal evolution of runoff from Haut Glacier d'Arolla, Switzerland and implications for glacial geomorphic processes. J Hydrol, 309(1-4), 133-148. Doi: 10.1016/j.jhydrol.2004.11.016

Tenthorey, G., 1992. Perennial néevées and the hydrology of rock glaciers. Permafrost Periglac, 3, 247–252. Doi: 10.1002/ppp.3430030313

Turner, K. W., Wolfe, B. B., Edwards, T. W., 2010. Characterizing the role of hydrological processes on lake water balances in the Old Crow Flats, Yukon Territory, Canada, using water isotope tracers. J Hydrol, 386(1-4), 103-117. Doi: 10.1016/j.jhydrol.2010.03.012

Vergara, W., Deeb, A., Valencia, A., Bradley, R., Francou, B., Zarzar, A., Grunwaldt, A., Haeussling, S., 2007. Economic impacts of rapid glacier retreat in the Andes. Eos, Transactions American Geophysical Union, 88(25), 261-264. Doi: 10.1029/2007eo250001

Winkler, G., Wagner, T., Pauritsch, M., Birk, S., Kellerer-Pirklbauer, A., Benischke, R., Leis, A., Morawetz, R., Schreilechner, M. G., Hergarten, S., 2016. Identification and assessment of groundwater flow and storage components of the relict Schöneben Rock Glacier, Niedere Tauern Range, Eastern Alps (Austria). Hydrogeol J, 24(4), 937-953. Doi: 10.1007/s10040-015-1348-9

Wolfe, B. B., Karst-Riddoch, T. L., Hall, R. I., Edwards, T. W. D., English, M. C., Palmini, R., McGowan, S., Leavitt, P. R., Vardy, S. R., 2007. Classification of hydrological regimes of northern floodplain basins (Peace-Athabasca Delta, Canada) from analysis of stable isotopes (d18O, d2H) and water chemistry. Hydrol Process, 21, 151–168. Doi: 10.1002/hyp.6229

Wu, H.W., Li, X.Y., Li, J., Jiang, Z.Y., Li, G.Y., Liu, L., 2015. Evaporative enrichment of stable isotopes ($\delta^{18}O$ and δD) in lake water and the relation to lake-level change of Lake Qinghai, Northeast Tibetan Plateau of China. J Arid Land, 7 (5), 623–635. Doi: 10.1007/s40333-015-0048-6

Yang, Y., Wu, Q., Yun, H., Jin, H., Zhang, Z., 2016. Evaluation of the hydrological contributions of permafrost to the thermokarst lakes on the Qinghai-Tibet Plateau using stable isotopes. Global Planet Change, 140, 1–8. Doi: 10.1016/j.gloplacha.2016.03.006

Yi, Y., Brock, B. E., Falcone, M. D., Wolfe, B. B., Edwards, T. W. D., 2008. A coupled isotope tracer method to characterize input water to lakes. J Hydrol, 350, 1–13.Doi: 10.1016/j.jhydrol.2007.11.008

Appendix

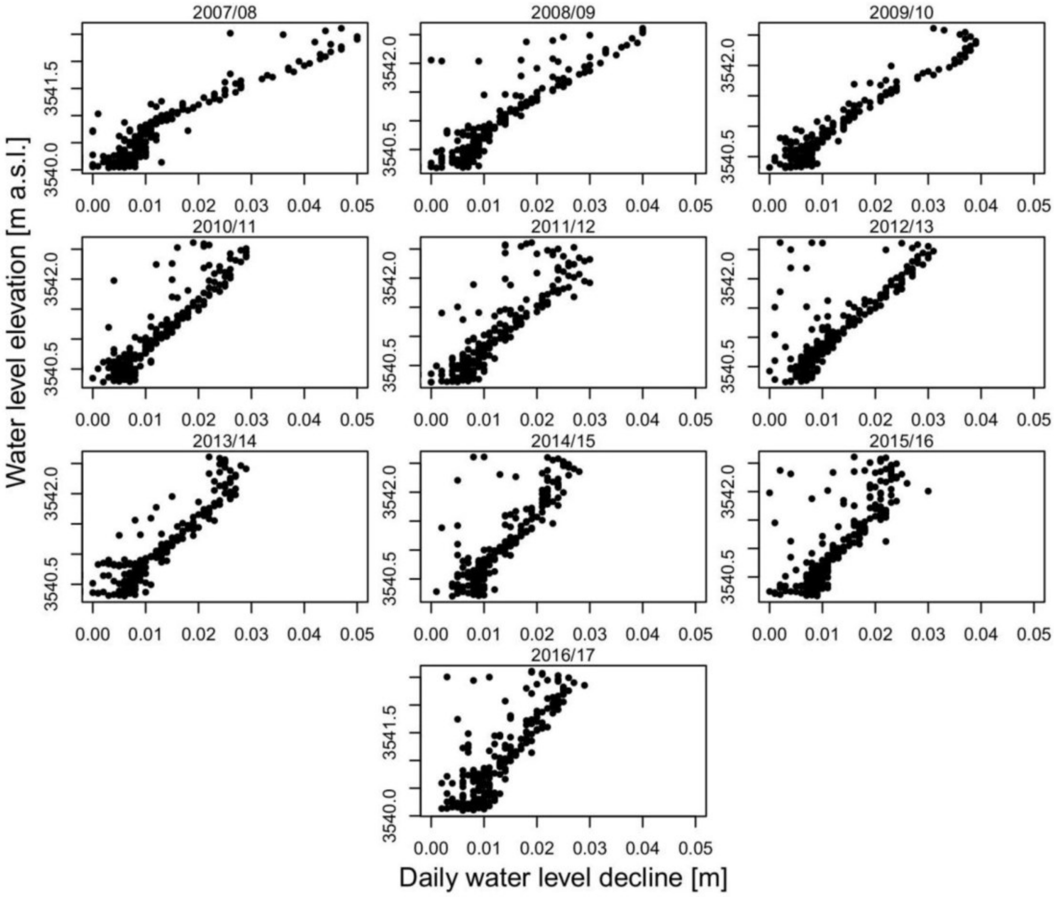

Figure A1. Daily water level decline of Lake 2 in the cold part of a year related to water level elevation in the 2007-2017. The analysed period (decline) is limited by the timing of the following two events for each season: water level dropping below the surface drainage channel and spring snow meltwater-induced water level rise.

6. Conclusions and discussion

The problematics of glacial lake development and hydrological conditions of a proglacial area was addressed within four scientific papers presented in the result section of this thesis. Here is the summary of results and linkage of the individual findings:

- Lake development is strongly influenced by the contact with a glacier and presence of buried ice (ice-rich debris) in basin bottom and sides. Lakes in intramorainic depression thus have their development linked to buried glacier remnants, ice blocks and lenses within debris accumulations.

- The most common type of glacial lake and also an outburst trigger (mechanism) varies across mountain areas. In the studied region, the typical lake is formed in an intramorainic depression and its sudden drainage is often caused by subsurface channel opening. In relation to that, an assessment procedure adapted to regional conditions is used to evaluate lake outburst susceptibility.

- Change of outburst susceptibility in future will be linked to permafrost degradation (slope failures), buried ice exposure and melting, formation of new lakes in overdeepenings of the exposed glacier bed, and also change in glacier runoff regime. The ablation season is expected to last longer, snow melting to occur earlier and shift exposure and melting of glacier ice to earlier time of a year. Lake stability may be influenced by varied meltwater inflow and temporal distribution of outburst cases will likely change its pattern.

- As the lakes are fed by glacier meltwater, monitoring of lake water level fluctuations provided useful information on daily and seasonal variations of glacier meltwater runoff. The proglacial lakes showed a typical glacial regime – during an ablation season, there was a distinct evolution of several water level fluctuation characteristics, namely daily amplitude, timing of daily peak, and time lag of daily peak after air temperature maximum. Water level fluctuation during a cold season, when inflow from glacier is very low, revealed properties of lakes' subsurface drainage system (changes in drainage channels capacity, their depth below surface).

- A dye tracer test helped to describe characteristics of water passage through the proglacial environment. According to the observed dye concentrations in the stream, the morainic landform involves a dual system – small part of incoming water is routed efficiently to the stream, larger part is delayed in the system. This system was also described in some moraine complexes and rock glaciers. By comparing the water isotopic composition of individual lakes, the influence of meltwater on a lake's balance was determined. Similarity of isotopic composition between several tarns in the moraine complex and glacier-fed lakes suggest that the tarns have subsurface connection to the meltwater.

There are several thematic areas that were not addressed thoroughly in this thesis, either due to the lack of data, time, resources, or due to the relatively wide thematic scope of this work. One of them is the role of permafrost in the proglacial lake development, their outburst susceptibility, and subsurface water routing. According to the scientific literature and the mean annual air temperature of the site, the upper part of the study site is very likely within the continual permafrost zone, the lower part (morainic complex at ~3500 m a.s.l.) is within the discontinual permafrost zone. However, to describe, for example, the role of permafrost in subsurface drainage system of the lower parts in more detail, precise borehole or geophysical data would be necessary.

Also, proper hydrological balance of the main lakes could not be determined as most of the lakes' inflow and outflow is below the surface and thus hard to quantify. The summer inflow rates of Lake 2 were estimated based on the discharge measurements of the lake's surface outflow and approximate capacity of the subsurface drainage channels (based on the cold season water level decline). However, without further knowledge of the lake's watershed (englacial meltwater routing), the share of glacial meltwater passing through this lake (and also the total glacier runoff) cannot be determined.

The comparison of stable water isotopes share in individual water bodies yielded certain results, showing distinctive differences among the tarns. These first-step findings could be build upon with further analyses of stable water isotopic changes within an ablation season and over the course of several years. The data could indicate possible changes in the drainage system of the moraine complex resulting from melting of buried ice or permafrost degradation.

References

Akhtar, M., Ahmad, N., & Booij, M. J. (2008). The impact of climate change on the water resources of Hindukush–Karakorum–Himalaya region under different glacier coverage scenarios. *Journal of hydrology*, 355(1-4), 148-163. doi: 10.1016/j.jhydrol.2008.03.015

Bajracharya, B., Shrestha, A. B., & Rajbhandari, L. (2007). Glacial lake outburst floods in the Sagarmatha region: hazard assessment using GIS and hydrodynamic modeling. *Mountain Research and Development*, 27(4), 336-344. doi: 10.1659/mrd.0783

Barry, R. G. (2006). The status of research on glaciers and global glacier recession: a review. *Progress in Physical geography*, 30(3), 285-306. doi: 10.1191/0309133306pp478ra

Benn, D. I., Wiseman, S., & Hands, K. A. (2001). Growth and drainage of supraglacial lakes on debris mantled Ngozumpa glacier, Khumbu Himal, Nepal. *Journal of Glaciology*, 47(159), 626–638. doi:10.3189/172756501781831729

Bliss, A., Hock, R., & Radić, V. (2014). Global response of glacier runoff to twenty-first century climate change. *Journal of Geophysical Research: Earth Surface*, 119(4), 717-730. doi: 10.1002/2013jf002931

Blöthe, J. H., Rosenwinkel, S., Höser, T., & Korup, O. (2018). Rock-glacier dams in High Asia. Earth Surface Processes and Landforms, doi:10.1002/esp.4532.

Bolch, T., Buchroithner, M. F., Peters, J., Baessler, M., & Bajracharya, S. (2008). Identification of glacier motion and potentially dangerous glacial lakes in the Mt. Everest region/Nepal using spaceborne imagery. *Natural Hazards and Earth System Sciences*, 8(6), 1329-1340. doi:10.5194/nhess-8-1329-2008

Bolch, T. (2015). Glacier area and mass changes since 1964 in the Ala Archa Valley, Kyrgyz Ala-Too, northern Tien Shan. *Lëd i Sneg*, 55(1), 28-39. doi: 10.15356/2076-6734-2015-1-28-39

Bolch, T., Peters, J., Yerogov, A., Pradhan, B., Buchroithner, M., & Blagoveshchensky, V. (2011). Identification of potentially dangerous glacial lakes in the northern Tien Shan. *Natural Hazards*, 59, 1691–1714. doi: 10.1007/s11069-011-9860-2

Bradley, R. S., Vuille, M., Diaz, H. F., & Vergara, W. (2006). Threats to water supplies in the tropical Andes. *Science*, 312(5781), 1755-1756. doi: 10.1126/science.1128087

Brenning, A. (2005). Geomorphological, hydrological and climatic significance of rock glaciers in the Andes of Central Chile (33–35 S). *Permafrost and Periglacial Processes*, 16(3), 231-240. doi: 10.1002/ppp.528

Buchli, T., Merz, K., Zhou, X., Kinzelbach, W., & Springman, S. M. (2013). Characterization and monitoring of the Furggwanghorn Rock Glacier, Turtmann Valley, Switzerland: results from 2010 to 2012. *Vadose Zone Journal*, 12(1). doi: 10.2136/vzj2012.0067

Clague, J. J. & Evans, S. G. (2000). A review of catastrophic drainage of moraine-dammed lakes in British Columbia. *Quaternary Science Reviews*, 19(17-18), 1763-1783. doi: 10.1016/s0277-3791(00)00090-1

Clark, I., & Fritz, P. (1997). Environmental Isotopes in Hydrogeology. Lewis Publishers, Boca Raton, 352 pp.

Cooper, R., Hodgkins, R., Wadham, J., & Tranter, M. (2011). The hydrology of the proglacial zone of a high-Arctic glacier (Finsterwalderbreen, Svalbard): Sub-surface water fluxes and complete water budget. *Journal of Hydrology*, 406(1-2), 88-96. doi: 10.1016/j.jhydrol.2009.09.020

Croce, F. A., & Milana, J. P. (2002). Internal structure and behaviour of a rock glacier in the Arid Andes of Argentina. *Permafrost and Periglacial Processes*, 13(4), 289-299. doi: 10.1002/ppp.431

Cuffey, K. M., & Paterson, W. S. B. (2010). The Physics of Glaciers. Academic Press, 4th edition, 704 pp.

Česák, J., & Šobr, M. (2005). Metody batymetrického mapování českých jezer. *Geografie*, 110, 141-151.

Déry, S. J., Stahl, K., Moore, R. D., Whitfield, P. H., Menounos, B., & Burford, J. E. (2009). Detection of runoff timing changes in pluvial, nival, and glacial rivers of western Canada. *Water Resources Research*, 45(4), W04426, doi:10.1029/2008WR006975

Ding, Y., & Liu, J. (1992). Glacier lake outburst flood disasters in China. *Annals of Glaciology*, 16, 180–184. doi: 10.3189/1992aog16-1-180-184

Duguay, M. A., Edmunds, A., Arenson, L. U., & Wainstein, P. A. (2015). Quantifying the significance of the hydrological contribution of a rock glacier–A review. *GEO Québec 2015: Challenges From North to South*, Québec, Canada.

Dussaillant, A., Benito, G., Buytaert, W., Carling, P., Meier, C., & Espinoza, F. (2010). Repeated glacial-lake outburst floods in Patagonia: an increasing hazard? *Natural hazards*, 54(2), 469-481. doi: 10.1007/s11069-009-9479-8

Emmer, A., & Vilímek, V. (2013). Lake and breach hazard assessment for moraine-dammed lakes: an example from the Cordillera Blanca (Peru). *Natural Hazards and Earth System Sciences*, 13(6), 1551. doi: 10.5194/nhess-13-1551-2013

Emmer, A., & Vilímek, V. (2014). New method for assessing the susceptibility of glacial lakes to outburst floods in the Cordillera Blanca, Peru. *Hydrology and Earth System Sciences*, 18(9), 3461-3479. doi: 10.5194/hess-18-3461-2014

Erokhin, S. A., Zaginaev, V. V., Meleshko, A. A., Ruiz-Villanueva, V., Petrakov, D. A., Chernomorets, S. S., Viskhadzhieva, K.S., Tutubalina, O.V., & Stoffel, M. (2018). Debris flows triggered from non-stationary glacier lake outbursts: the case of the Teztor Lake complex (Northern Tian Shan, Kyrgyzstan). *Landslides*, 15(1), 83-98. doi: 10.1007/s10346-017-0862-3

Erokhin S. A., & Zaginaev V. V. (2016). Forecasting of hazards from mountain lakes based on the catalogue. Report book of Kyrgyz Ministry of Emergency Situations, Part III, 542-555. (in Russian)

Fleming, S. W., & Clarke, G. K. C. (2005). Attenuation of high-frequency interannual streamflow variability by watershed glacier cover. *ASCE Journal of Hydraulic Engineering*, 131, 615–618. doi: 10.1061/(asce)0733-9429(2005)131:7(615)

Flowers, G. E., & Clarke, G. K. C. (2002). A multicomponent coupled model of glacier hydrology: 1. Theory and synthetic examples. *Journal of Geophysical Research*, 107(B11), 2287. doi: 10.1029/2001JB001122

Fountain, A. G., Jacobel, R. W., Schlichting, R., & Jansson, P. (2005). Fractures as the main pathways of water flow in temperate glaciers. *Nature*, 433 (7026), 618–621. doi: 10.1038/nature03296

Fountain, A. G., & Walder, J. S. (1998). Water flow through temperate glaciers. *Review of Geophysics*, 36, 299-328. doi: 10.1029/97rg03579

Frey, H., Huggel, C., Bühler, Y., Buis, D., Burga, M. D., Choquevilca, W., Fernandez, F., Hernández, J.G., Giráldez, C., Loarte, E., & Masias, P. (2016). A robust debris-flow and GLOF risk management strategy for a data-scarce catchment in Santa Teresa, Peru. *Landslides*, 13(6), 1493-1507. doi: 10.1007/s10346-015-0669-z

Frey, H., Haeberli, W., Linsbauer, A., Huggel, C., & Paul, F. (2010). A multi-level strategy for anticipating future glacier lake formation and associated hazard potentials. *Natural Hazards and Earth System Sciences*, 10(2), 339-352. doi: 10.5194/nhess-10-339-2010

GAPHAZ (2017). Assessment of Glacier and Permafrost Hazards in Mountain Regions – Technical Guidance Document. Prepared by Allen, S., Frey, H., Huggel, C. et al. Standing Group on Glacier and Permafrost Hazards in Mountains (GAPHAZ) of the International Association of Cryospheric Sciences (IACS) and the International Permafrost Association (IPA). Zurich, Switzerland / Lima, Peru, 72 pp.

Glazirin, G. E. (2010). A century of investigations on outbursts of the ice-dammed lake Merzbacher (central Tien Shan). *Austrian Journal of Earth Sciences*, 103(2), 171-179.

Haeberli, W., Schaub, Y., & Huggel, C. (2017). Increasing risks related to landslides from degrading permafrost into new lakes in de-glaciating mountain ranges. *Geomorphology*, 293, 405-417. doi: 10.1016/j.geomorph.2016.02.009

Hagg, W., Hoelzle, M., Wagner, S., Mayr, E., & Klose, Z. (2013). Glacier and runoff changes in the Rukhk catchment, upper Amu-Darya basin until 2050. *Global and planetary change*, 110, 62-73. doi: 10.1016/j.gloplacha.2013.05.005

Hegglin, E., & Huggel, C. (2008). An integrated assessment of vulnerability to glacial hazards: a case study in the Cordillera Blanca, Peru. *Mountain Research and Development*, 28(3), 299-309. doi: 10.1659/mrd.0976

Hock, R. (2005). Glacier melt: a review of processes and their modelling. *Progress in physical geography*, 29(3), 362-391. doi: 10.1191/0309133305pp453ra

Hubbard, B., & Glasser, N. F. (2005). Field techniques in glaciology and glacial geomorphology. John Wiley & Sons, Chichester, England, 425 pp.

Huggel, C., Haeberli, W., Kääb, A., Bieri, D., & Richardson, S. D. (2004). An assessment procedure for glacial hazards in the Swiss Alps. *Canadian Geotechnical Journal*, 41, 1068-1083. doi: 10.1139/t04-053

Huss, M., Jouvet, G., Farinotti, D., & Bauder, A. (2010). Future high-mountain hydrology: a new parameterization of glacier retreat. *Hydrology and Earth System Sciences*, 14(5), 815-829. doi: 10.5194/hessd-7-345-2010

Huss, M., & Hock, R. (2018). Global-scale hydrological response to future glacier mass loss. *Nature Climate Change*, 8(2), 135. doi: 10.1038/s41558-017-0049-x

Huss, M. (2011). Present and future contribution of glacier storage change to runoff from macroscale drainage basins in Europe. *Water Resources Research*, 47(7), W07511. doi:10.1029/2010WR010299

Huss, M., Farinotti, D., Bauder, A., & Funk, M. (2008). Modelling runoff from highly glacierized alpine drainage basins in a changing climate. *Hydrological processes*, 22(19), 3888-3902. doi: 10.1002/hyp.7055

Huss, M., Bauder, A., Werder, M., Funk, M., & Hock, R. (2007). Glacier-dammed lake outburst events of Gornersee, Switzerland. *Journal of Glaciology*, 53(181), 189-200. doi: 10.3189/172756507782202784

Immerzeel, W. W., Van Beek, L. P. H., Konz, M., Shrestha, A. B., & Bierkens, M. F. P. (2012). Hydrological response to climate change in a glacierized catchment in the Himalayas. *Climatic change*, 110(3-4), 721-736. doi: 10.1007/s10584-011-0143-4

IPCC (2014). Climate Change 2014: Synthesis Report. Contribution of Working Groups I, II and III to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change [Core Writing Team, R.K. Pachauri and L.A. Meyer (eds.)]. IPCC, Geneva, Switzerland, 151 pp.

Irvine-Fynn, T. D., Hodson, A. J., Moorman, B. J., Vatne, G., & Hubbard, A. L. (2011). Polythermal glacier hydrology: A review. *Reviews of Geophysics*, 49(4), 1-37, doi:10.1029/2010RG000350.1.INTRODUCTION

Ischuk, N. (2013). The Origin of the Mountain River Dams in Tajikistan. In: Margottini C., Canuti P., Sassa K. (eds): Landslide Science and Practice. Springer, Berlin, Heidelberg, doi: 10.1007/978-3-642-31319-6_2.

Ives, J. D., Shrestha, R. B., & Mool, P. K. (2010). Formation of glacial lakes in the Hindu Kush-Himalayas and GLOF risk assessment. Kathmandu: ICIMOD.

Janský, B., Šobr, M., & Erokhin, S. (2006). Typology of high mountain lakes of Kyrgyzstan with regard to the risk of their rupture. *Limnological Review*, 6, 135-140. doi: 10.1016/j.limno.2009.11.013

Janský, B., Engel, Z., Šobr, M., Beneš, V., Špaček, K., & Yerokhin, S. (2009). The evolution of Petrov lake and moraine dam rupture risk (Tien-Shan, Kyrgyzstan). Natural Hazards, 50(1), 83-96. doi: 10.1007/s11069-008-9321-8

Janský, B., Engel, Z., Kocum, J., Šefrna, L., & Česák, J. (2011). The Amazon River headstream area in the Cordillera Chila, Peru: hydrographical, hydrological and glaciological conditions. Hydrological Sciences Journal, 56(1), 138–151. doi: 10.1080/02626667.2010.544257

Jones, D. B., Harrison, S., Anderson, K., & Betts, R. A. (2018a). Mountain rock glaciers contain globally significant water stores. *Scientific reports*, 8(1), 2834. doi: 10.1038/s41598-018-21244-w

Jones, D. B., Harrison, S., Anderson, K., Selley, H. L., Wood, J. L., & Betts, R. A. (2018b). The distribution and hydrological significance of rock glaciers in the Nepalese Himalaya. *Global and Planetary Change*, 160, 123-142. doi: 10.1016/j.gloplacha.2017.11.005

Käss, W. (1998). Tracing Technique in Geohydrology. Balkema, Rotterdam, 581 pp.

Korup, O. & Tweed, F. (2007). Ice, moraine, and landslide dams in mountainous terrain, Quaternary Science Reviews, 26, 3406–3422. doi:10.1016/j.quascirev.2007.10.012

Kougkoulos, I., Cook, S. J., Edwards, L. A., Clarke, L. J., Symeonakis, E., Dortch, J. M., & Nesbitt, K. (2018a). Modelling glacial lake outburst flood impacts in the Bolivian Andes. *Natural Hazards*, 94(3), 1415-1438. doi: 10.1007/s11069-018-3486-6

Kougkoulos, I., Cook, S. J., Jomelli, V., Clarke, L., Symeonakis, E., Dortch, J. M., ... & Merad, M. (2018b). Use of multi-criteria decision analysis to identify potentially dangerous glacial lakes. *Science of the Total Environment*, 621, 1453-1466. doi: 10.1016/j.scitotenv.2017.10.083

Krainer, K., & Mostler, W. (2002). Hydrology of active rock glaciers: examples from the Austrian Alps. *Arctic, Antarctic, and Alpine Research*, 34(2), 142-149. doi: 10.2307/1552465

Langston, G., Bentley, L. R., Hayashi, M., McClymont, A., & Pidlisecky, A. (2011). Internal structure and hydrological functions of an alpine proglacial moraine. *Hydrological Processes*, 25(19), 2967-2982. doi: 10.1002/hyp.8144

Langston, G., Hayashi, M., & Roy, J. W. (2013). Quantifying groundwater-surface water interactions in a proglacial moraine using heat and solute tracers. *Water Resources Research*, 49(9), 5411-5426. doi: 10.1002/wrcr.20372

Levy, A., Robinson, Z., Krause, S., Waller, R., & Weatherill, J. (2015). Long-term variability of proglacial groundwater-fed hydrological systems in an area of glacier retreat, Skeiðarársandur, Iceland. *Earth Surface Processes and Landforms*, 40(7), 981-994. doi: 10.1002/esp.3696

Linsbauer, A., Paul, F., & Haeberli, W. (2012). Modeling glacier thickness distribution and bed topography over entire mountain ranges with GlabTop: Application of a fast and robust approach. *Journal of Geophysical Research: Earth Surface*, 117(F3). doi: 10.1029/2011jf002313

Lutz, A. F., Immerzeel, W. W., Shrestha, A. B., & Bierkens, M. F. P. (2014). Consistent increase in High Asia's runoff due to increasing glacier melt and precipitation. *Nature Climate Change*, 4(7), 587. doi: 10.1038/nclimate2237

Mathews, W. H. (1964). Water pressure under a glacier. *Journal of Glaciology*, 5(38), 235–240. doi: 10.3189/s0022143000028811

McKillop, R. J., & Clague, J. J. (2007a). Statistical, remote sensing-based approach for estimating the probability of catastrophic drainage from moraine-dammed lakes in southwestern British Columbia. *Global and Planetary Change*, 56, 153–171. doi:10.1016/j.gloplacha.2006.07.004

McKillop, R. J., & Clague, J. J. (2007b). A procedure for making objective preliminary assessments of outburst flood hazard from morainedammed lakes in southwestern British Columbia. *Natural Hazards*, 41, 131–157. doi:10.1007/s11069-006-9028-7

Mergili, M., & Schneider, J. F. (2011). Regional-scale analysis of lake outburst hazards in the southwestern Pamir, Tajikistan, based on remote sensing and GIS. *Natural Hazards and Earth System Sciences*, 11(5), 1447-1462. doi: 10.5194/nhess-11-1447-2011

Miller, J. D., Immerzeel, W. W., & Rees, G. (2012). Climate change impacts on glacier hydrology and river discharge in the Hindu Kush–Himalayas: a synthesis of the scientific basis. *Mountain Research and Development,* 32(4), 461-467. doi: 10.1659/mrd-journal-d-12-00027.1

Milner, A. M., Khamis, K., Battin, T. J., Brittain, J. E., Barrand, N. E., Füreder, L., ... & Hodson, A. J. (2017). Glacier shrinkage driving global changes in downstream systems. Proceedings of the National Academy of Sciences, 114(37), 9770-9778. doi: 10.1073/pnas.1619807114

Moore, R. D., Fleming, S. W., Menounos, B., Wheate, R., Fountain, A., Stahl, K., Holm, K., & Jakob, M. (2009). Glacier change in western North America: influences on hydrology, geomorphic hazards and water quality. *Hydrological Processes*, 23, 42-61. doi: 10.1002/hyp.7162

Narama, C., Duishonakunov, M., Kääb, A., Daiyrov, M., & Abdrakhmatov, K. (2010). The 24 July 2008 outburst flood at the western Zyndan glacier lake and recent regional changes in glacier lakes of the Teskey Ala-Too range, Tien Shan, Kyrgyzstan. *Natural Hazards and Earth System Sciences*, 10(4), 647-659. doi: 10.5194/nhess-10-647-2010

Nepal, S. (2016). Impacts of climate change on the hydrological regime of the Koshi river basin in the Himalayan region. *Journal of Hydro-environment Research*, 10, 76-89. doi: 10.1016/j.jher.2015.12.001

Nienow, P., Sharp, M., & Willis, I. (1998). Seasonal changes in the morphology of the subglacial drainage system, Haut Glacier d'Arolla, Switzerland. *Earth Surface Processes and Landforms*, 23(9), 825-843. doi: 10.1002/(sici)1096-9837(199809)23:9<825::aid-esp893>3.0.co;2-2

Nye, J. F. (1976). Water flow in glaciers: jökulhlaups, tunnels and veins. *Journal of Glaciology*,17 (76), 181–207. doi: 10.3189/s002214300001354x

O'Connor, J. E., Hardison, J. H., & Costa, J. E. (2001). Debris flows from failures of Neoglacial-age moraine dams in the Three Sisters and Mount Jefferson Wilderness areas, Oregon. US Geological Survey, Reston (Virginia), 93 pp. doi: 10.3133/pp1606

Pauritsch, M., Wagner, T., Winkler, G., & Birk, S. (2017). Investigating groundwater flow components in an Alpine relict rock glacier (Austria) using a numerical model. *Hydrogeology Journal*, 25(2), 371-383. doi: 10.1007/s10040-016-1484-x

Penna, D., Engel, M., Mao, L., Dell'Agnese, A., Bertoldi, G., & Comiti, F. (2014). Tracer-based analysis of spatial and temporal variations of water sources in a glacierized catchment. *Hydrology and Earth System Sciences*, 18(12), 5271-5288. doi: 10.5194/hess-18-5271-2014

Pepin, N., Bradley, R. S., Diaz, H. F., Baraër, M., Caceres, E. B., Forsythe, N., ... & Miller, J. R. (2015). Elevation-dependent warming in mountain regions of the world. *Nature Climate Change*, 5(5), 424. doi: 10.1038/nclimate2563

Quincey, D. J., Richardson, S. D., Luckman, A., Lucas, R. M., Reynolds, J. M., Hambrey, M. J., & Glasser, N. F. (2007). Early recognition of glacial lake hazards in the Himalaya using remote sensing datasets. *Global and Planetary Change*, 56(1-2), 137-152. doi:10.1016/j.gloplacha.2006.07.013,

Radić, V., Bliss, A., Beedlow, A. C., Hock, R., Miles, E., & Cogley, J. G. (2014). Regional and global projections of twenty-first century glacier mass changes in response to climate scenarios from global climate models. *Climate Dynamics*, 42(1-2), 37-58. doi: 10.1007/s00382-013-1719-7

Rangecroft, S., Harrison, S., & Anderson, K. (2015). Rock glaciers as water stores in the Bolivian Andes: an assessment of their hydrological importance. *Arctic, antarctic, and alpine research*, 47(1), 89-98. doi: 10.1657/aaar0014-029

Reynolds, J. M. (1998). High-altitude glacial lake hazard assessment and mitigation: a Himalayan perspective. Geological Society, London, Engineering Geology Special Publications, 15(1), 25-34. doi: 10.1144/gsl.eng.1998.015.01.03

Richardson, S. D., & Reynolds, J. M. (2000a). An overview on glacial hazards in the Himalayas. *Quaternary International*, 65/66, 31-47. doi: 10.1016/s1040-6182(99)00035-x

Richardson, S. D., & Reynolds, J. M. (2000b). Degradation of ice-cored moraine dams: implications for hazard development. In: Debris-covered Glaciers, edited by: Nakawo, M., Raymond, C. F., and Fountain, A., 187–197.

Rounce, D. R., McKinney, D. C., Lala, J. M., Byers, A. C., & Watson, C. S. (2016). A new remote hazard and risk assessment framework for glacial lakes in the Nepal Himalaya. *Hydrology and Earth System Sciences*, 20(9), 3455. doi: 10.3390/rs9070654

Röthlisberger, H. (1972). Water pressure in intra- and subglacial channels. *Journal of Glaciology*, 11 (62), 177–203. doi: 10.1017/s0022143000022188

Röthlisberger, H., & Lang, H. (1987). Glacial Hydrology. In: Gurnell, A. M., Clark, M. J. (eds.): Glacio-fluvial Sediment Transfer. 1st ed. New York (USA): Wiley, pp. 207-284.

Sakai, A., Nishimura, K., Kadota, T., & Takeuchi, T. (2009). Onset of calving at supraglacial lakes on debris-covered glaciers of the Nepal Himalaya. *Journal of Glaciology*, 55(193), 909–917. doi: 10.3189/002214309790152555

Schneider, D., Huggel, C., Cochachín Rapre, A., & García Hernández, J. (2014). Mapping hazards from glacier lake outburst floods based on modelling of process cascades at Lake 513, Carhuaz, Peru. Advances in Geosciences, 35, 145-155. doi: 10.5194/adgeo-35-145-2014

Shreve, R. L. (1972). Movement of water in glaciers. *Journal of Glaciology*, 11 (62), 205–214. doi: 10.3189/s002214300002219x

Singh, P., Spitzbart, G., Hübl, H., & Weinmeister, H. W. (1997). Hydrological response of snowpack under rain-on-snow events: a field study. *Journal of Hydrology*, 202(1-4), 1-20. doi: 10.1016/s0022-1694(97)00004-8

Singh, P., & Singh, V. P. (2001). Snow and Glacier Hydrology. Dordrecht (The Netherlands): Kluwer Academic Publishers, 742 pp.

Sorg, A., Bolch, T., Stoffel, M., Solomina, O., & Beniston, M. (2012). Climate change impacts on glaciers and runoff in Tien Shan (Central Asia). *Nature Climate Change*, 2(10), 725. doi: 10.1038/nclimate1592

Stahl, K., Moore, R. D., Shea, J. M., Hutchinson, D., & Cannon, A. J. (2008). Coupled modelling of glacier and streamflow response to future climate scenarios. *Water Resources Research*, 44(2), W02422. doi:10.1029/2007WR005956

Thayyen, R. J., & Gergan, J. T. (2010). Role of glaciers in watershed hydrology: a preliminary study of a Himalayan catchment. *The Cryosphere*, 4(1), 115-128. doi: 10.5194/tc-4-115-2010

Thompson, S. S., Benn, D. I., Dennis, K., & Luckman, A. (2012). A rapidly growing moraine-dammed glacial lake on Ngozumpa Glacier, Nepal. *Geomorphology*, 145/146, 1-11. doi: 10.1016/j.geomorph.2011.08.015

Vilímek, V., Emmer, A., Huggel, C., Schaub, Y., & Würmli, S. (2014). Database of glacial lake outburst floods (GLOFs)–IPL project No. 179. *Landslides*, 11(1), 161-165. doi: 10.1007/s10346-013-0448-7

Vuille, M., Francou, B., Wagnon, P., Juen, I., Kaser, G., Mark, B. G., & Bradley, R. S. (2008). Climate change and tropical Andean glaciers: Past, present and future. *Earth science reviews*, 89(3-4), 79-96. doi: 10.1016/j.earscirev.2008.04.002

Walder, J. S. (2010). Röthlisberger channel theory: its origins and consequences. *Journal of Glaciology*, 56(200), 1079-1086. doi: 10.3189/002214311796406031

Watanabe, T., Lamsal, D., & Ives, J. D. (2009). Evaluating the growth characteristics of a glacial lake and its degree of danger of outburst flooding: Imja Glacier, Khumbu Himal, Nepal. *Norwegian Journal of Geography*, 63, 255-267. doi: 10.1080/00291950903368367

Winkler, G., Wagner, T., Pauritsch, M., Birk, S., Kellerer-Pirklbauer, A., Benischke, R., Leis, A., Morawetz, R., Schreilechner, M. G., & Hergarten, S. (2016). Identification and assessment of groundwater flow and storage components of the relict Schöneben Rock Glacier, Niedere Tauern Range, Eastern Alps (Austria). *Hydrogeology Journal*, 24(4), 937-953. doi: 10.1007/s10040-015-1348-9

Worni, R., Huggel, C., & Stoffel, M. (2013). Glacial lakes in the Indian Himalayas—From an area-wide glacial lake inventory to on-site and modeling based risk assessment of critical glacial lakes. *Science of the Total Environment*, 468, S71-S84. doi: 10.1016/j.scitotenv.2012.11.043

Worni, R., Huggel, C., Clague, J. J., Schaub, Y., & Stoffel, M. (2014). Coupling glacial lake impact, dam breach, and flood processes: A modeling perspective. *Geomorphology*, 224, 161-176. doi: 10.1016/j.geomorph.2014.06.031

Yamada, T. (1998). Glacier lake and its outburst flood in the Nepal Himalaya. Japanese Society of Snow and Ice, Tokyo, 96 pp.

Yao, T., Pu, J., Lu, A., Wang, Y., & Yu, W. (2007). Recent glacial retreat and its impact on hydrological processes on the Tibetan Plateau, China, and surrounding regions. *Arctic, Antarctic, and Alpine Research*, 39(4), 642-650. doi: 10.1657/1523-0430(07-510)[yao]2.0.co;2

Zemp, M., Frey, H., Gärtner-Roer, I., Nussbaumer, S. U., Hoelzle, M., Paul, F., ... & Bajracharya, S. (2015). Historically unprecedented global glacier decline in the early 21st century. *Journal of Glaciology*, 61(228), 745-762. doi: doi.org/10.3189/2015jog15j017