FACTORS OF FORMATION AND DEVELOPMENT OF SUPRAGLACIAL LAKES AND THEIR QUANTIFICATION: A REVIEW

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ABSTRACT

Supraglacial lakes greatly affect the rate of glacier ablation and a potentially dangerous (GLOF – Glacier lake outburst flood) proglacial lake often forms through their development. The main part of the paper recapitulates the factors of the formation and drainage of supraglacial lakes, as well as the mechanisms of their development through a review of the scientific literature. In total there are five factors of the formation of supraglacial lakes and four factors (three of them alternative to one another) of the drainage. Three factors delimit the maximum extent of the emergence of supraglacial lakes, two of them determine the detailed distribution of localities suitable for hosting supraglacial lakes. The circumstances leading to the drainage mainly reflect the decisive role played by englacial voids. According to the current degree of scientific knowledge there are no factors controlling the development of supraglacial lakes. The complete process of the expansion of a supraglacial lake may be viewed as a positive feedback loop consisting of three major mechanisms. In the final part all of the factors are provided with quantitative intervals responding to the respective probability scales, which enable a relatively objective assessment of the probabilities of formation/drainage of supraglacial lakes. The most frequent application is the case of the assessment of the probability of the formation of a large supraglacial lake, due to its likely development into a proglacial lake.

Keywords: supraglacial lakes, glaciers, high mountain areas, GLOF

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1. Introduction

Supraglacial lakes are usually a small but very important phenomenon on a glacier, as shown in the following two paragraphs. In spite of their significance, only a few authors have discussed the evolution of supraglacial lakes in a global way and only Reynolds (2000) and Quincey et al. (2007) considered criteria for the assessment of supraglacial lakes. This study firstly aims to summarize, in the form of a review of scientific literature, the factors influencing the evolution of supraglacial lakes and the mechanisms and processes through which a supraglacial lake forms, expands and becomes extinct. In accordance with the theoretical background, the factors are given quantitative limit values that enable objective assessment of particular glacier reaches in respect of the emergence and development of supraglacial lakes.

Hutchinson (1957) regarded supraglacial lakes as transient, ephemeral phenomena that are interesting to a scientist only as limnological curiosities. He also quotes even older authors (e.g. Delebecque 1898; Collet 1925) who carried out a research of particular Swiss glaciers describing lake basins with a diameter reaching 130 m. Progressing climate change and the retreat of glacier termini showed the importance of supraglacial lakes and of the research carried out on them.

The formation and subsequent expansion of supraglacial lakes is regarded as a reaction of a glacier to the climate change (Benn et al. 2012; Xin et al. 2012). Supraglacial lakes and their outbursts substantially affect the net ablation rate of a glacier (Benn et al. 2000) having a great impact on its mass and water balance (Liu et al. 2013). The melting of ice cliffs exposed around supraglacial lakes accounts for large portions of the whole-glacier ablation rate, even though the ice cliffs cover only a small part of the surface of a glacier (Sakai et al. 1998, 2000). The evolution and expansion of a supraglacial lake finally results in its transition to a moraine- or bedrock-dammed proglacial lake (Komori 2008; Benn et al. 2012), which presents a significant threat to downstream situated areas with respect to many worldwide GLOF events (e.g. Lliboutry et al. 1977; Vuichard and Zimmerman 1987; Clague and Evans 2000). For this reason, Reynolds (2000) suggests identifying areas appropriate for the formation of large supraglacial lakes on glaciers with a negative mass balance, in order to start early remedial works that should prevent the storage of large volumes of meltwater.

1.1 Typology of glacial lakes

Glacial lakes may be classified with respect to many points of view. The simplest typology distinguishes scoured and dammed lakes. More detailed classification discriminates glacial lakes with respect to the material forming their dams: bedrock-, moraine-, and ice-dammed lakes (e. g. Emmer et al. 2014). For the purposes

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of this review the most appropriate approach is to classify glacial lakes according to their position relative to the glacier, i.e. proglacial lakes, located downstream of the glacier snout, and supraglacial lakes, that develop directly on the glacier (Gardelle et al. 2011). Logically, the second group should contain not only supraglacial lakes but also englacial and subglacial lakes emerging inside and under the glacier, respectively. In some cases, the lake may seem to be proglacial, but is still underlain by dead ice and should be classified as being supraglacial, such as Tsho Rolpa (Khumbu Himal, Nepal) (Chikita et al. 1999).

1.2 Typology of supraglacial lakes

There are two possible ways of classifying supraglacial lakes. Benn et al. (2012) distinguish two types of supraglacial lakes according to their relative elevation to the hydrological base of the glacier (the level at which meltwater leaves the glacier, usually the lowest point of the crest of the moraine dam) as perched and base-level lakes. Perched lakes are situated above the level of the hydrological base and persist only if their basin is formed by impermeable ice, otherwise their drainage takes place (Section 4.3). Their diameter seldom exceeds 100–200 m. The surface of base-level lakes, on the contrary, lies at the same level as the hydrological base. The existence of this type of supraglacial lakes depends on the stability of the dam and their length along the valley line may reach several kilometres (e. g. 3 km long Tsho Rolpa, Rolwaling Himal, Nepal) (Chikita et al. 1999).

A typology of Nakawo et al. (1997) or Takeuchi et al. (2012) is based on the existing/missing connection of supraglacial lakes to the glacier drainage system. Logically, base-level lakes can only be connected to the drainage system. In the case of perched lakes, both combinations are possible (Figure 1).

2. Methods

Information from various papers on high-mountain supraglacial lakes or on processes relating to them was gathered as a basis for this study. The absolute majority of the results of the field-work research published in the source articles was from on glaciers of high Asian mountain ranges (mainly the Himalayas, but also the Karakorum and the Tien-Shan), other destinations include the Southern Alps of New Zealand, the European Alps, and the mountain ranges of Alaska. The results of several laboratory and theoretical modelling studies were taken into account and a few review articles were also studied. The findings of the research of supraglacial lakes on the Greenland and Antarctic Ice Sheets were almost wholly omitted because substantial differences exist between lakes emerging on high-mountain glaciers and those forming on ice sheets. All of the information was analysed, critically assessed and categorized according to the main subject of the review.

Afterwards, each factor relevant for the formation/drainage was assigned a list of limit values. Published graphs and tables (ice density, surface gradient and debris cover thickness) were utilized when possible. The factors of surface depressions and surface outflow enable only a yes–no approach, which means they are either present or absent. Quantitative limit values of a particular factor generate a sequence of intervals with a corresponding succession of approximate probabilities. Particular probability scale bounds and also some of the limit values (e. g. the distance to the englacial void) were determined at least partly subjectively because of the lack of a background of empirical relations.

3. Factors of formation and distribution of supraglacial lakes on glaciers

There are three essential conditions for a supraglacial lake to form. First, the surface of a glacier must be impermeable for meltwater. During the process of firn compaction, the critical value of firn density is usually 0.80–0.83 g/cm^3 (Barnola et al. 1991). However, the firn-snow transition occurs at lower or higher densities in some cases, with the minimum and maximum density being 0.78 and 0.855 g/cm^3, respectively (Gregory et al. 2014). A lake also needs sufficient inflow of meltwater. Many supraglacial lakes emerge during the melt season and then freeze again in middle and high latitudes, whereas in the tropics the absence of thermal seasonality favours lake formation throughout the year. Finally, appropriate topography of the glacier surface enables the collection of meltwater. Shallow depressions serve as lake basins, so can the confluence of two or more glacier streams (Reynolds 2000), and if present debris cover can also trap water (Raymond and Nolan 2000). Occasionally, blocked englacial conduits may be exploited as a lake basin and sometimes a formerly drained lake basin can be re-filled by meltwater (Benn et al. 2000).

The maximum extent of the emergence of supraglacial lakes along the longitudinal profile of a glacier is constrained mainly by thermal conditions and debris cover. The highest position of the zero isotherm during the melt season defines the upper maximum level, and the glacier terminus the lower boundary of the possible formation of supraglacial lakes (Benn et al. 2012; Sakai 2012). Thick debris cover prevents melting on the lowermost part of glacier tongues. Thus, the lower end on debris-covered glaciers is shifted somewhat further from the terminus into the area, where the thickness of the debris layer gets thinner (Benn et al. 2012). For example, the initial development of Imja Tsho (Khumbu Himal, Nepal) in the 1960s followed this notion (Watanabe et al. 2009).

Liu et al. (2013) examined eight debris-covered glaciers in the Khan Tengri-Tomur Mountains (Tien-Shan,

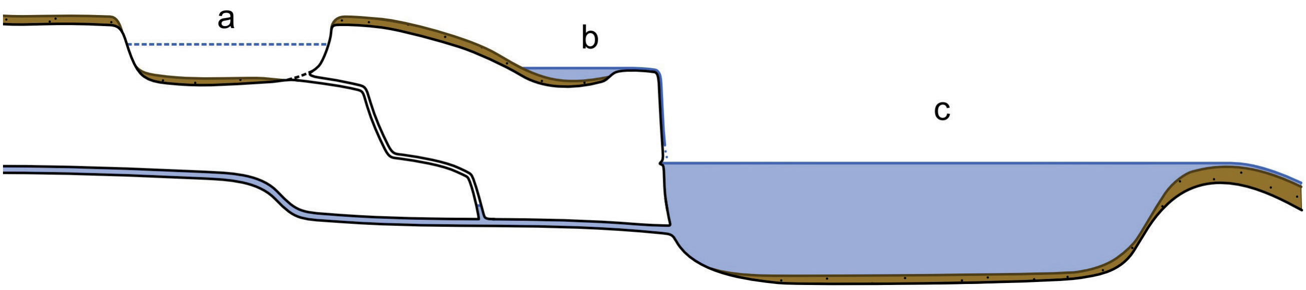

Fig. 1 Typology of supraglacial lakes. Drained perched lake formerly unconnected to the glacier drainage system (a). Perched lake connected to the drainage system (b). Base-level lake (c).

Central Asia) and their findings illustrate the impact of debris cover and mean air temperature on the distribution of supraglacial lakes quite well. In altitudinal distribution, supraglacial lakes appear 100 or 200 m above the glacier-snout level. The area of supraglacial lakes increases with altitude and thinning of the debris cover thickness, peaks near the altitude where small areas of clean ice emerge, and then gradually decreases with air temperature.

The distribution of supraglacial lakes on the glacier tongue is also influenced by the surface gradient. Based on the research carried out on several Bhutanese glaciers, Reynolds (2000) presents a simple relationship. A steeper gradient generally means a higher flow velocity, which then raises the probability of the opening of surface crevasses, and meltwater is more likely diverted off the glacier surface. Only a small number of supraglacial lakes form under such circumstances. When the values of the surface gradient exceed a particular limit no supraglacial lakes emerge since all of the meltwater is effectively drained downglacier. Reynolds (2000) also quantifies this relationship (Table 1).

More recent papers, however, question these intervals, especially the limit for the formation of supraglacial lakes. Salerno et al. (2012) applied the techniques of remote sensing to compile the database of all of the glacial lakes in the Sagarmatha National Park. An analysis of the obtained data suggests that supraglacial lakes may emerge on reaches sloping more than 10° but are usually short-lived. On the other hand, Salerno et al. (2012) confirm the value of 2° of the surface gradient as the upper threshold for the development of large supraglacial lakes. Liu et al. (2013) present similar results having used analogous methods in the Khan-Tengri Mountains. A small portion (14.3%) of supraglacial lakes was located on the glacier reaches, which sloped more than 10°. The majority (42.8%) of lakes lay in the zones with a surface gradient spanning from 2° to 6°. Baťka (2015) shows that the upper limit for the emergence of supraglacial lakes lies at 25°.

To summarize, there are relatively gentle reaches along the longitudinal profile of the glacier with the potential for the formation of supraglacial lakes alternating with icefalls where no lakes emerge regardless of other factors (temperature/debris cover).

Tab. 1 Relationship between the glacier surface gradient and supraglacial lake formation Reynolds (2000).

Surface gradient	Interpretation
0–2°	formation of large supraglacial lake over stagnant or very slow moving ice body forms from the merging of many smaller discrete ponds
2–6°	supraglacial ponds form, may also be transient locally, but sufficiently large areas affected by presence of ponds
6–10°	isolated small ponds may form, transient due to local drainage conduits opening and closing due to ice flow
>10°	all meltwater is able to drain away, no evidence of ponding

4. Mechanisms of growth of supraglacial lakes

4.1 Albedo change and ablation near the lake

The formation of a supraglacial lake means a change of surface material (from ice, snow, or debris to water surface, Table 2) and thus a drop in the value of albedo inside the polygon of this new lake (Reynolds 2000). The amount of absorbed heat rises through the higher input of the shortwave solar radiation, the main source of heat (Sakai et al. 2000). Due to this, the lake water temperature is kept a few degrees above the point of freezing for most of the year, and the lake actively ablates the ice that forms its basin and expands through the retreat of the sides of the basin (Reynolds 2000). Snowflakes melt in the lake water during snow storms, yet the nearby glacier surface becomes covered with a white blanket and the albedo difference peaks as the albedo of snow is the highest one. Supraglacial lakes also freeze later than their surroundings before or during winter and at the start of the ablation season the ice lid melts sooner than the adjacent surface (Reynolds 1981, 2000).

The change in albedo is important mainly in the case of debris-free glaciers, where supraglacial lakes present the principal means of glacier ablation (Komori 2008). On debris-covered glaciers, however, the difference in albedo of the water surface and debris is not as high as that of the former glacier type. Here, the thickness of the surface debris layer plays a substantial role as it lowers the albedo and isolates the underlying ice (Figure 2). The

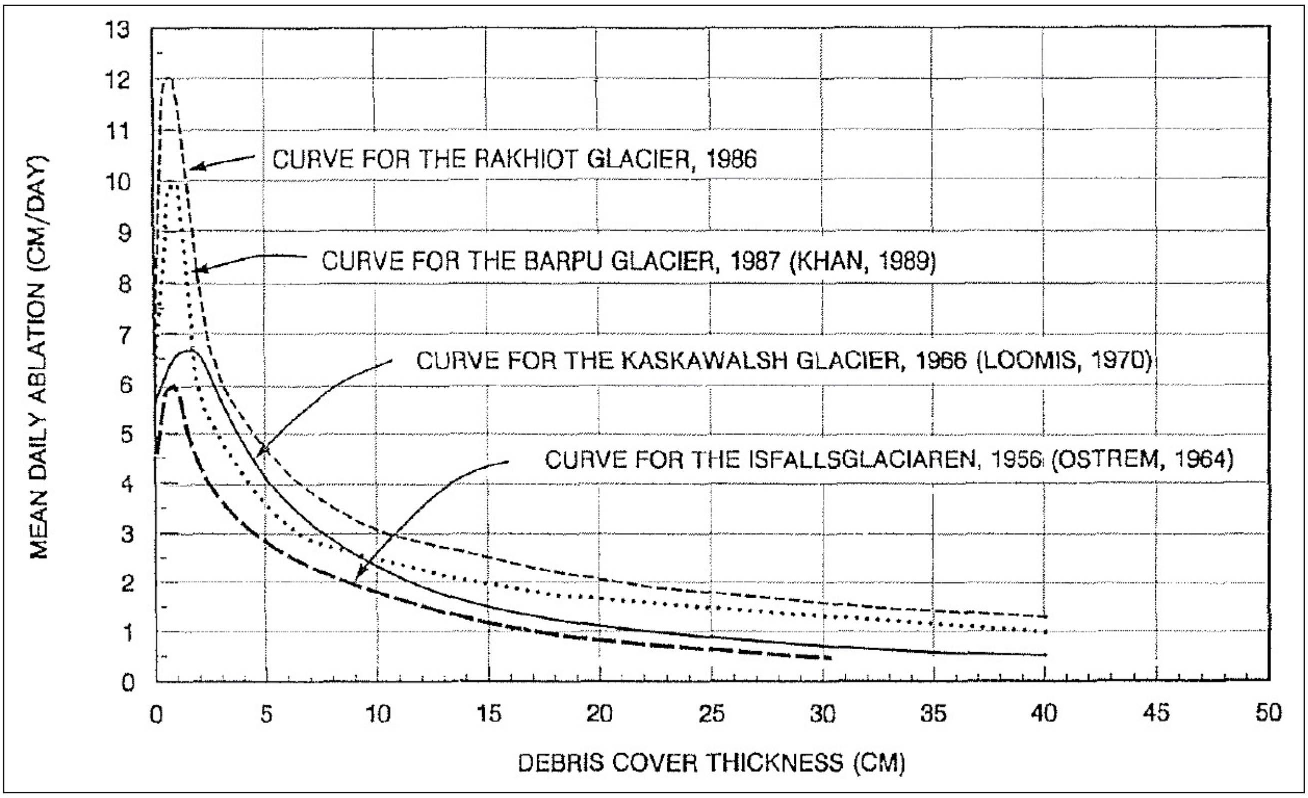

Fig. 2 Relationship between debris cover thickness and ablation rate (Mattson et al. 1993).

Tab. 2 Values for albedo of snow, firn, ice, water, and debris.

Surface type	Albedo	Source
Fresh dry snow	0.75–0.98	Cuffey and Paterson (2010)
Old clean wet snow	0.46–0.70	Cuffey and Paterson (2010)
Clean firn	0.50–0.65	Cuffey and Paterson (2010)
Debris-rich firn	0.15–0.40	Cuffey and Paterson (2010)
Clean ice	0.30–0.46	Cuffey and Paterson (2010)
Debris-rich ice	0.06–0.30	Cuffey and Paterson (2010)
Debris	0.15–0.25	Benn et al. (2012)
Water	0.10–0.35	Yamada (1998)

ablation rate firstly increases with the increasing thickness of the debris cover, reaching the maximum value at a thickness of approximately 1–2 cm (Mattson et al. 1993), because there is almost no insulation effect and the albedo of the wet and dirty ice can be as low as 0.06 (Cuffey and Paterson, 2010). After peaking, the ablation rate decreases exponentially because of the increasing influence of isolation by debris and the melt rate is always <1 cm/d under layers >0.5 m thick (Mattson et al. 1993). Every morning, a certain amount of time is needed to constitute stable heat flow through the debris cover so the period of day when ice ablation occurs is significantly shortened (Reznichenko et al. 2010).

These two counteracting effects of the debris cover mean that wherever the continuity of the debris layer is interrupted the melt rate substantially increases. This is the case of the nearby surroundings of supraglacial lakes. Immediately after its formation, the lake deepens and widens (due to the relatively low albedo of its surface). The slope angles of its basin are steepened to the point of the limit angle of debris repose, i.e. 30–40° (Sakai et al. 2000, Gardelle et al. 2011), the glacier surface gradually becomes debris-free and the lateral expansion of supraglacial lake accelerates.

There are two processes that somewhat slow the expansion. The former involves differential ablation of the slope (e.g. the slope of an ice cliff or, particularly in this paper, a slope forming the shore of a supraglacial lake). Due to topographic shading, places located higher on the slope receive greater amounts of incoming shortwave radiation than those near the lake surface. Thus, ablation is the highest at the top of the slope and decreases towards the bottom, which gradually reduces the slope angle back under the angle of repose. The process is most rapid on slopes oriented southeast to south in the northern hemisphere (Sakai et al. 1998, 2002). The latter of the two slowing processes is the cooling effect of meltwater inflow. Ablation around the supraglacial lake results in meltwater (temperature approximately 0 °C) flowing into the lake, where it mixes with the lake water and thus slightly reduces its temperature. However, the amount of heat from the absorbed incoming solar radiation substantially outpaces the rate of this cooling (Xin et al. 2012).

The rate of vertical expansion of a supraglacial lake on debris-covered glaciers progressively decelerates, as the ice initially forming the bottom of the lake becomes buried under the layer of debris formerly lying on the glacier

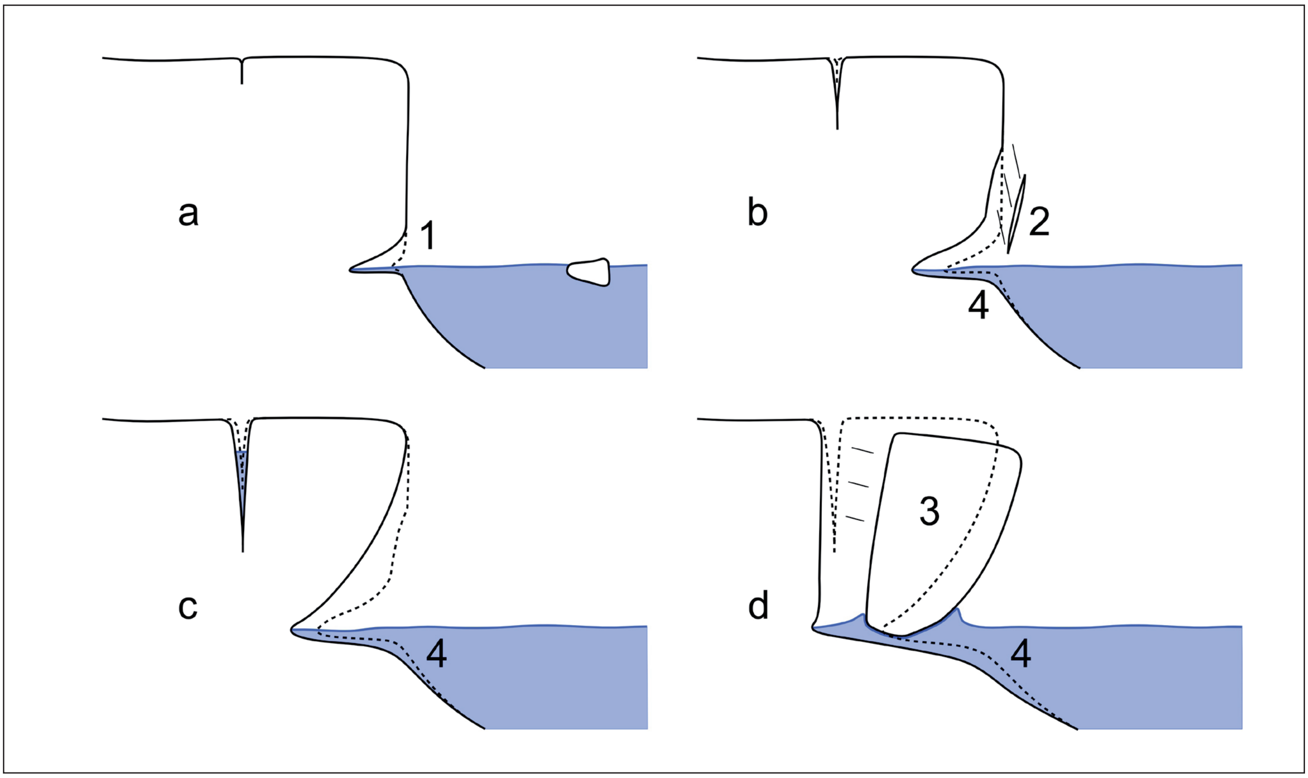

Fig. 3 Calving cycle (a–d) and calving types (1–4). Please, see text for description of sketches a–d. Calving at the waterline (1). Flake calving (2). Full-height slab calving (3). Subaqueous calving (4). (after Kirkbride and Warren 1997; Sakai 2012)

surface. In a similar way as the surface cover, the isolation effect of debris increases with increasing thickness. Chikita et al. (2000) described the probable final stadium of vertical expansion of a base-level lake. At the bottom of Imja Tsho (Khumbu Himal, Nepal) the medium consisting of debris, lake water and lake sediment separates the lake water and the dead ice. Very slow thermal conduction through the medium provides small amounts of heat for melting the dead ice. The meltwater mixes with the medium, pushes out approximately the same volume of water into the lake and the lake bottom subsides (Chikita et al. 2000).

4.2 Glacier calving and wind action

After attaining a diameter of approximately 30 m (Sakai et al. 2009) the lake volume is large enough for a new and more effective mechanism of expansion to commence, i.e. calving. After crossing this threshold, the expansion accelerates significantly (Röhl 2008; Benn et al. 2000). Kirkbride and Warren (1997) distinguish four calving types (Figure 3) in their study of the Maud Glacier (New Zealand): calving at the waterline, flake calving from the cliff face, full-height slab calving and subaqueous calving. The frequency decreases but the volume of calved ice increases from the waterline type towards subaqueous calving (Kirkbride and Warren 1997).

Calving at the water line forms and gradually widens a thermo-erosional notch that undercuts and destabilizes the ice cliff (**a** in Figure 3). Meanwhile, flake calving from the cliff face subsequently contributes to the process of notch enlargement, and a vertical fracture opens a few metres upglacier behind the cliff edge (**b** in Figure 3). Thus, the block is geometrically defined for full-height slab calving (**c** and **d** in Figure 3) (Kirkbride and Warren 1997). Diolaiuti et al. (2011) suggest the decisive role of fracture opening during calving, but Röhl (2006) and Xin et al. (2012) consider the horizontal notch the most important factor for subaerial calving.

Subaqueous calving, as described by Kirkbride and Warren (1997), involves detaching of subaqueous blocks and ramparts left by the subaerial calving cycles of full-height slabs, which is impossible in the case of supraglacial lakes. However, Röhl (2008) shows the importance of subaqueous melting of ice for the expansion of supraglacial lakes, also applying the term subaqueous calving for this process.

Continuing both lateral and vertical expansion brings a supraglacial lake to another threshold when the lake length exceeds approximately 50 m and the height of the usually present end moraine is sufficiently low. According to the model developed by Sakai et al. (2009), the wind speed along the water surface attains significant values (>1 m/s) and thus has a substantial impact on the lake water circulation. The valley wind makes the warmest surface water layer move towards the ice cliff (perched lake) or the active glacier front (base-level lake) where it accumulates, significantly accelerating the formation of the thermo-erosional notch and thus also calving (Sakai et al. 2009). The maximum height of the terminal moraine allowing this wind action probably depends on the length of the lake. Chikita et al. studied two great Nepalese supraglacial lakes – Tsho Rolpa (Chikita et al. 1999) and Imja Tsho (Chikita et al. 2000). Tsho Rolpa (3 km in length) was found to experience the circulation modelled by Sakai et al. (2009). In the case of Imja Tsho (1.3 km in length and the height of the end moraine 30 m), however, no such process was occuring.

4.3 Exploitation of englacial conduits

Thermal and mechanical incision of a supraglacial stream can produce an englacial cut-and-closure conduit when followed by roof closure. This tunnel persists if it is provided with a sufficient volume of meltwater by the source area of the glacier surface. Otherwise, glacier flow deforms, interrupts, and even fills certain reaches of the cut-and-closure conduit. The remaining unfilled parts become lines/zones of secondary permeability. Another possible way for secondary permeable structures to form is the downward propagation of water- or debris-filled crevasses, again followed by the roof closure (Gulley et al. 2009; Benn et al. 2012). Both cut-and-closure conduits and lines of secondary permeability are important means of drainage of perched supraglacial lakes (Section 5) and of base-level lake expansion.

After the formation of a base-level lake, the pattern of its expansion is determined by the presence of shallow englacial conduits (Benn et al. 2012). These conduits (and other voids) were originally located in somewhat bigger depths. Gradual downwasting of the glacier surface to the level of the hydrological base lowered the overlying ice and destabilized the ceilings of these structures. The roofs collapse and expose the debris-free ice on the sides of the conduits to incoming solar radiation. Former void spaces fill with meltwater and whole chains of new supraglacial lakes develop. After a certain period of time, all of the lakes coalesce into a single lake which then expands up- and downvalley through calving and dead-ice melt, respectively, and vertically through slow bottom subsidence (Section 4.1) (Benn et al. 2012).

5. Ways of the extinction of supraglacial lakes

Perched lakes drain off when their own expansion brings the basin to the vicinity of a permeable structure (englacial conduit, closed debris-filled crevasse etc. – Section 4.3) (lake "a" of Figure 1). The basin collapses and the lake water drains through the conduit into the lower-lying areas or into the englacial drainage system. In the case of crossing the line of secondary permeability, the partially closed voids are eroded by the relatively warm lake water and a new englacial conduit is formed. The probability of such a drainage increases with the expansion of the lake (Benn et al. 2012). After releasing most or all of the water, lake basins gradually cover with debris. If the cycle of formation and drainage of supraglacial lakes occurs frequently on a glacier, typical hummocky relief forms (Emmer et al. 2015).

The persistence of perched lakes connected to the glacier drainage system depends on the fragile balance between the downward lake expansion and erosion of the surface outflow channel, with the rate of meltwater inflow and outflow also being of importance (Raymond and Nolan 2000). Thermal erosion progressively enlarges the cross-section of the drainage channel. Unless the bottom of the lake basin subsides at a sufficient rate, the erosion of spillway results in an outburst (Raymond and Nolan 2000). In the case of englacial drainage conduit, the lake is likely be drained because of the concentration of relatively warm lake water in a narrow profile (Sakai et al. 2000).

Base-level lakes are incomparably longer lasting phenomena than mostly ephemeral perched lakes. Their life span is determined by the stability of the dam (Benn et al. 2012). A moraine dam may be overflowed by the lake water and/or incised by the water flowing through the outlet. This type of dam is therefore far less stable than those consisting of debris-covered dead ice, in which case only overflowing is possible (Benn et al. 2012; Hanisch et al. 1998).

6. Evolution to the stadium of a proglacial lake

As a rule, debris-free glaciers or those with a limited extent of debris-cover (usually cirque glaciers) cover a far smaller area than debris-covered ones, in most cases <10 km^2. Thus, there is enough space for expansion of only a single lake (Komori 2008). The lake grows through the processes described in Section 4, glacier downwasting may result in the transformation to a base-level lake and if a continuous end-moraine arc or a bedrock barrier is present in front of the glacier, the evolution continues until the stadium of a proglacial lake through the ice ablation in the entire zone (Benn et al. 2012). The whole process takes 20–30 years (Komori 2008).

Debris-covered glaciers, on the contrary, do not experience such a straightforward growth of a supraglacial lake. Komori (2008) discerns three stages of this development, which lasts about 50 years. Initially, many supraglacial lakes appear and expand on the lower part of the zone of ablation. Then, these lakes progressively coalesce into a single one. The evolution to this point lasts on average 10–20 years. Finally, the coalesced lake expands both up- and downglacier (Komori 2008).

Benn et al. (2012) outline a conceptual model of the evolution of the whole complex system, which a debris-covered glacier certainly is, also involving the expansion of supraglacial lakes. They define three regimes of glacier behaviour, transition to the next regime being caused by the crossing of a certain threshold. In Regime 1, the entire glacier is dynamically active, glacier flux

Fig. 4 Tsho Rolpa. The photograph was taken by Dr. N. Takeuchi in June 1994 (Sakai 2012).

compensates losses of ice in the ablation zone, meltwater is effectively drained away by the glacier drainage system and only ephemeral perched lakes emerge. Very few glaciers remain in Regime 1 in the Mount Everest region, one of them being the Kangshung Glacier (Benn et al. 2012).

Transition to Regime 2 is caused by the global warming. Such an external impulse is then amplified by several interconnected positive feedbacks. The upward shift of the rain-snow boundary changes the areas of ablation and accumulation zones, so that the ice inflow from upper portions of the glacier no longer balance the ablation in lower-lying areas. The resulting decline in the glacier surface gradient reduces the driving stresses forcing the ice flow into the lowermost parts of the ablation zone, which significantly stagnates. The decrease in the surface gradient also disrupts the previously efficient glacier drainage system. Both events enable the storage of large amounts of meltwater on the glacier and thus the formation of many perched lakes. Their gradual expansion (particularly by calving) again significantly raises the rates of ice-loss. Among the glaciers in the Sagarmatha region, the Ngozumpa Glacier shows most of the features characteristic of Regime 2 (Benn et al. 2012).

Further downwasting lowers the glacier surface below the level of the hydrological base and a base-level lake forms if a continuous moraine loop is present. Crossing this threshold also means a transition to Regime 3. A base-level lake rapidly expands both by calving and by exploiting the lines of shallow englacial conduits. Vertical expansion slowly reduces the thickness of the underlying (dead) ice, which eventually leads to a transition to a full-depth proglacial lake. The lower Imja Glacier and Trakarding Glacier are typical Regime 3-glaciers, hosting Imja Tsho and Tsho Rolpa, respectively (Figure 4) (Benn et al. 2012).

7. Quantification of factors

There are two papers focused, at least in part, on dividing the glacier surface into zones described by the distinct probability of the formation of supraglacial lakes. Reynolds (2000) considered only one factor – glacier surface gradient (Table 1). Quincey et al. (2007) developed on the previous notion by adding glacier velocity as an independent variable (Table 3) thus involving one of the processes responsible for the drainage of supraglacial lakes. These two ideas are very simple but strong, as we shall see soon.

The formation of supraglacial lakes is influenced by five factors: minimum air temperature, debris cover thickness, surface gradient of a glacier, ice density in situ, and the presence or absence of surface depressions (Table 4). The zones of glacier surface suitable for hosting supraglacial lakes are delimited by the position of the zero

Tab. 3 Relationship between glacier surface gradient, glacier velocity and supraglacial lake formation after Quincey et al. (2007).

	Surface gradient < 2°	Surface gradient > 2°
Stagnant ice	Minimal opportunity for reorganisation of drainage conduits, promoting large-scale lake development	No opportunity for reorganisation of drainage conduits through flow, but steeper hydraulic gradient aids drainage and lake development is unlikely
Measurable flow	Large lake likely but with a potential for drainage through the reorganisation of drainage conduits through ice flow	Opportunity for reorganisation of drainage conduits through flow and steeper hydraulic gradient aids drainage, resulting in most efficient drainage conditions so that lake development is least likely

Tab. 4 Factors of supraglacial lake formation and drainage; ice flow velocity and surface gradient are alternatives to the distance to englacial voids.

Factors of formation and drainage			Source of data
Formation	Surface gradient		remote sensing
	Debris cover thickness		remote sensing
	Minimum air temperature		field survey
	Ice density		field survey
	Presence/absence of surface depressions		field survey
Drainage	Distance to englacial voids		field survey
	OR	Ice flow velocity	remote sensing
		Surface gradient	remote sensing
	Presence/absence of surface outflow		remote sensing

isotherm, debris cover thickness, and surface gradient (Section 3). The data of debris thickness and surface gradient may be obtained through the analysis of remotely sensed images, whereas a presence in the field is required for obtaining the knowledge of the position of the zero isotherm. Inside previously defined zones, the detailed distribution of areas favourable for the emergence of supraglacial lakes is given by in-situ ice density and the presence of surface depressions – information accessible only through a time-expansive field survey.

No factors controlling the development of supraglacial lakes exist because the whole process of the growth from a minute pond to a great base-level lake may be considered as a complex positive feedback consisting of crossing sectional thresholds (Section 4). During the initial stage of development, the smallest ponds strictly follow the pattern determined mainly by the distribution of surface depressions and other meltwater traps, and also by ice density. When the lake area grows and supraglacial lakes coalesce, the importance of the position of the zero isotherm, debris thickness, and surface gradient rises. Thus, if one wants to find out the probable position of future large supraglacial lakes covering relatively great portions of a glacier, knowledge of the detailed structure of the zones defined by the three main factors is not required.

The development of a supraglacial lake may be interrupted or terminated at any moment by its partial or complete drainage, respectively. The circumstances relevant to the drainage are the distance from lake bottom to the nearest englacial void and the presence/absence of surface outflow channel (Table 4). The stability of the dam is not considered as it is usually one of the input characteristics for breach hazard assessment of a glacial lake. The data are obtained by field survey and analysis of satellite images, respectively. The distribution of englacial permeable structures may also be treated through considering ice flow velocity instead, which only requires satellite images. The argument runs that glacier flow subsequently reorganizes all the crevasses and voids and thus drives the drainage or survival of supraglacial lakes. As shown by Reynolds (2000), glacier flow velocity may then be represented by the surface gradient.

Quantification of factors may be applied in three cases. If one wants to determine the probability that supraglacial lakes emerge on a particular segment of a glacier currently free of lakes, the analysis includes most of or all the factors of supraglacial lake formation, depending on the temporal and financial circumstances. The features involved and their probability scales are summarised in Table 5.

When a supraglacial lake forms and develops further, an assessment of the probability of its drainage may be required. Except for the distribution of englacial voids in the vicinity of the lake, all of the data are collectable through relatively cheap remote sensing methods (Table 6).

When the probability of large supraglacial lake (usually base-level lake) formation is needed, as in the cases of Reynolds (2000) and Quincey et al. (2007), the analysis incorporates factors of both formation and drainage. As the researcher usually desires to use cost- and time-effective methods, the factors the information of which may be obtained through methods of remote sensing should only be used (Table 7). If there are other studies of the mountain range referring to the thermal conditions, the approximate position of the zero isotherm may be estimated and applied together with the other factors.

The theoretical likelihood of supraglacial lake formation/drainage is calculated through multiplication of individual factor probabilities, which are defined by the specific setting of these factors (i.e. their values) on a glacier.

8. Discussion

The majority of the above-mentioned factors affecting supraglacial lakes are quite well studied, such as the surface gradient of a glacier, debris cover thickness, or ice density, and some of them seem logical such as the mini-

Tab. 5 Analysis of supraglacial lake emergence: factors and probability scales, where Minimum T_{air} is minimum air temperature

Probability of lake formation	Ice density [g/cm³]	Minimum T_{air} [°C]	Surface gradient [°]	Debris cover thickness [m]	Surface depressions
1	>0.855	>0	<2	<0.05	present
0.75	0.830–0.855		2–10	0.05–0.10	
0.50	0.800–0.830	0	10–15	0.10–0.30	
0.25	0.780–0.800		15–25	0.30–0.50	
0	<0.780	<0	>25	>0.50	absent
Source	Gregory et al. (2014), Barnola et al. (1991)		Reynolds (2000), Baťka (2015)	Mattson et al. (1993)	Raymond and Nolan (2000)

Tab. 6 Analysis of supraglacial lake drainage: factors and probability scales, please remember that ice flow and surface gradient are alternatives to the distance to the nearest englacial void.

Probability of lake drainage	Distance to englacial void [m]	Ice flow [m/a]	Surface gradient [°]	Surface outflow
1	<0.5	>100	>10	
0.75	0.5–1	20–100	6–10	present
0.50	>1	5–20	2–6	
0	>5	0–5	<2	absent
Source	Benn et al. (2012)	Quincey et al. (2007), Cuffey and Paterson (2010)	Reynolds (2000)	Raymond and Nolan (2000)

Tab. 7 Analysis of large supraglacial lake formation: factors and probability scales.

Probability of lake formation	Surface gradient [°]	Debris cover thickness [m]	Ice flow [m/a]
1	<2	<0.1	0–5
0.50	2–6	0.1–0.3	5–20
0.25	6–10	0.3–0.5	20–100
0	>10	>0.5	>100
Source	Reynolds (2000)	Mattson et al. (1993)	Quincey et al. (2007), Cuffey and Paterson (2010)

mum air temperature or the impact of present surface outflow. The other influences bring significant uncertainties.

The impact of the glacier flow velocity on supraglacial lakes is only estimated. Quincey et al. (2007) discriminate between stagnant glaciers and glaciers with at least measurable motion, the threshold velocity lying at approximately 5 m/a. However, the behaviour of supraglacial lakes located in the area moving faster is only poorly understood, not to mention the maximum flow velocity at which the formation of supraglacial lake is possible. Cuffey and Paterson (2010) use the term fast-flowing for glaciers flowing at velocities exceeding 100 m/a.

Englacial permeable structures cause the vast majority of drainages of perched supraglacial lakes (Benn et al. 2012). The precise critical thickness of ice between lake bottom and the englacial void at which the lake bottom collapses depends highly on the strength and therefore on the internal structure of the ice. Thus, the defined limit values for the distance to englacial voids are only tentative and express the author's subjective opinion (based on the studied literature).

Supraglacial lakes emerge in previously formed depressions on the glacier surface. One of the leading roles is certainly played by glacier flow, which warps the glacier surface, especially near icefalls. Surprisingly, no study concerning such an essential process exists and only a few papers mention minor contributions to the formation of depressions that are made by surface debris cover (Raymond and Nolan 2000), blocked englacial conduits (Benn et al. 2000) and even confluences of glacier streams (Reynolds 2000).

9. Conclusions

Supraglacial lakes are a complex and, thanks to their linkage to glacier mass balance and their possible transition to potentially dangerous proglacial lakes, important phenomenon. However, the importance of supraglacial lakes is slightly underestimated and the processes relating to them somewhat poorly understood, as it has been demonstrated by the problems with precise quantification of certain thresholds.

The emergence of supraglacial lakes is controlled by at least five factors divided into two levels. Debris cover thickness, position of the zero isotherm and surface gradient define zones of the glacier surface generally suitable for the formation of supraglacial lakes, whereas ice density in situ and particularly the distribution of surface depressions determine the detailed pattern of possible future supraglacial lakes. In the initial phases of development of supraglacial lakes, the locations of lake surfaces strictly follow this pattern. However, great supraglacial lakes cover large areas of the former glacier surface and respect mainly three general factors. Drainage may terminate the development of supraglacial lakes at any moment, so the distribution of englacial voids (the distance from the lake bottom to the nearest englacial void) and the presence/absence of surface outflow also play a significant role in the overall analysis.

Quantification of factors that affect supraglacial lakes may be applied in three cases: the emergence and drainage of supraglacial lakes, and the formation of a large supraglacial lake. Reynolds (2000) and then Quincey et al. (2007) give possible ways of assessing the probability of the formation of a large supraglacial lake, only leaving out the factor of debris thickness. The thresholds of certain factors (i.e. the distance to englacial voids and glacier flow velocity) should only be treated as approximate values because there is currently no method for their precise quantification. This uncertainty constitutes only a minor obstacle as the distance to englacial voids may be represented quite well by the glacier flow velocity and in turn by the surface gradient.

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REFERENCES

BARNOLA, J.M., PIMIENTA, P., RAYNAUD, D., KOROTKEVICH, Y. S. (1991): CO2-climate relationship as deduced from the Vostok ice core: a re-examination based on new measurements and on re-evaluation of the air dating. Tellus 43B, 83–90. http://dx.doi.org/10.1034/j.1600-0889.1991.t01-1-00002.x

BAŤKA, J. (2015): The analysis of supraglacial lakes: Case study from the Cordillera Blanca, Peru. Prague, Bc thesis, Charles University in Prague.

BENN, D. I., WISEMAN, S., WARREN, C. R. (2000): Rapid growth of a supraglacial lake, Ngozumpa Glacier, Khumbu Himal, Nepal. In: Nakawo, M., Raymond, C.F. & Fountain, A. (eds.): Debris-Covered Glaciers, Proceedings of the Seattle Workshop, IAHS Publ. 264, 177–185.

BENN, D. I., BOLCH, T., HANDS, K., GULLEY, J., LUCKMAN, A., NICHOLSON, L. I., QUINCEY, D. J., 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, 156–174. http://dx.doi.org/10.1016/j.earscirev.2012.03.008

CHIKITA, K., JHA, J., YAMADA, T. (1999): Hydrodynamics of a Supraglacial Lake and its Effects on the Basin Expansion: Tsho Rolpa, Rolwaling Valley, Nepal Himalaya. Arctic, Antarctic, and Alpine Research 31(1), 58–70. http://dx.doi.org/10.2307/1552623

CHIKITA, K., JOSHI, S. P., JHA, J., HASEGAWA, H. (2000): Hydrological and thermal regimes in a supra-glacial lake: Imja, Khumbu, Nepal Himalaya. Hydrological Sciences Journal 45(4), 507–521. http://dx.doi.org/10.1080/02626660009492353

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

COLLET, L. W. (1925): Les Lacs, leur mode de formation, leurs eaux, leur destin, éléments d'hydrogéologie. Paris, Doin. (cit. in Hutchinson 1957)

CUFFEY, K. M., PATERSON, W. S. B. (2010): The Physics of Glaciers, 4th edition. Academic Press, Amsterdam.

DELEBECQUE, A. (1898): Les Lacs Français. Paris, Chamerot et Renouard. (cit. in Hutchinson 1957)

DIOLAIUTI, G., KIRKBRIDE, M. P., SMIRAGLIA, C., BENN, D. I., D'AGATA, C., NICHOLSON, L. (2005): Calving processes and lake evolution at Miage glacier, Mont Blanc, Italian Alps. Annals of Glaciology 50, 207–214. http://dx.doi.org/10.3189/172756405781813690

EMMER, A., VILÍMEK, V., KLIMEŠ, J., COCHACHIN, A. (2014): Glacier Retreat, Lakes Development and Associated Natural Hazards in Cordillera Blanca, Peru. In: Shan, W. (ed.): Landslides in Cold Regions in the Context of Climate Change. Environmental Science and Engineering, 231–252.

EMMER, A., LOARTE, E. C., KLIMEŠ, J., VILÍMEK, V. (2015): Recent evolution of the bent Jatunraju glacier (Cordillera Blanca, Peru). Geomorphology 228, 345–355. http://dx.doi.org/10.1016/j.geomorph.2014.09.018

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, 47–55. http://dx.doi.org/10.1016/j.gloplacha.2010.10.003

GREGORY, S. A., ALBERT, M. R., BAKER, I. (2014): Impact of physical properties and accumulation rate on pore close-off in layered firn. The Cryosphere 8, 91–105. http://dx.doi.org/10.5194/tc-8-91-2014

GULLEY, J. D., BENN, D. I., SCREATON, E., MARTIN, J. (2000): Mechanisms of englacial conduit formation and their implications for subglacial recharge. Quaternary Science Reviews 28, 1984–1999. http://dx.doi.org/10.1016/j.quascirev.2009.04.002

HANISCH, J., DELISLE, G., POKHREL, A. P., DIXIT, A. A., REYNOLDS, J. M., GRABS, W. E. (1998): The Thulagi glacier lake, Manaslu Himal, Nepal – Hazard assessment of a potential outburst. In: Moore, D. & Hungr, O. (eds.): Proceedings of the Eighth International IAEG Congress, 21st–25th September 1998, Vancouver, Canada, 2209–2215.

HUTCHINSON, G. E. (1957): A Treatise on Limnology. Volume 1: Geography, physics, and chemistry. New York, John Wiley & Sons, Inc.

KIRKBRIDE, M. P., WARREN, C. F. (1997): Calving processes at a grounded ice cliff. Annals of Glaciology 24, 116–121.

KOMORI, J. (2008): Recent expansion of glacial lakes in the Bhutan Himalayas. Quaternary International 184, 177–186. http://dx.doi.org/10.1016/j.quaint.2007.09.012

LIU, Q., MAYER, C., LIU, S. (2013): Distribution and recent variations of supraglacial lakes on detritic-type glaciers in the Khan Tengri-Tomur Mountains, Central Asia. The Cryosphere Discussions 7, 4545–4584. http://dx.doi.org/10.5194/tcd-7-4545-2013

LLIBOUTRY, L., ARNO, B. M., PAUTRE, A., SCHNEIDER, B. (1977): Glaciological problems set by the control of dangerous lakes in Cordillera Blanca, Peru. I. Historical failures of morainic dams, their causes and prevention. Journal of Glaciology 18(79), 239–254.

MATTSON, L. E., GARDNER, J. S., YOUNG, G. J. (1993): Ablation on Debris Covered Glaciers: an Example from the Rakhiot

Glacier, Punjab, Himalaya. In: Young, G. J. (ed.): Snow and Glacier Hydrology, Proceedings of the Kathmandu Symposium, November 1992, IAHS Publ. 218, 289–296.
NAKAWO, M., FUJITA, K., AGETA, Y., SHANKAR, K., POKHREL, A. P., TANDONG, Y. (1997): Basic studies for assessing the impacts of the global warming on the Himalayan cryosphere, 1994–1996. Bulletin of Glacier Research 15, 53–58.
RAYMOND, C. F., NOLAN, M. (2000): Drainage of a glacial lake through an ice spillway. In: Nakawo, M., Raymond, C. F., Fountain, A. (eds.): Debris-Covered Glaciers, Proceedings of the Seattle Workshop, IAHS Publ. 264, 199–207.
REYNOLDS, J. M. (1981): Lakes on George VI Ice Shelf, Antarctica. Polar Record 20(128), 425–432. http://dx.doi.org/10.1017/S0032247400003636
REYNOLDS, J. M. (2000): On the formation of supraglacial lakes on debris-covered glaciers. In: Nakawo, M., Raymond, C.F. & Fountain, A. (eds.), Debris-Covered Glaciers, Proceedings of the Seattle Workshop, IAHS Publ. 264, 153–161.
REZNICHENKO, N., DAVIES, T., SCHULMEISTER, J., MCSAVENEY, M. (2010): Effect of debris on ice-surface melting rates: an experimental study. Journal of Glaciology 56(197), 384–394. http://dx.doi.org/10.3189/002214310792447725
RÖHL, K. (2006): Thermo-erosional notch development at fresh-water-calving Tasman Glacier, New Zealand. Journal of Glaciology 52(177), 203–213. http://dx.doi.org/10.3189/172756506781828773
RÖHL, K. (2008) Characteristics and evolution of supraglacial ponds on debris covered Tasman Glacier, New Zealand. Journal of Glaciology 54(188), 867–880. http://dx.doi.org/10.3189/002214308787779861
SAKAI, A., NAKAWO, M., FUJITA, K. (1998): Melt rate of ice cliffs on the Lirung Glacier, Nepal Himalayas, 1996. Bulletin of Glacier Research 16, 57–66.
SAKAI, A., TAKEUCHI, N., FUJITA, K., NAKAWO, M. (2000): Role of supraglacial ponds in the ablation process of a debris-covered glacier in the Nepal Himalayas. In: Nakawo, M., Raymond, C. F., Fountain, A. (eds.): Debris-Covered Glaciers, Proceedings of the Seattle Workshop, IAHS Publ. 264, 119–130.
SAKAI, A., NISHIMURA, K., KADOTA, T., TAKEUCHI, N. (2009): Onset of calving at supraglacial lakes on debris-covered glaciers of the Nepal Himalaya. Journal of Glaciology 55(193), 909–917. http://dx.doi.org/10.3189/002214309790152555
SAKAI, A. (2012): Glacial Lakes in the Himalayas: A Review on Formation and Expansion Processes. Global Environmental Research 16, 23–30.
TAKEUCHI, N., SAKAI, A., KOHSHIMA, S., FUJITA, K., NAKAWO, M. (2012): Variation in Suspended Sediment Concentration of Supraglacial Lakes on Debris-covered Area of the Lirung Glacier in the Nepal Himalayas. Global Environmental Research 16, 95–104.
VUICHARD, D., ZIMMERMANN, M. (1987): The 1985 Catastrophic Drainage of a Moraine-Dammed Lake, Khumbu Himal, Nepal: Cause and Consequences. Mountain Research and Development 7(2), 91–110. http://dx.doi.org/10.2307/3673305
XIN, W., SHIYIN, L., HAIDONG, H., JIAN, W., QIAO, L. (2012): Thermal regime of a supraglacial lake on the debris-covered Koxkar Glacier, southwest Tianshan, China. Environmental and Earth Sciences 67, 175–183. http://dx.doi.org/10.1007/s12665-011-1490-1
YAMADA, T. (1998): Glacier Lake and its Outburst Flood in the Nepal Himalaya. Tokyo, Japanese Society of Snow and Ice.

RESUMÉ

Faktory vzniku a vývoje supraglaciálních jezer a jejich kvantifikace: rešerše

Supraglaciální jezera (jezera vznikající přímo na ledovci) se v posledních dvou desetiletích dostala do popředí vědeckého zájmu. Byl doceněn jejich význam pro odtávání (a tím i hmotovou bilanci) ledovců – jedná se tedy o možné indikátory změny klimatu. Díky jejich vývoji v úplnosti zakončeném stádiem proglaciálního jezera, potenciálně nebezpečného pro obyvatelstvo níže položených částí údolí, se začíná uvažovat o monitoringu plošně rozsáhlých supraglaciálních jezer. K tomuto aktuálnímu tématu (po obsáhlé rešerši vědecké literatury na téma faktorů ovlivňujících vznik, vývoj a zánik supraglaciálních jezer) přispívá i prezentovaný článek.

Vznik supraglaciálních jezera vyžaduje poměrně nízký sklon povrchu ledovce (do ~20°), teplotu nad bodem mrazu a ledovec bez suťového pokryvu (případně s málo mocným suťovým pokryvem, do 0,5 m). Uvedené tři okolnosti určují úseky ledovce příhodné pro tvorbu supraglaciálních jezer. Detailní rozložení jezer ovlivňují zejména kolísání hustoty ledovcového ledu/firnu a přítomné deprese na povrchu ledovce.

Ihned po svém vzniku supraglaciální jezero pomalu expanduje (voda díky nízkému albedu absorbuje více slunečního záření než okolní povrch). Pokud má jezero větší průměr než 30–100 m, zajišťuje rozšiřování jezerní pánve proces výrazně účinnější než ten předchozí – telení ledových bloků tvořících boky jezerní pánve. Za příhodné konfigurace okolního povrchu se od rozměrů jezera překračujících 500 m přidává i vliv větru urychlující telení. Jakmile hladina jezera klesne na úroveň hydrologické báze ledovce (definovaná lokalizací výtoku tavné vody z ledovce), dosáhne rychlost expanze supraglaciálního jezera svého maxima, jezero záhy pokrývá celou šíři ledovcového splazu a mění se na typ proglaciální (pokud je přítomen val čelní morény či skalní bariéra).

Podle relativní elevace vzhledem k úrovni hydrologické báze ledovce lze supraglaciální jezera rozdělit na jezera vyvýšená (orig. "perched lakes", nacházejí se nad úrovní báze) a jezera v úrovni hydrologické báze (orig. "base-level lakes", hladina v úrovni odtoku tavné vody). Rozhodujícím faktorem zániku supraglaciálních jezer vyvýšených je rozložení englaciálních tunelů, nedokonale uzavřených fraktur a dalších prostor uvnitř ledovce vyplněných jiným materiálem než ledovcovým ledem. Jakmile se dno pánve dostane během expanze jezera do blízkosti jedné z uvedených struktur, dojde k propadnutí dna a výtoku vody do nitra ledovce. Další okolností zvyšující riziko výtoku jezera je přítomnost povrchového či podpovrchového odtokového kanálu. Supraglaciální jezera v úrovni hydrologické báze přetrvávají, dokud nedojde k porušení jejich hráze (ledovcový led, morénový materiál, skalní bariéra).

Uvedené faktory vzniku a zániku supraglaciálních jezer byly v závěrečné části kvantifikovány prostřednictvím série mezních hodnot určujících pro každý faktor navazující posloupnost intervalů, ke kterým byly přiřazeny pravděpodobnostní stupnice. S pomocí provedené kvantifikace lze pro daný úsek ledovce realizovat objektivní zhodnocení pravděpodobností vzniku supraglaciálních jezer obecně a vzniku velkého supraglaciálního jezera (obvykle v úrovni hydrologické báze) a pro dané supraglaciální jezero pravděpodobnost jeho zániku (tj. výtoku). Zejména zhodnocení možnosti vytvoření velkého supraglaciálního jezera má význam při výběru oblastí pro monitoring, vzhledem k možnému vzniku jezera proglaciálního. Druhé dva způsoby použití kvantifikovaných faktorů lze zakomponovat např. do modelování ablace ledovce.

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