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Evolution of a High-mountain Thermokarst Lake in the Swiss Alps*

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Abstract

Thermokarst lakes , a characteristic landscape element of the Arctic , are rarely found outside arctic situations. Here, a 30-yr photogrammetric monitoring series of a thermokarst lake in the Gruben area, Swiss Alps, is presented. The lake, situated in an environment of dead-ice remains and creeping permafrost, reached a final size of ca. 10,000 m^2 in area and 50,000 m^3 in volume before it had to be drained artificially in 1995. Starting in the mid-1960s it grew with radial rates of ca. 1.5 to 5 m y^{-1}. Nonlinear coupling of lake diameter and energy turnover led to accelerated area growth. The development of the lake was presumably driven by thermal convection of water. By a dynamic model of lake growth we show that a change in climate conditions and/or the lake bottom topography could have significantly influenced the observed lake growth. The effective energy turnover used for ice-melt and subsequent lake growth was estimated to be in the order 10^0 to 10^1 W m^{-2}.

Introduction

The term "thermokarst" refers to characteristic landforms which result from thawing of ice-rich permafrost or the melting of massive ice (French, 1976; Washburn, 1980; van Everdingen, 1998). A thermokarst lake (also called thaw lake) occupies a depression formed by thaw settlement. Typically, such lakes may expand by thermokarst forming processes (Wallace, 1948; Hopkins, 1949; Williams and Smith, 1989). Even small disturbances, especially on the surface, can initiate thermokarst processes and create related lakes. Water pools in a depression and begins to thaw the permafrost beneath. As thaw continues along lake margins, the basin extends. Since underlying permafrost prevents percolation into underground aquifers, water is trapped until the lake drains. Thaw extension continues until intervening higher ground is breached, creating an outlet channel.

In connection with ground-ice occurrences, the freeze-thaw transition is a major threshold that, once crossed, may lead to the development of various landforms, some of which (e.g., thermokarst lakes) are irreversible at least on time scales of less than centuries. Thermokarst and related lakes are closely linked to the ground thermal regime and, thus, especially sensitive to changes in boundary conditions. The existence of thermokarst lakes is a clear expression of ground conditions, e.g., thermal regime and ice content. Even small changes in boundary conditions such as climate can be made clearly visible by thermokarst processes. Thermokarst lakes can, therefore, be seen as geo-indicators (Romanovskii et al., 1996).

Thermokarst lakes are most numerous, and often the characteristic landscape element, in areas with ice-rich ground and low relief (i.e., in arctic river flats under continuous permafrost conditions). Due to the rough relief and/or discontinuous permafrost conditions, such lakes are rarely found in non-arctic high-mountain environments. The largest concentration of such lakes outside the Arctic might be found on the debris-covered glaciers in the Himalaya (e.g., Benn et al., 2000; Sakai et al., 2000). Here we present the results of a 30-yr monitoring series of a thermokarst lake in the Swiss Alps and investigate its growth using a dynamic model.

Study Site

The thermokarst lake under study is situated in the Gruben area, Saas valley, canton of Valais, Switzerland (Fig. 1, 7°58' E, 46°10' N). Within a distinct cirque and roughly parallel to the orographic right side of the terminus of the polythermal Gruben glacier, a rock glacier (i.e., a body of all-year-round frozen and creeping debris) developed at an altitude of about 2900 m a.s.l. within thick accumulations of debris and morainic material in an environment of discontinuous mountain permafrost. Several lakes are located at the margin as well as in front of the glacier tongue (glacier terminus) and also on the rock glacier. In 1968 and 1970, outburst floods from a glacier-dammed lake (lake 3) in the contact zone between the glacier tongue and the rock glacier caused extensive damages to the nearby village of Saas Balen. In order to prevent further floods and debris flows, geophysical and geodetic investigations were carried out, and photogrammetric monitoring was started based on large-scale aerial photography annually taken by the Swiss Federal Office of Cadastral Surveys (Kääb et al., 1997; Kääb, 2001; Haeberli et al., in press).

During its Holocene and historical advances, Gruben glacier must have repeatedly overridden the upper part of the rock glacier (Kääb et al., 1997). During glacier retreat periods, debris-covered dead ice (i.e., ice decoupled from the originating glacier) was deposited on top of the rock-glacier permafrost, affecting the dynamics of the rock glacier in its upper and orographic left part. Characteristic mean annual surface temperatures of the rock glacier are close to –1°C today. The thickness of the permafrost is known to be up to 100 and more meters in the lower part, and characteristically a few tens of meters in the glacier-affected part (Haeberli, 1985; King et al., 1987).

In the mid-1960s, two depressions formed on the rock glacier, filled with water and progressively enlarged. One of these two thermokarst lakes, lake 6, only fills up during winter and spring. In late autumn, freezing water seals up the lake bottom and, subsequently, the lake fills with snow, water, and ice during winter and spring. In late spring, when the thermokarst lake is completely full with water, the frozen bottom becomes leaky and the lake drains slowly.

In contrast, permafrost and dead-ice damming the other thermokarst lake, lake 5 (Fig. 2), which is under study here, appeared to be impermeable up to the water level which remains nearly stable throughout the year. In 1994, lake 5 had a volume of about 50000 m^3. The annual growth rate had reached values of 7500 m^3 a^{-1} (Kääb et al., 1996). In view to the increasing risk of a lake outburst endangering the village of Saas Balen, most of the water within this thermokarst lake was pumped out in the late autumn of 1995 as part of a flood protection concept for the whole Gruben area (Kääb et al., 1996; Haeberli et al., in press). The slowly refilling lake 5 was definitively drained in 1997 by the construction of a roughly 10-m-deep V-shaped ditch. The formation of this ditch was initiated by artificial excavation of a slightly inclined channel down to the lake level and completed by downcutting through massive ground ice by the outflowing water (Haeberli et al., in press).

Observed Growth of the Thermokarst Lake

The push moraine (i.e., a frozen moraine deformed by glacier advance) at the former line of contact between the glacier and the rock glacier is creeping and sliding backwards into the topographic depression which was left by the glacier retreat during the past decades (Fig. 1; Kääb et al., 1997). Due to the resulting stresses crevasses formed in the mid-1960s at the upper margin of the dead-ice occurrences. One of these elongated depressions, filled with snow and water, initiated the investigated thermokarst lake 5. Photogrammetric mapping (Haeberli, 1980; Kääb, 1996) shows that the surface of the lake grew at an accelerating rate and asymmetrically towards the south, exposing there an increasingly high ice front (Figs. 2, 3). In 1994, the lake reached an area of 10,000 m^2 (Figs. 3, 4) and a volume (at this time derived from radio-echo soundings) of some 50,000 m^3. The radial growth rate of the lake is similar to the one observed for arctic thermokarst lakes (Burn, 1990). After lake drainage, a digital elevation model of the lake bottom was derived photogrammetrically (Fig. 4).

In the 1980s, the lake level smoothly dropped twice by several meters (Fig. 3) without leaving any traces of surface runoff. These changes in lake level pointed to the existence of crevasses or caves in the surrounding ice and permafrost enabling the partial drainage. In fact, an ice-cave west of the lake found after pumping out the lake in 1995, may have been related to the lake level lowering in the 1980s. However, the corresponding water losses did not significantly slow down the overall growth rate of the lake (Fig. 3). After being pumped out in 1995 the lake refilled to a large extent until 1997, presumably by precipitation and melt water input from the catchment area (approximately 20,000 m^2; cf. Fig. 4). Despite such water supply, the lake level was more or less stable before 1985 and after 1989 which points to a continuous natural subsurface outlet which obviously changed in the mid-1980s.

With increasing area the lake surface was more and more free of ice during the summer and the water warmed up under the influence of solar radiation and high air temperatures. As shown by strong undercutting of the ice front (Fig. 5) and drifting of small icebergs away from it (Fig. 2), the density increase of water due to warming between 0°C and up to 4°C presumably triggered a process of thermal convection: warmer and denser surface water sinks to the lake bottom, flows on the inclined bottom towards the ice front, cools by ice melting and undercutting the ice front, thereby reduces its density and rises again to the lake surface where it flows away from the ice front and warms up again (Fig. 5). In fact, after drainage of the lake, strong undercutting and vertical pear-like convexities of about 1 m height and some decimeters depth were found in the exposed ice front, presumably related to local thermal convection cells (Figs. 6, 7). The lower, sharply defined butts of the convexities were situated in the deeper part of the ice-front. The upper end consisted in channels smoothly thinning and flattening towards the former lake surface. As we assess from the overall melt rate of the ice-front (cf. also the following section), the water- (and energy-) flow of the related convection cells must have been more or less stable during some hours or few days to melt out the volume of one convex form. Older convexities were overlain by recently active ones (Fig. 6).

The above overall water convection was strengthened by a positive feed-back in that melting of ice leads to a larger surface area of the lake, thus to increased energy input and more melting, etc. This positive feed-back, together with the existence of massive ground ice, represents the main cause for the accelerating lake growth. During the early 1990s, the last years of its existence, the lake grew annually by about 1500 m^2 in surface area and 7500 m^3 in volume. Photogrammetrically measured differential melt, subsidence, and flow rates of the rock glacier (Fig. 1; Kääb et al., 1997) indicated that massive ground ice at the southern lake border continued right into the steeper slope where creep is accelerating (extending flow). Extrapolated lake growth would have reached this critical area with potential crevasse formation and fast lowering of the southern lake shore in the years 1998 to 2000 at the latest (cf. Fig. 3).

The growth of lake 5 was significantly more steady than the often erratic and faster development observed for thermokarst lakes situated on nonarctic active glaciers (supraglacial lakes; Benn et al., 2000; Sakai et al., 2000). This fact can presumably be attributed to thermal and dynamic stabilization by permafrost as present in our study (i.e., ground temperature < 0°C) in contrast to the temperate ice of most alpine glaciers (i.e., ice temperature = 0°C). Investigations on water fluxes within a supraglacial lake by Chikita et al. (1999) show that the above described solely thermally driven convection can be disturbed or even inverted by wind or density anomalies of sediment loaded water input.

Model of the Thermokarst Lake

A geometric model was set up in order to extrapolate the lake development in view of hazard prevention and to understand its accelerating growth. Thereby, the lake was approximated by a cylinder wedge with horizontal surface and the bottom inclined by a constant angle α (here: 10°, ca. 0.17 rad), and a radius r increasing with time (Fig. 8). The geometric lake volume V (eq. 1) equals the total amount of ice $V(t)$ melted until the time t (eq. 2):

$$V = \frac{2}{3} r^3 \alpha \qquad (1)$$

$$V(t) = \frac{1}{2} r^2 \pi \frac{Q_{eff}}{q\rho} t \qquad (2)$$

with the specific melt energy of ice q (334 kJ kg^{-1}), the density of ice ρ (0.9 kg dm^{-3}), and the effective melt energy Q_{eff} per unit time and per unit surface area. Considering that the lake bottom consists in permafrost and the lake borders are in ice-cemented debris or dead ice within permafrost, we assume in equation (2) that the melt-energy input into the lake (Q_{eff}) is transferred only via the lake surface (0.5 r^2 π).

Differentiating equation (1) for t provides

$$\frac{dV}{dt} = 2\alpha r^2 \frac{dr}{dt} \qquad (3)$$

and differentiating equation (2) for t

$$\frac{dV}{dt} = \frac{1}{2} r^2 \pi \frac{Q_{eff}}{q\rho} \qquad (4).$$

Both Equations (3) and (4) together give the differential equation

$$\frac{dr}{dt} = \frac{Q_{eff}\pi}{4q\rho\alpha} \quad (\approx c \ = const.) \qquad (5).$$

Assuming, in a first step, q, ρ, α, and Q_{eff} being constant with time the rate of lake growth dr/dt is constant with time (c). Under this condition, equation (5) can be solved analytically and gives the equations for lake radius $r(t)$, lake area $A(t)$ and lake volume $V(t)$:

$$r(t) = c\,t + r_0 \tag{6}$$

$$A(t) = \frac{1}{2} r(t)^2 \pi \tag{7}$$

$$V(t) = \frac{2}{3} r(t)^3 \alpha \tag{8}$$

with the initial radius r_0. Thus, approximating the lake geometrically as a cylinder-wedge gives theoretically a quadratic growth of lake area and a cubic one of lake volume, basically driven by the positive feed-back between lake area and energy transfer into the lake. Approximating the lake as a half cylinder with constant depth but variable radius would, for instance, give an exponential growth of the lake.

Determining the growth factor c by least-square fitting of equation (7) with the observed lake-area growth (Fig. 3) shows that it is not possible to obtain one single c for the entire temporal development of the lake. Therefore, the time of lake evolution was separated into the two intervals 1968-1985 and 1989-1994, omitting the period 1986-1988 where the lake level dropped twice. A growth factor of $c \approx 1.5$ m yr^{-1} was calculated for 1968-1985, and $c \approx 5$ m yr^{-1} for 1989-1994. According to this model the lake growth accelerated by more than three times in the second period. In the model, mainly two influences could cause such a change in growth rate (cf. eq. 5): First, a change in angle α of the cylinder wedge, i.e., a spatial change in lake bottom slope, or, second, a change in energy available for ice-melt (Q_{eff}) per time and area units. The calculated increase in c since 1989 corresponds to a decrease in lake bottom slope from 10° to 3°. Although the lake bottom slope is by far not constant (Fig. 4), such a decrease in slope can hardly be found.

Thus, the effective energy turnover in the lake may have increased since the mid-1980s. Possible explanations are:

(1) The period of insulating and reflecting snow- and ice-coverage markedly influences the energy balance of the lake (Gray and Male, 1981; Liston and Hall, 1995). In fact, visual inspection of the repeated aerial and amateur photography suggests that the period of open water has increased since the mid 1980s. The fact that the observed acceleration of lake growth corresponds with exceptionally negative glacier mass-balances of the nearby Gruben glacier (Kääb, 2001) points to climate influence on the change in lake ice conditions (Livingstone, 1997).

(2) A further positive feed-back process may have influenced the lake growth in that an enlarging lake surface and volume could accelerate the annual melt of surface ice and snow, or reduce the amount and period of ice and snow covering the lake.

(3) Increasing length and height of the dead ice front may have reinforced calving processes and, thus, intensified front retreat (cf. Haeberli and Röthlisberger, 1976).

(4) The temporal correlation between the lake-level droppings in the mid 1980s and the accelerated lake growth could suggest some connection, which is, however, not confirmed by additional observations (change in thermal convection, increased height of dead-ice front above water and enhanced melt water input?).

(5) While melt of the lake boundaries is well documented, the development of the lake bottom is not clear. Certainly, the lake growth was accompanied by some sublake talik formation (i.e., formation of a layer of unfrozen ground within permafrost; Burn, 1990). However, the influence of such permafrost melt on the lake-bottom topography is strongly dependent on the (unknown) ground ice content. In any case, the underlying permafrost has presumably melted much slower than the dead-ice front because of insulation by debris (cf. Sakai et al.,

2000). However, some temporal changes of bottom topography or properties may have also influenced the energy turnover in the lake.

From equation (5) an effective melt energy $Q_{eff} \approx 3$ W m^{-2} for 1968-1985 and $Q_{eff} \approx 10$ W m^{-2} for 1989-1994 can be derived. Calculating the energy needed to melt an ice volume equivalent to the volume of the lake depression formed since the mid-1960s (approx. 100,000 m^3) gives a total energy of approx. 30,000 W, and with an average lake area of 3500 m^2 an average effective melt energy for 1968-1994 of $Q_{eff} \approx 8$ W m^{-2} . This amount confirms the order of magnitude obtained above by statistic fitting of the geometric model to the observed lake growth. Gabathuler (1999) determined for the high-mountain Jöri lake in the Swiss Alps (altitude about 2500 m a.s.l.) an energy balance of a similar order of magnitude considering short- and long-wave radiation balance, sensible heat flux, latent heat flux and heat flux from precipitation. Although many local influences on the latter not ground-ice related Jöri lake are different from our thermokarst lake (cf. Livingstone et al., 1999), the good comparison of both energy balances suggests that a significant part of the energy, in the Jöri case consumed for lake water heating, is consumed by ice melt in case of the thermokarst lake. Sakai et al. (2000) obtained melt energies of roughly 15-20 W m^{-2} on average for several supra-glacial ponds in the Himalaya (cf. last paragraph of previous section).

Conclusions

The formation of thermokarst lakes over longer time periods requires special preconditions in ground thermal regime, ground-ice content and topography: area-wide permafrost to seal the lake bottom; super-saturation of ice-rich ground (itself often connected to permafrost) as a precondition for thaw settlement; flat topography to enable lake growth without corresponding melting of topographic barriers damming the lake. Such conditions are regularly found in Arctic landscapes but rarely in alpine environments.

The 30-yr photogrammetric monitoring series and the dynamic model of the geometric evolution of a high-mountain thermokarst lake on the Gruben rock glacier showed the following:

(1) Thermokarst-lake growth seems to be accompanied not only by heat conduction within the water body, but also by thermal convection of lake water towards the ice-rich underwater surface.

(2) Thermally driven water transport in the lake may concentrate on individual streams. In the presented case, such streams had a diameter in the order of 10^{-1} m. Some of the convection cells have apparently been stable in location for at least some hours (one day?).

(3) The spreading lake surface is able to trigger a nonlinear positive feedback process by causing an enhanced energy input through the growing surface and, thus, accelerating the lake growth.

(4) Variations in the ratio between lake-surface area and the area of ice-rich ground exposed to energy transport by lake water influence the growth rate of a thermokarst lake. The growth rate is significantly affected by undulations of the lake-bottom topography.

(5) Likewise, variations in energy input to the lake affect its development. An increase in growth-rate of Gruben thermokarst lake by a (modeled) factor of approximately three coincides with a change in meteorological conditions indicated by exceptional negative mass-balances of the nearby Gruben glacier, and with an increasing open-water period.

In a more general view on the coupling between high-mountain thermokarst lakes and climate variations we conclude from our study:

(1) The formation of thermokarst lakes under discontinuous permafrost can be connected to processes typical for warming climate (e.g. glacier retreat, or permafrost thaw). Thus, the likelihood of such lakes forming increases with warming.

(2) Climate variations have a significant influence on the growth rate of thermokarst lakes with time. However, individual local conditions (as given above) are able to markedly overlay such climate signal. Analyzing climate change from thermokarst lakes, therefore, requires careful consideration or modeling of such local influences not related to climate.

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Figures

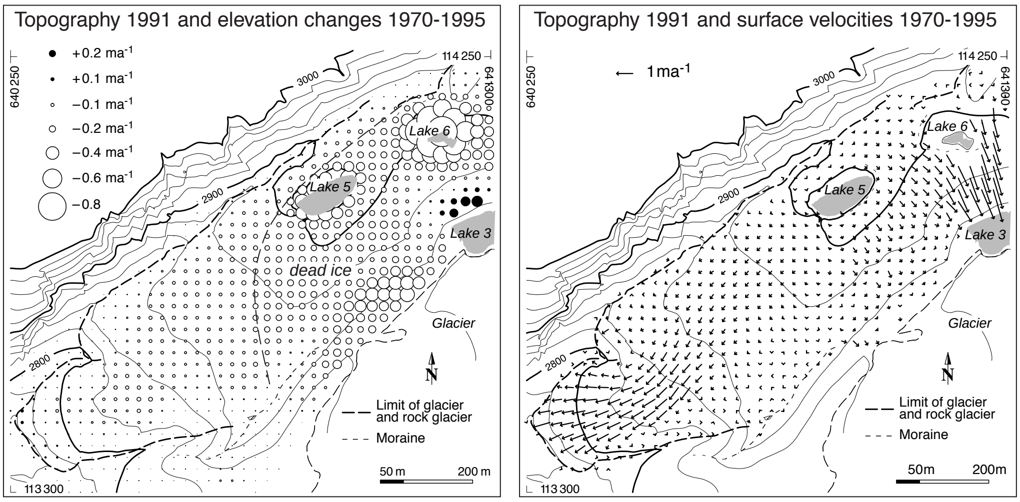

FIGURE 1. Surface topography and situation of 1991, elevation changes 1970-1995 (left) and surface creep velocities 1970-1995 (right) of the Gruben rock glacier, Swiss Alps, determined by photogrammetric techniques (Kääb et al., 1997). Changes in surface elevation indicate different subsidence rates in the periglacial part (left of dashed line) and the glacier-affected part containing massive remains of dead ice (right of dashed line). Vectors depicting the direction and the speed of surface displacements indicate different creep regimes of the periglacial part (speed up to 1 meter per year, m a^{-1}) and the glacier-affected part (speed up to 2 m a^{-1}). On the rock glacier the thermokarst lakes 5 and 6 have developed since the mid 1960s. Coordinates in meters and Swiss national system.

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FIGURE 2. Thermokarst lake 5 in August 1995 as seen from the North. Lake area is about 10000 m^2, the height of the ice-front to the left more than 10 m. The lake grew by accelerating melt into the dead-ice occurrences to the left. The right lake border containing rock glacier permafrost remained stable. In the ice-front a typical glacier-ice stratification can be seen (cf. Fig 6). Note the small icebergs drifting away from the ice-front and indicating some transport process.

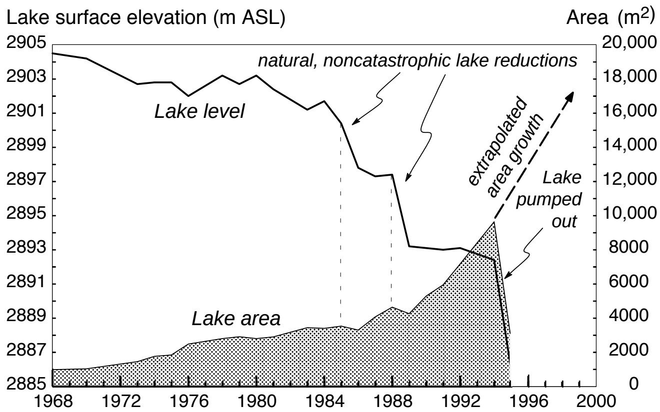

FIGURE 3. Level and area of the ice-dammed thermokarst lake 5 over the period 1968-1995 as determined from repeated photogrammetry. The lake level smoothly dropped some meters in 1985 and 1988. Assessment of future lake growth suggested an area of about 20,000 m^2 (or an volume of 100,000 m^3 respectively) in the year 2000 (dashed line with arrow). In 1995, the lake was pumped out due to its increasing hazard potential.

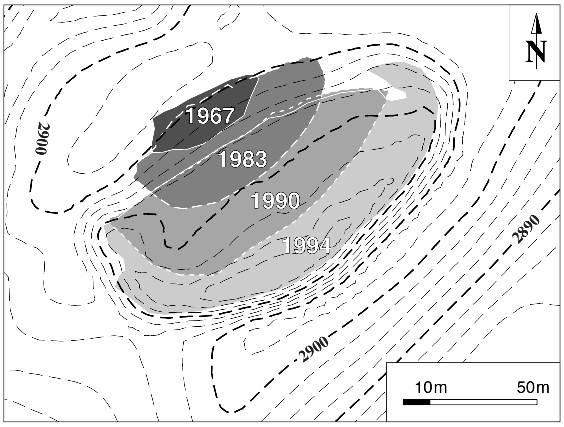

FIGURE 4. Planimetric representation of the outline of thermokarst lake 5 in selected years of the time period 1967-1995. Note asymmetric lake growth towards the massive ground ice in the south; the displacement of the northern lake shore is a consequence of the lake-level lowering. The topography of the lake bottom and the surrounding surface, represented by 2.5-m contour lines, was photogrammetrically determined after lake drainage in late 1995.

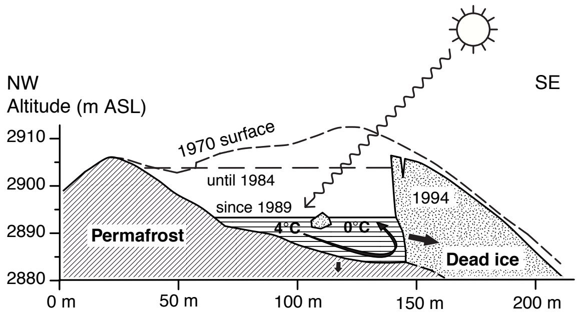

FIGURE 5. Schema of energy input by solar radiation, convective heat transport to the damming dead ice/permafrost, subsequent ice melt and water cooling of the lake water leading to a positive feed-back which drives the non-linear growth of thermokarst lake 5.

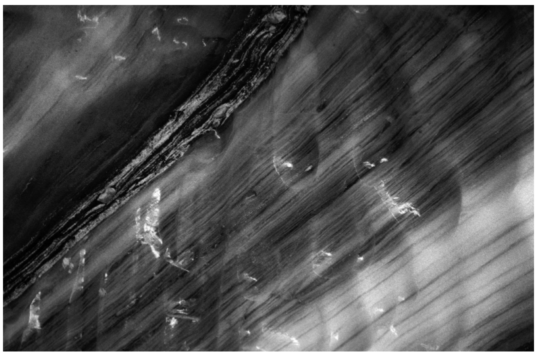

FIGURE 6. Section of the underwater ice-front shortly after lake drainage. Vertical pear-like convexities of about one meter height and maximum depth of a few decimeters were found in the exposed ice front, presumably related to local thermal convection cells (cf. Fig. 7). The frontal photograph depicts a section of about 2.5 m · 1.5 m.

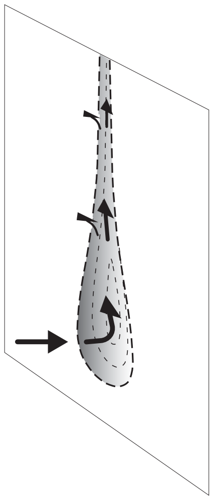

FIGURE 7. Oblique schematic view of a typical thaw convexity in the underwater ice-front of lake 5 and possible related water- (and, thus, energy-) transport. Depicted section is about 1.5 m high. Dashed lines represent approximate 0.1m contour lines of convexity depth.

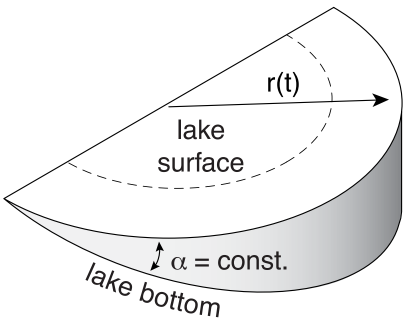

FIGURE 8. Simplified geometric model of the thermokarst lake with a horizontal semicircle surface and a constantly inclined lake bottom. The lake radius is increasing with time.