9 GLACIOLACUSTRINE PROCESSES

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9.1 INTRODUCTION

Lakes are basins that act as sinks for water and sediment and are sites for important biological and chemical processes. In the sedimentary record lakes are commonly associated with glacial processes where they form in deep basins produced by the action of glacial erosion or around the margins of ice masses (glaciers and ice sheets) where water is impounded by ice, moraines, or outwash gravel. The latter are often ephemeral features that fill quickly with sediment, drain once glaciers recede, or evolve into distal lakes upon glacier retreat. In contrast, ice-distal lakes persist for longer and may contain sedimentary records that span multiple glaciations.

It is common to identify two fundamentally different settings for glaciolacustrine processes: an ice-contact environment in which the lake is in direct contact or close proximity to glacier ice, and an ice-distal environment in which the lake is separated from the direct influences of glacial activity, typically separated physically by a system of outwash plains (Table 9.1). Ice-contact lakes are characterized by energetic processes, rapid accumulation, and rapid change in sedimentary processes that produce complex depositional environments characterized by abrupt lateral and vertical changes in sedimentary facies. Such lakes are also characterized by the presence of icebergs which introduce coarse sediments into environments dominated by accumulation from suspension and turbidity currents which result in the accumulation of ‘ice-rafted’ debris, as well as a range of scour marks produced by iceberg grounding. Although there is a significant difference in the nature and rate of depositional processes between the ice-contact and distal settings they are related to each other because frequently ice-contact lakes evolve into distal lakes during deglaciation.

In this chapter we examine how lacustrine processes determine depositional processes in lakes, review the range of depositional processes that characterize ice-contact lakes including developing research on subglacial lakes, and how glaciotectonic processes deform ice-contact lake sediments. We also review processes in ice-distal lakes and examine how such lakes have been used to reconstruct postglacial and Holocene environmental change.
9.2 PHYSICAL LIMNOLOGY AND SEDIMENTOLOGY

Physical and chemical processes in lakes determine how influent water interacts with lake water. The relationship between influent water and lake water is largely determined by the density structure of lakes, which is controlled by water temperature and sediment concentration, although in some circumstances lakes can also be chemically stratified (Sly, 1978).

Thermal stratification in lakes is the result of systematic changes in water density associated with temperature change. Water has its maximum density of 1000 kg m$^{-3}$ at 4°C. Below 4°C the density slowly reduces until the freezing point is reached, at which point there is a rapid reduction as ice crystals form. Similarly, above 4°C there is a systematic decrease in density as the water warms.

In temperate middle latitudes the normal summer conditions in lakes are characterized by solar warming of surface water that produces a warm surface layer, the epilimnion, that rests above a deeper cooler water layer, the hypolimnion (Fig. 9.1). Such surface warming can be enhanced in the case of lakes with high suspended sediment concentrations because the albedo of the lake can be reduced, resulting in efficient radiant heat transfer into surface waters of the lake. A typical summer temperature profile in such a stratified lake consists of a warm epilimnion and a colder, deeper hypolimnion connected by a thermocline in which the temperature changes most rapidly with depth. In this condition the lake is described as stable and there is minimal mixing between the warm surface layer and the deeper, colder layer. Such a condition is common in ice-distal lakes in summer. The stable temperature profile can be disturbed by the action of wind on the surface of lakes which mechanically mixes surface and deeper water through the development of circulation. In middle latitude locations the stable structure developed in summer often breaks down in autumn as the surface water experiences cooling to the extent where surface layers may reach the same temperature or cooler than the underlying water (Fig. 9.1). In this case the thermal structure breaks down and the now dense surface water overturns and moves downward under the

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influence of gravity. If surface cooling continues ice may form on the surface and the densest water at 4°C sinks to the bottom of the lake and is stratified again with a new thermocline in deeper water. This new stable configuration breaks down in spring with the thaw of surface ice and warming of surface water to 4°C from solar radiation. This spring turnover sees the reestablishment of the stratified configuration.

In addition to thermally derived stratification, lakes can also become density stratified as a result of unequal distributions in suspended sediment concentration. Although more gradual than thermal stratification the magnitude of the differences can be greater because of high variations in suspended sediment concentrations particularly in ice-contact lakes. For example, 

Gustavson (1975a,b) recorded suspended sediment concentrations of up to 200 mg L\(^{-1}\) in surface water in Lake Malaspina, while Crookshanks and Gilbert (2008) measured concentrations in Lake Kluane between 0.2 and 0.9 g L\(^{-1}\).

The thermal stratification of lakes acts as a direct control of the distribution of water and sediment that flows into lakes and on the nature of depositional processes principally by suspension and the formation of density flows. The relationship between the thermal structure outlined above and the density of the influent water produces three principal flow paths for sediment in lakes: overflows, interflows, and underflows.

*Overflows* occur when the influent water has a lower density than the lake water. In these circumstances a surface plume of river water spreads over the surface of the lake and that is redistributed by currents and wind acting on the surface. Sediment carried in a surface plume is deposited by settling-out of particles to the lake bottom. Although common in marine settings where the density of seawater can often be higher than that of influent floodwater a relatively small proportion of lake sediment comes from overflows because the inflowing water usually has a higher density than lake water, particularly during floods (Leeder, 1999).
Interflows occur where the density of inflowing water may be greater than that of the epilimnion but less than the hypolimnion. In this case the plume of water flows along the thermocline and sediment carried in the flow is deposited by rain-out from suspension to the lake floor. In a study of Lake Brienz in Switzerland, Sturm and Matter (1978) recorded large quantities of particulate matter that was injected into the thermocline when the density of inflowing water was greater than the epilimnion but less dense than the hypolimnion.

Material trapped in the thermocline can be widely distributed by currents and produce distinctive couplets of sediment that are generically called rhythmites or if individual rhythms can be attributed to a single year of accumulation, a varve (Talbot and Allen, 1996). A typical varve consists of a light silt or sand layer that is deposited in spring or summer from meltwater or river runoff and a dark clay-rich layer that is deposited in winter when the lake is ice-covered. The couplet represents a temporal separation in deposition of different grainsize fractions in lakes although the underlying causes of the temporal separation and hence the environmental significance are multifactorial. Methodologies for facies analysis of very finely laminated lacustrine turbidites with submillimetric layers have been developed and applied to a range of depositional environments (Zolitschka et al., 1992; Brauer, 2004; Mangili et al., 2005). These studies have shown that analysis of very finely laminated lake deposits can be used to correlate layers between sections and cores and can be used to determine the season of their deposition.

In Lake Brienz, Sturm and Matter (1978) presented a case for rhythmite couplets being the result of the variation of stratification conditions where coarse material trapped in the thermocline is deposited a dark silty layer during transportation with the finer material trapped in the thermocline until the stratification overturns at the end of the summer when a uniform fine silt and clay is deposited in the lake as a suspension blanket.

However, in almost all glacial lakes a substantial proportion of the sediment is deposited by underflows that frequently form turbidity currents. Consequently, rhythmically laminated lake sediment often has two distinct components—one driven by suspension sedimentation, which may have an annual cycle and the other by density flows, which may be sustained on a seasonal basis (Brauer, 2004). But in many lakes, particularly in middle latitude mountainous locations, laminated sediments are just as likely to be driven by individual events such as extreme flood events or by episodic mass wasting of slope and delta sediments.

Underflows occur where the inflowing water is denser than the hypolimnion in which case the influent water and suspended sediment flows along the lake floor. Such underflows form sediment density flows that are very important, volumetrically, in lacustrine environments. Most of our knowledge of sediment density flows comes from marine environments where it has been argued that they are the most important flow process moving sediment across the planet and produce the thickest sediment accumulations on earth (Talling et al., 2012). In marine settings subaqueous flow deposits have been used to develop records of submarine landslide dynamics, earthquake histories, river flooding, and glacial outburst floods (Talling et al., 2012). Such reconstructions are also common in lacustrine systems, e.g., recent work by Hilbe and Anselmetti (2014) using a combination of high-resolution bathymetry, seismic, and lithological data from cores identified subaqueous mass movements that are likely candidates to have generated tsunamis in Lake Lucerne. Their interpretation was based on the recognition of thick, deformed mass flow deposits overlain by ‘megaturbidites’, characterized by a graded, sandy base and a thick homogeneous mud unit deposited by suspension. However, there are very few records of sublacustrine sediment flows.
because they are difficult to instrument and their occurrence is difficult to predict (Gilbert and Crookshanks, 2009). The lack of studies of the processes has meant that most of our knowledge of subaqueous sediment flows is derived from inferential studies of geological outcrops or from sediment cores. Dependence on inferential studies that have been done in a wide range of environmental contexts with different purposes has led to the development of classifications and terminologies that can be confusing and difficult to reconcile.

A recent review of subaqueous sediment flow processes in marine settings by Talling et al., (2012) has produced a coherent classification framework that is directly applicable to lacustrine sediment flows. Their classification scheme uses the generic term subaqueous sediment density flows and divides the density flows into two types: turbidites that are size-segregated and deposited incrementally as a series of layers and debrites which are deposited en masse by abrupt deposition. Debrites are further divided on the basis of matrix mud content and whether the deposits contain outsized clasts. The subdivision of turbidites varies with this practice and is based on features that record whether the near-bed flow was relatively dense or dilute. The classification scheme of Talling et al., (2012) outlined above represents a major step forward in the systematic analysis of subaqueous mass wasting deposits and in the application of experimental and field-based studies to the interpretation of ancient deposits. Although application of the framework to lacustrine processes has yet to be thoroughly evaluated, it offers considerable potential as a means of providing a unified classification scheme as well as the prospect of developing the consistent use of terminology in an area where the terminology used to describe and interpret sediments is highly variable.

9.3 ICE-CONTACT LAKES

Ice-contact lakes are those in which glacier ice is in contact with the lake water. Such lakes can be partly dammed by ice and form at ice margins, they can exist on the surface of glaciers or they can be in a subglacial position between the glacier base and the substrate. They are characterized by a supply of water that is close to 0°C that inhibits the formation of thermal stratification, and a high supply of suspended sediment. Consequently, the lakes appear turbid much of the time (Fig. 9.2). The high supply of suspended sediment to ice-contact lakes can create conditions in which stratification from variations in density may occur. Such a situation was observed in Canada where Smith and Ashley (1985) observed an increase in suspended sediment concentration with depth that they attributed to upward diffusion from turbidity currents and downward settling of sediment from interflows.

Ice-contact lakes are also characterized by a wide range of sediment sources and high spatial and temporal variations in the supply of sediment. Fig. 9.3 shows Tasman Lake, which has developed from a series of supraglacial ponds in the glacier surface in the 1980s into a large proglacial lake that continues to increase in size as the terminal cliff of Tasman Glacier recedes (Burrows, 2005; Dykes et al., 2011). Similar behaviour can be recognized at the margins of current glaciers like the Tasman Glacier that retreat from the positions that they held when first mapped by Europeans in the late 18th century. In the case of Tasman Glacier, since 1890 several initially small supraglacial ponds have coalesced to form larger supraglacial ponds (Fig. 9.3A), which have formed a large proglacial
lake impounded by the outwash head of the Tasman River (Fig. 9.3B). Several of the ponds have formed along the paths of partially collapsed englacial drainage channels. The development of the proglacial lake has led to the rapid retreat of the terminus of Tasman Glacier principally by calving driven by the rapid development of a thermal erosion notch at the cliff base. Thus the turbid water of the lake acting as a sensible heat sink accelerates the rate of ice loss.

In this ice-contact configuration rhythmically-laminated fine sands and silts accumulate in the lake basin with a background of ice-rafted debris from the frequent calving events that generate icebergs that often carry substantial volumes of supraglacial debris into the lake basin. Coarser sediments accumulate in the lake as a result of collapsing lateral and terminal moraines that destabilize as the glacier thins and from debris deposited onto the glacier surface by episodic rock avalanches from the adjacent mountain range (Fig. 9.3C). These mass movements produce energetic mass flow deposits that can reach the deeper parts of the lake basin. Coarse sediments also reach the lake through englacial and subglacial channels that deposit subaqueous fans into the lake adjacent to the ice cliff (Fig. 9.3B). In addition, subaerial deltas form where tributary streams are captured by the lake as it expands, as was the case in Lake Tasman when the Murchison River breached the latero-terminal moraine belt and began to discharge directly into the lake (Fig. 9.3B). Syndepositional deformation of the accumulating sediment pile is common as the sediments are often deposited against or on top of melting ice (Fig. 9.4). Icebergs stranded along the shorelines of ice-contact lakes are also important agents of sediment deformation as they directly disturb sediment when they ground, which can produce deformation ranging from complete disaggregation to the preservation of keel marks (Eyles and Clark, 1988; Duncan and Goff, 2001). The key source of fine-grained sediment is englacial and subglacial channels that discharge into the lake from the terminus of Tasman Glacier.
FIGURE 9.3

Ice-contact lakes at the terminus of Tasman Glacier. (A) Terminus of Tasman Glacier in 1986 showing several supraglacial ponds. (B) Terminus of Tasman Glacier in 2009 showing a large proglacial lake (Lake Tasman) which reaches depths of greater than 120 m adjacent to the ice edge. The Murchison River drains into the lake forming a large delta (lower left) and the Tasman River is now entrenched into the floodplain by about 8 m. (C) Oblique photograph of Tasman Glacier and proglacial Lake Tasman in 2000. Episodic rock avalanches off the Mt Cook Range (left) carry rock debris onto the glacier surface. Aoraki is the highest peak and has an altitude of 3724 m.
The key point sources of sediment are meltwater streams which can form in subaerial or subaqueous positions, while the key area sediment sources are deposition from suspension from interflows and overflows, and by ice-rafted debris. Meltwater discharged into lakes is at sites of rapid deposition as large volumes of bed and suspended load are rapidly deposited. The coarsest
sediment is deposited close to the discharge point with finer material being transported further into the lake. In a subaerial setting the most common sediment and landform assemblage is a delta where coarse-grained sediments are deposited rapidly as the inflowing water experiences flow expansion. Fig. 9.3C shows an instance where a large outwash stream has cut through a lateral moraine to flow into an ice-contact lake. The delta of the Murchison River forms a Gilbert-type delta that consists of three main morphological elements—steep foreset beds that are built by episodic debris flows of bedload sediment with the suspended load transported further out into the lake basin by overflows, interflows, and underflows described earlier. Settling out of these sediments and deposition of underflows that form turbidity currents form the bottom set beds of such deltas. Finally, the deltas are capped by flat lying topset beds which are the product of bedload deposition as the delta progrades into the lake.

In addition to underflows that are generated by flood events, turbidity currents can be generated by episodic failures of the foreset slopes. In the case of the Murchison River delta shown in Fig. 9.3C there are two distinct topset slopes—one graded to the current lake level and the other graded to a high stand of the lake about 6 m above the present lake level which is concordant with the lake shorelines preserved on the distal slopes of the end moraine system.

In subaqueous settings deltas can emerge from grounding-line fans, which form where subglacial and englacial streams discharge into lake water, which is accompanied by rapid deposition. The operation of effluent jets which discharge at ice margins have not been well described in lacustrine settings but have been described in marine environments. Powell and Domack (2002) defined three different types of grounding line systems that include grounding line fans and ice-contact deltas, morainal banks (Holdsworth, 1973), and grounding line wedges. Grounding line fans are fan-shaped accumulations of sediment that occur at the mounts of subglacial and englacial conduits. They normally consist of accumulations of outwash gravels and sands that rapidly accumulate adjacent to ice margins. Rapid accumulation of coarse material on slopes leads to a range of mass movements and in lacustrine settings turbid meltwater discharges are likely to produce underflows that transport the finer-grained components into deeper water beyond the ice margin. Morainal banks are features that form a wide variety of sedimentary processes that result in the release of sediment underneath and immediately adjacent to ice margins including melt-out, lodgement, and glaciotectonic processes. These are referred to as morainal bank deposits because in marine settings they have a bank-like geometry similar to subaerial moraines that are deposited parallel to ice margins. Grounding line wedges, also called till deltas (Powell and Domack, 2002), are wedges of sediment deposited in a subglacial position by sediment gravity flows that have redistributed subglacial sediments released at the grounding line.

Ice-rafted debris is common in ice-contact lakes in situations where calving glacier ice contains or is overlain by debris. Icebergs calving directly into ice-contact lakes can redistribute debris almost anywhere in lake basins (Fig. 9.4A). The resulting sediments are characterized by anomalously large particles resulting in much finer sediments deposited in a suspension regime (Fig. 9.4). In cases where debris-covered icebergs can become top-heavy and rotate they release large clusters of ice-rafted debris onto lake floors.

A further complexity of ice-contact lakes and their sedimentary record is that elements of the sediment that accumulate in the lakes are deposited on or adjacent to glacier ice. As the glacier ice melts syn- and postdepositional deformation can occur resulting in a wide range of ductile and brittle deformation structures (Fig. 9.4).
Ice-contact lakes in the Arctic and Antarctica are generally characterized by lower sediment fluxes and have a lower preservation potential in the stratigraphic record. In Svalbard ice-contact lakes form an important element of the advance—retreat sequence of tidewater glacier such as Kronebreen/Kongsvegen complex (Bennett et al., 1999). The lakes, which were dammed by the advancing tidewater glaciers and by locally developed thrust moraines have poor preservation potential and the most prominent evidence of the formation is shorelines and meltwater channels.

A further extreme case is the proglacial ice-contact lakes that occur in the McMurdo dry valleys (Fig. 9.5). Over 20 lakes occur, 10 of which are currently in contact with alpine glaciers or outlet glaciers from the East Antarctic Ice Sheet. All except one hypersaline lake have perennial ice covers of 2.8–6 m thick (Doran et al., 1994). Most of the lakes are the remnants of larger glacial lakes that occupied the McMurdo Dry Valleys in the past. For example, the lakes in the Taylor Valley have evolved as remnants of Glacial Lake Washburn. Lake Washburn was an ice-contact lake that last occupied the Taylor Valley during the Last Glacial Maximum when an expanded Ross Ice Shelf sealed the eastern end of the valley leaving relict deltas and shorelines along the flanks of the valley and lacustrine and glaciolacustrine sediments in the current lake basins (Wagner et al., 2006).

After the Ross Ice Shelf retreated 8–9 ka BP the lake diminished in size mainly from evaporation, the last major evaporation event being 1–2.5 ka BP. Post 4 ka BP the sediment deposited in the lakes consists of laminated microbial mats interspersed with thick coarse sand horizons (Doran et al., 1994).

Sedimentation in these lakes is currently very low. Adjacent to the ice margins and small streams that flow into the lakes, land and fine gravels accumulate and deltas are dominated by sand-sized sediment with algae layers that have grown on topset sediments (Fig. 9.5) but in the deeper water, even immediately adjacent to the ice margins sedimentation is predominantly biological with the accumulation of algae (Doran et al., 1994). Several of the ice margins that have flowed into perennially ice-covered lakes have accreted ice-marginal lacustrine sediment that can subsequently form part of the basal ice sequence of the cold-based alpine glaciers (Fig. 9.5; Fitzsimons et al., 2008).

9.4 SEDIMENTARY FACIES IN ICE-CONTACT LAKES

Ice-contact lakes are complex depositional basins that experience rapid temporal and spatial variations in the location, rate, and nature of depositional processes. They are particularly difficult environments in which to observe or monitor contemporary depositional processes because they are often inaccessible due to turbid meltwater, icebergs, and collapsing ice margins. Studies of Pleistocene ice-contact lakes offer the possibility of a thorough understanding of the depositional processes and the spatial and temporal fluxing of sediments (e.g., Winsemann et al., 2007). However, there have been relatively few studies that have attempted to document and understand the three-dimensional architecture of such environments. An important study in this regard is the evaluation of a lake that repeatedly occupied the Copper River in Alaska. Bennett et al. (2002) defined seven facies associations, the majority of which are deposited by subaqueous debris
Joyce Glacier in the Garwood Valley, McMurdo Dry Valleys, Antarctica, flowing into perennially ice-covered Lake Buddha. (A) Shows the frozen surface of the lake. (B) Deltaic sediment with algae layers. (C) Ice-marginal ramp with sediment.
flows and hyperconcentrated sediment gravity flows deposited on subaqueous fans supplied by subaqueous and subaerial outwash streams. The facies architecture of the basin is dominated by subaqueous fans that have aggraded by basin-ward transportation of gravity-driven processes, most of which are attributed to destabilization and remobilization of rapidly deposited proximal sediment. Bennett et al. (2002) argued that the fans are inherently unstable because the spatial focus of deposition changes with the changing geometry of the ice margin which can evolve rapidly during deglaciation. The resulting facies architecture is a series of on-lapped features as the ice margin retreats (Fig. 9.6) in which the apexes of the fans are dominated by subglacial outwash, subglacial deformation, and deposition of ice-rafted debris. Upon retreat the subaqueous fans are incised and the focus of deposition is within the entrenched channels (Fig. 9.6B). Two scenarios are presented in the depositional model: (1) ice-contact deposition with fan apexes located at the mouths of subglacial conduits discharging directly into the lake and (2) deposition downstream from subaqueous or subaerial channels (Fig. 9.6). In both scenarios the bulk of the deposits is relatively coarse-grained debris and sediment gravity flows driven by energetic deposition on subaqueous fans. Through a glacial cycle the packages of gravel and sand in subaqueous fans are separated laterally and vertically by diamict and rhythmite interfan facies which are spatially and temporally separated from the next cycle by an erosion surface cut by rivers in the basin centre and at the basin edges by ice.

Despite this and many similar studies that capture the complexity of ice-contact depositional environments, the role and importance of glaciolacustrine processes in ice-marginal sedimentation remains poorly understood. Consequently, lacustrine environments are poorly represented in high-order landform/sediment assemblage models of glacial environments (Evans, 2003). In an attempt to improve understanding of the development of ice-contact landforms and sediments Bennett et al. (2000) undertook a systematic analysis of the late Holocene development of Hagavatn, an ice-contact lake adjacent to Hagafellsjökull Eystri in Iceland. The lake, which is now only 5 km² in area but was up to 12 km², formed as a result of a volcanic eruption blocking proglacial drainage.

Bennett et al. identified four major landform components that have developed as a consequence of lake sedimentation and the interaction between the glacier and lake sediments:

- Shorelines and overflow channels;
- Moraines;
- Morainal bank and ice-contact delta;
- Lake beds and sediments.

Shorelines and overflow channels were mostly depositional features formed in older sediments and overflow channels cut in cols where meltwater had been forced over drainage divides by ice damming. Bennett et al. described three groups of moraines: (1) constructional moraines from ice advances; (2) asymmetric moraines that are structural (glaciotectonic) features formed from imbricate slabs of a wide variety of glaciolacustrine sediments interpreted as being deformed as an ice margin advanced into a proglacial lake and against an adverse slope in a similar manner described by Fitzsimons (1992); and (3) large symmetrical moraine that contained thrust and highly sheared laminated sediment that were interpreted as overridden push moraines formed by deformation of ice-proximal and ice-distal glaciolacustrine sediments and subsequent infill by proglacial fluvial sediments.
FIGURE 9.6
Depositional model for a glaciolacustrine basin in Copper River, Alaska. See text for explanation.

Morainal banks were also interpreted as glaciotectonic features produced by bed-parallel thrust slices of glaciolacustrine sediment that ranged from completely homogenized to low-amplitude recumbent folds. They argued that the large folds were produced by deformation of subaqueous ice-marginal sediments during a substantial ice advance into the lake. The deformation that occurred was followed by the formation of an ice-contact subaqueous delta that evolved into a sub-aerial fan delta as it aggraded and prograded.

The final landform components were lake beds and sediments which formed on the floor of the lake and on the ice-proximal slopes of the moraine slopes. They consisted of fine-grained laminated sands, silts, and clays that contain numerous soft sediment deformation structures including convolute lamination and flame structure as a consequence of rapid accumulation. The principal sedimentary architecture was interpreted as partial Bouma sequences from underflows generated by meltwater influx into the lake. A Bouma sequence is a set of sedimentary structures that form as a result of the passage and deposition of turbidity currents in deep marine settings and the conceptual framework is widely applied to lacustrine environments (Shanmugam, 1997).

In this careful analysis of the spatial and temporal sequencing of the deposition and deformation of a proglacial environment dominated by the presence of glaciolacustrine sediment Bennett et al. argued that the grounding line landforms (morainal banks and ice-contact fans) acted as second-order controls of ice margin behaviour, particularly the location and intensity of calving processes as has also been noted in glaciomarine settings (Powell, 1990). This study underlines the importance of glaciolacustrine processes in the development of landform sequences/assemblages at the fluctuating glacier margins. The resulting landform assemblage has the characteristics of many deglacial and contemporary margins insofar as it is dominated by a variety of moraines.

What has not been as widely recognized is that the moraines are structural landforms formed by deformation of proglacial sediment which in this case have accumulated in an ice-marginal lacustrine basin. The deformation style, and ultimately the structure and morphology of the landscape, has been controlled by the presence of ductile laminated silt and clay overlying a bedrock basin (Bennett et al., 2000). The geotechnical characteristics of the glaciolacustrine sediment, and its spatial location with respect to the ice margin, control the development of a décollement surface, which is the location of the sole thrusts from which the glaciotectonic moraine ridges have developed. The laminated nature of the sediments facilitates thrusting (Butler, 1982) as does the presence of an adverse bedrock slope that creates locally high rates of shortening and compression of the proglacial sediment pile (Fitzsimons, 1992). Finally, the studies of Bennett et al. (2000, 2002) and Winsemann et al. (2007) form important reminders that simple lithostratigraphic and landform/sediment association models do not capture the complexity of ice-contact environments or the importance of glaciolacustrine processes and sediments in landscape development around glacier margins.

9.5 GLACIOTECTONIC DEFORMATION

The study of glaciolacustrine processes at Hagavatn described earlier emphasizes the importance of glaciotectonic deformation of sediment due to stresses imposed by glacier ice. There are two broad regimes that are widely recognized: subglacial deformation of material beneath
moving ice, and proglacial deformation of sediments deposited in front of and adjacent to ice masses (van der Wateren, 2002). Deformation of sediment adjacent to glacier ice occurs because glaciers can exert high stresses and because sediment bodies can have yield strengths less than the stresses exerted by glacier ice. Although glaciotectonic processes are described in detail in Chapter 13, Glacipectonics, lacustrine sediment is particularly susceptible to glaciectonic deformation and many ice-contact lake sediments bear the imprint of deformation by moving ice or from liquefaction and injection in subglacial environments (Phillips et al., 2013).

In a subglacial setting the ‘gravity spreading model’ of glaciotectonic deformation (Williams et al., 2001; van der Wateren, 2002) is widely used to explain how the static load of the glacier pushes sediment away from the load. In this model failure is likely where impermeable strata are compacted and water escape routes are restricted or pore water pressure is increased because of the presence of a confined aquifer such as fine-grained lake sediment. In proglacial settings, any unconsolidated sediment can be pushed from behind so sediment that accumulates in proglacial lacustrine basins is susceptible to deformation, particularly when the sediment is saturated. In a review of the causes of proglacial deformation, Benn and Evans (2010) identified four environmental characteristics that are conducive to deformation that correlate with the development of proglacial lakes and the accumulation of laminated sediments:

- The slope of the proglacial area, particularly if there are adverse slopes such as those that occur in lake basins;
- The presence of weaknesses (such as fine-grained lacustrine sediment that forms potential decollement layers);
- The nature of the ice—sediment contact, e.g., if glacier margins are partly buried by marginal sediment the glacier can drive a wedge of ice into strata;
- Subglacial and proglacial drainage.

The outcome of glaciotectonic deformation of glacial sediments ranges from lacustrine sediment that is completely homogenized and sheared to gentle small-scale folding and faulting (Fig. 9.6). Fig. 9.7 shows a glaciolacustrine delta adjacent to the margin of Joyce Glacier in Antarctica that has been deformed by glaciotectonic deformation as the glacier pushed the deltaic sediments from behind (left in Fig. 9.7A and 9.7B and right in Fig. 9.7C). The ridge closest to the glacier front is a 10m high sharp-crested ridge that consists of laminated deltaic sands and fine gravels with occasional layers of algae-rich sediment. The smaller outer ridge is a single high-angle thrust that has broken through to the surface of the ridge (Fig. 9.7B). An exposure of the ridge to the left side of the glacier margin (Fig. 9.7C) shows that the ridge is formed by a low-angle listric thrust fault that forms a décollement within the deltaic sediment. In this case the décollement has propagated along a 12 mm-thick layer of ice resting on an algae layer preserved within the deltaic sediment. Numerous high-angle normal faults and tension cracks have propagated from the décollement upwards into the laminated and cross-bedded sediment. Deposits immediately under the thrust have been brecciated and in front of the thrust they have been deformed into isoclinal folds. The area has a mean annual temperature of $-22^\circ$C and the active layer is only 100 mm thick so the deformation has occurred when the sediments were frozen. This assemblage shows that a wide range of glaciotectonic deformation structures, both brittle and ductile, can occur simultaneously even when the deforming sediment is permanently frozen.
Figure 9.7
Glaciotectonic deformation of deltaic sediments adjacent to Joyce Glacier in the Garwood Valley, Antarctica. (A) System of thrust-block moraines with an active thrust propagated to the surface of the outer ridge where it can be seen as a tension crack. (B) Thrust exposed in a stream channel cut through the outer moraine. (C) Listric thrust in planar and cross-bedded deltaic sediments. A small ridge has formed where the thrust breaks the surface and an isoclinal drag fold has developed under the thrust.
9.6 SUBGLACIAL LAKES

Subglacial lakes are a particular form of ice-contact lake that exist at the interface between glacier ice and its substrate. In the late 1960s the use of radio echo-sounding led to the discovery of flat-bottom reflectors in parts of the East Antarctic Ice Sheet that were interpreted as the upper surfaces of subglacial lakes (Robin et al., 1977).

The combined application of geophysical surveys and remote sensing have shown that subglacial lakes form elements of a dynamic subglacial hydrological network beneath the Antarctic Ice Sheet (Ashmore and Bingham, 2014; Fig. 9.8). The lakes occur where liquid water beneath ice sheets, generated by geothermal heating, leads to melting and the accumulation of the meltwater in basins within depressions in the substrate where water inflow is equal to or greater than water outflow. A series of inventories of subglacial lakes based on radio-echo sounding and remotely sensed imagery record the locations of lakes in Antarctica, the most recent of which records 379 lakes (Siegert et al., 2005; Fricker et al., 2007; Wright and Siegert, 2012). The discovery of subglacial lakes led to hypotheses that these water masses and potentially biological communities within them are refugia that have been out of contact with the atmosphere for hundreds of thousands of years (Priscu et al., 1999; Siegert et al., 2001) and that the accumulated sediment within the lakes may contain proxy data that can be used to interpret the history of the ice sheet (Bentley et al., 2011).

The hypothesis that the subglacial lakes form part of a subglacial drainage network was developed by Dowdeswell and Siegert (2003) and later satellite data were used to infer episodic filling and drainage of lakes beneath the West Antarctic Ice Sheet. A later study by Fricker et al. (2007) used satellite laser altimetry to record the connection between lakes in the Mercer and Whillans Ice Streams. Initial sampling of accreted ice that rests above Subglacial Lake Vostok showed the presence of a microbial community but the samples were contaminated with drilling fluid (Priscu et al., 1999). After a drilling programme in 2013 the WISSARD team sampled water from Lake Whillans some 800 m below the surface of the Whillans Ice Stream (Christner et al., 2014). These samples collected using clean drilling and sampling techniques contained over 4000 species of bacteria and archaea in a chemosynthetically driven ecosystem. Sedimentary processes in subglacial lakes and the sedimentology of deposits found in such lakes have yet to be systematically studied although recent and future efforts aimed at understanding these unusual ecosystems are likely to provide new insights into these processes.

9.7 ICE-DISTAL LAKES

Ice-distal lakes are glacial lakes that have become separated from glaciers during deglaciation or lakes that have formed in areas downstream and remote from glaciers. Ice-distal lakes are group of lakes that are removed from the immediate influence of glacial processes but are still influenced by the presence of glacier in the catchment. Typically, distal lakes are physically separated from glaciers by the presence of an outwash plain. Physical separation from ice margins mean that these environments do not experience the high variation in meltwater and sediment supply or the large range of sedimentary processes that characterize ice-contact lakes described above. The presence of an outwash plain and distance of travel from ice margins to lakes acts as a spatial and temporal filter, which results in diminished rates of change in sediment and meltwater supply. Consequently, the range of depositional processes is smaller, the rate of deposition generally less, and the depositional facies and facies assemblages simpler than those of ice-contact lakes.

As noted earlier ice-distal lakes often form from initially ice-contact lakes and become progressively more isolated from the direct influence of glaciers during deglaciation, indeed many modern lakes in glaciated basins are now devoid of glaciers in their catchments. Because of their isolation from glaciers, distal lakes are much more likely to be thermally stratified and experience cyclic changes in thermal stratification as described above (Ashley, 2002). Fig. 9.9 shows three large distal lakes in the central Southern Alps of New Zealand that are impounded behind the outwash sediments deposited during the Last Glacial Maximum. All the lakes are fed by large braided outwash plains that drain the highest topography of the Southern Alps. The catchment of Lake Ohau is almost entirely deglaciated, whereas the catchments of Lakes Pukaki and Tekapo are the most intensely glaciated catchments in the Southern Alps and they have large remnant valley glaciers that have proglacial ice-contact lakes that act as efficient sediment traps for sediment currently released by the glaciers. These fjord lakes (Eyles et al., 1991) are dominated by terrigenous sedimentation because the rapidly rising mountains supply relatively high volumes of sediment through the glacial outwash plains. However, the dominance of terrigenous sedimentation is by no means universal and many lakes in drier areas that were deglaciated at the end of the
Pleistocene are characterized by slow authigenic sedimentation dominated by the accumulation of chalks and marls (Charlet et al., 2008). The New Zealand fjord lakes are characterized by conspicuous moraines that occur around the outlet shorelines of these lakes but it is the hundreds of metres of outwash gravel that impound the lakes. The lateral and terminal moraines that enclose the southern ends of the lakes are characterized by glaciotectonically deformed rhythmically laminated silts and fine sands together with mass flow deposits, ice-rafted debris, and outwash sediments. These deposits have accumulated in proximal ice-contact lakes that formed as the ice margins oscillated during the Last Glacial Maximum and began to retreat.

New Zealand fjord lakes typically have a single simple basin with maximum depths between 120 and 400 m and steep lateral slopes. The mean inflow into Lake Tekapo is 120 m$^3$ s$^{-1}$ in spring as the seasonal snowpack melts and is lowest in winter when the inflow is around 40 m$^3$ s$^{-1}$ (Pickrill and Irwin, 1983). The lakes are dominated by single large deltas that prograde rapidly because of high sediment loads derived from rapidly uplifting mountain (Fitzsimons and Veit, 2001). Most of the bedload sediment is deposited on the delta topsets but frequent overloading results in collapses of the prodelta slopes that generate turbidity currents that travel several kilometres into the abyssal basin.

**FIGURE 9.9**
Fjord lakes of the central South Island of New Zealand Lakes Tekapo, Pukaki, and Ohau have formed in basins behind the outwash plains deposited during the last glaciation. Ice-contact lakes adjacent to Mueller, Hooker, and Tasman glaciers are visible in the Tasman Valley and in front of Godley and Maud Glaciers in the Tekapo Valley.
Ice-distal lakes have two main depositional systems: a deltaic system at the major influx points and a lake bottom system in which fine-grained, often laminated sediments dominate. Deltaic systems in ice-distal lakes are slightly different from their ice-contact counterparts because they are generally finer grained and prograde and aggrade at slower rates. In alpine deglaciated environments such as is shown in Fig. 9.9 they are predominantly Gilbert-type deltas. In a study of Lake Tekapo (Fig. 9.9), Pickrill and Irwin (1983) concluded that the delta of the main inflowing river accounted for 55% of the annual sediment accumulation in the lake and that the delta was dominated by the accumulation of sandy muds at the top of the foreset slope in winter and that the muds avalanched down the delta face during surficial slides. Deep-seated failures are also a feature of the delta which is subject to episodic rotational slumping. The scars of such failures are observable on multibeam imagery and the accumulation of debris from the mass movements is often visible on sub-bottom geophysical surveys. Fig. 9.10 shows a 3.5 MHz compressed high-frequency radar pulse (CHIRP) image of Lake Ohau, the westernmost fjord lake shown on Fig. 9.9. Here, a landslide has developed in sediments that have accumulated on the western slope of the lake and have been transported into the deeper part of the basin where the deposit overlies rhythmically laminated lake floor sediment.

Pickrill and Irwin (1983) estimated the volume of a single landslide the delta at $8.25 \times 10^6$ m$^3$ which is the equivalent of between 200 and 600 years of accumulation removed from the delta in a single event. Such landslide events produce turbidity currents capable of transporting large volumes of sediment into the lake basin and are likely the reason for an increase in sedimentation rate in the distal part of the basin. The main base of the lake is dominated by rhythmically laminated dark and light layers with 28–33 rhythms every 100 mm, although only 12 of these constitute major ‘varves’ that were used to determine an annual sedimentation rate of 0.3–0.36 cm a$^{-1}$.

In nearby Lake Ohau the lake sediments that have accumulated in the distal end of the lake at about 60 m depth have been classified as three different types of varves (Roop et al., 2015): type A consists of a fine silt basal layer that grades into a very fine silt layer, type B contains one or more grain-supported 0.5–1.5 mm-thick sublaminae within the primary coarse/fine stratigraphy. These ‘complex’ varves account for 56%, and type C consists of layers that are c. 9.0 mm thick and with a basal fine silt layer that grades into a relatively thick layers. Types A, B, and C constitute 34%, 56%, and 10% of the upper part of the cores, respectively. In the depocenter of the lake (c. 130 m deep) the same rhythmic laminated framework can be observed (Fig. 9.11A). However, the deeper parts of the basin are also characterized by episodic turbidites that range from 10 to 0.75 cm in thickness and grade from coarse sand to fine sandy silt (Fig. 9.11B). These turbidites have been deposited by episodic flood events and slope collapse events that have generated underflows that have traversed almost the entire length of the lake basin. Although they can be correlated into the stratigraphy of the distal end of the basin, the coarse elements of the turbidites are not represented beyond the deeper part of the basin. This is because they do not climb the adverse slope of the distal basin. Initial dating of the largest turbidite suggests that it may have been derived from the AD 1717 rupture of the Alpine Fault, which is recorded by similar turbidites on lakes on the West Coast of the South Island (Howarth et al., 2012). These event-based deposits provide a significant complexity for chronologies and environmental inferences that depend on the regularity of the quasiannual signal recorded in the rhythmically laminated sediments.
FIGURE 9.10
3.5 MHz CHIRP images of the sediment fill in Lake Ohau. (A) Sediment fill in the southeastern arm of Lake Ohau showing the postglacial laminated facies resting above an acoustically opaque postglacial facies deposited on the rock basement. (B) Central western slope of Lake Ohau showing a series of mass wasting deposits. The largest mass wasting deposit (MWD7) is about 3 km long and 9 m thick.
9.8 DISTAL LAKES AS ENVIRONMENTAL REPOSITORIES

As noted earlier many distal lakes have gone through an evolutionary sequence initially forming as ice-contact lakes and during glacier recession lakes can be increasingly isolated from glaciers and become ice-distal lakes. An example of this evolutionary sequence is captured by Fig. 9.9, which shows three fjord lakes that initially formed as ice-contact lakes during the LGM and are now separated from the

FIGURE 9.11
Cores from Lake Ohau showing the physical sedimentology of distal lakes. (A) Finely laminated silts. (B) Turbidite in the upper 0.52 m of the core resting on laminated silts.
modern glaciers of the Southern Alps by up to 60 km of outwash plain. Renewed glacier retreat since the Little Ice Age has led to the development of a new series of ice-contact lakes at the modern glacier margins which have grown progressively longer and deeper. If glacier retreat is sustained, they will become isolated from the ice margins that drain into them. Lakes that have undergone this environmental transition can provide excellent records of environmental conditions during the late glacial to Holocene transition, hydrological, vegetation and landscape changes during the Holocene, and the impact of seismic shaking on landscapes (Van Rensbergen et al., 1998; Moernaut et al., 2007; Fanetti et al., 2008; Charlet et al., 2008; Strasser et al., 2013; Hilbe and Anselmetti, 2014).

Many of the recent advances in using glacial lakes as environmental repositories have focused on lakes in Europe and in South America. A good example of the approach is the work of Van Rensbergen et al. (1998) on the sedimentary record preserved in Lake Annecy since the last glaciation. In this study they used high-resolution seismic stratigraphy and coring to examine the nature and variability of sediment supply since deglaciation. They concluded that the sediments recorded a systematic change in the sediment supply through deglaciation and identified three distinct styles of accumulation that represent different phases of lake evolution. The first phase occurs as the glacier thins and the subglacial lakes form during deglaciation and is dominated by a horizontal succession of grounding-line fans and deltas, which record the retreating of the glacier grounding line and lake enlargement. The second phase is the development of an open proglacial lake characterized by the rapid accumulation of laminated sediments from suspension and underflows accompanied by deposition of abundant ice-rafted debris. As the ice retreated further there is a switch from an underflow-dominated regime to an interflow-dominated regime that deposits silt-clay couplets that thin away from the major sediment sources as the sediment supply reduces. Finally, the sedimentary regime becomes predominantly authigenic as the vegetation reestablishes and the terrigenous sediment supply diminishes and the sediments are mainly chalks and marls as the lake evolves into its current configuration by about 5000 BP.

A similar pattern was observed in South America by Charlet et al. (2008), who divided the sedimentary fill of Lago Puyehue in Chile into five sedimentary units, based on seismic stratigraphy and a coring campaign. The oldest depositional unit (Unit 1) is characterized by transparent to chaotic high-amplitude reflectors overlain by irregularly stratified reflectors, which were interpreted as moraine deposits of the last glaciation. These deposits are overlain by Unit 2 which is characterized by a chaotic reflection-free to irregularly stratified seismic facies with low-amplitude reflectors. This unit was interpreted as the product of rapidly deposited glaciolacustrine deposits during glacier retreat either in a proglacial or subglacial lake. The glaciolacustrine deposits are overlain by fluvioglacial deposits with outwash channels and then more glaciolacustrine deposits (U4). The final stage of lake development is recorded by a drape of parallel continuous high-amplitude reflectors that record open lacustrine deposits, i.e., a distal glacial lake.

**9.9 CONCLUSION**

Glaciolacustrine environments are amongst the most complex on earth because they are characterized by a wide range of depositional processes and abrupt changes in processes and resultant sedimentary facies in horizontal and vertical dimensions. This complexity represents a significant challenge for future work in the development of reliable facies models for glacial lakes, particularly in ice-contact settings.
Other important challenges for future work include to strengthen the links between sedimentary processes and depositional products, both sediments and landforms. A great deal of our knowledge of glaciolacustrine processes remains largely inferential, that is, dependent on inferences about processes made from analysis of sediments and landforms from recent or ancient glacial lakes. In contrast there are a modest number of studies that have been able to establish direct links between sedimentological and hydrodynamic processes and resultant deposits. This is particularly evident in the context of ice-contact lakes and in the formation of turbidites and hyperpycnities in ice-distal lakes. Greater efforts are needed to monitor sedimentary and hydrodynamic processes and lakes and to characterize resultant deposits if we are to build more robust interpretations of depositional processes from sedimentary products.

REFERENCES


REFERENCES


