Glacial Lakes in the Himalayas:
A Review on Formation and Expansion Processes

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Abstract

In the recent several decades, glacial lakes have been expanding at the terminus of debris-covered glaciers associated with glacier shrinkage in the Himalayas. Accordingly, glacial lake outburst floods (GLOFs) have become serious natural disasters in the Himalayas. Bathymetrical surveys have been carried out at several glacial lakes, and the relationship between the lake area and water volume has been illustrated. At the initial stage, several small lakes are distributed near the glacier terminus, and then they coalesce and become one large lake, and finally, the large lake expands rapidly through calving of the glacial front. Surface inclination, surface velocity and surface lowering of glaciers are significant indicators of glacial lake expansion. Future prediction of topographical changes in debris-covered glaciers is important in forecasting the formation and expansion of glacial lakes.

Key words: ablation processes, calving, debris-covered glacier, glacial lake area, water volume

1. Introduction

There are many debris-covered glaciers in places such as the Himalayas, Andes, Italian Alps and Southern Alps in New Zealand. These debris-covered glaciers contains many glacial lakes in ablation areas (*e.g*., Lliboutry *et al*., 1977; Vallon, 1989; Kirkbride, 1993; Clague & Evans, 2000; Diolaiuti *et al*., 2006; Gardelle *et al*., 2011).

In the Himalayas, large valley glaciers several kilometers long occupy about 80% of the glacialized areas (Moribayashi & Higuchi, 1977) and have mostly debris-covered ablation areas. The fronts of debris-covered glaciers advanced down to approximately 4,500 m a.s.l. during the Little Ice Age. Some glacial lakes have started to expand at the terminus of debris-covered glaciers in the 1950s and 1960s (Yamada, 1998; Komori, 2008). Glacial lakes are surrounded and dammed by terminal- and lateral-moraines which formed during the Little Ice Age, and others are dammed by stagnant ice masses at the glacial front. The danger glacial lakes pose depends largely on the relative elevation from the base of the surrounding moraines to the lake surfaces (Benn *et al*., 2001; Sakai *et al*., 2007; Fujita *et al*., 2008).

Moraine dams generally fail on account of overtopping surges and incision by outlet streams. This overtopping is caused by large waves generated by huge masses falling into the lake, such as snow or glacier avalanches; or by moraine failures induced by earthquakes; or by advancing, sliding and calving (ice wastage caused by shedding of large ice blocks from a glacier's edge into the body of water) of the mother glacier (Ives, 1986; Vuichard & Zimmermann, 1987; Yamada, 1998; Iwata *et al*., 2002). Melting of ice cores in moraines and piping are other possible failure mechanisms (Clague & Evans, 2000). Thus, outburst floods are caused by breaking of the moraine damming glacial lakes due to the above reasons, and these events are called glacial lake outburst floods (GLOFs). At the beginning of the last century the number of GLOFs increased in the Himalayas (Richardson & Reynolds, 2000).

2. Measurements of Glacial Lakes

2.1 Bathymetric measurement of glacial lakes

The first series of the depth distribution measurements in the Himalayas was carried out at Abmachimai Co, Qangzonk Co, and Paqu Co in the Tibetan Plateau by LIGG / WECS / NEA (1988). In the Nepal Himalayas, Imja Glacial Lake was measured in 1992 by Yamada and Sharma (1993), and in 2002 by Sakai *et al*. (2003). The Tsho Rolpa, Lower Barun and Thulagi lakes have been also surveyed (Yamada, 1998). Recently, the Tsho Rolpa, Thulagi and Imja lakes were surveyed again in 2009 by ICIMOD (2011). Raphstren Tsho and Lugge Tsho in the Bhutan Himalayas were measured by the Geological Survey of India (1995) and Yamada *et al*. (2004), respectively. The location, area, water volume and maxi-

Table 1 List of locations, areas, volumes and maximum depths of glacial lakes, for which bathymetrical surveys have been carried out.

Lake name	Surveyed Date	Region	Latitude	Longitude	Area (km^2)	Volume ×10^6 (m^3)	Maximum depth (m)	Type	Reference
Abmachimai Co	Apr, 1987	Tibet	28°05'29"N	87°38'17"E	0.56	19.00	72	moraine-dammed glacial lake	LIGG/WECS/NEA (1988)
Qangzonk Co	Apr, 1987	Tibet	27°55'36"N	87°46'15"E	0.76	21.40	68	moraine-dammed glacial lake	LIGG/WECS/NEA (1988)
Paqu Co	Apr, 1987	Tibet	28°18'09"N	86°09'34"E	0.31	6.00	36	moraine-dammed glacial lake	LIGG/WECS/NEA (1988)
Lower Barun	Mar, 1993	Nepal	27°47'49"N	87°05'52"E	0.60	28.00	109	moraine-dammed glacial lake	Yamada (1998)
Imja	Apr, 1992	Nepal	27°53'58"N	86°55'18"E	0.60	28.00	99	moraine-dammed glacial lake	Yamada (1998)
Imja	Apr, 2002	Nepal	27°53'58"N	86°55'18"E	0.90	36.00	91	moraine-dammed glacial lake	Sakai *et al.* (2003)
Imja	May, 2009	Nepal	27°53'58"N	86°55'18"E	1.01	35.50	97	moraine-dammed glacial lake	ICIMOD (2011)
Tsho Rolpa	Feb, 1993 & Feb, 1994	Nepal	27°51'57"N	86°28'24"E	1.39	76.60	131	moraine-dammed glacial lake	Yamada (1998)
Tsho Rolpa	Aug-Sep, 2009	Nepal	27°51'57"N	86°28'24"E	1.54	85.94	134	moraine-dammed glacial lake	ICIMOD (2011)
Thulagi	Mar. 1995	Nepal	28°29'45"N	84°28'47"E	0.76	31.80	81	moraine-dammed glacial lake	Yamada (1998)
Thulagi	Jul. 2009	Nepal	28°29'45"N	84°28'47"E	0.94	35.37	80	moraine-dammed glacial lake	ICIMOD (2011)
Dig Tsho	-	Nepal	27°52'25"N	86°35'06"E	0.50	10.00	–	moraine-dammed glacial lake	Mool *et al.* (2001)
Tam Pokhari (Sabai Tsho)	-	Nepal	27°44'28"N	86°50'39"E	0.47	21.25	–	moraine-dammed glacial lake	Mool *et al.* (2001)
Ngozumpa	Oct. 1999	Nepal	27°57'13"N	86°42'01"E	0.05	0.24	12	thermokarst lake	Benn *et al.* (2000)
Ripimo	Nov. 1993	Nepal	27°53'24"N	86°28'15"E	0.02	0.18	25	thermokarst lake	personal communication from Dr. T. Kadota
Raphsthren	1984, 1986	Bhutan	28°06'14"N	90°14'45"E	1.38	66.83	88	moraine-dammed glacial lake	Geological survey of India (1995)
Lugge	Sep-Oct. 2002	Bhutan	28°05'39"N	90°17'44"E	1.17	58.30	126	moraine-dammed glacial lake	Yamada *et al.* (2004)

mum depth of the glacial lakes contacting glacier ice are summarised in Table 1. The relationships between areas and volumes of glacial lakes are shown in Fig. 1. The lake volume (V: ×10^6 m^3) and lake area (A: km^2) have following relationship:

$$V = 43.24 \times A^{1.5307}$$

Huggel *et al.*(2002) has also represented similar relationships from glacial lakes located in the Swiss Alps, including ice-dammed lakes.

The relationship between the area and maximum depth of glacial lakes is shown in Fig . 2. The relationship between the maximum depth of lakes (D_{max}: m) and lake areas can then be calculated as follows,

$$D_{max} = 95.665 \times A^{0.489}$$

The observed glacial lakes shown in Table 1 are in contact with glacier ice at the higher ends of the lakes, and the deepest points are located at upstream positions of almost all glacial lakes. Bathymetric re-measurements have been carried out only at Imja glacial lake by Sakai *et al.* (2003) and Fujita *et al.* (2009). The results show that as the glacial lake expands due to the glacier front retreating, the lake bottom near the terminal moraine undergoes little change. Takenaka *et al.* (2010) carried out electrical resistivity tomography near the terminal moraine of the Imja Glacial Lake. They concluded that dead ice remained under a thick debris layer, and only gradual dead ice melt occurred due to insulation by the thick debris.

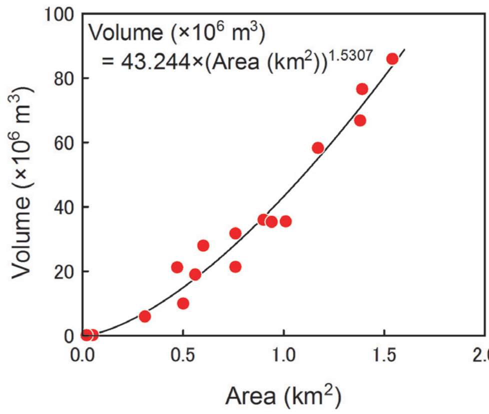

Fig. 1 Relationship between the areas and water volumes of glacial lakes in the Himalayas listed in Table 1.

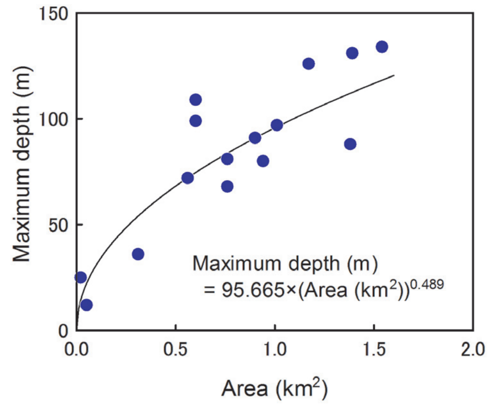

Fig. 2 Relationship between the areas and maximum depths of glacial lakes listed in Table 1.

Figure 3 shows changes in Tsho Rolpa Glacial Lake between 1994 and 2007. The pictures were taken from the mountain slope downstream. Some small islets near the terminal moraine in 1994 have disappeared or shrunk in 2007, although the water level of the lake was lowered by 2-3 m in 1999 by mitigative construction works (Reynolds, 1998). The reason of the islets disappearance would be the slow melting of dead ice under the thick debris layer. For this reason, it is necessary to assess the effect of dead ice melting in these glacial lakes, which would decrease the strength of the terminal moraine damming the lake water.

2.2 Records of glacial lake areal expansion

Expansion of glacial lakes has been recorded by topographic map and satellite images. Mool (1995), Yamada (1998), Ageta *et al*. (2000), Komori *et al*. (2004) and Komori (2008) have collected available maps or satellite images showing areal expansion of each glacial lake at different dates and summarized surface area expansion of glacial lakes in the Nepal and Bhutan Himalayas. At the initial stage, several small lakes were distributed in the ablation areas near the glacier terminus. Then those lakes coalesced and formed large lakes. Shore-lines at the upstream sides of such lakes tend to be straight lines, indicating expansion of the glacial lakes by calving of retreating glacier fronts. On the other hand, the change in shore-lines on the downstream side is relatively smaller than upstream (Watanabe *et al*. 2009).

Komori (2008) reported that the expansion rate of a glacial lake on the southern side of a debris-covered glacier in the Bhutan Himalayas was higher than that on the northern side, and the expansion on the northern side started earlier than that on the southern side. Gardelle *et al.* (2011) selected seven study sites along the east-west wide-spread mountain range from the Bhutan

Fig. 3 Tsho Rolpa Glacial Lake, taken in June 1994 by Dr. N. Takeuchi (Chiba University) (upper) and on Nov. 4, 2007 by Mr. Osafumi Sato (The Asahi Shimbun) (lower).

Himalayas to the Hindu Kush and mapped glacial lakes with LANDSAT satellite imagery acquired in 1990, 2000 and 2009. They reported that in the east part (India, Nepal and Bhutan), glacial lakes were bigger and more numerous than in the west, and have continuously grown, while glacial lakes have shrunk during that period in the Hindu Kush and in the Karakoram. This areal difference of glacial lake change corresponds to less glacier ice loss in the western Himalayas and mountains further west.

3. Elaborate Ablation Processes of Debris-covered Glaciers

Mattson *et al*. (1993) summarized the ablation rate of ice under the debris layer with various thicknesses, and reported that glacier ice covered with thick debris (more than 0.5 m) does not melt because of insulation by the thick debris. Generally speaking, retreats and advances of debris-free glaciers can be used as indicators of glacier response to climate change. While, in the case of debris-covered glaciers, frontal changes do not represent glacier response to climate changes since change in their terminus areas is controlled by the thick debris covers (Scherler *et al*., 2011).

At the Khumbu Glacier, the thickness of the debris layer increased gradually from just beneath the icefall of the glacier to the terminus, where the thickness of debris up to 5 km from the glacier terminus reached more than one meter (Nakawo *et al*., 1986). It can, therefore, be estimated that the Khumbu Glacier would not undergo melting or downwasting of the lower part in terms of the experimental relation shown by Mattson *et al*. (1993) because the thickness of the debris covering the lower part was larger than 0.5 m. However, actually, the glacier surface has lowered 5-18 m from 1978 to 1995 according to Kadota *et al*. (2000) and the estimated annual ablation rate using thermal resistance from satellite data (Nakawo *et al*., 1993) is more than one meter (at most 3 m) (Nakawo *et al*., 1999). Inoue and Yoshida (1980), Sakai *et al*. (1998, 2002) and Röhl (2008) reported that the melt rate of ice cliffs surrounding supraglacial lakes was much higher than that of areas of ice covered with thick debris. Thus the relatively high melt rate at the lower part of the Khumbu Glacier, covered with a debris-layer of more than one meter in thickness, can be explain by rapid ablation at the ice cliff surface and the presence of supraglacial lakes.

As for supraglacial lakes themselves, Kirkbride (1993) inferred that supraglacial lakes aligned along an englacial conduit may be created by the collapse of the conduit roof. Actually, Gulley and Benn, (2007), Gulley *et al*. (2009), Benn *et al*. (2009) have directly observed entrances or insides of conduits connected with englacial conduits at the Khumbu Glacier. Furthermore, Sakai *et al*. (2000) suggested that the roof of the conduit could collapse, leading to the formation of ice cliffs and ponds, which would accelerate the ablation of the debris-covered glacier. Iwata *et al*. (2000) has supported the above presumption, indicating that large relief areas, occupied by a lot of ice cliffs and supraglacial ponds, expanded both upward and downward from the middle parts of the Khumbu Glacier ablation area from 1978 to 1995.

4. Supraglacial Lake Expansion

Kirkbride (1993) has summarized glacial lake expansion processes in terms of topographical views based on his observation at the Tasman Glacial Lake in New Zealand. He concluded that 1) debris-covered ice melts slowly under the debris mantle, 2) collapse of conduit roofs perforates the supraglacial mantle and allows more rapid melting of bare-ice walls, and 3) lakes coalesce and increase in depth by bottom calving with a potentially unstable ice floor. Röhl (2008) also indicated that subaqueous calving occurred in supraglacial lakes of the Tasman Glacier, and 4) disintegration of the ice floor increases water depth and initiates a rapid calving retreat. In the case of the Himalayas, terminal moraines are ice-cored, and glacier ice near the terminus may be filled with meltwater, since ice-cored moraines have less permeability, as seen in 1) in Fig. 4 (Yamada *et al*., 2002). Figure 4 shows a schematic diagram of glacial lake expansion processes, taking into account the dead ice in the terminal moraine, as modified by Kirkbride (1993).

Kirkbride (1993) and Röhl (2008) focused on the transition from melting to calving glacier termini, since

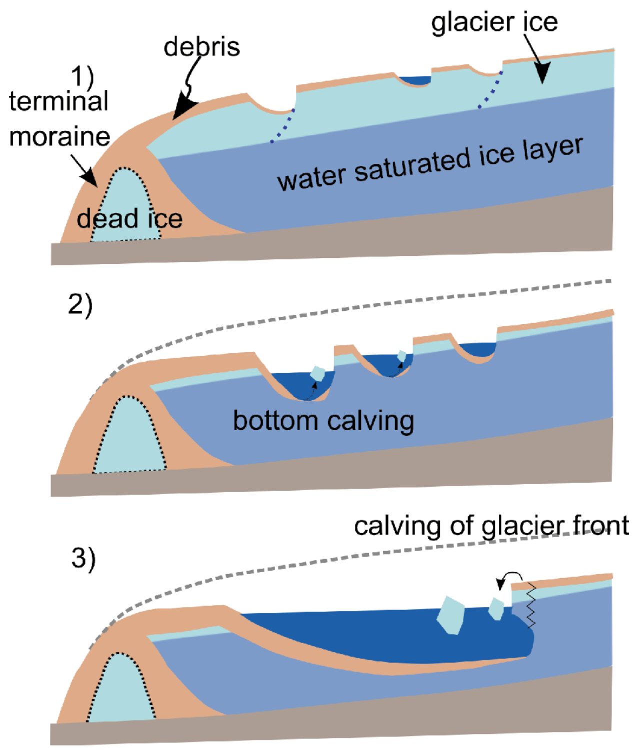

Fig. 4 Schematic diagram of the glacial lake expansion process on debris-covered glaciers, indicating the longitudinal cross section of the glacier terminus. (1) Slow melting under thick debris layer. Glacier surface is still high. (2) Growth of lakes by calving from the lake bottom. (3) Initial rapid lake expansion by calving of glacier front during retreat.

the onset of calving at glacier termini represents the commencement of rapid expansion of glacial lakes. Most calving research has focused on tidewater glaciers (Van der Veen, 1996, 2002; Vieli *et al*., 2001), but lake-calving has received much less attention (Benn *et al*., 2007). Glacial lakes expand rapidly due to glacier ice calving, therefore, calving processes have to be studied to establish a glacial lake expansion model.

It was found that glacier ice calving has been caused by flotation of glacier ice at several tidewater and lacustrine glaciers, such as the Mendenhall Glacier (Motyka *et al*., 2002; Boyce *et al*., 2007) and Bering Glacier (Lingle *et al.*, 1993) in Alaska, Rhonegletscher in Switzerland (Tsutaki *et al*., 2011), Upsala Glacier (Naruse & Skvarca, 2000; Skvarca *et al*., 2003) and Glacier Nef (Warren *et al*., 2001) in Patagonia. Meanwhile, recently several studies have found that thermo-erosional notch growing rates at the waterline are important determinants of the rate-controlling mechanism during sub-aerial calving (Kirkbride & Warren, 1997; Benn *et al*., 2001; Warren & Kirkbride, 2003; Diolaiuti *et al*., 2005; Röhl, 2006, 2008; Benn *et al*., 2007). Röhl (2006), who has directly measured thermal undercutting at the Tasman Glacial Lake, concluded that the calving rate was directly controlled by the rate of thermal undercutting, and implied that the processes of thermal undercutting play a central role in the transition from mere melting to calving termini. Warren and Kirkbride (1998), Benn *et al*. (2001) and Haresign and Warren (2005) have also observed thermal undercutting rates. The above results by several researchers suggest that thermal-undercutting of subaqueous glacier ice acts as a trigger for lacustrine calving.

In the high Himalayan mountains, valley winds are strong during the daytime (Inoue, 1976; Chikita *et al*., 1997, 1999). The average daily maximum wind velocity can reach more than 10 m s^{-1} during the melt season (June to September) (Yamada, 1998). To examine the contribution of water currents driven by valley winds in the variable topography of pro-glacial lakes, to the onset of calving due to thermal undercutting, Sakai *et al*. (2009) calculated wind velocity distributions according to varying fetches, as well as wind-driven lake surface currents. The results suggest that the onset of calving due to thermal undercutting is controlled by water currents driven by winds at the surface of the lake, which develop in proportion to expanding water surface. Chikita (2007) calculated the wind velocity distribution over the lake surface at Tsho Rolpa and Imja Glacial Lake. He concluded that the wind speed over the lake surface at the Imja Glacial Lake was less than that at Tsho Rolpa due to the difference in end moraine heights, which act as an obstacle in the path of valley winds. Actually the rate of expansion of the lake at Imja (40 m yr^{-1}) was less than that at Tsho Rolpa (70 m yr^{-1}) from the 1960s to the 1990s (Yamada, 1998). The difference in rates of expansion between the Imja and Tsho Rolpa lakes is due to the difference in wind speeds over the lakes.

5. Formation Conditions of Glacial Lakes

Reynolds (2000) reported that in the Bhutan Himalayas, glacial lakes form in those parts of the glacier where the inclination of the glacier surface is less than 2°. Quincey *et al*. (2007) reported that glacial lakes develop upon debris-covered glaciers with low slopes and a surface speed less than 10 m a^{-1}. Suzuki *et al*. (2007) calculated the thermal resistance of debris-covered glaciers and found that those with relatively thin debris layers tend to develop glacial lakes at their terminus. This finding suggests that glaciers with high rates of ablation tend to develop glacial lakes. In summary, the available evidence indicates that glaciers that record a relatively large lowering of their surface are likely to develop glacial lakes at their terminus. Indeed, Kirkbride (1993) reported that lowering of the glacier surface resulted in the formation of supraglacial ponds and the commencement of calving at the glacial lake. Lamsal *et al*. (2011) have reported that the glacier surface declined about 100 m during the Imja Glacial Lake expansion from several small lakes in 1964 to a large lake in 2006.

Usually, the ablation rate reached a maximum at the glacier terminus in the case of debris-free glaciers. In general, though, large glaciers are covered with debris at the lower part, and the thickness of the debris increases from equilibrium line altitude, reaching more than one meter at the terminus of the glacier. Thus, ablation rates at the terminus are negligibly small and increase up to several kilometers from the terminus. After reaching ablation maximum at the middle part of the ablation area, ablation rates decrease with altitude due to decreasing air temperature. Therefore, small surface lowering at the glacier terminus and large surface lowering at the middle part of the ablation area reduce inclinations at the lower part of debris covered glaciers.

Naito *et al*. (2000) calculated the longitudinal glacier surface change of the Khumbu Glacier by establishing a new model coupling the mass balance of debris-covered glaciers and glacier flow models. The model predicted the formation of depressions in the lower ablation area, leading to glacial lake formation in the depressions. Bolch *et al*. (2008) confirmed intensive surface lowering by comparing digital elevation models produced from Corona with observations from ASTER (Advanced Spaceborne Thermal Emission and Reflection radiometer) at the Khumbu Glacier. Nuimura *et al*. (2011) analysed glacier surface elevation changes using digital elevation models in 1978, 1995 and 2004, which were taken by field observations, and they detected a remarkable acceleration of surface lowering in the middle part of the ablation area at the Khumbu Glacier.

Sakai and Fujita (2010) identified the formation conditions of glacial lakes by analysing two easily measurable topographic parameters: inclination of the glacier surface and the difference in relative height between the glacier surface and lateral moraine ridges (herein, this difference is referred to as DGM, which is an indicator of glacier surface lowering on debris-covered glaciers). They concluded that for all glaciers with large supraglacial lakes, values of DGM exceed 60 m and the average inclination of the glacier surface is less than 2.0°. Low inclination indicates relatively small ice flux to the lower part of the glacier, which induces glacier shrinkage, and a large DGM represents the proximity of the glacier surface to the water level in glacier ice, which yields stable supraglacial lakes as shown in Fig. 4.

Benn *et al*. (2001) believed that the Khumbu and Ngozumpa glaciers were close to the threshold for moraine-dammed glacial lake formation. Those two glaciers definitely have low inclinations but there have been no appropriate indicators for the possibility of formation of glacial lakes. Consequently, glaciers with lakes and without lakes can be clearly distinguished by taking into account not only inclination but also DGM in the Nepal and Bhutan Himalayas as stated by Sakai and Fujita (2010).

6. Concluding Remarks

Debris-covered glaciers both with lakes and without lakes are found under identical meteorological conditions, for example, at the Khumbu Himal glacier, and the Lunana glacier in the Bhutan Himalayas (Fig. 5). What is the difference between these glaciers with lakes and without lakes? Twenty years before, we did not have an answer to this question, but now we can give some explanations. In the past two decades, formation and expansion processes have been elucidated.

Glacial lakes expand on a debris-covered glacier, which has low inclination (little ice flux from upstream) and large surface lowering. Both formation conditions of glacial lakes are caused by one phenomenon, glacier shrinkage. Glacier shrinkage is caused by less precipitation or large ablation. Ablation processes of debris-covered glaciers are more complicated than those of debris-free glaciers. Not only there is a debris mantle of various thicknesses and sizes, but also unstable supraglacial lakes, which constitute heat absorbing spots on debris-covered glaciers, preventing us from establishing a model of ablation processes of debris-covered glaciers at present. In the next twenty years of research on Himalayan debris-covered glaciers, however, we should concentrate on the study of ablation processes of debris-covered glaciers taking into account that of rapid changing supraglacial lakes in order to predict fluctuations of debris-covered glaciers and formation of glacial lakes.

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Fig. 5 Lunana region, Glacier lakes in the Bhutan Himalayas, taken by Dr. Koji Fujita (Nagoya University) on Oct. 17, 2004. Bechung Glacier, Raphstren Tsho, Thorothormi Glacier from the left. There are several glacial lakes at different expansion stages on the debris-covered glaciers.

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