CHAPTER

Glacier-related outburst floods

14

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Abstract

Water bodies impounded by glaciers, moraines, and ice jams on rivers can drain suddenly, with disastrous downstream consequences. Lakes can form at the margins of an alpine glacier or ice cap, on its surface, or at its base. Smaller pockets of water may also be present within some glaciers. In all cases, these water bodies might drain by enlarging subglacial tunnels or by mechanical collapse of the glacier dam. Many formerly stable glacier lakes have failed over the past century, in some cases repeatedly, as Earth's atmosphere has warmed and glaciers thinned and receded. The peak discharge, duration, and volume of a subglacial outburst flood depend mainly on (1) the geometry and rate of development of the tunnel at the base of the glacier and (2) the size and geometry of the impounded water body. Discharge commonly increases exponentially during the outburst, but ends quickly when the lake empties or when the drainage tunnel is plugged by collapse of the tunnel roof or closes due to plastic ice flow. Some glacier outburst floods result from the mechanical collapse of the ice dam. In such cases, the peak flow is achieved rapidly during the collapse. Outburst floods from glacier lakes attenuate due to temporary storage of floodwaters in channels and on valley floors.

Many hazardous lakes are dammed by lateral and end moraines that formed in the past two centuries when valley and cirque glaciers retreated from advanced positions reached during the Little Ice Age. Moraine dams are susceptible to failure because they are steep and relatively narrow, because they comprise loose poorly sorted sediment, and because they may contain ice cores or interstitial ice. These dams generally fail by overtopping and incision. The triggering event may be a heavy rainstorm, strong winds, or an ice avalanche or landslide into the lake that generates waves that overtop the dam. Melting of moraine ice cores and piping are other possible failure mechanisms. Outflow from a moraine-dammed lake increases as the breach enlarges and then decreases as the level of the lake falls. The moraine breach may become armored, preventing further incision, or the hydraulic gradient at the breach may decrease to a point that erosion ceases.

Outburst floods from glacier- and moraine-dammed lakes typically entrain, transport, and deposit large amounts of sediment. If the channel is steeper than about 0.10-0.15 and contains abundant loose sediment, the flood likely will transform into a debris flow. Such flows may be larger and more destructive than the flood from which they formed. A period of protracted warming is required to trap lakes behind moraines and create conditions that lead to dam failure. The warming also forces glaciers to retreat, prompting ice avalanches, and landslides that have destroyed many moraine dams.

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14.1 Introduction

About 10% of our planet is covered by glacier ice, and about 99% of this ice is in Greenland and Antarctica. The other 1% forms ice fields, ice caps, and cirque, valley, and piedmont glaciers, mainly in mountains of northwest North America, Arctic Canada, Asia, southern South America, and in Iceland.

Glaciers provide many benefits; for example, meltwater from alpine glaciers augments runoff during summer, which is important for agriculture, mining, municipal water supply, and hydroelectric power generation (Huss et al., 2017). A recent study concluded that 1.9 billion people rely on water from glaciers (Immerzeel et al., 2020). Glaciers are also an aesthetic and recreational resource. For example, Athabasca Glacier in the Rocky Mountains of Alberta is visited by more than 2 million people each year.

However, processes associated with glaciers can also be hazardous, and some of these processes may be amplified by climate change. This chapter explores hazards and risks associated with outburst floods caused by failures of lakes impounded by glaciers and their moraines. We review flood sources and dam failure mechanisms, discuss the current situation and possible future scenarios in a warming world, and briefly discuss evolving risk associated with the hazard.

During the Pleistocene, outbursts from water bodies impounded by glaciers produced some of the largest floods on Earth (Fig. 14.1; Bretz, 1923a, b; O'Connor et al., 2002; O'Connor and Costa, 2004;

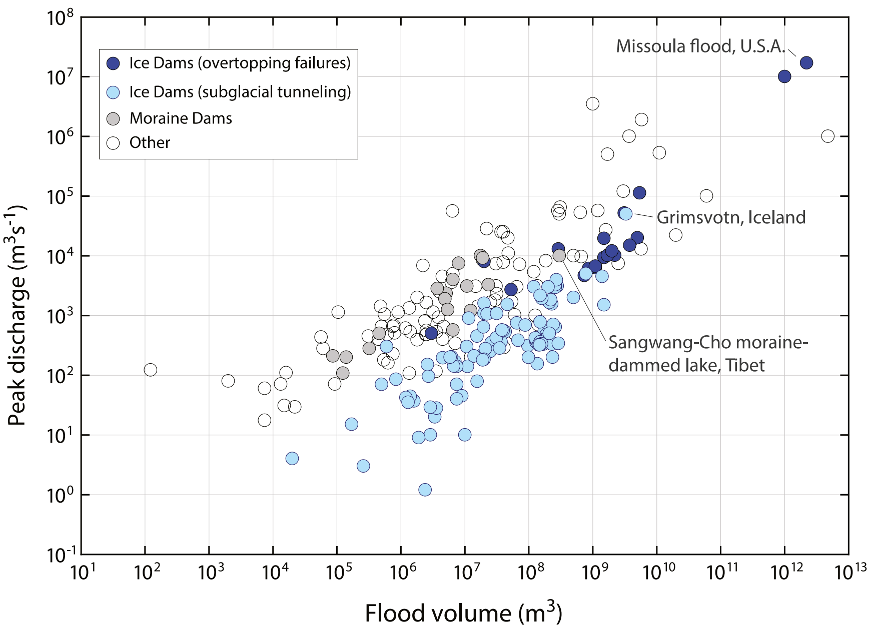

FIG. 14.1

Outburst floods for which flood volume and peak discharge are known. The largest floods of each type are labeled. *Adapted from Fig. 1 of O'Connor, J.E., Clague, J.J., Walder, J.S., Manville, V., Beebee, R.A., 2013. Outburst floods. In Shroder, J., Wohl, E. (Eds.). Treatise on Geomorphology, vol. 9, Fluvial Geomorphology. Academic Press, San Diego, CA, pp. 475–510; data from Walder, J.S., Costa, J.E., 1996. Outburst floods from glacier-dammed lakes: the effect of mode of lake drainages on flood magnitude. Earth Surf. Process. Landf. 21, 701–723 and O'Connor, J.E., Beebee, R.A., 2009. Floods from natural rock-material dams. In: Burr, D. M., Carling, P.A., Baker, V.R. (Eds.): Megafloods on Earth and Mars. Cambridge University Press, Cambridge, UK, pp. 128–171.*

Herget, 2005; Baker, 2009, 2013; Bohorquez et al., 2019; Benito and Thorndycraft, 2020). These floods inundated large tracts of land and produced long-lasting geomorphic effects. In contrast, outburst floods that have happened in recent centuries are much smaller than those of the Pleistocene; even so, they have caused much damage and injury (Lliboutry et al., 1977; Yesenov and Degovets, 1979; Hewitt, 1982; Haeberli, 1983; Eisbacher and Clague, 1984; Ruren and Deji, 1986; Vuichard and Zimmerman, 1987; Costa and Schuster, 1988; Haeberli et al., 1989; Reynolds, 1992; O'Connor and Costa, 1993; Evans and Clague, 1994; Watanabe and Rothacher, 1996; Cenderelli, 2000; Richardson and Reynolds, 2000; Cenderelli and Wohl, 2003; Vilímek et al., 2005; Korup and Tweed, 2007; Carrivick and Tweed, 2013; Kropáček et al., 2015; Allen et al., 2016). Most mountain ranges where these floods occur were formerly sparsely populated, but in recent times have experienced explosive population growth accompanied by increased tourism. The European Alps, for example, are presently home to about 14 million people, and about 120 million people visit the Alps each year. As a result, the risk from outburst floods and other natural hazards has greatly increased.

14.2 Flood sources

14.2.1 Glacier-dammed lakes

A glacier can impound a water body at its margins, on its surface, at its base, or within it (Fig. 14.2). Some lakes at the margins of alpine glaciers contain more than 200 million m^3 of water (e.g., Hauser, 1993), although most are smaller. These lakes are small compared to the more than 400 lakes that currently exist at the base of the East Antarctic Ice Sheet, the largest of which, Lake Vostok at 400 km^3, holds more than three times the water in Lake Ontario, one of the five Great Lakes in North America.

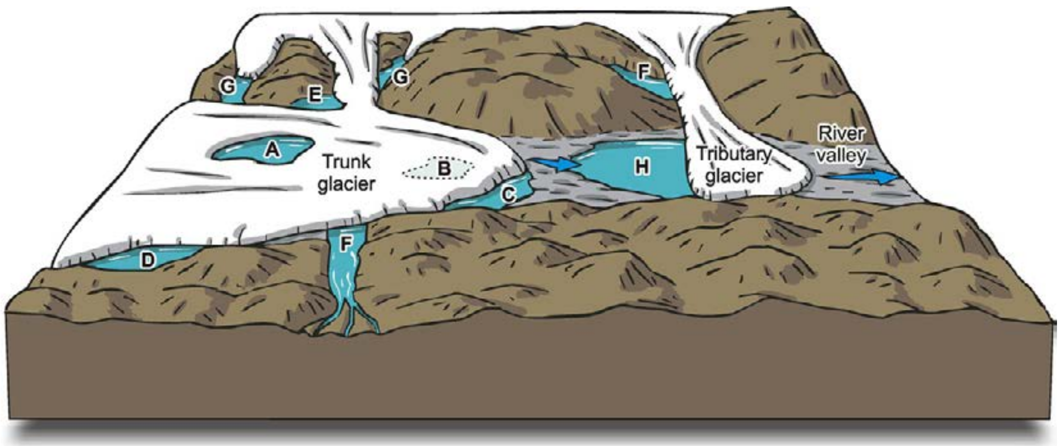

FIG. 14.2

Schematic diagram showing locations where a water body can be trapped by a glacier: (A) supraglacial, (B) subglacial, (C) proglacial, (D) embayment in slope at glacier margin, (E) area of coalescence of two glaciers, (F) tributary valley adjacent to a trunk or tributary glacier, (G) the same as (F) except that glaciers dam both ends of the lake, and (H) main valley adjacent to a tributary valley. Toned areas are land; unpatterned areas are ice (Clague and Evans, 1994).

Lakes at the base of the Antarctic Ice Sheet, however, are stable, and in any case do not pose a risk to people. In contrast, lakes impounded by alpine glaciers and ice caps are susceptible to sudden emptying or to a large overflow due to a displacement wave and consequently are a threat to residents and infrastructure in valleys below.

Relations between valley-blocking ice and the surrounding landscape are complex, leading to a variety of unstable situations. Lakes can exist where glaciers block a tributary or trunk valley (Figs. 14.3–14.5). They also can form during glacier retreat when formerly confluent glaciers separate, leaving space between them (Costa and Schuster, 1988; Tweed and Russell, 1999; Geertsema and Clague, 2005). Lakes of these types have been described in mountainous areas around the world, including Alaska (Post and Mayo, 1971), western Canada (Marcus, 1960; Clague and Mathews, 1973; Clague and Evans, 1994, 2000; Geertsema and Clague, 2005), Scandinavia (Liestøl, 1956), Iceland (Roberts, 2005; Björnsson, 2009), Europe (Haeberli, 1983; Eisbacher and Clague, 1984; Haeberli et al., 1989, 2001; Kääb et al., 2004; Emmer et al., 2015), Asia (Hewitt, 1968, 1982; Ives, 1986; Ding and Liu, 1992; Richardson and Reynolds, 2000; Xin et al., 2008; Hewitt and Liu, 2010; Bolch et al., 2011; Mergili and Schneider, 2011; Rounce et al., 2016; Zhang et al., 2019), and South America (Lliboutry et al., 1977; Hauser, 1993; Dussaillant et al., 2010; Portocarrero, 2014; Anacona et al., 2015; Emmer et al., 2016).

Unstable glacial lakes also exist beneath ice caps in Iceland. During eruptions of Grímsvötn and other active Icelandic volcanoes (Fig. 14.6), subglacial lakes rapidly enlarge and then drain suddenly, producing large outburst floods termed "jökulhlaups" (Björnsson, 1974, 1992, 2003, 2009, 2010;

FIG. 14.3

Lakes dammed by Brady Glacier in southeast Alaska. The largest of the lakes is Abyss Lake, which, when full, is over 250 m deep at the glacier dam (Google Earth image, May 2016).

FIG. 14.4

Glacial Lake Melbern in northwest British Columbia. The lake is 19 km long and is impounded behind confluent Melbern and Grand Pacific glaciers northwest of Glacier Bay, Alaska.

Photo by John Clague.

Gudmundsson et al., 1997; Tómasson, 2002; Waitt, 2002; Alho et al., 2005; Roberts, 2005; Carrivick, 2007, 2009; Russell et al., 2010). The 1918 Katla jökulhlaup had a peak discharge of about 300,000 m^3/s (Tómasson, 1996) and was one of the largest floods on Earth during the historic period. An eruption of Grímsvötn volcano beneath Vatnajökull in 1996 had a peak discharge of about 50,000 m^3 and released 3.3 km^3 of water (Björnsson, 2003, 2009). Not all Icelandic jökulhlaups, however, are associated with volcanic eruptions; some result from geothermal melting of ice at the glacier bed.

Small, but potentially damaging outburst floods and debris flows also result from the sudden draining of "water pockets" beneath and within cirque and valley glaciers (Richardson, 1968; Haeberli, 1983; Driedger and Fountain, 1989; Walder and Driedger, 1994). Haeberli (1983) documented more than 26 outburst floods from water pockets in glaciers in the Swiss Alps. Dozens of similar outburst floods have happened on Mount Rainier, Washington, in the 20th century (Driedger and Fountain, 1989; Walder and Driedger, 1994). Some of the Mount Rainier floods happened during rainstorms, suggesting a possible trigger, but many occurred during dry warm periods. Driedger and Fountain (1989) inferred that the floods on Mount Rainier result from sudden emptying of one or a few water-filled cavities into lower pressure, subglacial drainage channels. Pressurized water-filled cavities can arise from glacier flow over steep or stepped beds (Haeberli, 1983; Driedger and Fountain, 1989), and such outbursts can happen without warning and pose a particular hazard to tourists visiting the termini of glaciers, for example, in New Zealand and Europe.

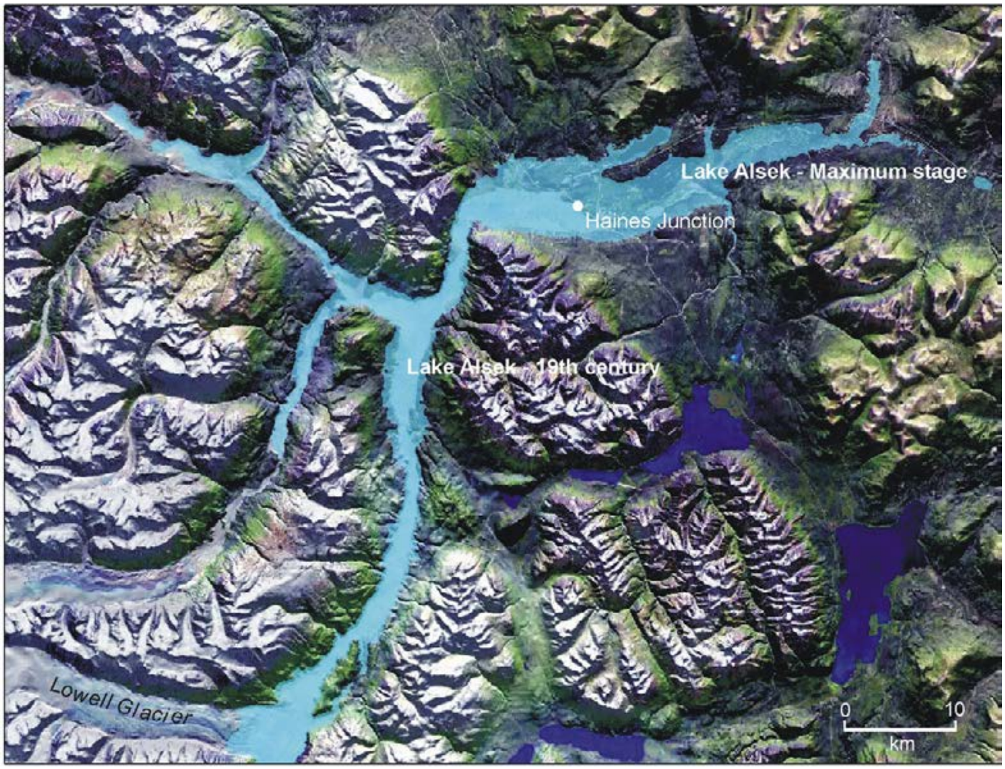

FIG. 14.5

Glacial Lake Alsek formed repeatedly during the Little Ice Age when Lowell Glacier, a large valley glacier in the St. Elias Mountains in southwest Yukon Territory, advanced across the Alsek River valley and blocked the flow of the river. This figure depicts the lake in the mid-19th century, when it reached to the present site of Haines Junction on the Alaska Highway, and earlier when the site of Haines Junction was inundated by water about 70 m deep. The lake no longer exists.

Courtesy of Jeff Bond, Yukon Geological Survey.

14.2.2 Moraine-dammed lakes

Alpine glaciers have thinned and retreated over the past century, and lakes have formed inside the Little Ice Age moraines of many of these glaciers (Figs. 14.7 and 14.8; Costa and Schuster, 1988; O'Connor and Costa, 1993; Clague and Evans, 1994, 2000; Emmer et al., 2016; Wang, 2016; Wang et al., 2017; Kougkoulos et al., 2018). In the high mountains of Asia, heavily debris-covered glacier tongues are downwasting and receding, and lakes continue to expand behind end moraines (Fig. 14.8).

Moraine dams typically are steep-sided, consist of loose sediment, and are sparsely vegetated; some are ice-cored. As a consequence, they are potentially unstable and vulnerable to failure. Irreversible rapid incision of moraine dams may be caused by a large overflow triggered by an ice avalanche or

FIG. 14.6

Eruption through the Vatnajökull ice cap, Iceland, in October 1996. A large amount of water pooled beneath Grímsvötn volcano during the eruptive sequence and emptied during a large jökulhlaup with a peak discharge of 45,000 m^3/s on November 5, 1996.

Courtesy of Oddur Sigurdsson.

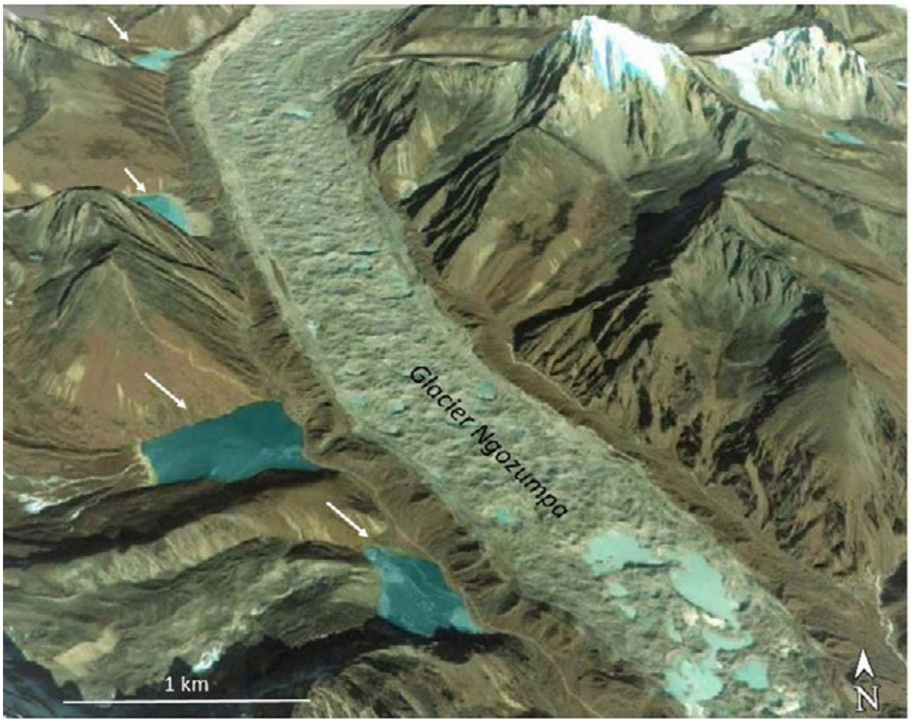

FIG. 14.7

Four moraine-dammed lakes (arrowed) on the west side of Ngozumpa Glacier in Nepal. The glacier is 19 km long and flows south from Cho Oyu, the sixth highest mountain in the world. A large lake is likely to form behind the glacier's Little Ice Age end moraine of the glacier (just beyond the southern limit of this image) later in this century (Google Earth, NASA/Copernicus image, April 2009).

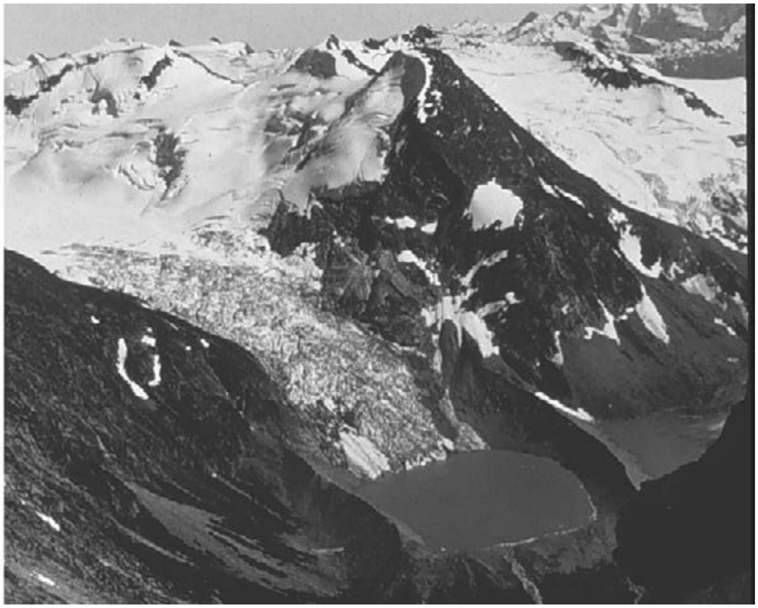

FIG. 14.8

Moraine-dammed Nostetuko Lake in the southern Coast Mountains of British Columbia (photo taken in July 1977 by J.M. Ryder). The large moraine impounding the lake was built during the Little Ice Age by Cumberland Glacier, which has subsequently thinned and retreated. Compare with Fig. 14.9, a photo taken after the outburst flood of July 19, 1983.

a rockfall (Fig. 14.9; Blown and Church, 1985; Costa and Schuster, 1988; Clague and Evans, 2000; Kershaw et al., 2005). Such breaching can also happen during periods of rapid snowmelt or intense rainfall when large amounts of water flow over the dam, initiating outlet incision. Other failure mechanisms include earthquakes, slow melt of buried ice (Reynolds, 1992), and removal of fine sediment from the dam by groundwater ("piping"). Outbursts from these lakes are sudden and rapid; they can produce impressive floods far from their sources. Maximum historical breakout volumes have approached 50,000,000 m^3 with breach depths of up to 40 m (O'Connor and Beebee, 2009).

Outburst floods from moraine-dammed lakes were first documented comprehensively in the Cordillera Blanca of Peru, where they have caused much damage and loss of life (Lliboutry et al., 1977). The most deadly of these events destroyed much of the city of Huaraz in 1941, killing at least 1800 people (Fig. 14.10; Wegner, 2014). Cenderelli (2000), Clague and Evans (2000), Richardson and Reynolds (2000), Kattelmann (2003), and Harrison et al. (2018) provide recent summaries of floods from moraine-dammed lakes.

Because moraine-dammed lakes form in the wake of retreating glaciers, the relation of the hazard to atmospheric warming over the past century is a subject of considerable interest (Lliboutry et al., 1977; Liu and Sharma, 1988; O'Connor and Costa, 1993; Clague and Evans, 1994; O'Connor et al., 2001; Kattelmann, 2003; Veh et al., 2019). Most moraine dams that have breached in recent decades formed during the Little Ice Age when glaciers achieved their maximum Holocene extents. Terminal and

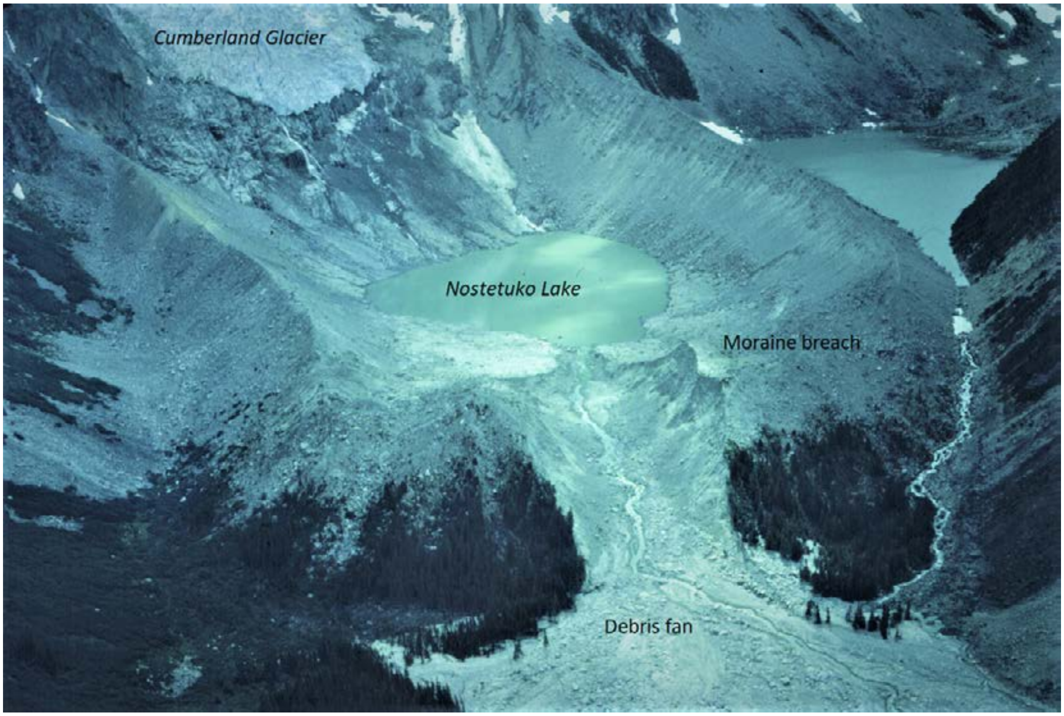

FIG. 14.9

Breached Little Ice Age moraine of Cumberland Glacier and the remains of Nostetuko Lake in the southern Coast Mountains of British Columbia. About 6.5 million m^3 of water flowed out of Nostetuko Lake on July 19, 1983, after an ice avalanche from the toe of Cumberland Glacier entered the lake and produced a displacement wave that overtopped and incised the moraine. The level of the lake fell 28 m during the outburst.

Photo by John J. Clague.

lateral moraines constructed during this period are up to 100 m high and thus are formidable barriers. During the 20th century, valley glaciers thinned and retreated from their maximum Little Ice Age positions, allowing lakes with volumes up to 100,000,000 m^3 and depths of nearly 100 m to form in the basins between the moraines and the retreating ice (Yamada and Sharma, 1993). Hundreds of these lakes are present in the mountains of Tibet (Liu and Sharma, 1988), Nepal and the Hindu Kush-Himalayan region (Yamada and Sharma, 1993; Richardson and Reynolds, 2000; Gardelle et al., 2011; Wang and Jiao, 2015; Wang et al., 2017; Veh et al., 2019), the South American Andes (Lliboutry et al., 1977; Reynolds, 1992; Worni et al., 2012; Emmer et al., 2016), the European Alps (Haeberli, 1983), British Columbia (Clague and Evans, 1994; McKillop and Clague, 2007a, b), and the Cascade Range of Oregon and Washington in the United States (O'Connor et al., 2001). Many of the lakes are still growing in size as glaciers continue to thin and retreat.

In some cases, a moraine dam breaches soon after the lake forms behind it (O'Connor et al., 2001), but most dams fail years or decades later (Clague and Evans, 2000; Harrison et al., 2018). As a general rule, the likelihood of breaching increases as a lake increases in size and the glacier retreats out of the

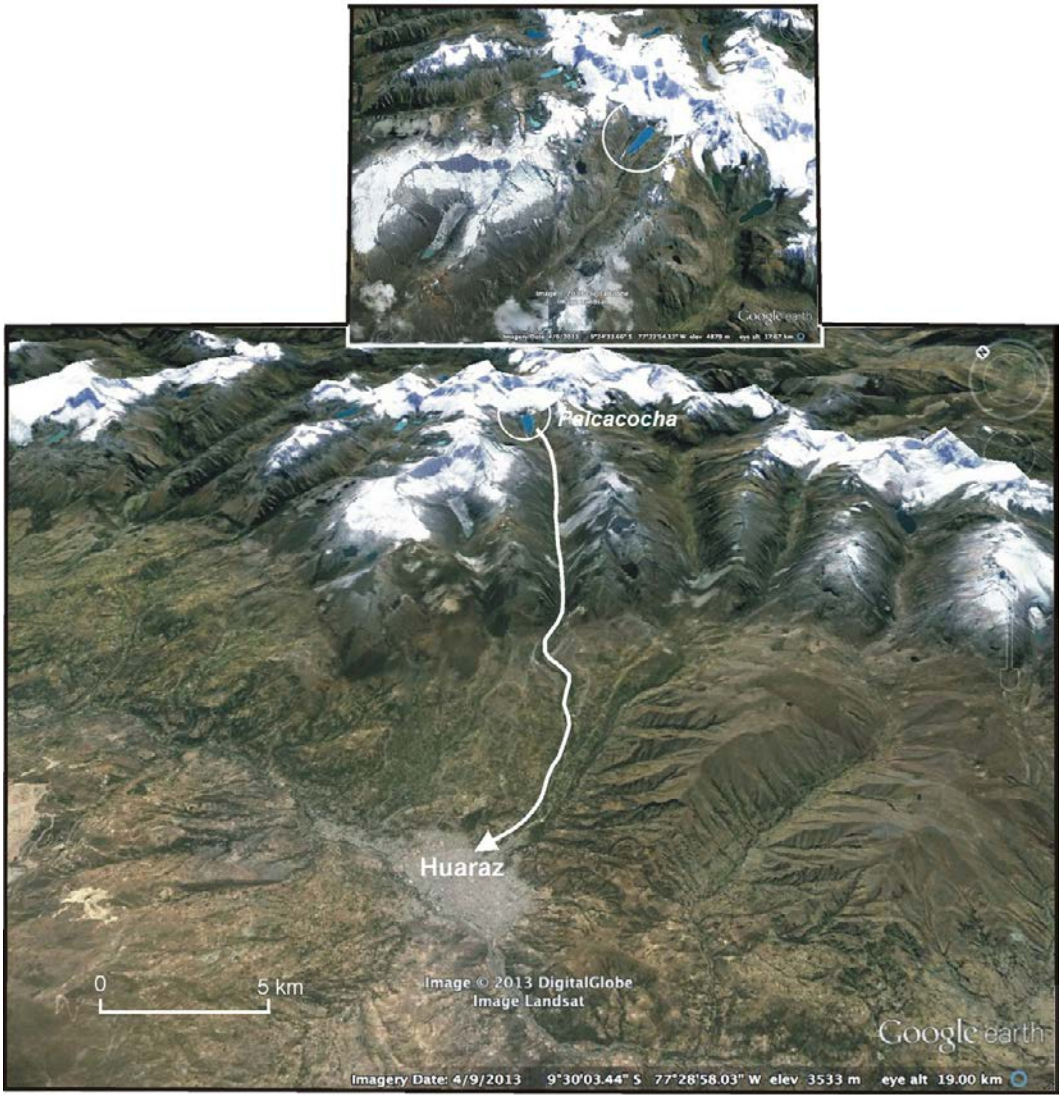

FIG. 14.10

Path of the 1941 aluvión, which killed more than 1800 people in the city of Huaraz, Peru. The disaster resulted from the breach of the Little Ice Age moraine impounding Palcacocha (Lake Palca) *(circled)*. The path of the aluvión is shown by the *white line* (Google Earth image).

lake and up a steep bedrock slope. Later, as the glacier continues to retreat farther from the lake, the likelihood that ice avalanches will trigger overtopping and breaching decreases (Clague and Evans, 2000). In many mountain ranges, moraine dams that are most likely to fail have already done so, reducing the candidate population. For example, 11 of the 14 moraine-dammed lakes in the Oregon Cascade Range that formed between 1924 and 1956 partially or completely drained between 1934 and

1971 (O'Connor et al., 2001). Only two partial breaches from a single, small lake occurred after 1981 (Sherrod and Wills, 2014), indicating that breaches in the Oregon Cascades are becoming more rare with time. The incidence of moraine breaches in the Himalaya, however, remains steady and may increase in the future as lakes continue to grow behind moraines and remain in contact with glacier ice (Liu and Sharma, 1988; Mool, 1995; Richardson and Reynolds, 2000; Kattelmann, 2003; Harrison et al., 2018; Veh et al., 2019).

Not all moraine dams fail. The large number of existing lakes impounded by Pleistocene moraines attests to long-term stability of some moraine dams. In addition, many Little Ice Age moraine dams have persisted for more than a century without breaching. McKillop and Clague (2007a) found that only 10 of 175 lakes in their inventory of moraine-dammed lakes in southern British Columbia had partly or completely drained at the time of their survey. Armoring of outlet channels with boulders can stabilize a moraine dam, and some dams are stable because they are broad and have low-gradient outlet channels (Clague and Evans, 2000). Moraine dams may be stable even in very large earthquakes. For example, the Gorkha (Nepal) earthquake in 2015 caused no moraine dam collapses in Nepal or China (Kargel et al., 2016). Nevertheless, even a wide moraine dam can be breached if overtopped by a wave caused by a large landslide or ice avalanche (Kershaw et al., 2005). Moraine dams formed on stratovolcanoes are particularly susceptible to breaching because they are narrow and steep-sided, are situated on steep slopes, and consist of easily eroded volcanic rock debris. Over 60% of the moraine-dammed lakes on stratovolcanoes in Oregon have failed (O'Connor et al., 2001), far surpassing the 6% that have failed in the mostly crystalline rocks of the British Columbia Coast Mountains (McKillop and Clague, 2007a).

Occasionally, a large outburst flood occurs without the moraine dam being breached. The peak discharge of such a flood relates to the amount of water that passes over the dam, which itself is a function of the size of the overtopping wave. The latter, in turn, is dictated by the size of the landslide or ice avalanche that produces the wave. An example of such an event is the "Steinholtshlaup" in Iceland in 1967. In January of that year, a rockslide fell onto Steinholtsjökull, a glacier flowing onto the coastal plain of southwest Iceland. The rockslide debris entrained a large amount of ice, creating a mass flow that ran off the glacier and into the proglacial lake, Steinsholtslón (Kjartansson, 1967). The mass flow displaced water from Steinsholtslón and overtopped the moraine dam that impounds the lake, causing a downstream debris flood that transported boulders and ice blocks far beyond the dam. A second example is the 2002 Laguna Safuna Alta outburst in the Cordillera Blanca, Peru (Hubbard et al., 2005). During that event, a large landslide fell onto Glaciar Pucajirca and then entered moraine-dammed Laguna Safuna Alta, creating a displacement wave that increased in height as it traveled down the lake. The moraine dam had 80 m of freeboard, but the wave still overtopped it, and the overflowing water ran into Laguna Safuna Baja, another moraine-dammed lake about 350 m downvalley from Laguna Safuna Alta. The moraine impounding the lower lake had about 15 m of freeboard, sufficient to completely contain the inflow of water from the upper lake.

Another example of an overtopping flood, but in this case accompanied by breaching, is the 1997 Queen Bess event in the British Columbia Coast Mountains (Kershaw et al., 2005). In July 1997, about 2 million m^3 of ice detached from the toe of Diadem Glacier, located on a steep rock slope about 200 m above Queen Bess Lake. The ice plunged into the lake and produced a large surge wave over 30 m high that overtopped the moraine dam and ran down the valley below the lake (Fig. 14.11). The overtopping wave initiated a breach that lowered the level of the lake by 8 m and produced a second flood peak. The flood surged 20 km to Homathko River and thence west to tidewater at Bute Inlet. The flood attenuated as it moved downvalley, but it still generated a marked spike on a hydrograph 100 km from the source.

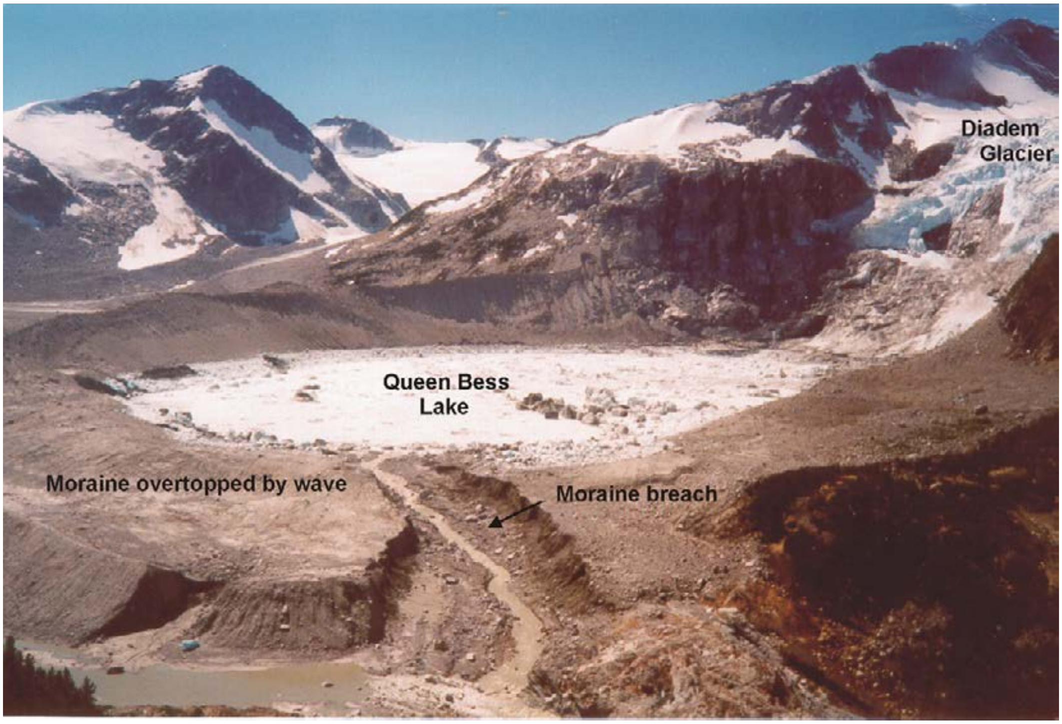

FIG. 14.11

Queen Bess Lake and its breached Little Ice Age moraine. This photo was taken one day after the outburst flood, which happened on August 13, 1997. An ice avalanche from the toe of Diadem Glacier triggered a large displacement wave that overtopped a 600-m length of the end moraine by 15–25 m and breached it. The level of Queen Bess Lake fell 8 m during the breach phase of the outburst.

Photo courtesy of Interfor Forest Products.

14.3 Outburst mechanisms and flood magnitude

14.3.1 Glacier dams

Glacier-dammed lakes drain by overtopping or flow through subglacial tunnels; the latter mechanism is the more common of the two. Initial outflow via a subglacial tunnel may be triggered by flotation of a part of the ice dam that is in contact with the lake, but flotation does not appear to be a requirement for the initiation of drainage (Fowler, 1999; Flowers et al., 2004; Ng and Liu, 2009). As water begins to flow along a subglacial channel, the channel walls enlarge by both thermal and mechanical erosion (Liestøl, 1956; Nye, 1976; Clarke, 1982; Roberts, 2005; Björnsson, 2010). The channels can, however, narrow or close during or after the flood by ice deformation caused by glacier flow, especially if the ice is thick.

Much of our understanding of glacial outburst floods has come from analyses of Icelandic events, beginning with Thórarinsson (1939) who recognized that a glacier dam might float due to hydrostatic stresses once the depth of the impounded water reaches a threshold value. Floating of the edge of the glacier dam may explain why few dams fail by overtopping (Björnsson, 1974, 1976; Walder and Costa, 1996), but it does not fully explain the hydrographs of observed glacial outburst floods. Glen (1954) and Liestøl (1956) proposed that subglacial tunneling may be an important process, and Mathews (1973) inferred that tunnels enlarge by thermal erosion, with the energy supplied by the potential energy of the impounded lake.

Nye (1976) first explained the physics of glacier tunnel enlargement. Although his model has been refined (Spring and Hutter, 1981, 1982; Clarke, 1982, 2003), the basic principles remain the same: (1) Sensible heat derived from water flowing through a tunnel within or at the base of a glacier is transferred to the tunnel walls; (2) the tunnel enlarges, further increasing the flow of water through it; and (3) more sensible heat enlarges the tunnel, as does viscous heat derived from friction associated with turbulent water flow. This positive feedback process produces an exponential rise in water discharge (Clarke, 1982; Björnsson, 1992, 2010). Outbursts produced by this process generally develop more slowly than those caused by breaching of moraine dams involving similar impoundment volumes (Fig. 14.1). Peak discharges are controlled by lake volume, dam geometry, the material properties of the dam, and downstream topography and sediment availability.

The tunnel enlargement process has solid theoretical underpinnings, but the initiation of outflow is not well understood. Outflow can begin at much lower lake levels than required for hydrostatic flotation, indicating that other factors are important (Roberts, 2005). Similarly, the end of a glacial outburst flood is not entirely predictable or understood. In some instances, a flood ends when the water supply in the lake is exhausted (Clarke, 1982), but in many others, ice deformation may seal the outlet before the lake completely empties (Nye, 1976; Roberts, 2005). Furthermore, in the case of thin glacier dams, the roof of a tunnel may collapse, ending outflow (Mathews, 1973; Sturm and Benson, 1985).

Some englacial or subglacial water bodies drain too rapidly to be the result of simple tunnel enlargement (Björnsson, 1977, 1992, 2009, 2010; Haeberli, 1983; Walder and Driedger, 1995; Walder and Fountain, 1997; Roberts, 2005). These floods are not well understood but may involve sudden rupture of pressurized water-filled englacial cavities (Walder and Fountain, 1997), fracturing of the base of the glacier due to high hydrostatic pressures (Glen, 1954; Fowler, 1999), or transient hydraulic conditions (Roberts, 2005; Björnsson, 2009). Some glacial outburst floods, notably those resulting from subglacial eruptions, may involve sheet flow over a large area of the glacier bed rather than channelized flow (Björnsson et al., 2001; Johannesson, 2002; Flowers et al., 2004).

Some glacier-dammed lakes fail by overtopping or ice-marginal drainage, much like outburst floods associated with other types of natural dams (Haeberli, 1983; Walder and Costa, 1996). These floods typically have much larger peak discharges than floods resulting from thermal erosion of subglacial channels. Their character is similar to floods associated with failures of constructed dams. The similarity is intuitive because peak discharge from an overtopped ice dam approximates critical flow through a breach that is enlarging by mechanical erosion. Historical examples of this type of flood are the 1986 and 2002 Russell Lake, Alaska, outbursts. The earlier of the two floods had a peak flow of about 10,000 m^3/s (Mayo, 1989). Lake Alsek in Yukon Territory also may have drained in this way (Clague and Rampton, 1982).

A relation has been noted between the size of glacier-dammed lakes and the peak discharge of the floods they produce—the largest lakes generally produce the largest floods (Fig. 14.1; Clague and Mathews, 1973). This observation has led researchers to develop empirical regression equations based on variables such as impoundment volume and impoundment depth to estimate peak discharge from glacier-dam failures (Clague and Mathews, 1973; Haeberli, 1983; Evans, 1986; Costa and Schuster, 1988; Walder and Costa, 1996; Walder and O'Connor, 1997; Cenderelli, 2000). However, the regression equations can only provide rough estimates of peak discharge because glacier dams differ markedly in length, thickness, and other characteristics, and because outflow is strongly influenced by hydraulic factors and breach erosion rates (Walder and O'Connor, 1997). Another complication is the difficulty of accurately estimating peak discharges of the floods on which these regression equations are based. Flood discharges at the breach are rarely known; typically discharges used to derive empirical equations are estimated at different distances downstream of the breach and are underestimates of peak discharge.

14.3.2 Moraine dams

Once a lake forms behind a moraine, several processes facilitate breaching of the dam. First, a hydraulic gradient is established across the dam, promoting groundwater flow through the sediments forming the moraine. Second, seepage of groundwater from the downstream face of the moraine may increase the likelihood of mass movements that destabilize the dam (Massey et al., 2010). Third, retrograde erosion of sapping channels in permeable sediment may incise the barrier and trigger overflow. Fourth, a rapid inflow of water or the generation of high waves by wind, landslides, or ice avalanches may increase water flow in the outlet channel and initiate incision.

Most outbursts from moraine-dammed lakes, like those from landslide-dammed lakes and artificial reservoirs, result from overtopping and breaching of the barriers. Outflow from the lake increases once a stream begins to incise a moraine dam. The increased outflow further incises the channel, leading to increased outflow in a self-enhancing process. The breach commonly grows through lakeward migration of one or more knickpoints on the downstream face of the dam. At the same time, the breach widens by slumping and sliding from the steep banks of the deepening outlet channel (Lee and Duncan, 1975; Plaza-Nieto and Zevallos, 1994; Dwivedi et al., 2000). The breach continues to enlarge until either the level of the lake drops to a critical level with an associated reduction in the hydraulic gradient or the outlet channel becomes sufficiently armored to halt erosion. In some instances, incision ceases when the channel is lowered onto bedrock. Once the outlet stabilizes, outflow continues at a diminishing rate until the lake drops to the level of the stable outlet.

Peak discharge during a breach event is controlled by the interplay between erosion and enlargement of the outflow channel and drawdown of the impounded lake (Fig. 14.12). The important factors are as follows: (1) the speed at which the breach grows; (2) the final depth of the breach; and (3) downstream water and sediment interactions that affect the volume, peak discharge, and type of flow.

Parameterized and physical models used to estimate outflow and peak discharges during breaching of natural and constructed dams are based on coupling between breach development and the drawdown of the impounded water body (Ponce and Tsivoglou, 1981; Fread, 1987; Froehlich, 1987, 1995, 2008, 2016; Singh et al., 1998; Webby and Jennings, 1994; Webby, 1996; Walder and O'Connor, 1997; Manville, 2001; Marche et al., 2006; Begam et al., 2018; Wang et al., 2018). The models typically assume critical flow through a growing breach, and breach growth is parameterized using either shape and time functions (e.g., the DAMBRK model of Fread, 1988) or physically based sediment transport and mass movement rules (e.g., the BREACH model of Fread, 1987, the BEED model of Singh et al., 1998, and the ERODE model of Marche et al., 2006). The modeler iteratively fits parameters until the measured and calculated hydrographs are in agreement, although the large number of parameters rarely permits a unique fit.

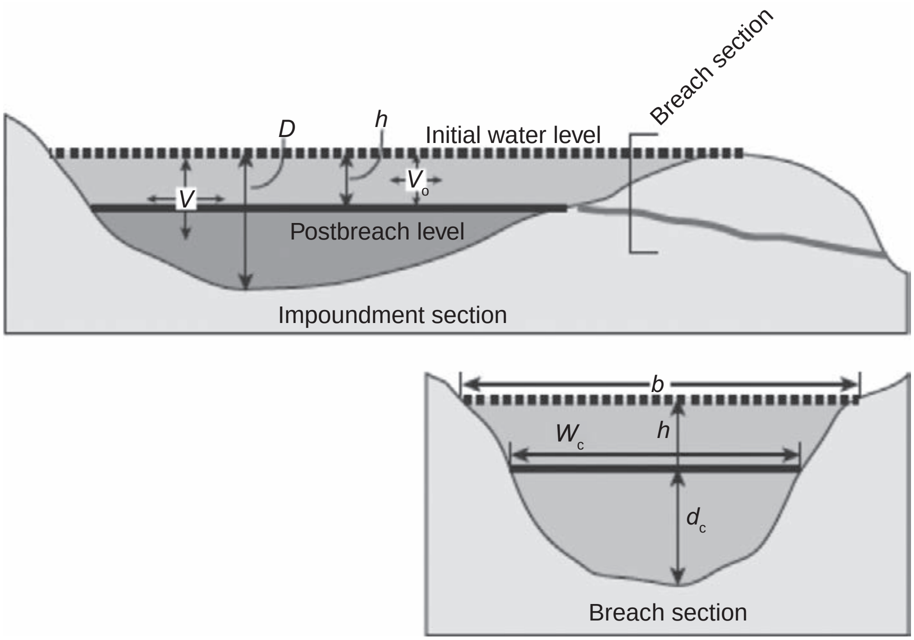

FIG. 14.12

Definition diagram for impoundment and breach geometry. V is the total impoundment volume; V_o is the volume released during the breach event; D is the maximum impoundment depth; h is the change in impoundment level during the breach event and is approximately equal to the maximum possible specific energy of flow through the breach; b is the breach width at the maximum impoundment elevation; W_c is the breach width at flow stage d_c associated with critical flow [$V=(gd)^{0.5}$] through the breach section with specific energy h (O'Connor et al., 2013).

Recent experimental studies have provided new insights into earthen and moraine dam failures applicable to moraine dam breaching (Coleman et al., 2002; Hanson and Cook, 2005; Morris et al., 2007; Zhang et al., 2009; Pickert et al., 2011; Begam et al., 2018). The studies show that parameterized erosion laws and assumptions about critical flow poorly replicate breach evolution in observed events. As a result, several investigators (e.g., Wang and Bowles, 2006; Faeh, 2007; Wu et al., 2009) have developed models that differ from the parametric models mentioned above. The new models assume that embankment breaching can be replicated by calculating fluid flow along an erodible channel, where only a small part of the flow passes through the embankment. From this perspective, little difference exists between dam-breach modeling and flood routing.

When considering breach formation, it is useful to discriminate between large and small impoundments (Walder and O'Connor, 1997). In the case of a "large impoundment," as the term is used by Walder and O'Connor (1997), the breach develops fully before the level of the lake drops significantly, either because the lake is large relative to the breach or because the breach develops rapidly. In either case, peak flow is approximately equivalent to the critical flow through the fully developed breach. If we assume that the breach cross-section is about as wide as it is deep, $Q \approx g^{0.5}h^{2.5}$, where Q is the peak outflow discharge, g is gravitational acceleration, and h is the height difference between the level of the impounded water body and the bottom of the breach at the point of outflow. If the breach is significantly wider than it is deep, peak discharge will be larger by a factor of approximately b/h, where b is the breach width. The strong dependence of peak discharge on breach depth explains why deep lakes can produce such large outflows. In the case of "small impoundments," the surface of the lake falls significantly as the breach develops and thus evolving breach geometry controls peak discharge. The peak in discharge generally occurs before the breach fully develops and is dependent largely on the vertical erosion rate.

Final breach depth is an important factor for both peak flow and the total volume of water discharged (Webby, 1996). Escaping waters may erode only part way through a moraine dam, to the base of the dam, or, in rare cases, even deeper. Most breaches, however, extend only part way through the dam (O'Connor and Beebee, 2009). The character of the dam materials is important in controlling the depth of breaching. The interplay between outflow rates and breach deepening, widening, and armoring will dictate the rate of breach growth and the ultimate geometry of the breach (Clague and Evans, 1992; Worni et al., 2012). These aspects make it difficult to accurately predict outburst-flood magnitude because (1) materials forming moraines are heterogeneous; (2) a variety of processes contribute to breach enlargement (Walder and O'Connor, 1997); and (3) predictions are sensitive to the bed-load transport formulae that are used to parameterize models (Cencetti et al., 2006).

14.4 Downstream flood behavior

Outburst floods from glacier- and moraine-dammed lakes entrain, transport, and deposit huge amounts of sediment, in rare cases hundreds of kilometers from the flood source (Fig. 14.13). Erosion and deposition by the floodwaters can themselves significantly affect flow behavior. Outburst floods commonly broaden floodplains, destroy preflood channels, and alter river planform. The changes can persist for years or decades after the flood, although rivers quickly reestablish their preflood gradients by incising flood deposits.

Many outburst floods from moraine-dammed lakes change into debris flows by entrainment of sediment from the breached moraine and by incorporation of downstream bank and bed materials (Fig. 14.14; Lliboutry et al., 1977; Eisbacher and Clague, 1984; Clague et al., 1985; Clague and Evans, 1994; Gallino and Pierson, 1985; Schuster, 2000; Huggel et al., 2004; McKillop and Clague, 2007b; Procter et al., 2010; Carey et al., 2012). Such transformations also have occurred during glacier outburst floods, particularly on volcanoes (Walder and Driedger, 1994, 1995). Downvalley sediment bulking can increase peak discharge by an order of magnitude or more. An extreme example is the 1985 Dig Tsho outburst in Nepal, where about 4 million m^3 of water redeposited 3.3 million m^3 of sediment over a distance of 40 km downstream from the dam (Vuichard and Zimmerman, 1987). Analysis of the flow transformation process is difficult because few observations of flow rheology have been made. Nevertheless, it is clear that steep channel slopes (>0.10–0.15) are required to sustain debris flows and hyperconcentrated flows (Clague and Evans, 1994; O'Connor et al., 2001; McKillop and Clague, 2007b; Procter et al., 2010). Most flow bulking happens within the first 10 km of the breach, because entrainment requires steep channels that are most common in watershed headwaters where many moraine dams are located.

Debris flows or hyperconcentrated flows will deposit sediment in unconfined reaches of a valley or where the gradient of the valley floor decreases to less than ~0.15 (Clague and Evans, 2000). Entrainment and deposition of sediment and woody plant debris by floodwaters has important implications for hazard appraisal, because debris flows are more destructive than floods of the same magnitude.

FIG. 14.13

Valley of the west fork of the Nostetuko River about one year after the outburst flood from moraine-dammed Queen Bess Lake, British Columbia, in August 1997. The flood removed forest from the valley floor and left a sheet of gravel and boulders in its wake. Flow is toward the viewer.

Photo by John J. Clague.

Downstream attenuation of outburst floods controls the distribution and size of the resulting flood features. In general, a given flood will attenuate more rapidly if there is substantial downstream space for temporary storage of floodwaters or where the valley gradient is low. Floods of small volume and short duration attenuate more rapidly than floods of large volume and long duration. The latter do not attenuate as rapidly because the first floodwaters fill available space in the valley prior to the passage of the peak.

The characteristics of flood-produced features and the caliber of the transported sediment depend on the flow type and size, flow attenuation rates, and channel morphology and materials (O'Connor, 1993; Cenderelli and Wohl, 1998, 2003; Kershaw et al., 2005). The size of erosional and depositional features, including flood bars, "scablands," megaripple bedforms, and flood-transported boulders, scales with cross-sectional flow geometry, whereas maximum clast size scales with flow strength. Assessments of the relation between flood landforms and flow are based on measures of flow strength, notably shear stress and stream power. Both the caliber of entrained sediment and the distribution of erosional features are governed by the distribution of shear stresses and stream-power magnitude

FIG. 14.14

An outburst flood from moraine-dammed Klattasine Lake in the British Columbia Coast Mountains transformed into a debris flow that traveled 8 km to the fan in the foreground of this photo. Up to 20 m of debris accumulated on the fan, temporarily stemming the flow of the river flowing toward the bottom of the photo.

Photo by John J. Clague.

(Baker, 1973; O'Connor, 1993; Benito, 1997; Cenderelli and Wohl, 2003; Carrivick, 2009; Denlinger and O'Connell, 2010). Shear stress, τ, is the tangential force applied to the bed per unit area—for hydrostatic conditions, $\tau = \rho g d_f S$, where p is the fluid density, d_f is the flow depth, and S is the local flow energy gradient, which for steady and uniform flow corresponds to channel slope. Unit stream power, ω (Bagnold, 1966), can be expressed as $\omega = \rho g d_f v S = \tau v$, where v is the flow velocity. These formulae are indices of geomorphic work, although only a portion of available mechanical energy is expended as geomorphic work, with the rest dissipated in other forms of energy. The formulae do not consider vertical and horizontal accelerations that can produce stresses of the same magnitude as the hydrostatic forces (Iverson, 2006).

Outburst floods generate shear stresses and stream powers greatly exceeding those of "normal" meteorological floods because of their large flow depths and locally steep energy gradients (Baker and Costa, 1987). Cenderelli and Wohl (2003) concluded that floods generated by breaching of moraine dams produce stream power values several times greater than those from the largest meteorological floods affecting the same reaches. Montgomery et al. (2004) suggested that stream power values up to $5 \times 10^6\,\mathrm{W\,m^{-2}}$ might have been attained during Holocene floods from breached glacier dams in

FIG. 14.15

Large boulders transported in the hyperconcentrated phase of the outburst flood from Queen Bess Lake (British Columbia) in August 1997. Horizontal black line at lower left corresponds to 1 m.

Photo by John J. Clague.

the Tsangpo River gorge, Tibet. These values are two orders of magnitude higher than those reconstructed from discharge measurements of flash floods in steep basins (Baker and Costa, 1987).

Outburst floods commonly transport huge boulders, some more than 10 m across (Fig. 14.15). Theory and observations (Baker and Ritter, 1975; Costa, 1983; Williams, 1983; Komar, 1987; O'Connor, 1993) demonstrate that entrainment and tractive transport of boulders increase with local flow strength. Empirical analyses by Costa (1983) and Williams (1983) indicate that a boulder 1 m in diameter can be moved tractively when local stream power reaches about 500 $W\,m^{-2}$. Although subject to many uncertainties (Jacobson et al., 2003), such flow competence relations have been used in many paleohydrologic analyses of large floods (Lord and Kehew, 1987; Waythomas et al., 1996; Manville et al., 1999; Hodgson and Nairn, 2005; Carrivick, 2009; Lamb and Fonstad, 2010). It is unclear, however, whether the commonly used flow competence relations apply in the case of hyperconcentrated flows, in which large clasts are transported on a mobile basal carpet of sediment. Such flows may be responsible for many extremely coarse outburst flood deposits.

14.5 Outburst floods and climate change

Glaciers around the world repeatedly grew and shrank in response to climate warming during the Holocene. Most alpine glaciers in the Northern Hemisphere achieved their maximum Holocene size in the past three centuries (Grove, 1988). Since the end of the 19th century, however, alpine glaciers have thinned and receded, and today, ice cover in most mountain ranges is one-half to two-thirds of what it was in the middle of the 19th century (Zemp et al., 2017; Hock et al., 2019)

A change in climate can affect the stability of glacier dams. Fewer glacier-dammed lakes drain when climate is stable than when climate changes markedly, causing glaciers to advance or retreat. During the Little Ice Age, new lakes formed where glaciers advanced across streams and blocked drainage. When these lakes first formed, the glacier dams may have been weak and many of the lakes perhaps drained repeatedly. As these glaciers continued to advance, the dams stabilized. Many Little Ice Age lakes in northwest North America drained one or more times in the 20th century when their dams weakened due to glacier thinning and retreat (Costa and Schuster, 1988; Clague and Evans, 1994).

A glacial outburst flood cycle can develop during a prolonged period of glacier retreat. For an individual lake, the cycle is characterized by recurrent outburst floods separated by times when the lake refills (Evans and Clague, 1994). Outbursts may occur annually or on shorter or longer timescales depending on interactions between the blocking glacier and filling of the reservoir (Hulsing, 1981; Mathews and Clague, 1993; Depetris and Pasquini, 2000; Walder et al., 2006; Huss et al., 2007). Flood character changes with overall glacier conditions—outbursts commonly decrease in peak discharge and volume as the glacier thins and retreats (Clague and Evans, 1994; Evans and Clague, 1994). Many former glacier-dammed lakes no longer exist because the glaciers that dammed them have thinned and retreated so much that they no longer impound water. However, lakes have formed in new locations at the margins of some receding glaciers, typically at higher elevations than former lakes, and pose new risks to downvalley areas (Fig. 14.16; Geertsema and Clague, 2005).

Atmospheric warming through the remainder of this century will further decrease the extent of permanent snow and glacier ice in all presently glacierized mountain ranges (Huss et al., 2017; Hock et al., 2019). Researchers have shown that new lakes will form within closed basins beneath many present-day glaciers (Frey et al., 2010; Linsbauer et al., 2012, 2016; Colonia et al., 2017). Ice avalanches or landslides into these newly formed lakes may trigger displacement waves that overtop the lake sills and cause downstream flooding (Haeberli et al., 2016, 2017; Emmer, 2017a, b).

A relation also exists between climate change and the stability of moraine dams. Most existing moraine-dammed lakes formed in the 20th century when glaciers retreated from bulky Little Ice Age end moraines. The lakes soon began to fail as climate warmed. If warming and glacier retreat continue, the supply of moraine-dammed lakes susceptible to failure in many mountain ranges will decrease and the threat they pose will diminish (Clague and Evans, 2000). An exception may be moraine-dammed lakes in the high mountains of South Asia. In parts of the Himalaya, for example, moraine-dammed lakes are still evolving due to downwasting and fragmentation of ice tongues behind moraines.

14.6 Risk assessment and reduction

Assessment and mitigation of an outburst flood hazard require knowledge of both the hazard and the physical and socioeconomic vulnerability downstream (ICIMOD, 2011). The potential for damage and loss of life can only be reduced, and resilience enhanced, after the possible outburst has been characterized and its impact reliably assessed.

Elements at risk from outburst floods include people and their property, infrastructure, systems such as tourism and trade that support people's livelihood, and environmental resources such as forest, pasture and grazing land, and fisheries. The capacity of people to cope with an outburst flood, as any disaster, depends largely on individual and community assets and access to information, technology, services, and institutions.

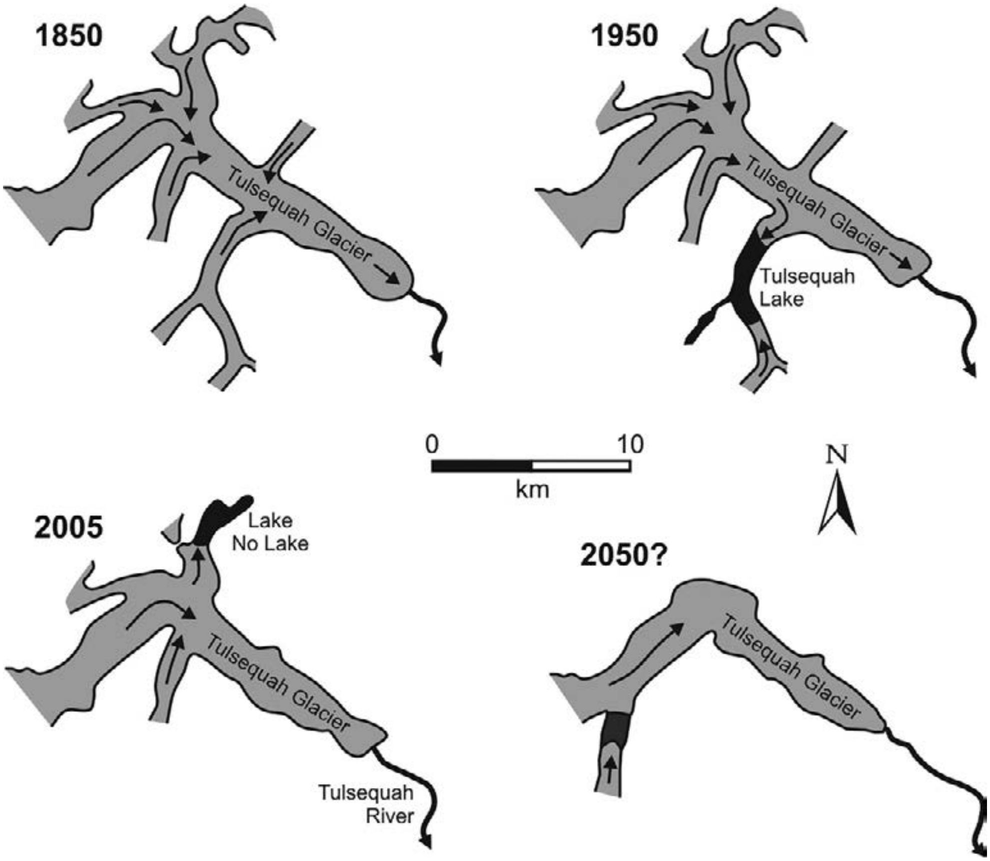

FIG. 14.16

Evolution of Tulsequah Glacier and its ice-marginal lakes from the end of the Little Ice Age to ca. AD 2050. During the first half of the 20th century, Tulsequah Lake produced large annual outburst floods *(upper right)*. By the early years of the 21st century, Tulsequah Glacier had downwasted and receded, and Tulsequah Lake no longer held enough water to generate significant outburst floods. During the same period, however, Lake-No-Lake formed in a tributary valley about 5 km farther upvalley and is now the primary source of large outburst floods. If Tulsequah Glacier continues to recede, Lake-No-Lake itself will cease to produce outburst floods and the locus of outburst activity might shift farther up the glacier.

An assessment of outburst flood hazard and risk requires the following: (1) mapping and classifying glacier- and bedrock-dammed lakes using satellite imagery and aerial photographs; (2) field inspection of lakes that are hazardous; (3) assessment of the likelihood and size of an outburst flood; and (4) assessment of socioeconomic and environmental vulnerability in the hazard zone (Hegglin and Huggel, 2008). In the case of high-risk lakes, it is advisable to monitor them to document changes that might signal an imminent outburst. Monitoring could involve repeat mapping of the lakes using remote sensing, field inspection of seepage, degradation of ice cores of moraines, slope instability in moraine dams, measurements of water inflow into lakes, and inspection of adjacent glaciers and slopes for signs of instability.

Much research has been done on identifying, mapping, characterizing, and monitoring glacier-dammed lakes using aerial photographs and satellite images (e.g., Buchroithner, 1996; Huggel et al., 2002, 2003; Iwata et al., 2002; Kääb et al., 2005; Quincey et al., 2005, 2007; Randhawa et al., 2005; McKillop and Clague, 2007a, b; Bolch et al., 2008, 2011; Fujita et al., 2008, 2009; Mergili and Schneider, 2011; Wang et al., 2012a, b, 2014; Mergili et al., 2013; Xie et al., 2013). The marked increase in the number and variety of active satellite sensors and improvements in image resolution now allow researchers to monitor hazardous lakes on a continuous basis.

In recent years, a major advance in predicting the downstream impacts of outburst floods and debris flows has been the coupling of different physically based numerical models to simulate process cascades (Worni et al., 2012, 2013, 2014; Schneider et al., 2014; Somos-Valenzula et al., 2016) Schneider et al. (2014), for example, used a physically based, mass-movement model (RAMMS) to model the 2010 ice avalanche into Laguna 513 in the Cordillera Blanca, Peru, and a hydrodynamic model (IBAER) to simulate the resulting impact wave that overtopped the moraine dam. They tested the model results against an observational dataset for the event. Numerical models have also been applied, for example, to outbursts in the Indian Himalaya (Worni et al., 2013) and the Argentine Andes (Worni et al., 2012).

Another noteworthy development in outburst flood hazard assessment is the use of glacier hypsometry to predict subglacial closed basins where lakes will develop if valley and cirque glaciers continue to thin and retreat through the remainder of this century, as expected. These lakes could be new sources of future outburst floods if impacted by landslides or ice avalanches from adjacent slopes (Haeberli et al., 2016, 2017, Emmer, 2017a, b).

High-risk situations also require preparedness and precautionary measures to minimize damage and loss of life should the lake empty (Carey et al., 2012). One element of preparedness is land-use planning and structural mitigation. Another is early warning—the provision of information on a potential threat in sufficient time to allow an appropriate response. Early warning systems include video cameras and water-level recorders that detect changes in lake level. Data are transmitted in real time to communities that might be impacted by outburst floods, and sirens or other devices provide warning of floods or debris flows. Early warning systems must be simple to operate, easy to maintain or replace, and reliable. The linked communication network must be capable of relaying the warning to appropriate authorities and is effective only if placed in the hands of the local communities with appropriate response systems.

Early warning systems have met with limited success. For example, an automatic early warning system was set up at and below Tsho Rolpa, a moraine-dammed lake in Nepal, in 1998 before mitigation work was undertaken to reduce the risk of an outburst (Reynolds, 1999). However, by 2002 the system was no longer operating, in part because local residents assumed the lake had been lowered to a safe level.

An important structural mitigation measure for reducing the chance of an outburst flood is to lower the level of the lake by pumping or siphoning water, or tunneling through the moraine barrier or under an ice dam (Lliboutry et al., 1977; Kääb et al., 2004; Vincent et al., 2010). The hazard can also be mitigated by constructing an outlet-control structure, armoring the outlet of a moraine-dammed lake to prevent breaching, or controlled breaching of the moraine dam. Downstream infrastructure can be protected against floods, for example, by designing bridges with decks higher than levels likely to be reached by an outburst flood. Land-use zoning should also be considered, as it reduces the number of structures at risk.

14.7 Summary

Outburst floods from glacier- and moraine-dammed lakes can be large and dangerous, with effects tens to hundreds of kilometers from the source. The largest floods have been caused by failures of glacier dams during the Pleistocene Epoch, but great floods also have occurred during the Holocene and future floods pose a serious threat to downvalley communities and infrastructure in many mountain ranges.

The peak discharge, duration, and volume of outburst floods are interrelated and depend mainly on the geometry and rate of development of breaches (moraines) and subglacial channels (glaciers), and also on the size and geometry of the impounded water body. The largest floods derive from large water bodies. In the case of moraine dams, flows from large lakes through rapidly developing breaches have peak discharges that approximate critical flow through the final breach. Floods from small lakes have peak discharges that are strongly dependent on the breach rate.

The characteristics of outburst floods, and the deposits and landforms they produce, depend on the dam failure process, the rate and duration of flow out of the lake, and downstream interactions with sediment and the valley floor. Outbursts that incorporate large amounts of sediment, either from a moraine dam or from downstream channels, may evolve into debris flows. Some debris flows have peak discharges several times the peak outflow at the breach and substantially increase the flow volume. However, debris flows attenuate rapidly downstream by deposition, and they commonly come to rest where channel gradients fall below 0.15. Outburst floods attenuate because of channel and valley storage. Large floods with relatively low peak discharges may attenuate at a slower rate than small, high-discharge floods where the peak flow is large relative to the length of the flood. Erosional and depositional features left by large floods reflect the depth and breadth of inundation and the large forces applied by the deep, fast-moving floodwaters. Floods from natural dam failures can achieve local stream power values one to two orders of magnitude greater than those generated by the large meteorological floods.

Climate is an important determinant of the stability of moraine and glacier dams. Most moraine-dammed lakes formed in the past century or two as glaciers thinned and retreated from end moraines constructed during the Little Ice Age. These lakes began to fail as climate warmed. With continued atmospheric warming and glacier retreat, some moraine-dammed lakes will grow in size, increasing the threat they pose. In some mountain ranges, however, the supply of moraine-dammed lakes that are susceptible to failure will decrease, as the glaciers to which they are presently connected retreat upvalley or disappear altogether. Over the past century, many glacier-dammed lakes have gone through a period of cyclic or sporadic outburst activity lasting up to several decades. The outburst floods from any one lake ended when the glacier dam weakened to the point that it could no longer trap water behind it. However, with continued glacier retreat, the locus of outburst activity, in some cases, shifted up-glacier to sites where new lakes developed in areas that became newly ice-free.

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