Sedimentology and depositional model for glaciolacustrine deposits in an ice-dammed tributary valley, western Tasmania, Australia

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ABSTRACT

Quaternary sedimentary successions are described from the Linda Valley, a small valley in western Tasmania that was dammed by ice during Early and Middle Pleistocene glaciations. Mapping and logging of exposures suggest that an orderly sequence of deposits formed during ice incursion, occupation and withdrawal from tributary valleys. Four principal sediment assemblages record different stages of ice occupation in the valley. As the glacier advanced, a proglacial, lacustrine sediment assemblage dominated by laminated silts and muds deposited from suspension accumulated in the valley. A subglacial sediment assemblage consisting of deformed lacustrine deposits and lodgement till records the overriding of lake-bottom sediments as the glacier advanced up the valley into the proglacial lake. As the glacier withdrew from the valley, a supraglacial sediment assemblage of diamict, gravel, sand and silt facies formed on melting ice in the upper part of the valley. A lacustrine regression in the supraglacial assemblage is inferred on the basis of a change from deposits mainly resulting from suspension in a subaqueous setting to relatively thin and laterally discontinuous laminated sediments, occurrence of clastic dykes, and increasing complexity of the geometry of deposits that indicate deposition in a subaerial setting. A deltaic sediment assemblage deposited during the final stage of ice withdrawal from the valley consists of steeply dipping diamict and normally graded gravel facies formed on delta foresets by subaqueous sediment gravity flows. The sediment source for the delta, which prograded toward the retreating ice margin, was the supraglacial sediment assemblage previously deposited in the upper part of the valley. A depositional model developed from the study of the Linda Valley may be applicable to other alpine glaciated areas where glaciers flowed through or terminated in medium- to high-relief topography.

INTRODUCTION

Glacial lakes can be divided into ice-contact lakes which develop beneath, on, or adjacent to active or stagnant ice, and distal lakes which are some distance from the ice and are fed primarily by outwash streams (Eyles & Miall, 1984; Ashley, 1988). Studies of ice-contact lakes and lake sediments have emphasized the episodic nature of deposition and the wide variety of sedimentary facies that can form in this environment (Ashley, 1975; Shaw, 1977; Shaw & Archer, 1978; Shaw et al., 1978; Cohen, 1983; Eyles et al., 1987; Ashley, 1988; Donnelly & Harris, 1989; Hicks et al., 1990; Persson & Lagerlund, 1990). Several recent studies of Quaternary ice-contact lacustrine deposits have stressed the importance of diamict and gravel facies deposited by high-energy subaqueous sediment gravity flows (Cohen, 1983; Eyles, 1987; Eyles et al., 1987). These studies describe depositional environments in lakes with relatively stable ice margins or during deglaciation when ice margins are either retreating or melting in situ. This paper reconstructs the sedimentary environments of a small ice-contact lake basin in western Tasmania that contains a relatively complete record of the deposits that formed when the valley was dammed and occupied by a mobile ice margin. In this depositional setting a distinctive assemblage of subaqueous, subglacial, supraglacial and deltaic deposits accumulated at different stages of ice occupation of the valley.
Although the sediment assemblages have been described previously, their distribution and association in the Linda Valley are evidence of a particular glacial subenvironment that occurs when valley glaciers block and flow up tributary valleys. A depositional model for sedimentation in ice-dammed tributary valleys is presented. This is based on the spatial and inferred temporal relationships between the sediment assemblages and distinguishes between conditions in the valley during damming, ice advance, retreat, and in situ melting.

Sediments in the Linda Valley (Fig. 1) were mapped, described and sampled. The major sections in the Linda Valley were measured and logged in the field using the scheme described by Miall (1978) and Eyles et al. (1983). A wide range of data on sedimentary structures and characteristics of the sediments was recorded in order to characterize the lithofacies. The

Fig. 1. Location map of the Linda Valley. The shaded area is land over 400 m and the dotted line is the southern limit of Jurassic dolerite in Quaternary sediments.
data included measurements of particle size, clast shape, lithology, sedimentary structures and pebble fabric. Pebble fabric in diamicts was measured from the orientation and dip of prolate-shaped clasts greater than 2 cm in length and with axial ratios \( b/a = 2/3 \) and \( c/b = 2/2 \). Measurements were plotted on lower hemisphere Schmidt Nets and contoured according to the method of Kamb (1959). The data were analysed using the eigenvector method of Mark (1973) where eigenvector \( V_1 \) gives the direction of maximum clustering and the eigenvalue \( S_1 \) gives the strength of clustering about \( V_1 \). Particle size analyses on samples from the major facies types were performed using sieves and a hydrometer.

**REGIONAL SETTING AND AGE OF THE DEPOSITS**

The Linda Valley is a small tributary valley of the King Valley in western Tasmania (Figs 1 & 2). The valley is surrounded by steep, incised slopes of Cambrian volcanic rocks and Ordovician conglomerates. The town of Gormanston is built on a large, dissected Early Pleistocene moraine at the head of the valley (Figs 1 & 2). Downstream, the valley floor is flat and then closes into a small gorge where waterfalls are formed before Linda Creek flows onto the eroded floor of the King Valley (Fig. 3).

During the Quaternary, the King Valley was glaciated on several occasions (Fitzsimons et al., 1990). An outlet glacier flowed south down the King Valley from an ice cap formed on the Tyndall Plateau and Eldon Range (Fig. 1). At its maximum extent, the King Glacier split into four distributary lobes that flowed south down the King Valley, south and east down the Nelson Valley, and west up the Comstock and Linda valleys (Fig. 1). The maximum extent of the ice in the King Valley is marked by a limit in the dispersal of Jurassic dolerite erratics derived from the Eldon Range (Fig. 1). In the upper part of the Linda Valley, Karlsons Gap (Fig. 2) forms a low saddle close to the maximum altitude of glacial deposits (Figs 2 & 3). Although the similar altitude of the gap and the deposits suggests ice may have passed through the drainage divide, this is offset by the absence of lithologies derived from the King Glacier system west of the divide.

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*Fig. 2. Map of the Linda Valley showing the location of the sections discussed in the text. The stages indicate the position of the ice margin when the main sediment assemblages were deposited. The proglacial lacustrine sediment assemblage was deposited at Stage 1, the subglacial sediment assemblage at Stage 2, the supraglacial sediment assemblage at Stage 3 and the deltaic sediment assemblage at Stage 4 and Stage 5 as the ice retreated from the valley.*
Glaciogenic deposits described in this paper formed during Middle and Early Pleistocene glaciations (Fitzsimons et al., 1990; Fitzsimons & Colhoun, 1991). Middle Pleistocene deposits in the lower part of the valley are moderately chemically weathered, have a normal detrital magnetization (Pollington, 1991) and lie 7 km outside the limits of the Last Glaciation which peaked at about 18 000 yr BP. Early Pleistocene deposits occur in the middle and upper parts of the valley where they form the large Gormanston Moraine (Figs 2 & 3). These deposits are intensely chemically weathered, have a reversed detrital magnetization and rest unconformably on Late Tertiary fluviatile deposits (Fitzsimons, 1988).

LITHOFACIES

The lithofacies classification used in this study is based on the scheme devised by Miall (1978) and Eyles et al. (1983). Facies descriptions and their relative abundances in the valley are summarized in Table 1. The single dominant facies type is laminated silt and mud but combined values for fine-grained and diamict facies show these to be equally abundant (Table 1).

Rhythmically laminated silts: Fl, Fld

Description
Rhythmically laminated silts and clays are the commonest facies in the Linda Valley and are associated with all other facies (Table 1, Figs 4–6). The facies ranges from horizontally bedded, finely laminated silt and mud to highly deformed facies in which bedding is difficult to recognize. The dominant facies is finely laminated silt and mud, often with abundant dropstones a few millimetres to over 1 m in size (Figs 4 & 8). Particle size analysis of individual laminae shows that the thicker (5–20 mm) laminae are silts and the thinner laminae (1–2 mm) are muds (Fig. 7).

Laminated silt and mud facies also occur as steeply dipping silt and mud drapes that overlie gravel and diamict facies (Figs 5 & 10c). The silt and mud layers form thin continuous bands within the gravels and diamicts and form distinct horizons that separate episodes of high-energy deposition. Many of these laminated bands show evidence of scouring, loading and faulting that indicates disturbance during deposition of the overlying gravel and diamict facies.

Deformed silt and mud facies range from mud with small-scale deformation such as convolute lamination, faulting and other loading structures (Fig. 9a), to highly deformed silt that has been sheared and shows evidence of brecciation and folding (Fig. 9b).

Interpretation
Laminated silts and muds were deposited from suspension in the deeper parts of an ice-contact lake and represent the lowest energy conditions in the lake basin. Gilbert & Shaw (1981) attributed fine silt laminae to diurnal fluctuations in sediment supply and Smith et al. (1982) described seasonal and subseasonal laminations. In the Linda Valley, alteration of silt and mud laminae was probably due to fluctuations in the volume of sediment input over
Table 1. Description of facies types, coding and relative importance in the Linda Valley expressed as a percentage of total thickness logged.

<table>
<thead>
<tr>
<th>Facies type</th>
<th>Facies description</th>
<th>Facies code</th>
<th>% Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mud</td>
<td>Laminated silt and mud</td>
<td>Fl</td>
<td>24</td>
</tr>
<tr>
<td></td>
<td>Laminated silt and mud with dropstones</td>
<td>Fld</td>
<td>5</td>
</tr>
<tr>
<td></td>
<td>Total mud</td>
<td></td>
<td>29</td>
</tr>
<tr>
<td>Diamict</td>
<td>Massive, matrix-supported diamict with boulder to silt particle sizes</td>
<td>Dmm</td>
<td>18</td>
</tr>
<tr>
<td></td>
<td>Matrix-supported, stratified diamict with cobble to silt particle sizes</td>
<td>Dms</td>
<td>9</td>
</tr>
<tr>
<td></td>
<td>Massive, sheared, matrix-supported diamict with boulder to silt particle sizes</td>
<td>Dmm(s)</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>Total diamict</td>
<td></td>
<td>29</td>
</tr>
<tr>
<td>Gravel</td>
<td>Stratified gravel with normal grading, moderately to moderately well sorted</td>
<td>Gg</td>
<td>13</td>
</tr>
<tr>
<td></td>
<td>Stratified gravel with inverse grading</td>
<td>Gi</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>Massive or crudely bedded gravel, poor to moderate sorting</td>
<td>Gm</td>
<td>9</td>
</tr>
<tr>
<td></td>
<td>Total gravel</td>
<td></td>
<td>24</td>
</tr>
<tr>
<td>Sand</td>
<td>Horizontally laminated, moderately well sorted to well sorted sand</td>
<td>Sh</td>
<td>8</td>
</tr>
<tr>
<td></td>
<td>Massive, poorly to moderately sorted sand</td>
<td>Sm</td>
<td>5</td>
</tr>
<tr>
<td></td>
<td>Planar-bedded, steeply dipping, laminated sand</td>
<td>Sp</td>
<td>5</td>
</tr>
<tr>
<td></td>
<td>Total sand</td>
<td></td>
<td>18</td>
</tr>
</tbody>
</table>

Fig. 4. Graphic logs of sections dominated by fine-grained facies. The contour interval of the Schmidt nets is two standard deviations. \( V_i \) and \( P_i \) give the azimuth and plunge of the principal eigenvector respectively and \( S_i \) gives the strength of clustering about the principal eigenvector. The location of the sections is shown on Fig. 2.
short periods and is not likely to represent annual variations in sedimentation.

The thickest sequences of laminated silts occur at the mouth of the valley and at the base of sections in the upper part of the valley (Figs 4 & 6). Formation of the laminated sediments in the lower valley requires the formation of an ice dam which was at most a few hundred metres from the site of deposition of the sections indicated in Fig. 2. Given the proximity of the ice and the fine-grained nature of the sediments, this facies was probably deposited during a period when the ice margin was relatively stable, melting rates were relatively low and the supply of sediment was low. These conditions appear to be consistent with a slowly advancing ice margin and the initial formation of an ice-dammed lake. The lack of ice-rafted material and coarser material in the laminated silt facies support this interpretation, although previous analyses of ice-contact lakes have suggested that ice-rafted deposits may be rare despite the presence of icebergs (Gilbert & Desloges, 1987).

Laminated silt facies in the upper part of the valley (Fig. 6) were also deposited from suspension or by turbidity currents in ice-contact lakes. Two depositional environments can be deduced on the basis of sedimentary structures, geometry of deposits and facies association. The thicker, laterally continuous sequences at the base of the sections were deposited in a relatively large ice-contact lake and the thinner, complex sequences associated with abundant gravel and sand facies formed in relatively small ephemeral ponds. The changes in the geometry, complexity and facies associations from the bottom to the top of the sections suggest that a lacustrine regression occurred and there was a change from dominantly subaqueous supraglacial deposition to subaerial deposition where

Fig. 5. Graphic logs of sections dominated by steeply dipping normally graded gravel and massive diamicts. Note the difference in scales between (a) and (b). Other details as in Fig. 4.
the small layers and lenses of laminated sediment accumulated in small ponds. Abundant evidence of syndepositional warping and folding suggest that the sediments accumulated on melting ice.

Laminated silt and mud facies that occur in steeply dipping beds at the mouth of the valley were deposited by turbidity currents on the foreset beds of a delta. Beds of sand, silt and mud are characteristic of glaciolacustrine deltas (Ashley, 1975; Gustavson et al., 1985; Cohen, 1979; Gilbert & Shaw, 1981; Smith et al., 1982), although they usually form on bottomset sediments and are not generally recorded as foreset deposits. Similar laminated facies have been described by Cohen (1983) in high-energy deltaic environments at ice margins dominated by mass flows. Scouring and deformation of the upper surface of the laminated silts suggest they are foreset deposits that accumulated during quiescent periods of delta aggradation and were eroded as the overlying diamict and normally graded gravel facies were deposited.

**Diamict facies: Dmm, Dms, Dmm(s)**

**Description**

Diamict facies occur throughout the valley in association with several other facies, in particular the laminated silt facies (Figs 4–6). Particle size analyses of all diamict facies show that they have a high silt content (Fig. 7), are well graded and often have a multi-modal particle size distribution. The dominant
facies is massive, matrix-supported diamict (Dmm) with a silt matrix and particles up to 1-5 m in diameter. The pebble fabrics of massive, matrix-supported diamicts are generally parallel to the axis of the valley. Steeply dipping diamicts (26–30°) that occur at the mouth of the valley contain intraformational blocks of laminated silt and mud (Fig. 5a) and are interbedded with steeply dipping laminated silts. Pebble fabrics of these diamicts are aligned parallel to the depositional slope rather than the direction of ice movement (Fig. 5). In two locations the noses of individual flows were observed.

Less common are stratified diamicts (Dms) which occur mainly in the upper part of the valley (Fig. 6). The stratification in the diamicts consists of a crude layering from textural differentiation and/or the inclusion of layers and lenses of sorted and stratified material. Often the layers and lenses of included sediment are folded and faulted. In several locations stratified diamicts have a near-vertical dip and show evidence of collapse and the development of sedimentary dykes with distinct vertical walls (Fig. 10b). The dykes are wedge shaped, up to 1-5 m wide, 4 m deep, and occur in swarms with a consistent parallel strike. Near-vertical layering in the dyke fills may be the result of several periods of filling. The pebble fabric of the host sediment is weak and the fill sediment has a moderately strong, steeply dipping concentration indicating filling from above (Fig. 6b). Similar dykes have been described in the King Valley by Fitzsimons & Colhoun (1989). Pebble fabrics of stratified diamicts are generally weak and do not reflect the direction of ice movement (Fig. 6b).

Another less common facies are massive diamicts with prominent shear planes (Fig. 9b). These sheared diamicts are highly consolidated, have a matrix with a high silt content and lie above deformed laminated silt facies (Fig. 4). The pebble fabric of this facies is strong and parallel to the direction of ice movement.

**Interpretation**

Massive, matrix-supported diamicts (Dmm) were deposited as subaqueous debris flows. In the lower valley where this facies mainly occurs (Figs 4 & 5a), the planar, steeply dipping geometry of the diamict beds and the presence of intraformational silt blocks at the base of the diamicts suggest accumulation as high-energy subaqueous sediment flows deposited as foreset beds during the early stages of the formation of a delta. Several descriptions of Pleistocene glacio-lacustrine deltas stress the importance of avalanching and mass flow mechanisms on delta foresets (e.g. Cohen, 1979; Thomas, 1984). Cohen (1983) describes foreset beds dominated by subaqueous mass flows associated with rhythmites which strongly resemble the association of massive diamicts and laminated silts in the Linda Valley (Fig. 5a). The texture, poor sorting, preservation of flow noses and ripped-up silt clasts suggest derivation from an unsorted sediment, rapid movement, and very rapid deposition or 'freezing' of the traction carpet of the flows similar to that described by Eyles et al. (1987). Pebble fabric of the diamicts is typical for sediment flows with a moderately strong cluster (S, from 0-49 to 0-60) and an α-axis orientation that dips downslope. The occurrence of silt and mud drapes on the sediment flows suggests deposition from suspension after flow events and indicates that the sediment flows were episodic.
Most stratified, matrix-supported diamicts (Dms) have been deposited as subaerial or subaqueous sediment gravity flows. The dominantly subaerial deposition of this facies is indicated by the occurrence of deformation structures produced during deposition and collapse on melting ice and the presence of washed layers in the diamicts which indicate deposition by meltwater flows on the surfaces of the sediment flows. Sedimentary dykes that occur in stratified diamicts (Fig. 10b) also indicate subaerial deposition because they formed as tension cracks that were filled from above (Fitzsimons & Colhoun, 1989). Tension cracks form in supraglacial sediments when mass movement occurs at the ice–sediment interface and has been linked to the lateral migration of thaw lakes that cause destabilization of the supraglacial sediment cover. Development of dykes in supraglacial sediments on Icelandic glaciers has been discussed by Eyles (1979). The pebble fabrics of the stratified diamicts are similar to data for subaerial sediment flows described by Lawson (1979) and show weak to moderately strong concentrations (S, values from 0.49 to 0.60) in directions unrelated to those of ice flow (Fig. 6b).

Massive, matrix-supported diamicts that are sheared (Dmm(s)) are interpreted as lodgement tills that formed under pressure as the glacier advanced into the valley. Low-angle shear planes (Fig. 9b) were produced as the ice advanced into the ice-contact lake and over laminated sediments that were deposited at the bottom of the lake. Although the ice would have placed a considerable load on the laminated facies, folding and faulting is limited to 2 m below the prominent shear plane shown in Fig. 9(b) because the deformation was predominantly plastic. The sediments were probably preserved because the advancing ice would have been partly floating and would have exerted a relatively low pressure on the overridden sediments (R. Gilbert, personal communication, 1991). Both the pebble fabric and the highly consolidated nature of sheared diamicts are also consistent with subglacial deposition. The pebble fabric is

![Fig. 8. Laminated silt and mud facies: (a) laminated mud with ice-rafterd gravel; (b) deformed laminated mud with abundant ice-rafterd material and a diamicton between the two cobbles.](image-url)
These graded units occur as multiple stacked units or rest between laminated sand and silt facies (Fig. 5b). Where the gravels rest on laminated silts the basal contacts are scoured and eroded and contain rare silt clasts eroded from the underlying silt layers. The sandy parts of the graded units consist of poorly sorted medium to coarse sand and contain numerous ‘floating’ clasts up to 150 mm in diameter. Measurements of the pebble fabric of clasts in the sandy parts show clearly defined a-axis orientations that are parallel to the beds and have a downslope dip (Fig. 5b). Rare beds of inversely graded gravels occur within the normally graded gravel beds (Fig. 5b). Individual beds are 0.8–1.2 m thick and consist of gravelly sand that grades into moderately well sorted gravel.

Although most of the normally graded gravel beds form continuous sheets, several decrease in thickness and terminate as shown in the lower left corner of Fig. 10(c). These terminations appear to record individual episodes of gravel deposition followed by syndepositional slumping of the underlying sediments. Massive glacial facies can be divided into two subfacies on the basis of particle size and associations with other facies. The most common subfacies are poorly sorted, pebble gravels (with clasts up to 70 mm in diameter) that occur in thin beds of laminated silt, diamict and sand facies in the upper part of the valley (Fig. 6).

The second massive gravel subfacies is exceptionally coarse gravel (with particles up to 1.5 m in diameter) that has a highly variable matrix content and packing. This subfacies forms lenses up to 3 m thick at or near the base of sequences of laminated silt facies, particularly in the lower part of the valley (Fig. 4). Many exposures of this facies show a clast-supported gravel with a coarse sandy matrix, no sedimentary structures and numerous intraformational blocks of laminated and massive silt.

**Interpretation**

Normally graded gravel facies (Gg) were deposited as sediment flows at the base of high-velocity turbidity currents on a steep foreset delta slope. This facies (Fig. 10c) resembles the deposits described by Postma & Roep (1985) on a Pliocene Gilbert-type fan delta and are similar to mass flow deposits described by Postma (1984) and Eyles et al. (1987). Although the facies is dominated by normally graded beds, there are some inversely graded beds (Fig. 5b). Eyles et al. (1987) suggest that inversely graded gravels are of considerable significance because they are evidence for
of subaqueous mass flow processes. Although not fully understood, the origin of the inverse grading has been attributed to dispersive pressure involving the upward dispersion of large clasts (Walker, 1975; Lowe, 1982) and to 'kinetic sieving' which involves the settling of finer particles into voids, the upward displacement of larger particles and the formation of a traction carpet. Eyles et al. (1987) argue that irrespective of the origin of inverse grading, it is critical evidence of mass flow processes and very rapid deposition of a traction carpet at the base of a turbidity current. The presence of clasts 'floating' in sandy gravel above the base of each normally graded unit may be explained by recent experiments on high-density turbidity currents. These experiments suggest that the floating clasts are transported along rheological interfaces that develop within the flow and that they are preserved by very rapid deposition of the suspended layer of the flows (Postma et al., 1988). Clast fabrics of the normally graded facies show moderately strong to weak concentration about the mean axis ($S_1$ values from 0.72 to 0.79) and are similar to debris flows described by Lawson (1979), Eyles et al. (1988) and Eyles & McCabe (1989). The $a$-axis alignment and dip of clasts are consistently parallel to and down the depositional slope. Although $a$-axis dip in debris flows is usually upslope, Eyles & Kocis (1988) report both up- and downslope dips in subaerial debris flows on an alluvial fan and $S_1$ values between 0.45 and 0.73.

The fine-grained massive gravel subfacies (Gm) were deposited by subaerial meltwater flows. Sorting, scouring at the base of the gravel, imbrication and rare bedding in some of the fine-grained massive gravel facies suggest transportation by water. Collapse structures in the gravel and associated laminated silt and stratified diamict facies (Fig. 6) suggest that some deposits formed on melting ice. Meltwater flow is common in supraglacial environments where thin meltwater sheets and small channels develop (Eyles, 1979; Lawson, 1979). The limited thickness and lateral extent of the fine-grained subfacies (Fig. 6) indicates deposition by small ephemeral stream channels.

Fig. 10. Diamict and gravel facies: (a) massive diamict with intraformational blocks of laminated silt to the left of the hammer head; (b) stratified diamicton with vertical sedimentary dyke (outlined and arrowed); (c) steeply dipping, normally graded gravel facies interbedded with laminated sand and mud; (d) slightly deformed laminated sand with thin silt laminae and dropstones.
Coarse-grained massive gravel facies are the eroded remnants of diamict facies. Particles up to 1-5 m in diameter and inclusions of lenses of finer sediments and clasts of massive silt suggest that the gravel is a lag deposit. The lag is a remnant of sediments that were partially eroded and redeposited by high-velocity streams or mass movements that have resulted in mixing of gravel, diamict and silt facies. The location of the massive gravels at the base of lacustrine sequences (Fig. 4) suggests that they may have formed during catastrophic drainage of the ice-dammed lake during the earlier stages of lake formation or as subaqueous fans.

Horizontally bedded, massive and planar bedded sand: Sh, Sm, Sp

Description

Sandy facies mainly occur in the upper part of the valley where they are associated with diamict and silt facies (Fig. 6). Horizontally laminated sands are well sorted medium to fine sands that commonly contain thin silt laminae and rare dropstones. Massive sand facies range from poorly sorted, gravelly coarse sand to moderately well sorted medium sand. This facies typically contains lenses and/or layers of laminated fine sands and silt that are deformed, although the surrounding sand usually shows no sign of deformation (Fig. 10d). Both the horizontally laminated and massive sand facies occur in relatively thin units (less than 3 m in thickness) that form lenses and beds with limited lateral extent.

A relatively minor planar-bedded sand facies (Sp) occurs within the steeply dipping normally graded gravel facies (Fig. 5b). The planar-bedded facies form continuous sediment drapes and consist of poorly to moderately well sorted, laminated, coarse to medium sand with occasional silt laminae. Planar-bedded sand facies commonly grade into laminated silt and mud (Fig. 5b).

Interpretation

The sand facies are thought to have been deposited by sediment gravity flows, as traction deposits of small streams and/or from suspension.

The fine particle sizes and grading indicate that the horizontally bedded sand facies were probably produced by the traction and suspension elements of thin grain flows. Minor elements of the horizontally laminated sand facies are also associated with stratified diamictons deposited in supraglacial positions (Fig. 6a). This association suggests that grain flows may have formed on the surface of subaerial sediment flows possibly as excess porewater was expelled. Similar massive to graded flows have been described by Lawson (1979) who attributed his type IV flows to low-density grain flows with a high water content that are at most 0-3 m thick.

Massive sand facies are interpreted as the product of relatively low-density, turbulent, sediment gravity flows. Similar deposits were described by Eyles et al. (1987) who inferred several possible mechanisms of flow support including dispersive pressures created by grain collision, fluidization and buoyant lift provided by fine-grained sediment. Association of massive sand with normally graded gravel and diamict facies (Fig. 5) suggests that the sand was transported in turbulent suspension in the upper part of the mass flows that produced diamictons and normally graded gravel facies. The laminated silt that lies above the massive sand was deposited from suspension on top of the flow deposits.

Planar-bedded sand facies occur above normally graded gravel facies (Fig. 5a) and were formed from the tails of turbidity currents or have built up as successive grain flows during relatively quiescent periods of delta aggradation. Similar layering associated with massive sand was attributed to successive ‘freezing’ of individual grain flow pulses within a single high-volume surging flow (Eyles et al., 1987). Interpretation of the massive and planar-bedded sand facies as elements of turbidity currents is consistent with their association with normally graded gravel facies deposited as mass flows which generate turbidity currents (Hampton, 1972; Postma et al., 1988).

SEDIMENT ASSEMBLAGES AND DEPOSITIONAL MODEL

The depositional setting in which the sediments were formed is a valley that was dammed and subsequently occupied by a distributary lobe of a valley glacier. During the Early and Middle Pleistocene ice advances, the thickness of the King Glacier was greater than 400 m (Fitzsimons, 1988); this was sufficiently large to impede the local drainage patterns, dam and flow up several tributary valleys. In the blocked valleys, ice-contact lakes fed by meltwater and drainage from the surrounding rock hillsides collected in the ice-dammed basins in an environment dominated by lacustrine sedimentation.
The abundance of laminated silt facies together with its association with all other facies indicates that sedimentation in the valley was dominantly lacustrine and that many of the other facies formed against a background of lacustrine deposition. Together with the presence of steeply dipping diamicths, normally graded gravels and ice-rafted debris, this indicates that the deposits of the Linda Valley formed in a subaqueous environment characterized by rapidly changing depositional conditions.

Analyses of the spatial relationships between facies that occur in the valley suggest the primary control on the sedimentary environments was the relationship between the location of the ice edge and the relief of the valley. On the basis of this relationship, four principal sediment assemblages can be identified: proglacial lacustrine, subglacial, supraglacial and deltaic. A depositional model based on the reconstructed sedimentary environments (Fig. 11) attempts to place the formation of the sediment assemblages into the spatial and temporal context of ice advance into a tributary valley. The groupings of the sediments are called sediment assemblages to distinguish them from the morphogenetic landsystems and sediment associations described by Boulton (1976), Boulton & Paul (1976) and Eyles (1983).

Proglacial lacustrine sediment assemblage

The proglacial lacustrine sediment assemblage consists of gravels, diamicths, and laminated silts and clays deposited in a proglacial position as ice entered and dammed the valley forming a proglacial lake (Fig. 11a). The area and depth of the lake was controlled by the thickness and location of the ice front in the valley. The nature of the sediments deposited appears to have been controlled by the activity of the ice margin and the proximity of the depositional site to the ice margin.

As the ice entered the valley, deposition was dominated by massive gravel and diamicth facies that probably formed adjacent to the ice edge before a stable proglacial lake formed. Similar deposits that

Fig. 11. Depositional model for an ice-dammed tributary valley: (a) initial damming of the valley by the ice lobe and sedimentation dominated by suspension and turbidity currents with some ice rafting; (b) ice advance up the valley, into the proglacial lake, and formation of the subglacial sediment assemblage; (c) retreat of ice from the valley, decoupling and in situ melting of the stagnant ice, and the development of the supraglacial sediment assemblage; (d) further retreat and/or thinning of the ice dam causing a drop in the lake level and a lacustrine regression which changes the subaqueous supraglacial sediment assemblage to subaerial; (e) ice retreat from the valley causing removal of the ice dam and draining of the lake. The ice-contact delta collapses and may become unrecognizable.
form at the lateral and latero-frontal margins of valley glaciers during stationary and advance phases have been described by Boulton & Eyles (1979). In the early stages of ice damming, episodic drainage of the lake into the King Valley may have occurred. Such episodic drainage events may explain the formation of massive, chaotic gravels with intraformational blocks of laminated silt and the scourred surfaces with large clasts protruding into laminated silt facies (Fig. 4). When the ice dam sealed the mouth of the valley, there was an abrupt change to sedimentation dominated by suspension and turbidity currents (Fig. 4), and laminated silts and muds were deposited above the massive gravels and diamicts. Laminated silt deposits that are relatively uninterrupted by coarser sediments indicate deposition at the bottom of the lake either when the ice margin was stable or in a location some distance from the ice margin. Deposits that are interrupted by abundant ice-rafted material and/or subaqueous debris flows (Figs 4c & 9) indicate deposition close to the ice margin and are similar to proximal glaciolacustrine facies described by Ashley (1988). As the ice thickened and/or continued to advance into the valley, a lacustrine transgression occurred and the depositional site of lacustrine facies migrated up the valley (Fig. 11a). The depth of the lake may have changed, depending on the local topography and thickness of the ice dam.

Subglacial sediment assemblage

The second part of the model (Fig. 11b) records the advance of the ice lobe up the valley and into the icecontact lake. At this stage the proglacial lacustrine and preglacial deposits were overridden, perhaps at least partly eroded and then buried by subglacial sediments.

The resulting subglacial sediment assemblage consists of lodgement till and lacustrine deposits that have been subglacially deformed (Figs 4a & 11b). Subglacial deformation is indicated by folding, faulting and injection structures due to loading and shearing (Figs 4b & 9b).

Although the load and movement of ice over the silts has produced numerous normal and reversed faults, most deformation was plastic and confined to within 2 m of the top of the silts (Fig. 4a). Movement of ice across the floor of the lake has resulted in large isoclinal folds where the silt has been dragged along and upwards (Fig. 9b). Small faults and boudinage structures occur where the silts experienced tensional stresses and small folds and parasitic faults have developed where the silts were compressed. Above the warped silt a thin melange zone with prominent shear planes occurs where the structure of the silts has been destroyed and silt has been mixed with subglacial sediments (Fig. 9b). The melange grades into a layer of highly compact lodgement till that dips up-glacier. The style of the subglacial deformation is similar to till deformation described by Dredge & Grant (1987) where the silt for the most part is deformed plastically but has sufficient strength for numerous small, parasitic faults to develop. Although the sediments were observed only at the mouth of the valley where they formed during the last ice incursion, they could potentially occur at any location in the valley.

Supraglacial sediment assemblage

In the third part of the model (Fig. 11c) the glacier retreats from the upper part of the valley leaving a mass of stagnant ice. The isolated ice begins to melt in situ and a supraglacial sediment association begins to form (Boulton & Eyles, 1979; Eyles, 1979).

Sections through the Gormanston Moraine (Fig. 6) indicate that the earlier parts of the supraglacial assemblage are characterized by subaqueous sediments deposited by suspension and interrupted by diamicts and sands that formed as sediment gravity flows. In the upper parts of the moraine the decrease in the frequency, thickness and lateral extent of beds of laminated sediments, the occurrence of sedimentary dykes and the increasing structural and geometric complexity of the deposits indicate a change from subaqueous to subaerial deposition (Fig. 6). The change from dominantly subaqueous to subaerial sedimentation suggests that during formation of the early part of the assemblage a high lake level was maintained and the supraglacial sediment assemblage was submerged. As the ice continued to retreat from the valley or became thinner, the lake level decreased causing a lacustrine regression and the development of a subaerial supraglacial sediment assemblage (Fig. 11c,d).

The sharp edge of the moraine (Fig. 3), together with the surface form, structure and sedimentary properties, suggest it evolved as a partly ice-cored kame that formed where sediment was trapped between the walls of the valley and the ice edge (Figs 2 & 3). The sedimentary setting appears to be similar to elements of the subaqueous supraglacial deposition described by Shaw (1977) and Eyles et al. (1987).
Sedimentology of deposits in an ice-dammed valley

Deltaic sediment assemblage

The fourth part of the depositional model is characterized by further retreat of the ice, decrease in the lake level, stabilization of the moraine up the valley, continued lacustrine sedimentation and the formation of a deltaic sediment assemblage (Fig. 11d). As the ice retreated from the valley a mass of stagnant ice became a source of sediment that was transported down the valley. Dip of the foreset beds to the southeast and the a-axes of pebbles (Fig. 5) indicate that the delta prograded down the valley, toward the retreating ice margin and was probably derived from the supraglacial sediment and/or local streams that reworked glacial sediments. The stacked, sheet-like geometry of the diamict and gravel foreset deposits suggests that deposition was episodic and that the planar-bedded sand and laminated silt facies were deposited during quiescent periods as sandy sediment flows or from suspension. Episodic sedimentation on the delta may be linked to the lacustrine regression which could have triggered mass movements by increasing porewater pressures during draw-down events as was suggested by Shaw (1977).

Although both diamict and gravel foreset deposits are regarded as ice proximal, textural differences between the two may be due to deposition in different positions with respect to the sediment source and/or to different properties in the sediment being delivered to the delta. During the initial period of sediment accumulation, the sediments appear to have been derived from mass movements in the supraglacial assemblage and probably fluvial reworking of the sediments, transported as debris flows and deposited as steeply dipping diamictics. As the delta prograded, the sediments derived from the supraglacial sediment assemblage were increasingly modified by fluvial processes during transport across the delta top. The sorting of the sediments increased and they were transported as gravelly mass flows and grain flows on the foreset slope of the delta and deposited as normally and inversely graded gravel and sand facies. Alternatively, the deposits could be part of the same foreset beds with textural differentiation as a result of downslope evolution of fine-grained mass flows into fully turbulent flows due to dilution, acceleration and progressive deposition of the flow as it moved down the foreset slope (Postma, 1986; Eyles & McCabe, 1989). Progressive downslope winnowing of fines from the sediment flows may therefore have resulted in the deposition of diamictic on the upper part of the foreset beds and gravel in the lower parts of the beds.

Although mapping of the two deltaic deposits suggests they were deposited at different times of delta progradation, accumulation of the deposits at different locations on the same foreset beds cannot be completely discounted. A similar ice-contact delta may form against the retreating ice edge (Fig. 11d) and laminated silt and mud facies accumulate on the floor of the lake.

The fifth part of the depositional model (Fig. 11e) records removal of the ice dam, collapse of the ice-contact delta, if present, and stabilization of the deposits. The presence of a section of highly deformed sediments near the mouth of the Linda Valley suggests that an ice-contact delta did form, but that it collapsed and the sediments can no longer be recognized. Stabilization and preservation of the sediments was at least in part due to their burial by deposits from adjacent hillsides immediately after deglaciation. Burial of glaciogenic sediments by such paraglacial sediment fluxes (Ryder, 1971; Church & Ryder, 1972) is common in Tasmania and is a characteristic by-product of deglaciation in areas of medium to high relief.

DISCUSSION

Interpretation of the glacial and glaciolacustrine facies described above records and illustrates the development of an ice-contact lake in a tributary valley that was dammed and occupied by an ice lobe. All the sediment associations have been previously described and occur in several glacial environments (Shaw, 1975; Boulton & Paul, 1976; Boulton & Eyles, 1979; Eyles, 1979, 1983; Cohen, 1983). Their association in the Linda Valley is evidence for a particular kind of glacial environment that forms where tributary valleys are blocked by ice and sedimentation is controlled by the position of the ice margin in the valley and the geometry of the lake basin.

In the Linda Valley, laminated silt facies, which represent the lowest energy conditions, and diamict facies, which represent the highest energy conditions, are the two dominant facies (Table 1). Although the valley sequence contains a mixture of low- and high-energy deposits, the entire valley must be regarded as ice proximal because of its small size (Fig. 2). Abrupt changes in energy conditions have resulted in the migration of lacustrine environments, lacustrine transgressions and regressions. Although ice-contact lakes are unlikely to demonstrate any consistent
spatial or temporal trends in grain size or sorting (Ashley, 1988), interpretation of the sedimentary environments of the Linda Valley indicates a high degree of spatial and temporal organization. This organization is directly related to the interaction of a mobile glacier margin and the geometry of the valley (Fig. 11). The model of ice entry into the valley is an attempt to summarize how this interaction controls the formation of the different facies types and sediment assemblages.

The sediments of the Linda Valley are comparable with a glaciolacustrine sequence in the Okanagan Valley of Canada studied by Shaw (1977) and Shaw & Archer (1978). Shaw (1977) describes a sequence of diamicts including lodgement till, stratified and cross-stratified gravel and sands that are overlain by laminated silts and clays. Sedimentary processes deduced from the sediments indicate that they were deposited during a lacustrine transgression and deglaciation. The reconstructed sequence of events in the Okanagan Valley is stagnation of an ice lobe, in situ downwasting, development of a subaerial supraglacial sediment assemblage, lacustrine transgression and formation of a subaqueous supraglacial sediment assemblage. In contrast, the sequence deposited during deglaciation of the Linda Valley consists of subaqueous supraglacial sediments followed by subaerial supraglacial sediments and deltaic sediments. The events that produced these sediments were stagnation of ice, development of a subaqueous sediment assemblage on the downwasting ice and lacustrine regression. The lacustrine regression was caused by the retreat and thinning of the glacier in the valley (Fig. 11c,d). As the lake level dropped, the subaqueous supraglacial sediment assemblage emerged from the lake and became a subaerial sediment assemblage (Fig. 11c). This change in the nature of sedimentation in the upper part of the valley may have regulated the supply of sediment to the deltaic sediment assemblage.

In the Okanagan Valley, deltaic sediments have experienced considerable rotation and faulting because they were underlain by melting ice (Shaw, 1977). Although the deltaic sediments of the Linda Valley were derived from supraglacial sediments and possibly from other reworked sediments in the valley, there is no evidence of rotation or faulting. The implication of the absence of such evidence is that the delta did not rest on melting ice and that the retreating ice margin may have become detached from the stagnant ice, leaving a gap between the ice margin and the stagnant ice further up the valley (Fig. 11c). Possible explanations for this detachment are the influence of the topography of the valley on ice flow, and faster melting of ice beneath the lake than at the subaerial surface. Although the sharp step in the topography of the lower valley (Fig. 3) may have created conditions conducive to the detachment and isolation of the stagnant ice mass, it does not completely explain how the ice was removed. It is possible that the ice in this location melted rapidly because it was submerged, as was recorded in different geographical settings by Johnston & Brown (1964) and Shaw & Archer (1978). In contrast, the ice further up the valley may have melted relatively slowly because subaerial supraglacial debris insulates underlying ice and slows melting rates. If the debris layer reached the thickness of the frozen layer, surface melting may have ceased while the ice beneath the lake continued to melt.

In a regional context, the deposits are not unique to the Linda Valley. Comparable ice-contact sediment associations occur throughout the West Coast Range of Tasmania and may also occur in other areas of alpine glaciation where the glaciers have terminated in areas of complex topography. The preservation potential of such deposits is high because they occur in small valleys that acted as sediment traps that may have become isolated from valley glaciers and subsequent ice advances. If the ice re-enters the valley, elements of the sequence may be eroded and/or repeated, in which case the depositional model presented here may assist in the recognition and interpretation of the deposits.

ACKNOWLEDGMENTS

This work was supported by an Australian Research Grants Committee grant, the Hydro Electric Commission of Tasmania and the Department of Geography and Oceanography, Australian Defence Force Academy. I thank Dr Dave Gillieson and Val Horvath for reviewing the manuscript, Paul Ballard for drawing the diagrams and Eileen Hampson for typing the text. Critical comments on the manuscript by Drs Robert Gilbert and Gail Ashley were greatly appreciated.

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*Manuscript received 26 June 1991; revision received 13 January 1992*