3-D sedimentary architecture of a Quaternary gravel delta (SW-Germany): Implications for hydrostratigraphy

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Abstract

This paper investigates a Quaternary Gilbert-type gravel delta that was formed in an ice-marginal environment at the end of the last glacial period.

Outcrop, sediment core and ground-penetrating radar (GPR) studies reveal the sedimentary facies and depositional architecture of the delta that comprises three major units: (1) a 2–5 m thick, gravelly topset with an erosional base, formed by accretion of bedload sheets in a braided river; (2) an up to 40 m thick, steeply inclined (13–35°) foreset, dominated by gravelly lithofacies being the product of cohesionless debris flows and debris falls as well as gravity slides while sandy lithofacies was deposited by traction currents; and (3) a 10–20 m thick, sandy bottomset comprising low-density turbidites. Syn-to postdepositional deformation of parts of the bottomset deposits largely resulted from rapid deposition of overlying gravels and differential loading of the prograding foreset beds. The development of the delta was most likely controlled by a high sediment supply and lake level fluctuations. The overall coarsening-upward succession reflects delta progradation and aggradation into a glaciolacustrine environment.

Outcrop sedimentology served as a direct analogue in order to characterise the three-dimensional sedimentary and hydraulic architecture of the nearby gravel-delta aquifer. Applying a multidisciplinary approach, sedimentological, geophysical, and hydrogeological data were integrated within the 3-D modelling package Gocad (Earth Decision Sciences) to develop high-resolution 3-D aquifer models.

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1. Introduction

Gilbert-type deltas (Gilbert, 1890) with their characteristic tripartite structure comprising topset, foreset and bottomset are common in a variety of depositional environments (e.g. Nemec and Steel, 1988; Colella and Prior, 1990). In glacial settings
they may form directly in front of glacier margins where only a short subaerial distributary plain may be developed (ice-contact deltas; Feenstra and Frazer, 1988; Lønne, 1993) or they are fed by glacial meltwater streams transported across a well-defined proglacial outwash plain (glaciofluvial deltas; Lønne, 1995; Benn and Evans, 1998). In contrast to glaciofluvial deltas, the sedimentology and internal structure of ice-contact deltas reflect the close proximity to glacier ice. Redeposited diamicton sediments derived from the glacier surface may be interbedded within delta topset and foreset deposits (Shaw, 1988). Ice-front oscillations may cause syndepositional glaciotectonic deformation of the delta sediments being among the most diagnostic features of ice-contact deltas (Lønne, 1995). Typical deformation structures including slump folds and normal faults can also result from the melt-out of buried isolated blocks of glacier ice. Glaciofluvial deltas may be difficult to distinguish from Gilbert-type deltas of non-glacial depositional environments. Their origin, however, may be inferred from associated glacial facies.

The Gilbert-type gravel delta investigated in the present study was formed at the margin of the retreating Pleistocene Rhine glacier in the northern alpine foreland of SW-Germany (Schreiner, 1978). The delta developed at the mouth of the river Argen that today is a small river flowing into Lake Constance. The lower delta succession represents an important local aquifer that has partly been excavated by quarrying activities. This study concentrates on (1) the analysis of lithofacies in outcrop sections, (2) the 3-D mapping of the delta architecture based on ground-penetrating radar, sediment cores, borehole tomography data and outcrop sections, and (3) the simulation of 3-D hydraulic conductivity fields of the aquifer based on a high-resolution sedimentary model.

2. Study area

The alpine foreland surrounding Lake Constance was influenced by various advances of the Rhine glacier during the Pleistocene. Older Pleistocene deposits resting on structured Tertiary Molasse bedrock formed a moderate relief at the onset of the Würmian glaciations (Ellwanger et al., 1995). During this glacial period the present-day topography was largely modified by at least two pronounced readvances of the Rhine glacier and its meltwater streams.

The study area is situated within a terrace-shaped morphology at the eastern margin of the Schussen valley south of the city of Tettnang (Fig. 1). The studied deposits were formed after the last maximum glacier extent when the Rhine glacier retreated southward along the Schussen valley. An ice-dammed lake was formed in the Tettnang area (Heinz, 2001) and two periods of delta progradation are mapped in two different terrace niveaus indicating that former lake levels were approximately 70 m and 40 m above the present-day lake level of Lake Constance, respectively. The lower coarse-grained Gilbert-type delta was formed in a depositional environment mostly affected by ice decay in front of an active glacier as documented by dropstones interbedded in the delta deposits and large-scale kettle holes caused by buried ice blocks. Quarrying activities in three large coexisting gravel pits expose the sedimentary architecture of this gravel delta, which is the object of this study.

3. Lithofacies types

Well-exposed outcrops of the Tettnang delta were studied along three-dimensional sections to analyse the lithofacies types and their vertical and lateral variations. The lithofacies were distinguished on the basis of grain size, texture, roundness, sorting, clast fabric and stratification (Fig. 2), using a coding scheme (slightly modified from Keller, 1996; Heinz et al., 2003). These are described and interpreted below.

3.1. Poorly sorted pebble gravel (cfGcm,i)

3.1.1. Description

This lithofacies occurs exclusively in the topset and is poorly to very poorly sorted with grain sizes ranging from silt to cobbles with few scattered boulders (Fig. 2). The clasts are commonly rounded to well rounded and coarse clasts are typically imbricated with their a-axis transverse to flow (a(t)) and their b-axis imbricated (b(i)). Bed boundaries are vaguely developed and single beds are a few deci-
metres thick. Individual subhorizontal beds extend for several metres to a few tens of metres and are discernible through changes in maximum grain size and clast orientation.

3.1.2. Interpretation

The orientation and imbrication of coarse clasts suggest bedload transport (Harms et al., 1975, 1982). The movement of clasts in gravel-bed rivers happens in pulses and takes place at high flow stages primarily in form of thin bedload sheets (Reid and Frostick, 1987; Whiting et al., 1988). Deposition occurred rapidly as indicated by the lack of sorting. The relatively high matrix content mainly comprising silt and fine-grained sand indicates proximity to glacier ice and probably originated from reworked dia-mictic deposits. This lithofacies has often been described from coarse-grained braided-river deposits (e.g. Steel and Thompson, 1983; Smith, 1990; Siegenthaler and Huggenberger, 1993).

Fig. 1. The study area is located close to Lake Constance and exposes coarse-grained delta deposits of the last glaciation in three large gravel pits (location indicated with Gauss-Krüger coordinates). Different methods were used to investigate the delta sediments and to construct 3-D high-resolution models of the delta aquifer.
3.2. Alternating gravel ((c)Gcg,a)

3.2.1. Description

This lithofacies occurs in the topset and the foreset with different geometries. It generally consists of a bipartite unit with a lower zone of clast-supported, matrix-filled gravel ((c)Gcm,(b)) and an upper zone of clast-supported openwork gravel ((c)Gcg,o). Both zones are often normally graded and therefore are thought to represent one genetic unit (Fig. 2). The lower zone is characterised by a bimodal (foreset) or polymodal (topset) grain size distribution dominated by pebbles and cobbles embedded in a sandy matrix. The matrix-free upper zone comprises well sorted gravel. All components are rounded to well rounded. Openwork gravel that occurs in the topset and in the

<table>
<thead>
<tr>
<th>lithofacies-code</th>
<th>photo</th>
<th>characteristic grain size distribution</th>
<th>description &amp; process interpretation</th>
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<tbody>
<tr>
<td>cfGcm,i</td>
<td><img src="image1.png" alt="Image" /></td>
<td><img src="graph1.png" alt="Graph" /></td>
<td>POORELY SORTED PEBBLE GRAVEL occurs exclusively in topset - clast-supported - rounded to well-rounded clasts - poor to very poor sorting - imbrication of larger clasts a(t), b(i) - several dm thick beds tractional bedload</td>
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<td>(c)Gcg,a</td>
<td><img src="image2.png" alt="Image" /></td>
<td><img src="graph2.png" alt="Graph" /></td>
<td>ALTERNATING GRAVEL dominantly found in foreset, - clast-supported - normally graded - well-rounded to rounded clasts - upper zone: matrix-free (Gcg,o) - lower zone: bimodal (Gcm,b) - often cobble-dominated (cGcg,a) - bed thickness: 10-500 cm combination of debris fall &amp; cohesionless debris flow</td>
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<td>Gcm,d</td>
<td><img src="image3.png" alt="Image" /></td>
<td><img src="graph3.png" alt="Graph" /></td>
<td>DEFORMED GRAVEL occurs exclusively in foreset - clast-supported - rounded to well-rounded clasts - moderate sorting - gradually steeper clast fabric a(p), a(i) upwards - bed thickness: 20-50 cm sediment slide with minor internal deformation</td>
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Fig. 2. Summary of the major lithofacies types occurring in the coarse-grained delta deposits in Tettnang.
<table>
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<tr>
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<tr>
<td><strong>Gcx</strong></td>
<td><img src="image" alt="Image" /></td>
<td><img src="graph" alt="Graph" /></td>
<td><strong>STRATIFIED GRAVEL</strong> &lt;br&gt; occurs dominantly in foreset, continuous lateral transition from Gcx → Gmx → GS-x &lt;br&gt; - clast-supported &lt;br&gt; - well-rounded to rounded clasts &lt;br&gt; - moderate sorting &lt;br&gt; - stratification often indistinct &lt;br&gt; - preferred clast orientation a(p), a(i) &lt;br&gt; - some beds inversely graded &lt;br&gt; - bed thickness: 40-120 cm</td>
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<td><strong>Gmx</strong></td>
<td><img src="image" alt="Image" /></td>
<td><img src="graph" alt="Graph" /></td>
<td>- matrix-supported &lt;br&gt; - well-rounded to rounded clasts &lt;br&gt; - moderate sorting &lt;br&gt; - stratification parallel to bedding &lt;br&gt; - preferred clast orientation a(p) &lt;br&gt; - bed thickness: 40-90 cm</td>
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<td><strong>GS-x</strong></td>
<td><img src="image" alt="Image" /></td>
<td><img src="graph" alt="Graph" /></td>
<td>- sand to gravel mixtures &lt;br&gt; - well-rounded to rounded clasts &lt;br&gt; - well to moderate sorting &lt;br&gt; - stratification parallel to bedding, larger clasts may show a(p) fabric &lt;br&gt; - mm-thick sandy interlayers &lt;br&gt; - bed thickness: 40-90 cm</td>
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<tr>
<td><strong>S-x</strong></td>
<td><img src="image" alt="Image" /></td>
<td><img src="graph" alt="Graph" /></td>
<td><strong>STRAFORMED SAND</strong> &lt;br&gt; dominant occurrence in bottomset &lt;br&gt; - typical Bouma T&lt;sub&gt;C&lt;/sub&gt; sequence &lt;br&gt; - water-escape structures &lt;br&gt; - intercalated granules to fine pebbles &lt;br&gt; - bed thickness: 5-20 cm</td>
</tr>
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**cohesionless debris flows**

"Fig. 2 (continued)."
upper part of the foreset (close to the topset) may not be completely matrix-free and some of the interspaces may be filled with silt while single clasts are coated by fines (clay and silt).

Within the topset the alternating gravel forms cross-beds with sets of several centimetres to a few decimetres that extend laterally for 1–5 m. This lithofacies is much more prominent in the delta foreset and occurs as thin to massive (10–500 cm) layers that generally show a lobe-shaped geometry. Single lobes vary in size from a few metres in diameter to the entire length of the exposed foreset (>40 m). Maximum particle size is positively correlated with bed thickness and increases downslope. Although the two zones that comprise this lithofacies mostly occur together end-members can be found consisting of only one zone (lower or upper zone).

3.2.2. Interpretation

Different mechanisms have been proposed to explain the formation of alternating gravels associated with the migration of barforms at comparatively low bedload transport rates. Steel and Thompson (1983) considered this alteration to be formed in response to waning flows and clast segregation over the surface of a bar, whereas Carling and Glaister (1987) and Carling (1990) showed that alternating gravel can be produced without flow fluctuations downstream of a negative step (e.g. slope of a gravel bar or scour pool). A particle-overpassing mechanism and flow separation upstream of a bar accretion front initiate a grain size segregation (grading) while at the base of the slope sand supplied from suspension and bedload is mixed into the basal zone by counter flows. Reverse flows are weaker in the upper part of the slip face trapping less sand in the interstitial spaces producing an openwork texture. Downstream movement of such bedforms typically produces horizontally alternating beds, but when they migrate across an oblique plane inclined gravel alternations can be formed (Siegenthaler and Huggenberger, 1993).

While similar mechanisms may account for the formation of the alternating gravel lithofacies in the topset higher sediment rates (e.g. due to slope failure) are more likely to occur along foreset slopes, which is recorded in deposits of alternating gravel several metres thick. Mass-movement mechanisms are more important and in the foreset the alternating gravel lithofacies is interpreted as the result of avalanching processes mainly controlled by debris-fall mechanics. This is indicated by the normal grading of individual units, the downslope increase in maximum grain size and the openwork texture, which principally makes up a major part of debris-fall deposits (Nemec, 1990). The sandy matrix of the bimodal lower zone most likely is incorporated during the avalanching process by reworking of a sandy substratum as indicated by underlying sandy relicts in few places. Consequently, matrix-free deposits without a bimodal lower zone are formed when the substratum lacks sandy material. The massive assemblages of this lithofacies type suggest that clasto-clast collisions may have also played a certain role in creating mass-flow mobility. Lobes of clast-supported, essentially openwork coarse gravel that show a combination of depositional attributes of debris falls and cohesionless debris flows have been described by Colella et al. (1987). Silt coatings and infills in openwork gravels in the delta topset and in the upper part of the delta foreset are probably the result of an incorporation of suspension-rich water during the avalanche process or by later infiltration of fine-grained fractions.

3.3. Deformed gravel (Gcm,d)

3.3.1. Description

This lithofacies occurs mainly in the foreset comprising clast-supported, moderately sorted, massive gravel with rounded to well rounded clasts from medium sand to coarse pebbles. A significant clast fabric is expressed by the vertical change of imbrication angles (a-axis parallel to flow and imbricated: a(p), a(i)) that become gradually steeper from the base to the top of a bed dipping in an upslope direction (Fig. 2). Beds are a few decimeters thick and show a sharp lower boundary and non-uniform upper boundary. Single beds extend for several metres to a few tens of metres.

3.3.2. Interpretation

Imbrication of clasts with the long axes parallel to flow is usually attributed to a dominant grain-to-grain momentum transfer that forces clasts parallel to shear (Rees, 1968; Todd, 1989). Rotation of the longest clast axes towards the vertical causing a steepening-upward clast fabric is due to compression during final deposition (Massari, 1984) as displayed in sediment slides. Absence of other shear features suggests
deposition of a coherent mass that experienced minor internal deformation.

3.4. Stratified gravel (Gcx, Gmx, GS-x)

3.4.1. Description

This lithofacies type comprises clast-supported (Gcx) to matrix-supported (Gmx) stratified gravel and gravel-sand mixtures (GS-x) (Fig. 2). The grain size ranges from medium sand to gravel, cobbles are rare. Clasts are usually rounded to well rounded showing a preferred orientation parallel to bedding (a(p)) while larger clasts may be imbricated (a(i)). Stratification in clast-supported gravel is rather indistinct and predominantly apparent by clast fabric while some layers are inversely graded. Stratification in matrix-supported gravel and gravel to sand mixtures originates from grain size changes and clast fabric. Bed contacts are usually flat with a slightly non-uniform upper relief due to postdepositional erosion. Beds are a few to several decimeters thick and commonly pinch out up- and down-slope showing an inferred sheet-like geometry. A transition from a clast- to a matrix-supported gravel and finally to a gravel to sand mixture is revealed by an increasing proportion of sandy matrix that correlates with a decrease in maximum grain size. Isolated millimeter-thick sandy interlayers that laterally extend for several decimeters to metres are common in stratified gravel to sand mixtures. The varying matrix content within the clast-supported gravel may result in an open-work texture that, however, only embraces a few particles.

3.4.2. Interpretation

Preferred clast orientation parallel to flow is most likely due to clast interactions and dispersive pressure (Lewis et al., 1980) implying high-concentration flows, where clasts are not fully free to move (see above). The effect of dispersive pressure is further documented by inverse grading, which is held responsible for pushing coarser clasts upwards within the flow (Bagnold, 1954; Lowe, 1976) while smaller particles may percolate downwards to the base of the flow by kinematic sieving (Middleton, 1970; Scott and Bridgwater, 1975). Thus, this lithofacies is assumed to be the product of grain flows (Lowe, 1979), analogous to cohesionless debris flows of other authors (Nemec and Steel, 1984; Postma, 1986; Nemec, 1990). Stratification is more pronounced in matrix-supported gravel and sand to gravel mixtures reflecting segregation of clasts, which may be assigned to surging mechanisms (Nemec and Steel, 1984). The varying proportion of sandy matrix implies that the water-to-sediment ratio was slightly fluctuating from flow to flow and sediment concentration was highest in the clast-supported gravel type.

3.5. Stratified sand (S-x, S-x,d, S-x,cr)

3.5.1. Description

This lithofacies type (1) builds up the delta bottomset and most frequently appears in lower parts of the delta foreset, (2) less frequently forms horizontally stratified layers in the upper parts of the foreset where they preferentially occur on top of stratified gravels, and (3) rarely occurs in the topset as cross-beds. The grain size varies from fine- to medium-grained sand.

(1) In lower parts of the foreset and particularly in the bottomset individual beds sometimes show an erosive base and are mostly graded. A massive base is commonly overlain by horizontal stratification that passes into climbing ripples on top (S-x) (Fig. 2). Bed thickness is in the order of several centimetres but does not exceed 20 cm with a lateral extent of several metres to a few tens of metres. Water-escape structures are common. Single clasts (rounded to angular) or layers of granules and fine pebbles may be intercalated within the stratified sand lithofacies. Small-scale faulting or folding structures may occur and overprint the original fabric (S-x,d). This lithofacies rarely occurs as climbing ripples (S-x,cr) that migrate in an upslope direction extending laterally for several metres.

(2) In the upper parts of the foreset this lithofacies type occurs as non-erosive, horizontally stratified thin layers (S-x) up to 1 dm in thickness and 2–15 m in lateral extent. The preservation potential of these sandy deposits within the gravel-dominated proximal foreset beds is relatively low, which is documented by a more patchy distribution. In topset the stratified sand lithofacies forms cross-stratified trough fills (S-x) that interfinger with sets of inclined alternating gravels. Single sand layers are typically 5–20 cm thick and a few decimetres to metres in length.

3.5.2. Interpretation

(1) Normal grading, basal scour and sedimentary structures (typical T a–c succession of Bouma, 1962)
suggest deposition from sediment gravity flows, in which the sediment is supported by fluid turbulence (low-density turbidity currents according to Lowe, 1982). Intercalated gravel layers represent avalanche deposits typically from cohesionless debris flows. Single embedded clasts are thought to be either outrun components most likely associated with debris falls or dropstones extracted from melting glacier ice. Syn-to postdepositional slumping and faulting of the sandy deposits was possibly triggered by the upward escape of pore fluid as displayed by water-escape structures. Climbing ripples migrating in an upslope direction (ripple progradation and aggradation) are interpreted as the product of backflow currents on the distal delta slope modifying the top zones of turbidites. (2) The horizontally stratified sand layers suggest deposition from tractive currents flowing down the frontal delta slope. (3) Tractional bedload processes also account for the infilling of scoured depressions in topset when sand was deposited during falling stage conditions.

4. Gilbert-type deltaic elements

The three elements of a Gilbert-type delta are characterised by a distinctive lithofacies assemblage (Fig. 3), internal geometry and external form that enables the reconstruction of the sedimentary dynamics within the depositional system.

4.1. Topset

4.1.1. Description

The topset deposits mainly consist of poorly sorted pebble gravel (>80%). The horizontal to subhorizontal stacking of several beds results in a crude subhorizontal stratification that is typical for the topset structure. These stacked beds are several decimetres thick with a lateral extension up to several tens of metres. They are dissected in few places by small-scale channel to larger scale pan-like structures generally several decimetres in thickness and 2–15 m in lateral extent, which are characterised by an erosive concave lower and a flat upper bounding surface. Their internal structure consists of cross-bedded lithofacies sets of alternating gravel (10–15%) and lenses of stratified sand (1–2%) that are common at the inferred downstream margin of this unit.

The total thickness of the topset deposits ranges from 2 to 5 m at the Tettnang site. The topset erosively covers the underlying foreset beds, which is reflected by a distinct truncation surface. The upper surface of the topset outlines the smooth

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Fig. 3. Lithofacies distribution within the delta. The topset beds are mainly comprised by poorly sorted gravel lithofacies (cfGcm,i) while the bottomset beds are dominated by stratified sand lithofacies (mainly S-x and S-x,d).
Fig. 4. Description and interpretation of core KB 5/01 showing glacial diamicton overlain by a coarsening-upward succession, from prodelta, delta-bottomset to lower delta-foreset deposits. This succession indicates progradation into a glaciolacustrine lake (core photos provided by Antragsgemeinschaft Tettnanger Wald).
present-day topography, which is dissected by ice-decay features (kettle holes) tracing through the substratum as well as flat sheets of possibly eolian sand dunes.

4.1.2. Interpretation

The subhorizontally stratified units are interpreted as the vertical accretion of pulsatory bedload sheets within the active river channel. Hein and Walker (1977) related the growth of diffuse gravel sheets to bar formation. According to their model, a crude horizontal stratification typically with imbricated clasts develops under conditions of rapid gravel transport when the gravel sheets lengthen downstream faster than they aggrade. These low-relief barforms are very common in modern outwash rivers, where they form fundamental depositional units (e.g. Gustavson, 1974; Smith, 1974; Rust, 1978).

Locally, these in-channel accumulations are partly eroded by cut-and-fill structures, which is clearly indicated by the curved, basal erosional surface (cf. Ramos and Sopena, 1983; Morison and Hein, 1987; Smith, 1990). The erosional phase is followed by filling of the carved structures by foreset deposition at the upstream end and lateral accretion along the flanks (Siegenthaler and Huggenberger, 1993). Channel-like forms possibly represent scour pool fills that occur at channel bends or intersections of minor channels (Ashmore, 1982). Larger pan-like structures may represent broader depressions in river channels that may result from a less channelised discharge pattern (Heinz and Aigner, 1999). The occurrence of stratified sand lenses at the downstream margin of cut-and-fill structures implies decreasing flow conditions at the final stages of their formation.

The internal structure of the topset suggests the development of a highly dynamic river system that is characterised by unstable and low-relief channels and a high sediment discharge pattern typical of braided streams. Within such a system prominent scour pools are commonly not well developed and gravel sheet deposits predominate.

4.2. Foreset

4.2.1. Description

The foreset deposits are characterised by steeply inclined (13–35°) beds of alternating gravel (25–40%) and stratified gravel (55–70%) with minor occurrence of deformed gravel (<5%) and stratified sand (35%). Dip orientations of the foreset beds vary from a southwesterly to northerly (230–340°) direction implying some curvatures in the depositional front. Single angular clasts up to boulder size are scattered within the foreset succession. Diamicton deposits have not been observed in outcrops, but are documented in cores where they are preferentially interbedded within lower foreset beds (Fig. 4). Direct correlation of outcrop-analogue data with subsurface data was possible along one exposed section indicating the continuation of the foreset succession in the subsurface (Fig. 5). As described above, single foreset beds show different geometries ranging from lobate to sheet-like and commonly pinch out upslope or down-slope. Distinct erosional surfaces are common features within the foreset structure that sharply truncate individual foreset beds. These features were furthermore resolved by GPR measurements that allowed to infer the internal structure of the delta deposits in great detail. Direct comparison of the GPR profile with the outcrop face illustrates the close similarity of radar reflection patterns with the exposed sedimentary structures (Fig. 6). The average grain size and the dip angle of foreset beds generally decrease downslope defining a typical proximal to distal trend within the foreset. While massive, alternating gravels (cGcm,b; cGcg,o) predominate in proximal parts, stratified gravels (Gcx,Gmx,GS-x) and stratified sands (S-x) are more frequent in distal parts of the foreset (Fig. 7).

The total thickness of the foreset deposits as revealed from outcrops and cores is as much as 40 m. Whereas the upper erosive contact to the topset is roughly flat, the interface between the foreset and bottomset is incised in places.

Fig. 5. Field sketch highlighting the internal geometry and lithofacies assemblage of the proximal delta foreset in the study area. The proximal foreset beds are dominated by thick openwork gravel (cGcg,o) that typically form broad lobes a few tens of metres in diameter. Stratified gravel (predominantly Gcx) displays a more sheet-like geometry with single beds commonly pinching out up-and downslope. Close to this section the sediment core A2/02 allowed to correlate the outcrop with the non-exposed subsurface revealing an overall total thickness of the foreset of 40 m, which is underlain by sand-dominated bottomset beds.
4.2.2. Interpretation

Sedimentation processes on the foreset slopes were complex resulting in a heterogeneous foreset structure. Resedimentation and transportation of sediments on the foreset slope was dominated by various types of sediment gravity flows. The main contributing processes forming the foreset beds were by cohesionless debris flows and debris falls, which were particularly prominent on the proximal foreset slope, with only minor influence of gravity slides and traction currents. Interbedded angular clasts within commonly well-rounded avalanche deposits are interpreted as dropstones derived from melting ice blocks. Interbedded diamicton deposits reflect direct input from glacier debris most likely by cohesive debris flows although little is known about the distribution and geometry of these deposits. Shearing and loading forces of moving, overpassing sediment masses partly reworked the substratum thus generating prominent erosional surfaces. Low-density turbidity currents mainly bypassed these high-angle slopes and played a major role in transporting sandy sediments to the delta toe area.

4.3. Bottomset

4.3.1. Description

The bottomset deposits are characterised by flat-lying beds of stratified sand (＞85%) with interfingered stratified gravel (＜15%) and display a sheetlike geometry with lateral dimensions of at least several tens to hundreds of metres (Fig. 8). Locally, the beds are faulted and folded. Their transition to underlying silt-bearing sands that most likely indicate prodelta sedimentation is gradual and the total thickness of the bottomset deposits can only roughly be estimated to range between 10 and 20 m. Erosion by subsequent prograding foreset deposition locally altered the thickness of the bot-
Fig. 7. Field sketch illustrating the transition between distal delta foreset and bottomset. The downslope decrease in foreset dipping angle and overall grain size coincides with an increase in stratified gravel (Gcx, Gmx, GS-x) and stratified sand (S-x) in comparison to proximal parts of the foreset. Note the change in orientation angle of the section at 0 m.
Fig. 8. Field sketch of the delta bottomset. The bottomset beds mainly consist of stratified sand (S-x, S-x,d and S-x,cr) and show a sheet-like geometry with a lateral extent of several tens to a few hundreds of metres. Preservation of the bottomset sediments is largely controlled by the degree of erosion of overlying, prograding foreset beds that commonly truncate the underlying sandy deposits (as indicated in W/E section).
tomset deposits, which are therefore assumed to vary laterally.

### 4.3.2. Interpretation

The primary sediment-transport process that carried fine-grained sediments downslope and deposited the sand-dominated bottomsets was by turbidity currents. The collapse of the front of the low-density turbidity currents (Lowe, 1982) was likely triggered by a break in the delta slope accompanied by minor changes in flow velocities that resulted in successive accumulations of incomplete (“top cut-out”) turbidites. Minor occurrence of intercalated stratified gravel implies that only few cohesionless debris flows were competent enough to descend the delta slope and reach the distal bottomset zone. Rapid deposition of gravels and differential loading by the prograding foreset beds on top of the water-saturated sandy bottomset beds may have contributed to syn- and post-depositional deformation of the latter.

### 5. GPR facies

Ground penetrating radar was used to map the subsurface structure of the delta. Twenty profiles of several tens to hundreds of metres were measured with a GSSI SIR 2000 (Geophysical Survey Systems Inc.) using shielded 100 MHz antennas. Basic data processing steps included dewow, DC-shift, time zero adjustment and amplitude scaling with AGC using the program Reflexw (Sandmeier Scientific Software, Germany). The depth scale was based on average near-surface velocities of 0.12 m/ns (unsaturated zone) and 0.07 m/ns (saturated zone), which were determined from common midpoint measurements.

Fig. 9. (A) Discontinuous, subhorizontal reflectors that are locally truncated by semi-circular and oblique tangential reflectors characterise the delta topset GPR facies. This reflection pattern truncates the underlying foreset reflectors recording a major erosional surface. (B) The foreset GPR facies is distinguished by continuous, inclined reflectors that show onlap and truncation patterns, which indicate reworking of foreset beds. Onlapping of subhorizontal, parallel reflectors that characterise the bottomset GPR facies onto inclined foreset reflectors, may point to lowstand deposition initiated during a lake level fall. 3-D GPR surveys are needed to clarify this interpretation. Vertical exaggeration of GPR profiles is a factor of two.
Three major radar facies were identified (cf. Smith and Jol, 1997; Aspiron and Aigner, 1999).

5.1. Topset GPR facies

The reflection pattern of the topset is characterised by discontinuous, subhorizontal to wavy reflectors of generally moderate amplitude (Fig. 9). They are locally truncated by semi-circular and oblique tangential reflectors of similar amplitude, which are interpreted to represent small-scale cut-and-fill structures (Huggenberger, 1993). Subhorizontal reflectors are thought to reflect bedload sheets. The erosive base to the underlying foreset is clearly indicated in the GPR profiles.

5.2. Foreset GPR facies

GPR signatures from the foreset are characterised by inclined (10–25°), continuous (5–40 m) reflectors of low to high amplitudes (Fig. 9). The variation in amplitude is caused by changing lithofacies types. Major reflectors represent distinct lithofacies transitions such as the base of openwork gravel in contact to bimodal gravel (Becht, 2004). However, a correlation of internal reflection patterns with individual beds and specific lithofacies types is unclear. The reflectors are steeper (20–25°) in the upper part of the foreset and may flatten to nearly horizontal in the lower part transitional to the bottomset. An erosional truncation against the overlying topset GPR facies marks the upper termination of the foreset reflectors. Internally, truncation of single reflectors is common illustrating the erosion of single foreset beds by succeeding deposits.

5.3. Bottomset GPR facies

The reflection pattern of the bottomset is distinguished by parallel, subhorizontal reflectors of moderate to high amplitude. Typically, these reflectors are of high continuity (several tens of metres), but in places may be truncated by inclined foreset reflectors. Onlap terminations of horizontal bottomset reflectors against the steeper inclined foreset reflectors were locally observed in GPR profiles (Fig. 9). The interpretation of this distinct reflection pattern from 2-D profiles alone is ambiguous. However, this reflection pattern may possibly reflect a lowstand deposit, which accumulated during a lake level fall. It is also possible that this pattern was caused by a directional shift in delta growth.

6. Delta architecture

Outcrop, sediment core and GPR data were integrated to resolve the overall delta architecture. The gravel delta covers an area of 2–3 km² and forms an isolated body generally 40–60 m thick. It exhibits a fan-like external shape with a steep marginal slope. Progradation of the delta into the ice-dammed lake occurred for about 1–2 km length from the inferred point of sediment origin in the northeast producing a typical coarsening-upward succession (Fig. 4). The upper boundary surface of the delta is represented by the present-day topography, while particularly in distal parts of the delta, the lower contact to prodelta sediments is gradational.

The fine-grained prodelta deposits underlying the delta sediments partly smoothed out the relief that was generated by the melting Rhine glacier at the end of the last glaciation. Deposition of fines, thick diamicton deposits and glaciofluvial gravels prior to the delta formation took place in a depositional environment characterised by ice-decay processes. Kettle holes record the melt-out of locally buried ice-blocks. The stream-driven gravel deposits generally form elongated ridges indicative of esker or kames formation. In proximal parts of the delta, they are directly overlain by the delta deposits that are the focus of this study.

The transition between gravelly foreset beds and underlying sandy bottomset beds generally rises with increasing distance from the original source of sediment input. As a result, bottomset deposits are exposed in the same topographic level as foreset beds across the gravel pits in several hundred metres distance. Delta growth therefore was both progradational and aggradational, which indicates a generally high supply of sediments (cf. Homewood et al., 2000). Sediment discharge was largely influenced by the melting of glacier ice that accounted for high loads of sediments. Aggradation of bottomset beds may be related to a phase of minor lake level fall, which caused erosion and bypassing of sediment in the
topset zone and accumulation of lowstand deposits in the bottomset zone (as indicated in GPR profiles). A fall in lake level thus may explain the erosive nature between topset and foreset deposits. Generally, a falling lake level played a major role in shifting delta growth from the north (upper delta terrace) to the south (investigated younger delta terrace) in the Tettnang region.

There is no evidence to suggest that glaciotectonics influenced delta development. Identification of a synsedimentary fault within the foreset was the only indication of glaciotectonic activity in the study area apart from small-scale soft sediment deformation structures of bottomset beds. A glaciotectonic control on the sedimentary architecture of the delta can therefore be neglected.

**7. 3-D high-resolution models and hydrostratigraphy of delta foresets**

The lower succession of the delta sediments forms a local groundwater aquifer. In order to predict groundwater flow and contaminant transport in such heterogeneous systems a three-dimensional distribution of hydraulic properties is required. Based on the genetic understanding of the depositional system described above a high-resolution 3-D sedimentary model of the delta aquifer was developed by integrating outcrop-analogue, core, and geophysical data in the program Gocad. The sedimentary model was used as an accurate representation of the subsurface architecture to simulate 3-D images of the hydraulic conductivity distribution, which were quantified by flowmeter measurements.

The database used to construct the 3-D sedimentary model consists of five sediment cores, GPR profiles and crosshole tomography data (Fig. 10, for location see Fig. 1). The cores were analysed and interpreted in terms of lithofacies types by applying a standardised scheme considering outcrop-analogue results (Fig. 11). GPR profiles were measured at the site with a line spacing of 1 m in dip direction and a line spacing of 7 m in strike direction covering an area of 24 m times 30 m (see Fig. 1). Crosshole radar-tomographic measurements were acquired in vertical sections between all wells to obtain tomographic images of the velocity distribution in the subsurface.

![Fig. 10. (A) The database used to construct the 3-D aquifer models of the Tettnang delta consists of sediment cores, ground-penetrating radar profiles, crosshole tomographic sections, and hydraulic conductivity (K) logs. (B) Based on these data and outcrop-analogue studies, different lithofacies were identified and mapped by tracing their boundary surfaces. (C) Detail highlighting the irregular cell grid structure that underlie the aquifer models. The flexible stratigraphic grid (SGrid) conforms to lithofacies boundaries and ensures that small-scale (cm-dm scale) changes in lithofacies distribution are adequately represented.](image-url)
The tomography survey was carried out with a Ramac borehole radar system (MALA Geosciences, Sweden) and two 100 MHz antenna probes.

7.1. Model construction

The overall size of the 3-D models is 13.5 m × 12 m with a thickness of 26 m covering an area that is outlined by the five wells. The model is composed of irregular cells that are defined by their corner points forming a flexible stratigraphic grid (SGrid, Fig. 10). The SGrid structure conforms to controlling boundary surfaces (here lithofacies boundaries) and consequently single cells do not share the same shape. The deformed cells still exhibit the same horizontal grid size of 15 cm times 17 cm but vary in thickness. Variations in cell thickness typically range from 2 cm to 15 cm and occur not only between different cells but also within single cells. The chosen grid size represents the maximum resolution within Gocad that still produces reliable results taking into account the generally large grain sizes of the deposits and the measurement accuracy of geophysical and sedimentological investigation methods. The flexible grid structure ensures that small-scale changes in sedimentary architecture can be reproduced in great detail. This is particularly important when modelling inclined structures such as a delta foreset with variable dip angles.

The sedimentary model was generated in two steps comprising (1) the mapping of lithofacies types and (2) the modelling of grain size distributions within individual lithofacies. Based on this lithological framework, (3) the hydraulic conductivity distribution was simulated (Fig. 12). The various realisations of hydraulic conductivity fields serve as input data for subsequent three-dimensional flow modelling.

7.2. Mapping of lithofacies

The main lithofacies types in the delta foreset as mapped from outcrop were also identified in the sediment cores B1–B5 (Fig. 1). The only exception is the deformed gravel lithofacies (Gcm,d) that could not clearly be distinguished in cores, because of its similar grain size distribution to the stratified gravel lithofacies (Gcx, Gmx) and the disturbed texture during the drilling process.

The spatial distribution of lithofacies between the cores were correlated by GPR and tomography data due to changes in relative permittivity between different lithofacies types. The high density of GPR and tomography profiles and their calibration with core data allowed to determine the lithofacies boundaries and to trace them through the model region. In this way, various triangulated boundary surfaces were mapped defining distinct lithofacies types (see Fig. 10).

7.3. Modelling of grain size distribution

Based on the grain size distribution determined from core data different grain size classes were defined on the basis of the dominant grain size. The spatial distribution of these grain size classes within
different lithofacies was modelled using geostatistical analyses. In order to remove dipping effects, the geostatistical analyses were carried out in the XYW space by transforming the vertical, real world Z coordinate into a normalised W coordinate. The W coordinate runs parallel to distorted cells of the SGrid ranging from 0 to 1. Generally, the spatial trends in a distributed data set (e.g. cores) can be described by a 3-D variogram that is calculated from different directional variograms (combination of vertical (1-D) and aerial (2-D) variograms). The computation of a theoretical model to the 3-D variogram defines the variogram model and characterises the distribution of variances observed in the experimental variograms. However, due to the small number of cores and the relatively large distance between the cores in the model area compared to the generally smaller correlation length of the property (here grain size classes), the calculation of precise aerial variograms was limited and often not possible. In such a situation a trial and error approach was applied by generating a number of models and choosing the model that revealed the geologically most realistic pattern.

The spatial distribution of grain size classes were modelled using indicator kriging. Indicator kriging is a variogram-based method that estimates the weighted spatial mean of neighbouring samples derived from the variogram model considering the distance and the spatial distribution of the samples (e.g. Isaaks and Srivastava, 1989). Indicator kriging tries to find an optimum combination of the weights to minimise local errors. In this way clustering effects can be prevented although there exists a tendency to underestimate the full range of property values.

7.4. Simulation of hydraulic conductivity distribution

Based on the sedimentary model that describes the spatial distribution of grain size classes within individual lithofacies, various hydrofacies were determined (cf. Poeter and Gaylord, 1990). These interconnected hydrogeological units display relatively homogeneous hydraulic properties (e.g. hydraulic conductivities), which were quantified in the field by flowmeter measurements and were theoretically calculated from grain size distributions (Kozeny, 1927; Carman, 1937; Beyer, 1964) (Table 1).

Applying the geostatistical relationships established in the previous model step (see above), the Sequential Gaussian Simulation (SGS) method was used to simulate the variability of hydraulic conductivities within single hydrofacies types. SGS is a variogram-based geostatistical approach that specifies not only statistical anisotropy, but models heterogeneities by adding a random factor. This factor is expressed by the mean and variance of hydraulic conductivity values of single hydrofacies types and their geostatistical distribution (e.g. Gaussian distribution). In addition, the SGS approach allows to quantify minimum and maximum estimates that restrict the possible range of hydraulic conductivity values. In contrast to indicator kriging the SGS method can produce an infinite number of realisations of equally

<table>
<thead>
<tr>
<th>Lithofacies</th>
<th>Hydrofacies</th>
<th>Porosity [-] (extended after Heinz et al., 2003)</th>
<th>( K ) [m/s] measured (flowmeter data)</th>
<th>( K ) [m/s] calculated</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gg,a</td>
<td>Gg,o</td>
<td>0.26 ± 0.02</td>
<td>( 2.6 \cdot 10^{-2} \pm 2.3 \cdot 10^{-2} )</td>
<td>( 2.4 \cdot 10^{-2} \pm 8.4 \cdot 10^{-3} )</td>
</tr>
<tr>
<td>cGg,o</td>
<td>0.26 ± 0.02</td>
<td></td>
<td>( 1.3 \cdot 10^{-1} \pm 7.4 \cdot 10^{-2} )</td>
<td>( 2.8 \cdot 10^{-1} \pm 5.4 \cdot 10^{-1} )</td>
</tr>
<tr>
<td>Gcm,b</td>
<td>0.20 ± 0.08</td>
<td></td>
<td>( 7.4 \cdot 10^{-4} \pm 6.0 \cdot 10^{-4} )</td>
<td></td>
</tr>
<tr>
<td>cGcm,b</td>
<td>0.20 ± 0.08</td>
<td></td>
<td>( 5.0 \cdot 10^{-4} \pm 2.0 \cdot 10^{-4} )</td>
<td>( 9.8 \cdot 10^{-6} \pm 8.6 \cdot 10^{-6} )</td>
</tr>
<tr>
<td>Gcx</td>
<td>Gcx</td>
<td>0.18 ± 0.03</td>
<td>( 4.0 \cdot 10^{-2} \pm 2.8 \cdot 10^{-2} )</td>
<td>( 1.0 \cdot 10^{-3} \pm 7.7 \cdot 10^{-4} )</td>
</tr>
<tr>
<td>Gmx</td>
<td>Gc–Gmx</td>
<td>0.17 ± 0.07</td>
<td>( 1.1 \cdot 10^{-3} \pm 9.3 \cdot 10^{-4} )</td>
<td>( 7.5 \cdot 10^{-4} \pm 1.8 \cdot 10^{-4} )</td>
</tr>
<tr>
<td>GS-x</td>
<td>GS-x</td>
<td>0.27 ± 0.07</td>
<td>( 3.1 \cdot 10^{-3} \pm 1.2 \cdot 10^{-3} )</td>
<td>( 2.9 \cdot 10^{-4} \pm 5.1 \cdot 10^{-5} )</td>
</tr>
<tr>
<td>S-x</td>
<td>S-x</td>
<td>0.36 ± 0.04</td>
<td>( 6.2 \cdot 10^{-4} \pm 3.2 \cdot 10^{-4} )</td>
<td>( 2.5 \cdot 10^{-4} \pm 8.4 \cdot 10^{-5} )</td>
</tr>
</tbody>
</table>

\( ^a \) Based on Kozeny-Carman equation (Kozeny, 1927; Carman, 1937).

\( ^b \) Calculated based on \( K = K \) (Gcm; sand) \((1 - V(C/B))\).

\( ^c \) Based on empirical equation according to Beyer (1964).

likely hydraulic conductivity fields, which provide a quantitative measure of uncertainty. The realisations were conditioned on measured in-situ data, thus honouring hydraulic conductivity estimates derived from flowmeter measurements at the aquifer site.

7.5. Characteristics of model results

The 3-D sedimentary model is characterised by steeply inclined (9–27°) beds of alternating gravel lithofacies that consists of generally 1.5–3 m thick openwork layers (cGgc,o) and 1–2 m thick bimodal gravel layers (cGcm,b) (Fig. 13). Normal grading within this lithofacies is recorded in the model by a grain size transition from large cobbles to cobbles. Stratified gravel lithofacies (Gcx, cGcx, and rarely Gmx) occur throughout the succession but are particularly prominent in the lower half of the model while stratified sand lithofacies (S-x) locally forms a small lense within the gravel lithofacies. Sand to gravel mixtures (GS-x) build the lowermost bed, which is abruptly overlain by gravel lithofacies.

Within the lower two third of the succession the sequence shows an overall coarsening-upward trend from coarse sands to large cobbles, which is interpreted as the transition from sandy bottomset beds to gravel-dominated foreset beds. The decrease in average grain size in the upper third of the succession may reflect decreasing energy conditions (temporally restricting the supply of coarse material to the delta slope) or a directional shift of delta growth away from the investigated site (increasing distance to the distributary channels). Direct correlation of the nearby outcrop face exposing proximal delta-foreset beds with the model results suggests that the modelled lithofacies represent distal deposits on the delta-foreset slope. Variations in dip direction of the foreset beds from a southwesterly to westerly (245–285°)

Fig. 13. The 3-D sedimentary model is characterised by an overall coarsening-upward trend from coarse sand to large cobbles in the lower two third of the succession indicating progradation of delta-foreset beds over delta-bottomset beds. Normal grading of the alternating gravel lithofacies is particularly well defined for the illustrated cGgc,a-layer, which consists of cobble-dominated openwork gravels in their upper half representing high conductivity zones.
orientation is in accordance with outcrop measurements and may document minor fluctuations in the course of deposition. Care has to be taken however, when interpreting the overall development of the delta merely based on the small-scale model results.

In order to produce three-dimensional hydraulic conductivity fields of the gravel aquifer, hydrofacies types were defined based on the sedimentary framework described above. By applying the SGS approach hydraulic conductivity estimates were assigned to each grid cell within a particular hydrofacies type considering not only the variance of the hydraulic property but also spatial geostatistical relationships described through a variogram model. In this way, the heterogeneity due to grain size variations that have the greatest influence on flow pattern and the variability of hydraulic conductivity values of individual hydrofacies can be honoured.

The hydraulic conductivity field is characterised by inclined layers of high conductivity (Fig. 14). The thickest and laterally most continuous conductivity layers \( (K=10^0-10^{-2} \text{ m/s}) \) correspond to openwork gravels. Cobble-rich stratified gravels may form zones of relatively high conductivity \( (K=10^{-1}-10^{-2.5} \text{ m/s}) \) that are locally restricted. The high conductivity layers are interbedded within zones of relatively low conductivity \( (K=10^{-2}-10^{-4} \text{ m/s}) \) with the lowest values typically corresponding to bimodal gravels. Generally, hydraulic conductivity estimates display only minor variations laterally (typically up to one order of magnitude as a maximum), but are very variable vertically (up to two orders of magnitude). This anisotropy is caused by the vertical stacking of different hydrofacies that are relatively continuous over the entire length of the modelled area. The different equally probable realisa-

\[ \text{log} k > 2 \text{ m/s} \]

\[ 0 -1 -2 -3 -4 -5 \]

Fig. 14. (A) 3-D image of the hydraulic conductivity distribution of the gravel aquifer. (B) The most prominent high-conductivity layers (log \( k > -2 \) m/s) correspond to thick openwork gravels (cGgc,o) while stratified gravels (cGcx, Gcx) may form less continuous, high-conductivity areas with slightly lower conductivity estimates.
tions of hydraulic conductivity fields serve as input data for three-dimensional flow modelling and allow to quantify groundwater flow patterns and contaminant transport.

8. Conclusions

The Tettnang gravel delta was deposited in an ice-marginal lake at the end of the last glaciation. Proximity to the Rhine glacier is documented by kettle holes, dropstones, diamicton deposits and to a minor degree by synsedimentary faults that most likely were initiated by glaciodynamic processes. These features are indicative of an ice-contact delta although the sedimentology and internal structure of the delta is similar to Gilbert-type deltas of non-glacial origin (e.g. Hwang and Chough, 1990; Sohn et al., 1997).

The delta topset was formed by a braided river forming low-relief gravel bars and local shallow cut-and-fill structures. Foreset sediments were mainly transported by cohesionless debris flows and debris falls depositing lobe-shaped alternating gravels with thick openwork beds and sheet-like stratified gravels. Gravity slides accounting for deformed gravels and traction currents forming stratified sands are less common in the delta foreset. Low-density turbidity currents frequently bypassed the delta slope and deposited stratified sands that dominantly comprise the bottomset. The bottomset beds were locally deformed showing small-scale folds, faults and water-escape structures.

Delta growth was mainly progradational with an aggradational component due to high sediment input, which represents the dominant control on delta development. Melting of glacier ice is thought to account for high discharge rates that generated abundant sediment supply, but also caused fluctuations in lake level that may explain the erosive nature between topset and foreset.

The integration of outcrop, sediment core, GPR, crosshole tomography and flowmeter data within the 3-D modelling tool Gocad provides a mean for the three-dimensional characterisation of sedimentary and hydraulic heterogeneities in coarse-grained aquifers. The investigated delta aquifer is characterised by continuously inclined high conductivity layers that mainly correspond to openwork gravels. These layers display hydraulic conductivities that are generally 2–3 orders of magnitude higher than for other gravel and sand lithofacies. As a consequence, groundwater flow is largely affected by these interbedded high conductivity layers that have a tendency to promote channeling of flowpaths.

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