Middle Pleistocene (Saalian) lake outburst floods in the Münsterland Embayment (NW Germany): impacts and magnitudes

Janine Meinsen a, Jutta Winsemann a,*, Axel Weitkamp a, Nils Landmeyer a, Andreas Lenz b, Manfred Dölling b

a Institut für Geologie, Leibniz Universität Hannover, Callinstr. 30, D-30167 Hannover, Germany
b Geologischer Dienst Nordrhein-Westfalen, De-Greiff-Straße 195, D-47803 Krefeld, Germany

A R T I C L E   I N F O

Article history:
Received 3 September 2010
Received in revised form 28 April 2011
Accepted 19 May 2011
Available online 5 July 2011

Keywords:
Glacial Lake Weser
Lake outburst floods
Megafloods
Plunge pools
Streamlined hills
Megaflutes
Saalian
Drenthe glaciation
Hondsrug ice stream
Strait of Dover

A B S T R A C T

During the late Saalian Drenthe glaciation ice-damming of the Upper Weser Valley led to the formation of glacial Lake Weser. The lake drained catastrophically into the Münsterland Embayment as the western ice dam failed, releasing up to 110 km³ of water with a calculated peak discharge of 2.5 × 10⁶ m³/s to 1.3 × 10⁷ m³/s. Geographic information systems (GIS) and high-resolution digital elevation models (DEM) were used to map streamlined landforms and channel systems in front of lake overspills. Geological maps, 2450 boreholes and the DEM were integrated into the 3D modeling program GOCAD to reconstruct the distribution of flood-related deposits, palaeotopographic surfaces and the internal facies architecture of streamlined hills. The drainage pathways are characterized by the occurrence of deep plunge pools, channels, streamlined hills and 4 km long and 12 m deep V-shaped megaflutes. Plunge pools are deeply incised into Mesozoic basement rocks and occur in front of three major overspill channels. The plunge pools are up to 780 m long, 400 m wide and 35 m deep. Approximately 1–10.5 km downslope of the overspill channels fan shaped arrays of streamlined hills are developed, each covering an area of 60–130 km², indicating rapid flow expansion. The hills commonly have quadrilateral to elongated shapes and formed under submerged to partly submerged flow conditions, when the outburst flood entered a shallow lake in the Münsterland Embayment. Hills are up to 4300 m long, 1200 m wide, 11 m high and have characteristic average aspect ratios of 1:3.3. They are separated by shallow, anabranching channels in the outer zones and up to 30 m deep channels in the central zones. Hills partly display V-shaped chevron-like bedforms that have apices facing upslope, 1.6–2.5 km long, 3–10 m high, 0.8–1.2 m from limb to limb, with limb separation angles of 20–35°. These bedforms are interpreted as mixed erosional depositional features. It is hypothesized that the post-Saalian landscape evolution of the Münsterland Embayment has considerably been influenced by catastrophic floods of glacial Lake Weser, creating large and deep valleys, which subsequently became the new site of river systems. The outburst floods probably followed the east-west-trending Saalian Rhine-Meuse river system eventually flowing into the North Sea, the Strait of Dover and the Bay of Biscay. It is speculated that the Hondsrug ice stream may have been enhanced or even triggered by the formation and outburst of glacial lakes in the study area.

1. Introduction

The largest known terrestrial floods, with immense geomorphological impacts, are associated with glaciation and are linked to flood events from sub- and proglacial lakes, such as the Late Pleistocene outburst of glacial Lake Missoula and glacial Lake Agassiz in the North-West USA (Baker, 2002a, b; Teller et al., 2002) and the Altai flood in Siberia (Rudoy, 2002; Carling et al., 2009, 2010). During the Holocene, outburst floods have been triggered by volcanic activity underneath the Icelandic glaciers (Björnsson, 2009).

All these superfloods have had an enormous impact on landscape evolution and show extreme hydraulic variables, including total volumes of tens to thousands of cubic kilometers and peak discharge of 0.1–20 × 10⁶ m³/s. Valleys formed by these floods significantly influenced the courses of rivers and groundwater flow paths (Gupta et al., 2007; Jørgensen and Sandersen, 2009).
volume megafloods are thought to have even forced abrupt climatic change by disrupting ocean circulation pattern (Baker, 2002b; Teller et al., 2002; Murton et al., 2010).

Although large and deep proglacial lakes formed in central Europe at the margin of the Middle Pleistocene ice sheets (Thome, 1983; Kostermann, 1992; Junge, 1998; Eissmann, 2002; Winsemann et al., 2007, 2009, 2011), little is known about the drainage history of these lakes, the pathways of outburst floods and their impact on landscape evolution. The outburst of a middle Pleistocene glacial lake in the North Sea area probably caused the formation of the Strait of Dover (Belt, 1874; Smith, 1985; Gupta et al., 2007).

During the late Saalian Drenthe glaciation ice-damming of the Upper Weser Valley led to the formation of glacial Lake Weser (Figs. 1 and 2). During the maximum lake-level highstand ~120 km³ water was stored in the lake basin (Fig. 2). The lake drained catastrophically southwestward into the Münsterland Embayment as the western ice dam failed, releasing up to ~110 km³ of water (Winsemann et al., 2011, in press).

In this paper we will document lake-outburst related streamlined landforms and channel systems that formed during the catastrophic drainage of glacial Lake Weser. Geological maps, boreholes and digital elevation models (DEM) were integrated into the 3D modeling program GOCAD to reconstruct the distribution of flood-related deposits, flood-related palaeotopographic surfaces and the internal facies architecture of streamlined hills. The main objective of this work is to provide a synthesis of lake outburst-related erosional and depositional features that may add to the recognition and interpretation of pre-Weichselian lake outburst events.

1.1. The work of glacial outburst floods

Glacial outburst floods (jökulhlaups) occur in all glaciated regions of the world with varying magnitudes and destructive energies depending on peak discharge. Three types of jökulhlaups can be distinguished (Alho, 2005). “Minor” jökulhlaups with a peak discharge of 10³–10⁴ m³/s, “major” of 10⁴–10⁶ m³/s (Alho, 2005) and “megafloods” with a peak discharge of more than 10⁶ m³/s (Baker, 2009a). Recent floods on Iceland can be classified as minor jökulhlaups with a peak discharge of up to 10⁷ m³/s (Russell et al., 2006; Björnsson, 2009), nevertheless having a large destructive potential and require organized hazard management.

Previous research was conducted on the Late Pleistocene Missoula flood (Bretz, 1923; Baker, 1978, 2007, 2009, Benito, 1997), the Bonneville flood (Keihew and Lord, 1986) and the Altai flood (Carling, 1996; Rudoy, 2002; Herget, 2005; Carling et al., 2009, 2010). These floods are categorized as megafloods with peak discharges of up to 18 × 10⁶ m³/s (Rudoy, 2002; Baker, 2009a). Regions characterized by catastrophic glacial outburst floods of large sub- or proglacial lakes show typical erosional and depositional bedforms.

1.1.1. Erosional bedforms

Deep gorges or valleys commonly indicate the major drainage routes of megafloods (Keihew and Lord, 1986; Carrivick et al., 2004; Gupta et al., 2007; Keihew et al., 2009; Denlinger and O’Connell, 2010) and are often combined with dry waterfalls, deep plunge pools and potholes (Rathburn, 1993; Rudoy, 2002; Carrivick et al., 2004). The most spectacular landscape on Earth related to the catastrophic outburst of glacial Lake Missoula is the Channeled Scabland in Washington State, where more than 2500 km² of water was discharged (Baker, 2002b). The erosion of deep channels and plunge pools is assigned to high-energy flood erosion, such as plucking, abrasion and cavitation (Whipple et al., 2000; Baker, 2009b). Plucking mainly occurs in jointed rocks (Whipple et al., 2000), whereas abrasion is the main mechanism in massive rocks. Both erosional processes may be supported by cavitation, which is often associated with the formation of horse-shoe vortices that incise into the bedrock (Whipple et al., 2000).

Streamlined hills are residual landforms that commonly are preserved in marginal areas of high-magnitude flows (Keihew and Lord, 1986). The best studied examples of streamlined hills are located in the Channeled Scablands (Baker, 1978). These hills are up to 12 km long, 8 km wide, 40 m high and consist of loess, underlain by basalt (Baker, 1978). Typical associated features are flow obstacles of resistant sediments, upstream crescent-like scour marks and oblique channels cutting through small divides at the crest of the streamlined forms (Baker, 1978). In the Northern Great Plains streamlined hills consist of till, flanked by boulder lags (Keihew and Lord, 1986). Three distinct shapes of streamlined hills can be distinguished, thought to have formed under various flow depth conditions (Patton and Baker, 1978). These hills have aspect ratios between 1:1.3 and 1:5.8, on average 1:3.2. Hills that formed under submerged flow conditions display the smallest and best streamlined forms. They show indefinite scars at their flanks and a vague, smooth topography (Patton and Baker, 1978; Keihew and Lord, 1986). These observations are supported by the results of flume tank experiments (Fig. 3A; Komar, 1983), conducted under varying flow discharges, water depths and a constant Froude number Fr < 1. A water level, which overtops a bar or hill, leads to the most significant streamlined features, caused by the development of a hydraulic jump on the lee side. This hydraulic jump generates high erosion rates on the back of the hill resulting in an erosional shaping (Fig. 3B). In the final stage the hill shape resembles a lemniscate loop with aspect ratios between 3 and 4 (Komar, 1983), characterized by a steep blunter end and a more pointed end downstream (Baker, 1978; Komar, 1983; Sjogren and Rains, 1995). Hills that form under partly submerged flow conditions have shallow channels at their tops, partly filled with reworked loess or gravel, and display incomplete streamlined. Gravel bars at their flanks are common. Emerged hills are associated with high deposition rates on the islands’ lees with less erosion and streamlined (Fig. 3A). Streamlined hills may also show tails at their downstream end, which decrease in height in the downstream direction and are composed of upstream eroded material.

1.1.2. Depositional bedforms

Large bars may be developed in outburst related regions with examples reported e.g. from features along the Big Lost River in central Idaho (Rathburn, 1993), the Channeled Scablands (Smith, 2006; Waitt et al., 2009) and in Siberia (Altai Mountains; Carling et al., 2002, 2009; Komatsu et al., 2009). The boulder bars in Siberia are up to 120 m high, 5 km long and have steep margins (Carling et al., 2002). The bar deposits record several individual floods, depositing cross-bedded sediment successions with imbricated boulder gravel (Carling et al., 2009). Boulder bars are commonly deposited at sites of flow deceleration, e.g. downstream of cataracts (Rathburn, 1993), along outburst channels at widening sections (Waitt et al., 2009) or in tributary valleys (Carling et al., 2002, 2009). Bar gravel range from pebble to boulder size up to 3 m in diameter (Smith, 2006) and the sediment may consist of poorly sorted massive, matrix-supported gravel, deposited from hyperconcentrated flows (Lord and Keihew, 1987) or well-washed gravel and sand, deposited in zones of abrupt flow deceleration (Rudoy, 2002). Long boulder axis are often orientated parallel to the channel axis and imbricated clasts show an a(p) a(l) fabric (Rathburn, 1993; Carling et al., 2002; Smith, 2005). Large sand and gravel dune fields occur adjacent or on top of large boulder bars in outburst channels, on top of fluvial terraces or in sheltered regions (Carling, 1996; Carling et al., 2002; Rudoy, 2002; Waitt et al., 2009).
Fig. 1. Location and topography of the study area. A) Topographic map of northern Europe, showing the maximum extent of Pleistocene ice sheets (modified after Ehlers et al., 2004). B) Hill-shaded relief model of the study area based on SRTM data.
These subaqueous dune forms display a system of elongated poorly sinuous ridges in plan view, which are orientated sub-perpendicularly to the valley trend (Rudoy, 2002). Dunes may form during high-stage flows (Carling, 1996) or waning flow stages (Baker, 2009b).

Slackwater deposits often occur in side valleys that have been backflooded during outburst floods (Baker, 1978; Rudoy, 2002; Benito and Thorndycraft, 2005; Waitt et al., 2009; Denlinger and O’Connell, 2010). These deposits show a lateral fining with distance from the outburst channel and are used as palaeostage water surface indicators (Benito and Thorndycraft, 2005).

2. Geological setting and previous research

The Münsterland Embayment is located south of the North German Lowlands (Fig. 1). It is bounded to the north by the Teutoburger Wald Mountains that form an up to 400 m high NW-SE trending ridge. Toward the south it is bounded by the up to 850 m high Rhenish Massif and toward the west the Embayment passes into the Lower Rhine area (Drozdzewski, 1995).

The NE–SW trending folded Paleozoic basement is unconformably overlain by Mesozoic sedimentary rocks. Structurally the Münsterland Embayment is a syncline that formed during the Late Cretaceous (Drozdzewski, 1995; Grobe and Machel, 2002). It has an asymmetric geometry with steeply dipping to overturned beds of Cretaceous and Jurassic rocks at the northeastern and northern margin (Teutoburger Wald Mountains), gently sloping beds of Cretaceous rocks at its southern margin (Haarstrang) and Triassic rocks at its eastern margin (Fig. 1B).

The Mesozoic basement rocks are overlain by Pleistocene and Holocene deposits (Lotze, 1951; Seraphim, 1979; Skupin et al., 1993; Skupin, 2002). Up to 25 m thick pre-glacial fluvial deposits of probably late Holsteinian to early Saalian age lie unconformably on the basement rocks (“Unterer Schneckensand”, cf. Thiermann, 1970 and “Oberer Schneckensand” cf. Hesemann, 1950), which are preserved in NW–SE trending channel systems. Above these fluvial deposits up to 30 m thick late Saalian meltwater deposits and an up to 15 m thick diamicton, interpreted as a basal till occurs (Skupin et al., 1993; Skupin and Staude, 1995). In the central Münsterland Embayment the diamicton is partly covered by up to 12 m thick late Saalian meltwater deposits (Skupin, 2002).

During the maximum extent of the late Saalian Drenthe glaciation the Münsterland Embayment was probably completely covered by an ice lobe with a thickness of approximately 130–200 m (Figs. 1 and 2; Thome, 1983; Skupin et al., 2003). Glacioclastic deformation structures and clast composition of glaciogenic deposits indicate that the ice-lobe advanced from a northwesterly direction into the Münsterland Embayment (Seraphim, 1979; Thome, 1980; Klostermann, 1992; Skupin et al., 1993, 2003; Herget, 1997; Lenz, 2003; Dölling, 2005). The position of the retreating ice margin is probably indicated by the occurrence of erratic boulder accumulations, which mainly occur north of the Teutoburger Wald Mountains (Seraphim, 1962, 1972). In the Münsterland Embayment ice-marginal deposits are sparse and it is difficult to define the position of the retreating ice margin (Skupin et al., 1993; Herget, 1997). The Saalian glaciogenic deposits are partly covered by younger Weichselian fluvial and Holocene aeolian deposits (Skupin, 2002; Lenz, 2003).

2.1. Glacial lakes

2.1.1. Glacial Lake Münsterland

Ice-damming of the eastern Münsterland Embayment led to the formation of a glacial lake (Thome, 1998; Herget, 1998, 2002; Skupin, 2002), in the following referred to as “glacial Lake Münsterland”. This glacial lake probably had a maximal lake level of ~350 m a.s.l., indicated by the occurrence of scattered erratic clasts with a Scandinavian/Baltic provenance (Herget, 2002). During highstand the lake had a water volume of ~45 km³ with an average depth of 60 m. The maximum depth at the western ice margin was probably up to 170 m. The early drainage history of this lake is not known. During initial ice-lobe retreat the lake probably drained southwestwards...
into the Mühle and Ruhr valley through outlet channels, located at the southwestern lake margin (Figs. 1B and 2; cf. Thome, 1980, 1983; Herget, 1998). The remaining water volume was about 10 km³.

2.1.2. Glacial Lake Weser

Ice-damming of the Upper Weser Valley led to the formation of a large and deep glacial lake (Fig. 2), referred to as glacial Lake Rinteln, glacial Lake Weserbergland or glacial Lake Weser, respectively (Spethmann, 1908; Thome, 1983; Klostermann, 1992; Winsemann et al., 2004, 2009). The spillway system of glacial Lake Weser was a series of valleys in the Teutoburger Wald Mountains, over an altitude range of 40–205 m a.s.l. through which the proglacial lake drained southwestward into the Münsterland Embayment (Fig. 4; Thome, 1983; Klostermann, 1992; Winsemann et al., 2009, 2011). These overspill channels increase in altitude toward the east (Fig. 4B) and were successively closed during ice advance (Skupin et al., 1993) leading to a long-term transgression and a lake-level highstand of ~200 m a.s.l. The location of these valleys is probably predefined by Variscian faults intersecting with the Teutoburger Wald Mountains (Baldschuhn et al., 2001).

The principle lithologic evidence for a large and deep glacial lake in the Upper Weser Valley is the occurrence of subaqueous ice-marginal deposits, fine-grained lake bottom sediments, and ice-rafterd debris far beyond the former ice margin (Winsemann et al., 2009). The lake-level curve of glacial Lake Weser has been mainly defined by foreset–topset transitions of deltas (Winsemann et al., 2011).

Following the long-term transgression a series of rapid lake-level drops occurred. The first lake-level drop of up to 65 m is attributed to the opening of the 135 m a.s.l. overspills in the Teutoburger Wald Mountains releasing ~90 km³ of water into the Münsterland Embayment (Winsemann et al., 2011, in press). The second lake-level drop in a range of ~35 m is related to the opening of a 95 m a.s.l. overspill. Approximately 20 km³ of water drained into the Münsterland Embayment. A subsequent ice lobe advance (Hondsreg ice stream, cf. Van den Berg and Beets, 1987; Skupin et al., 1993) led to the renewed closure of the 95 m a.s.l. spillway and a related lake-level rise of the Weser Lake. Rapid destabilization of this ice-lobe caused the final drainage of the Weser Lake (Winsemann et al., 2011, in press).

2.2. Streamlined hills

On the southern slope of the Teutoburger Wald Mountains streamlined landforms occur, which have been first recognized by Elbert (1905) and interpreted as fan-shaped arrays of hillside kames. These streamlined hills occur in front of the Bielefeld, Borg-holzhausen and Ihburg pass (Fig. 4) and were later mapped in more detail by Seraphim (1973), who assumed a subglacial drumlin formation by a north-eastward approaching ice-lobe. The drumlin origin was questioned by later workers, who assumed a Late Pleistocene formation of streamlined hills by fluvial erosion of a once complete till plain (Skupin, 2002; Lenz, 2003; Dölling, 2005). However, new data from glacial Lake Weser as discussed further in the text, indicate, that these streamlined landforms may have formed as a result of a catastrophic lake outburst flood representing eroded remnants of a once continuous cover of diamict and meltwater deposits (Meinsen et al., 2009).

2.3. Fluvial systems of the Münsterland Embayment

There are two main river systems in the Münsterland Embayment. The River Ems has a length of 370 km with an average discharge of 80 m³/s (NLWKN, 2005a) and drains into the North Sea. All smaller rivers, for example the Rivers Bever, Hessel and Lutter, in the northern Münsterland Embayment merge into the River Ems (Fig. 4B). The River Lippe in the south of the Münsterland Embayment is 255 km long, has an average discharge of 45 m³/s (NLWKN, 2005b) and flows into the River Rhine (Fig. 1B).

The fluvial systems of the River Ems and Lippe have been studied since the 1950s (Hesemann, 1950; Lotze, 1954; Thiermann, 1974; Speetzen, 1990; Herget, 1997; Skupin, 2002). During the early Middle Pleistocene the River Alme, Lippe and Ems probably merged into a broad anabranching river system, draining to the NW. The fluvial sediments consist of gravel and sand derived from the Rhenish Massif and Teutoburger Wald Mountains (Skupin, 2002).

During the late Saalian Drenthe glaciation the Münsterland Embayment completely became ice-covered. After deglaciation the two main river valleys of the River Ems and River Lippe developed. The Ems has an up to 12 km broad and 10–30 m deep valley (Figs. 1

Fig. 3. A) Shapes of streamlined islands formed in flume tank experiments (modified after Komar, 1983). Flow direction is from left to right. B) Schematic sketch of a flow over a submerged island, showing the development of a hydraulic jump on the island lee side (modified after Komar, 1983). C) Characteristic shapes of chevrons (modified after Scheffers et al., 2008). D) Typical shapes of streamlined hills in the study area.
Fig. 4. A) Close-up view of the study area. The hill-shaded relief model shows the Teutoburger Wald Mountains and the northeastern part of the Münsterland Embayment. B) Interpreted relief model showing the recent river system, lake overspill channels and streamlined hills. The DEM is based on data from the Bezirksregierung Köln (Germany).
The basal valley-fill consists of up to 7 m thick coarse-grained sand and pebble-sized gravel, referred to as “Knochenkies” (“bone gravel” cf. Menzel, 1912). These deposits contain clasts with a Scandinavian/Baltic provenance as well as wood and plant remains. The basal deposits probably record an intense reworking of Saalian glacigenic sediments (Skupin, 2002; Lenz, 2003). The age is still controversially discussed ranging from late Saalian (Herget, 1997), Warthian (Schmitz, 1990) to early Weichselian (Skupin, 1983). The upper Ems valley-fill is interpreted as Weichselian in age and consists of up to 10 m thick sand and silt alternations (Lenz, 1997), showing a fining-upward trend. It is covered by up to 8 m thick fine- and medium-grained sand with a silt matrix (Skupin, 1987; Lenz, 1997). The recent river valley is 1–2 km broad and 2–5 m deep, incised into the older valley-fill and filled with fluvial Holocene sand. The modern river channel is up to 20 m wide and flanked by Holocene dunes (Skupin, 2002).

The River Lippe has an up to 7 km broad and 15 m deep valley (Fig. 1B). The today’s river valley is 1–2 km wide, 5 m deep and filled with Holocene sandy silt. The modern river channel is up to 25 m wide. As in the Ems valley, the basal fill consists of up to 5 m thick “Knochenkies” (“bone gravel”) with sand remains, overlain by 3 m thick silt and silt alternation and up to 7 m thick fine- and medium-grained sand (Skupin, 2004). The river history of the Lippe has been studied by Herget (1997). He showed that the recent Lippe course probably already developed in Saalian times.

3. Database and methods

Streamlined hills in front of the mountain passes were mapped and analyzed in a geographic information system (ArcGIS) from a high-resolution digital elevation model DEM (10 m grid, vertical resolution: ±0.5). The DEM was combined with information from geological maps (1: 25 000, 1: 100 000), outcrops and 2450 borehole logs (Fig. 5) to document the regional pattern and character of deposits in front of lake overspill channels (Iburg, Dissen, Borgholzhausen, Hesseln and Bielefeld) in the Teutoburger Wald Mountains.

Based on borehole data 9 depositional units were defined on the basis of grain size and clast composition (Table 1). These depositional units were defined as markers for usage in the 3D model.

In a next step borehole data and DEM were integrated into the 3D modeling program GOCAD to reconstruct the spatial distribution of sedimentary facies in front of lake overspill channels. The model in front of the Dissen and Borgholzhausen overspills are based on 502 borehole logs. The model in front of the Bielefeld overspill is based on 1890 boreholes. Each model area has a size of ~160 km² (Figs. 4 and 5). The 3D modeling program GOCAD uses geological data to reconstruct stratigraphic horizons based on the DSI logarithm created by Jean-Laurent Mallet (Mallet, 2002). Each stratigraphic surface was created separately from a pointset based on the identified, significant marker. The DEM was integrated as reference surface. Subsequently the modeled surfaces were used to define the thickness of each depositional unit, to identify the Mesozoic basement relief and to create cross-sections showing the spatial distribution of sediments.

Geographic information systems (GIS) were used for the paleogeographic reconstruction of glacial Lake Münsterland, superimposing water planes onto a land surface DEM. To calculate the reservoir geometry of the lake palaeotopographic surfaces were generated. The maximum lake surface elevation is taken as 350 m a.s.l., based on the occurrence of ice-rafted debris and the topographic height of the lake outlet channel (Herget, 1998, 2002).
Table 1  Depositional units.

<table>
<thead>
<tr>
<th>Age</th>
<th>Depositional unit</th>
<th>Lithology</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Late Saalian</td>
<td></td>
<td>Silt and fine-grained sand. Reworked peat, wood and plant remains are common.</td>
<td>Flood-related deposits</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Fine-grained sand to pebbly sand with a silt matrix. Clasts consist of “Plänerkalkstein”, Cretaceous sandstone and Scandinavian/Baltic material.</td>
<td>Flood-related deposits</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Pebble-sized gravel. Clasts mainly consist of Cretaceous limestones (“Plänerkalkstein”).</td>
<td>Flood-related deposits</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Matrix-supported gravel with a silt matrix. Clasts consist of Cretaceous limestone and Scandinavian/Baltic material.</td>
<td>Flood-related hyperconcentrated flow deposits? (Diamicton III)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Large deformed sand clasts are common.</td>
<td>Flood-related deposits</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Matrix-supported gravel. The matrix is sandy silt. Clasts consist of Cretaceous limestone and Scandinavian/Baltic material. Large deformed sand clasts are common.</td>
<td>Till (Diamicton I)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Fine-grained sand to pebbly sand with erratic clasts.</td>
<td>Meltwater deposits</td>
</tr>
<tr>
<td>Early Saalian to Holsteinian</td>
<td></td>
<td>Silt and pebbly sand. Clasts consist of Cretaceous “Plänerkalkstein” and sandstone. Rare occurrence of molluscs and wood remains.</td>
<td>Pre-glacial lacustrine to fluvial deposits (“Oberer Schneckensand”)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Pebble-sized gravel with a sand matrix. Common occurrence of molluscs and wood remains.</td>
<td>Pre-glacial fluvial deposits (“Unterer Schneckensand”)</td>
</tr>
<tr>
<td>Cretaceous</td>
<td></td>
<td>Fine-grained sand to pebbly sand. Wood remains.</td>
<td>Mesozoic basement</td>
</tr>
</tbody>
</table>

The discharge of the outburst flood was calculated using two approaches. The first method uses the continuity equation, calculating the maximum discharge for an overspill channel, based on cross-sections and characteristic flood velocities in the range of 5–15 m³/s (Baker, 2002b). The second approach uses the method proposed by Walder and Costa (1996) and Carling et al. (2010), calculating the minimum discharge based on lake volume and dam height for an overspill channel.

4. Geometry and internal facies architecture of streamlined hills and associated channel systems

4.1. Geometry of streamlined hills and channel systems

Streamlined hills occur along a 50 km wide area at the southern slope of the Teutoburger Wald Mountains. They are arranged into three main fields in front of the Iburg, Borgholzhausen and Bielefeld overspill channels, covering an area of 60–130 km², each (Fig. 4). The mapped streamlined hills are 2–11 m high, 600–4300 m long and 170–1200 m wide, mostly trending NNE–SSW to NE–SW. Their geometry is quadrilateral to elongate, partly V-shaped forms (chevrons) with an aspect ratio between ~1:1 and 1:8, on average 1:3.3. Hills commonly have a steep blunter end in the NE and a more pointed end in the SW. Commonly, the streamlined hills are separated by channels, which are 50–700 m wide, 2–30 m deep, and display an anabranching pattern.

4.1.1. Streamlined hills and channel systems in front of the Iburg pass

A small field of streamlined hills, covering an area of ~60 km², occurs in front of the Iburg pass (Fig. 6). The pass has an altitude of 152 m a.s.l. and incises into Lower and Upper Cretaceous basement rocks. The hill field has a fan-shaped geometry in plan view, radiating from the Iburg pass. It consists of 19 hills that trend NE–SW and occur over an altitude range of ~60–80 m a.s.l. (Figs. 4 and 6). The hills are 700–3000 m long, 300–950 m wide and 2–5 m high with an aspect ratio of 1.6–5.4 (on average 1:2.8) and an elongated or rounded form. The highest elevation of each hill occurs in the northeastern region (Fig. 7A). The hills directly in front of the pass are separated by 300–2000 m broad valleys occupied by underfed streams draining to the SW. The exact depth of the valleys is not known, due to a lack of borehole data.

4.1.1. Interpretation. The NE–SW trending fan-shaped array of hills give the general impression of features radiating from a channel mouth in the northeast (Iburg pass) and extending southwest. South-westward directed palaeoflows are also indicated by the geometry of hills with the highest elevations in the northeast (Fig. 3A, B and 7A; Komar, 1983). Most hills are small and well streamlined, commonly attributed to high discharges and a formation under fully submerged flow conditions (Patton and Baker, 1978; Komar, 1983). The restricted hill field indicates rapid flow deceleration when the flow entered the Münsterland Lake.

4.1.2. Streamlined hills and channel systems in front of the Hesseln, Borgholzhausen and Dissen pass

The second field of streamlined hills occurs over an altitude range of ~70–105 m a.s.l. in front of the Dissen, Borgholzhausen and Hesseln pass (Figs. 7B and 8). The Dissen and Hesseln pass have altitudes of 178 and 175 m a.s.l., while the smooth flanked, broad Borgholzhausen pass has an altitude of 135 m a.s.l.
48 streamlined hills were mapped, showing a very narrow, nearly parallel alignment and a northeast to southwestward trend (Fig. 8). The whole field covers an area of approximately 125 km². The hills have lengths between 600 m and 4330 m, are 170–1170 m wide and 4–11 m high with aspect ratios range from 1:1.8 to 1:6.0, on average 1:3.6 (Fig. 8). Most hills have elongated shapes with rounded, blunted ends in the NE and a more pointed ends in the SW (Fig. 7B). The highest elevation of each hill is in the northeast (Fig. 7B).

Fig. 6. A) Close-up view of Fig. 4 showing streamlined hills in front of the Iburg overspill channel. B) Mapped streamlined hills with aspect ratios. The DEM is based on data from the Bezirksregierung Köln (Germany).
The hills directly in front of the Borgholzhausen overspill have very large, elongated shapes and show a narrow spacing (Fig. 8B). Some of these hills have tails, representing nearly one sixth to one eighth of the total hill length. A few small hills in front of the Dissen and Hesseln pass have a V-shaped geometry (Fig. 8).

The channels separating the hills are very shallow, commonly 2–5 m deep, and have width of 50–400 m. They show an anabranching channel pattern. The River Hessel has an up to 1500 m wide, 9–12 m deep valley (Fig. 4B; Dölling, 2005).

4.1.2.1. Interpretation. The hills in front of the Borgholzhausen pass show a narrow alignment, arranged into a fan-shape, which slightly widens to the SW. The highest elevation of hills is in the NE indicate flow directions from NE to SW. Some hills have tails that resemble features produced by currents around obstacles. In this case the obstacle is a plug of more resistant diamicton and the tail is composed of sand. The tails are an indicator for the beginning of streamlining in an early stage of flooding due to high discharges and submerged flow conditions (Komar, 1983; Kehew and Lord, 1986; Kehew et al., 2009). The hills in the west and east of the Borgholzhausen pass show a well streamlined form, possibly representing an area of continuous long-term flooding (Kehew et al., 2009). The hills directly in front of the Borg-holzhausen overspill are very large and have a rough relief. The formation of these hills is attributed to rapid waning discharges of the flood, resulting in incomplete streamlining and channel erosion. Evidence is given by the shallow depth of channels separating these hills. The development of tails suggests a short forming process (Komar, 1983).

4.1.3. Streamlined hills and channel systems in front of the Bielefeld and Oerlinghausen pass

The steep flanked Bielefeld and Oerlinghausen passes have altitudes of 135 m and 205 m a.s.l., respectively, with the mountain range in this area having altitudes of approximately 200–260 m a.s.l. (Figs. 4 and 9). They are incised into Jurassic and Cretaceous basement rocks.

29 hills were mapped, covering an area of approximately 130 km² over altitude ranges between ~80–115 m a.s.l. (Figs. 4 and 9). These hills trend NE–SW, are 650–3500 m long and 160–1150 m wide, with elevations of 4–9 m. The aspect ratios are between 1.9 and 5.4, on average 1:3.7. The highest point of most

---

**Fig. 7.** Shape, geometry and morphology of selected streamlined hills in front of the A) Iburg, B) Borgholzhausen and C) Bielefeld overspill channels.
hills is in the NE (Fig. 7C). Hills are commonly widely spaced. Near the overspill channels in the NE the hills have distances of 100–1000 m, in the SW the distance between hills increases across stream to more than 2000 m leading to a patchy hill distribution. All hills have a steeper blunter end at the northeast and the majority of the hills have an elongated, oval shape (Fig. 7C). A few hills are characterized by smaller tails at their downstream end, representing nearly one third of the hill length and the lowest points of the hills. Other hills in the east have a V-shaped form. These bedforms have apices facing upslope, are 1.6–2.5 km long, 3–10 m high, 0.8–1.2 km from limb to limb, with limb separation angles of 20–35°. Hills are separated by small, up to 5 m deep channels, increasing downslope to 8–10 m. The width increases from 250 to 500 m in the NE in front of the passes to more than 1000 m in the SW. In the SW the channels merge into the River Lutter (Fig. 4). Its underfit valley has a width of up to 2000 m and a depth of up to 22 m, deeply incised into the Mesozoic basement (Lenz, 2003).

A large, up to 15 km long, 0.5–5 km wide and 25 m high, horseshoe-shaped hill is located 17 km to the south of the
Oerlinghausen pass, trending NE-SW and referred to as "Delbrücker Rücken" (Figs. 4B; Lotze, 1951; Skupin, 1983, 1987, 2002).

4.1.3.1. Interpretation. The hills are very large, have a wide spacing, diverse shapes and show characteristic aspect ratios of fluid-formed streamlined hills (Komar, 1984). Smaller hills in the NW are well streamlined and attributed to fully submerged flow conditions (Patton and Baker, 1978; Komar, 1983). Larger hills probably formed under partly submerged or emerged flow conditions resembling hills in the Channeled Scablands being emerged during flood (Patton and Baker, 1978).

Fig. 9. A) Close-up view of Fig. 4 showing streamlined hills in front of the Bielefeld overspill channel; location of modeled 2D cross-sections (Fig. 12: AA’ – DD’ and Fig. 19B: ZZ’). B) Mapped streamlined hills with aspect ratios. V-shaped hills are indicated by a light yellow overlay. For key see Fig. 6. The DEM is based on data from the Bezirksregierung Köln (Germany). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
V-shaped chevron-like streamlined bedforms occur in channel–lobe transition zones of deep-sea fans. These depositional forms are attributed to supercritical flow conditions, possibly associated with thinning flows (Morris et al., 1998). V-shaped chevron-like bedforms have also been described from coastal areas affected by tsunamis (Hearty et al., 1998; Kindler and Strasser, 2000; Hearty and Neumann, 2001; Scheffers et al., 2008). These forms are up to 10 km long and attributed to strong erosion in the intertidal zone by the approaching tsunami and a subsequent rapid redeposition by high-energy backwash flows (Hearty and Neumann, 2001). The V-shaped hills in the study area are much smaller and resemble streamlined islands conducted in flume experiments under short and submerged flow conditions, representing mixed erosional depositional bedforms (Komar, 1983).

The geometry of streamlined hills and V-shaped forms is in accordance with palaeoflows from the NE, flowing through the Bielefeld and Oerlinghausen overspill channels and partly flowing over lower reaches of the Teutoburger Wald Mountains with altitudes up to 220 m a.s.l. (Fig. 9).

The origin of the large “Delbrücker Rücken” is not clear. It has previously been interpreted as an end moraine (Bärtling, 1921) or a landform, caused by late Pleistocene fluvial erosion (Skupin, 2002). The typical NE-SW trending horse-shoe shape may point to a formation during the outburst flood, resembling V-shaped forms in flume tank models by Komar (1983). This is also supported by the facies architecture, which will be discussed further in the text.

4.2. Facies architecture of streamlined hills and channel-fills

The internal facies architecture of streamlined hills and channel-fills has been reconstructed from 1500 boreholes. Detailed 2D cross-sections are shown for the hill fields in front of the Borgholzhausen and Bielefeld pass (Figs. 11 and 12). There are limited boreholes available for the Iburg field and no detailed cross-sections could be drawn.

The thickness distribution and palaeotopographic surfaces of 9 depositional units have been modeled (Figs. 13 and 14, Table 1). Cretaceous basement rocks are over lain by Holsteinic to early Saalian lacustrine and fluvial deposits (Table 1). The Holsteinic sediments (“Unterer Schneckensand”) consist of pebble-sized gravel with a sand matrix; the early Saalian deposits (“Oberer Schneckensand”) consist of silt and pebbly sand. Both depositional units contain molluscs and wood remains (Table 1). They are unconformably overlain by Saalian meltwater deposits and diamicton (Lenz, 2003; Dölling, 2005). The diamicton I has a patchy occurrence and locally reaches a very high thickness of up to 15 m. Parts of the diamicton in front of the mountain passes, however, may only represent flood-related mass-flow deposits (diamicton II), containing mainly local clasts, derived from the adjacent bedrock and large deformed sand blocks (Seraphim, 1973). The diamicton is erosively overlain by fine-grained sand to pebbly sand, interpreted as flood-related deposits. Within the pass areas flood-related pebble-sized gravel and silt occurs.

4.2.1. Facies architecture of streamlined hills and channel systems in front of the Borgholzhausen pass

The streamlined hills in front of the Borgholzhausen pass consist of meltwater deposits, overlain by diamicton and/or fine-grained sand to pebbly sand (Fig. 11 AA/BB). The area-wide occurring pre-flood meltwater deposits have a thickness of up to 25 m and consist of fine- to medium-grained sand and pebbly sand (Figs. 11 and 13A). The maximum thickness occurs slope-parallel, at a distance of approximately 2–3 km in front of the Teutoburger Wald Mountains (Fig. 13A).

The diamicton is only patchily preserved, mainly on top of the hills. In front of the Borgholzhausen and Hesseln pass the diamicton is absent. The thickness of the diamicton is commonly 2–8 m, but may locally reach up to 15 m (Fig. 13B). A few hills are entirely built up by diamicton and may have sand tails at their downstream end. In the Borgholzhausen overspill channel the uppermost section consists of a mixture of silt, sand and gravel with a thickness of up to 15 m thick (Fig. 13C). Some hills are covered by up to 10 m thick silty fine- to medium-grained sand (Fig. 13D). These deposits only occur in a patchy distribution in front of the Teutoburger Wald Mountains.

4.2.1.1. Interpretation. Hills commonly consist of pre-flood meltwater sand and diamicton, partly overlain by fine-grained flood deposits. The thickness distribution of meltwater deposits points to
flows from the NE. The till (diamicton I) is missing in front of all main overspill channels and is only preserved on top of the streamlined hills in the marginal areas. This indicates the erosion and formation of a broad, up to 4 km wide and at least 10 m deep channel in the central zone of the hill field (Fig. 11). The distal part of this central channel is widely unfilled. Within the uppermost proximal part of the channel and the Borgholzhausen pass up to 15 m thick diamicton occurs (Fig. 13C). These sediments are interpreted as mass flow deposits (diamicton II), which formed by sediment mobilization and redeposition in the final stage of flood. Hills directly in front of the Borgholzhausen pass are separated by shallow channels and consist entirely of these mass flow deposits. It is suggested that the Borgholzhausen pass was blocked by large ice masses early during the outburst flood. Therefore, the streamlined hills in front of the pass show an incomplete streamlining and the deep channel partly remained unfilled.
4.2.2. Facies architecture of streamlined hills and channel systems in front of the Bielefeld and Oerlinghausen pass

The hills in front of the Bielefeld pass consist of pre-flood meltwater deposits, overlain by till (diamicton I; Fig. 12 BB’-DD’). Palaeoflow data, obtained from cross-bedding of the basal meltwater sand ~7 km SW of the Bielefeld pass, indicates a flow direction from NE to SW (Seraphim, 1973). This meltwater sand covers almost the whole study area, having a thickness of up to 15 m, reaching a maximum thickness of 25 m (Figs. 12 BB’-DD’ and 14A). In front of the Bielefeld overspill the meltwater deposits are absent in a trumpet shaped area, widening to the SW (Figs. 12 and 14A). In the east a NE-SW trending zone with a lower sand thickness (5 m) occurs (Fig. 14A). The diamicton I is patchily preserved with a thickness commonly ranging from 1 to 10 m. It is absent in front of the overspill channel (Figs. 12 and 14B). In the southernmost area (8 km SW of the Bielefeld overspill channel) a V-shaped bedform occurs. This bedform has an apex facing upslope to the east (Fig. 14C). It is up to 4 km long, 3–12 m deep, and 2.0–2.7 km from limb to limb, with limb separation angles between 40 and 45° (Figs. 14C and 15). The internal facies architecture shows a typical flute-like scour, shallowing in flow direction (Fig. 15B). The pre-flood meltwater deposits and diamicton (I and II) are erosively overlain by 2–30 m thick fine-grained sand to pebbly sand, covering most of the hills, and commonly displaying convolute bedding and dewatering structures (Figs. 14D and 16). The trumpet shaped area in front of the Bielefeld pass is filled with up to 30 m thick sand, decreasing in thickness (5–15 m) toward the SW (Figs. 12 AA’ and 14D).

The facies architecture of the large horse-shoe shaped “Delbrücker Rücken” is shown in Fig. 10. The basal succession consists of pre-glacial Holsteinian lacustrine and fluviatile deposits (“Oberer Schneckensand”). These deposits are covered by meltwater
deposits and till (diamicton I). The meltwater deposits and diamicton occur only in the hill and form the ridge. They are missing in the surrounding area. The till is only patchily preserved and builds the highest elevations of the “Delbrücker Rücken” (Fig. 10). At the top Holocene aeolian fine-grained sand occurs (Skupin, 2002).

4.2.2.1. Interpretation. As in the Borgholzhausen field the streamlined hills consist of meltwater sand and till patches (diamicton I), partly erosively overlain by flood-related fine-grained sand to pebbly sand (Figs. 12 and 14).

In front of the Bielefeld pass a NW–SE trending, trumpet-shaped, up to 30 m deep channel is incised into the pre-flood Pleistocene deposits and the Cretaceous basement rocks (Figs. 12 and 14). In this central zone till and meltwater deposits are absent and have been completely stripped off (Fig. 14A/B). Large parts of the northern area are covered by up to 20 m thick flood-related fine-grained sand to pebbly sand (Figs. 12 and 14D). These deposits, previously interpreted as meltwater deposits (Lenz, 2003), erosively overlie the till and are interpreted as fine-grained flood sediments, deposited during the waning stage of flood. Frequent dewatering structures and convolute bedding indicate rapid deposition and high sedimentation rates (Fig. 16; Manville and White, 2003). The highest sediment thickness is recorded from the central channel in front of the Bielefeld pass (Fig. 14D), filled with up to 30 m thick gravel, sand, silt and mud (Figs. 12, 14 and 17).

In front of this deep trumpet-shaped channel an approximately E–W trending V-shaped chevron-like bedform occurs, which comprises diamicton II and flood-related sand. The cross-section (Fig. 15) clearly indicates an erosional formation of this chevron-like bedform, resembling a typical flute-like scour, similar to megaflutes known from the base of turbidite channels (Elliott, 2000) and the channel–lobe transition zone of deep-water fans (Wynn et al., 2002; Kane et al., 2009). Megaflutes are also known from highly turbulent, erosive, unidirectional flows, which are attributed to tsunamis (Bryant et al., 1996; Le Roux et al., 2004; Le Roux and Vargas, 2005), and filled with hyperconcentrated flow deposits or redeposited coastal sediments. However, these megaflutes are much smaller in size (meter-scale; cf. Elliott, 2000; Le Roux and Vargas, 2005). Chevron-like scours described from deep-water channel–lobe transition zones reach lengths of up to 2500 m and widths of 2000 m (Wynn et al., 2002), comparable with the chevron-like bedform in front of the Bielefeld channel. The erosive bedform indicates strong and turbulent flows eroding the pre-existing meltwater sand during initial outburst flood and subsequent fill of the scour by hyperconcentrated flows (Massari and D’Alessandro, 2000; Manville and White, 2003).

The hill formation probably took place during the final stage of flood. Evidence is given by flood-related fine-grained sand to pebbly sand, which partly covers the streamlined hills (Fig. 12 CC’). A couple of smaller hills to the west of the Bielefeld pass show
sandy tails in the lee of diamicton plugs, where the resistant till acted as an obstacle causing the deposition of tails in the lee of the obstacles during waning flows (Komar, 1983).

The modeled facies architecture of the large horse-shoe shaped “Delbrücker Rücken” (Fig. 10) indicates strong erosion of pre-glacial and glacigenic deposits, pointing to a formation by a large water mass during an outburst flood. Since the deposits mainly consist of sand-mud-alternations and there is no evidence for glaciotectonic deformation an end moraine origin seems unlikely as stated before by Bärtling (1921). A formation of the “Delbrücker Rücken” by late Pleistocene fluvial erosion is unlikely, because i) the low erosive power of the Lippe and Ems River in that area and ii) the absence of major fluvial valleys near the “Delbrücker Rücken”.

The shallow channels separating the streamlined hills are filled with fine- and medium-grained sand and silt (Lenz, 2003). Coarse-grained gravel lags, referred to as “Knochenkies”, are common at the base of larger valleys of the Rivers Lutter, Ems and Lippe (Skupin, 1983; Lenz, 1997, 2003) and are interpreted as flood-related lag deposits. However, the age of the “Knochenkies” is poorly constraint and no absolute age determinations have been carried out. The channels dividing the streamlined hills are nowadays occupied by small underfed streams (Fig. 4B). Underfed streams are known from other outburst related regions and are characterized by a channel width exceeding the recent rivers and their floodplains (Baker, 1978). The streams use the existing channels formed during catastrophic flood to have a minimal drag. The channels indicate the southwestward drainage pathway during flood, following the topographic gradient of the Münsterland Embayment.

5. Bedrock erosional features

Based on borehole data two 3D models were created to reconstruct the topographic surface of the Cretaceous basement rocks (Fig. 18). The modeled basement surface gently slopes from the Teutoburger Wald Mountains toward the south and southwest (Fig. 18).

In front of the Borgholzhausen, Hesseln and Bielefeld pass deep erosional features are cut into the basement rocks. These features occur in Upper Cretaceous marl- and limestones and Lower Cretaceous sand- and marlstones (Fig. 19). The erosional features have a round to oval shape and are up to 35 m deep.

The erosional feature in front of the Borgholzhausen pass is 780 m long, 400 m wide and up to 35 m deep (Fig. 18A). The one in front of the Hesseln pass has a length of 480 m, a width of 330 m and a depth of up to 30 m. Both erosional forms are filled with gravel, silt and sand (Fig. 11 EE'). The basal succession consists of up to 15 m pebble-sized gravel containing clasts with a Scandinavian/
Baltic provenance. These deposits are overlain by interbedded silt and fine-grained sand and matrix-supported gravel.

The third erosional feature, in front of the Bielefeld pass has a length of 350 m, a width of 330 m and a depth of 26 m (Fig. 18B). It is filled with ripple cross-laminated silt and fine-grained sand, overlain by ripple cross-laminated fine- to coarse-grained sand and pebbly sand (Figs. 12 AA' and 17). The deposits frequently contain reworked plant remains and peat clasts (Lenz, 2003). The upper part consists of fine-grained sand with slump structures. The whole channel infill is interpreted as flood-related deposits. 800 m further upstream in the pass area another 10 m deep erosional feature occurs filled with flood-related interbedded silt, fine-grained sand and pebble-sized gravel, containing reworked plant remains (Fig. 12 AA').

5.1. Interpretation

The large and deep bedrock erosional features in front of the main lake overspill channels suggest very high-energy flows and their shapes resemble plunge pools, produced by waterfalls (Comiti et al., 2005; Lamb et al., 2007; Lamb and Dietrich, 2009). The formation of plunge pools are commonly assigned to fluvial erosion processes in particular plucking of bedrock (Whipple et al., 2000; Baker, 2009b). The dimension of these plunge pools is similar to those formed by the Niagara Falls (Hayakawa and Matsukura, 2009). The second erosional feature in the Bielefeld pass resembles potholes known from outburst floods (Carrivick et al., 2004). Other examples of deep plunge pools are at the Channeled Scablands and the Altai region and are attributed to the erosive work of megafloods and the formation of waterfalls (Baker, 1978; Rathburn, 1993; Rudoy, 2002; Lamb and Dietrich, 2009). Gupta et al. (2007) analyzed the bathymetry of the English Channel and described crescent-like scours that taper upstream into headcuts. These features are interpreted as formed by headward recession of waterfalls (Lamb and Dietrich, 2009).

Plunge pools, formed during the Altai flood are up to 1200 m long, 400 m wide and 70 m deep (Rudoy, 2002). Dry falls, up to 120 m high and 1800 m wide, and plunge pools up to 90 m deep, related to the Missoula flood are even larger (Waitt et al., 2009). The location of these plunge pools are connected with overspill channels, which show high knickpoints between the overspill and the adjacent river channel (Rudoy, 2002; Waitt et al., 2009).
2D-hydrological modeling shows that plunge pools form in zones of high discharges and velocities. The high velocities occur upstream, especially in narrow passages like spillways, but abruptly decelerate downstream when the flow expands (Denlinger and O’Connell, 2010). The consequence is a water accumulation in front of the narrows, retarding the flood, and lower water levels in the adjacent area (Denlinger and O’Connell, 2010).

It is therefore likely that the erosional bedrock features in front of the Borgholzhausen, Hesseln and Bielefeld pass are related to a catastrophic outburst flood of glacial Lake Weser, during which high waterfalls formed in the pass area of the Teutoburger Wald Mountains, pouring into the Münsterland Embayment (Fig. 20). No data are available for the Iburg and Oerlinghausen pass.

The plunge pools in front of the Borgholzhausen and Hesseln pass have a round to oval form, indicative of missing or beginning backward erosion (Lamb and Dietrich, 2009). The fact that headward erosion is not well-developed suggests a rapid formation of the plunge pools in the study area with a high water discharge, which inhibited the complex formation of several steps (Comiti et al., 2005; Lamb et al., 2007). The more elongated plunge pool in front of the Bielefeld overspill together with a second erosional feature occurring upstream, interpreted as a pothole, may indicate beginning backward erosion during flood suggesting a longer active phase (Fig. 18B).

6. Discussion

6.1. Evidence for lake outburst related megafloods

The analysis of streamlined hills and channel systems in the Münsterland Embayment clearly indicates characteristic erosional and depositional bedforms associated with glacial lake outburst floods. Related features are potholes and up to 35 m deep plunge pools, incised into Cretaceous bedrocks, in front of major overspill channels. All these bedforms suggest high-magnitude floods as known from lake outbursts (Baker, 1978; Kehew and Lord, 1986; Rathburn, 1993; Carrivick et al., 2004; Smith, 2006; Gupta et al., 2007; Kehew et al., 2009; Carling et al., 2010; Denlinger and O’Connell, 2010).

A drumlin origin for the streamlined hills, as assumed by Seraphim (1973, 1979) and Skupin et al. (1993) is not likely. Drumlins are subglacial bedforms, often occurring in fields (Newman and Mickelson, 1994; Clark et al., 2009). They are commonly characterized by smooth, oval-shaped hills of glacial drift with steep, blunter ends in the up-ice direction. The down-ice direction is characterized by a gentle slope (Menzies, 1979). Characteristic length to width ratios are up to 7:1, on average 1:2 to 1:3.5 (Fig. 21). Drumlin core composition is variable, depending on the formation processes. It may consist of diamicton, mud, sand and gravel (Newman and Mickelson, 1994) or a glaciofluvial core covered by a thin layer of till (Zelcs and Dreimanis, 1997). The core composition often suggests a formation by accretion of basal till in the subglacial environment (Boyce and Eyles, 2000). Other workers assume a drumlin formation by subglacial meltwater erosion (Shaw, 2002).

Our reconstruction shows that the streamlined hills in the Münsterland Embayment are mixed erosional depositional in origin and formed by high-energy floods, previously interpreted as Weichselian low-energy fluvial forms (Skupin, 2002; Lenz, 2003; Dölling, 2005). Palaeofloods were from the NE and opposite to the main direction of ice-advance. The shapes and aspect ratios of hills are very similar to streamlined hills formed in flume tank experiments (Komar, 1983), formed by lake outburst floods, e.g. in the Great Plains (Kehew and Lord, 1986; Kehew et al., 2009) and the Channeled Scablands (Baker, 2009b) and depositional chevron-like
Fig. 18. A) Modeled Mesozoic basement surface in front of the Borgholzhausen and Hesseln overspill channel. B) Modeled Mesozoic basement surface in front of the Bielefeld overspill channel. The deep erosional bedrock features in front of the overspill channels are interpreted as plunge pools.
bedforms described from the channel–lobe transition zone of deep-sea fans (Morris et al., 1998) or coastal areas affected by tsunamis (Bryant et al., 1996; Le Roux and Vargas, 2005). It should be noted that the today’s hill geometry is probably partly reshaped by erosion processes since the Saalian glaciation. A large, up to 12 m deep and 4 km long V-shaped chevron-like erosional bedform, filled with diamicton resembles megaflutes described from deep-sea channels (Elliott, 2000; Wynn et al., 2002), indicating that this diamicton (II) represents a hyperconcentrated flood deposit and not a till.

Fig. 19. Cross-sections used to calculate the maximum peak discharge using the continuity equation. A) Borgholzhausen pass and B) Bielefeld pass. For location see Figs. 8 and 9.

Fig. 20. Schematic sketch illustrating the formation of the 26 m deep plunge pool in front of the Bielefeld overspill channel. The plunge pool was eroded into the Cretaceous basement rocks during the outburst of glacial Lake Weser, when an up to 50 m high waterfall temporary formed in the overspill area.
peak discharge for the Bielefeld overspill channel ranges from 2.5 \times 10^5 \text{ m}^3/\text{s} (v = 5 \text{ m/s}) to 7.5 \times 10^5 \text{ m}^3/\text{s} (v = 15 \text{ m/s}; \text{Fig. 19B}).

The second approach uses the method proposed by Walder and Costa (1996) and Carling et al. (2010), calculating the minimum discharge based on lake volume and dam height for an overspill channel:

\[ Q_p = 0.27C_1\left(\frac{f_R\rho_wV_i}{\rho_iL^2}\right)^{1/2}gh_1^{3/2} \]  

(2)

Where \(Q_p\) is the peak discharge in the dam (\text{m}^3/\text{s}), \(f_R\) is a roughness coefficient (typically 0.05 to 0.09), \(\rho_i\) and \(\rho_w\) are the densities of ice and water (taken here as 900 kg/m\(^3\) and 1000 kg/m\(^3\) respectively), \(V_i\) is the initial lake volume (120 km\(^3\)), \(h_i\) is the initial height of the ice dam (e.g. Bielefeld and Borgholzhausen = 65 m) and \(g\) is the acceleration due to gravity (9.8 m/s). The ‘effective’ heat of fusion is given as \(L = \rho_w(h_0 - \theta_i)\) where \(L\) is the heat of fusion (3.35 \times 10^6 \text{ J/kg}), \(\rho_w\) is the specific heat of water (4.2 \times 10^3 \text{ J/kg}) and \(\theta_i\) are the temperatures of the lake water and of the ice, respectively. The term \(C_1(p)\) typically varies between 0.33 and 0.5 for values of a lake shape parameter \(p\) between 1 and 3, respectively. The water temperature of the palaeolake is assumed to have been 2 °C, whilst the ice temperature is taken as 0 °C (Carling et al., 2010).

Calculated peak discharges, using the second approach, are lower, and in the range of 3.1 \times 10^5 to 4.1 \times 10^5 \text{ m}^3/\text{s} for the Borgholzhausen and Bielefeld overspills. These discharges agree with flood magnitudes reconstructed for Holocene and Pleistocene floods in Iceland (Alho, 2005) and Pleistocene megafloods, e.g. in the Great Plains (Keew et al., 2009) and are capable of rapidly eroding the Pleistocene sediments and the underlying poorly resistant fractured marl- and limestone of the Cretaceous basement (Fig. 19) within and in front of the main lake overspill channels. The calculated discharges using approach one are much higher and different for the overspill channels, but still correlate with discharges in the Great Plains (Keew et al., 2009).

The different peak discharges for the overspills, using approach one, are attributed to the size of the overspill channels. Bielefeld shows a narrow pass area with several steps interpreted to have been incised during the flood. Steps one and two probably indicate incision from an initial overspill height of 150 m a.s.l. to 135 m a.s.l. (Fig. 19B). The other overspills have no distinctive steps (Fig. 19A). The plunge pools in front of the Borgholzhausen and Hesselin pass have a round to oval form indicative of rapid erosive formation (Lamb and Dietrich, 2009). The plunge pool in front of the Bielefeld-overspill channel is the largest one correlating with the highest calculated discharges of up to 1.3 \times 10^6 \text{ m}^3/\text{s} (maximum model).

### 6.3. Overall geological model

During the late Saalian Drenthe glaciation ice-damming of the Upper Weser Valley led to the formation of glacial Lake Weser. At the maximum lake-level highstand of 200 m a.s.l. ~120 km\(^3\) water was stored in the lake basin (Winsemann et al., in press). The reconstructed lake-level curve of glacial Lake Weser shows an overall lake-level rise, followed by two high-amplitude lake-level falls caused by the successive opening of lake outlets in the Teutoburger Wald Mountains during ice-lobe retreat. During the first lake-level fall from ~200 m–135 m a.s.l. the water drained catastrophically through the Derlinghausen, Bielefeld, Borgholzhausen and Iburg pass into the Münsterland Embayment (Figs. 4 and 22A). All overspills present natural spillways in mountain ridges, opened after ice-lobe retreat/collapse (Winsemann et al., 2011). Probably ~90 km\(^3\) of water was released during this first lake outburst within a few days or weeks (Winsemann et al., 2011, in press).
During the initial outburst Lake Weser drained into the Münsterland Embayment, via the Knochenkies ("bone gravel") flood high-energy, turbulent flows. Glacial Lake Münsterland drained to the Delbrücker Rücken and Oerlinghausen pass V-shaped chevron-like hills are arranged slope parallel, approximately 4–9 km downslope of the major overspill channels (Fig. 9B). These mixed erosional depositional V-shaped hills have a diamict core and tails of fine-grained flood deposits and resemble streamlined hills conducted in flume experiments under submerged and short flow duration (Komar, 1983). The formation of these streamlined hills therefore probably indicates very short flood duration through the high (205 m a.s.l.) Dörenschlucht and Oerlinghausen passes.

An unsolved question is the timing of the opening of overspill channels. Probably the easternmost passes opened first (Dörenschlucht und Oerlinghausen pass), indicated by the formation of V-shaped hills. Subsequently a rapid opening of the passes further west occurred, when the ice dam failed. It is likely that the Bielefeld, Borgholzhausen and Iburg overspills opened simultaneously because streamlined hill fields occur in front of all these overspill channels and the Bielefeld and Borgholzhausen pass have the same altitude of 135 m a.s.l.

In front of the major overspill channels (Bielefeld and Borgholzhausen) the expanding flows became rapidly channelized, carving out the deep inner channels, and streamlined hills, separated by a shallow channel network, formed in the outer zones (Figs. 9, 13 and 23B/C; Kehew and Lord, 1986; Kehew et al., 2009).

It is suggested, that the Borgholzhausen overspill channel was closed, probably by large ice masses, in an early flood stage. Evidence is given by the shallow incision of the inner channel in front of the pass, the missing infill of fine-grained deposits and the incomplete streamlining of the hills, separated by very shallow channels. The shape of the Borgholzhausen plunge pool does not show headward erosion, which is indicative for a rapid formation process of the
plunge pool (Comiti et al., 2005). In contrast, the Bielefeld plunge pool and pass area give evidence for a longer active phase of the Bielefeld pass indicated by the plunge pool shape, which shows backward erosion (Comiti et al., 2005). The deep plunge pool in front of the Bielefeld and Borgholzhausen overspill channels formed by waterfall erosion during lake outburst, where plucking processes rapidly carved out the fractured marl- and limestone of the Mesozoic basement (Fig. 20; Rathburn, 1993; Whipple et al., 2000; Lamb and Dietrich, 2009). The shallower, broader channel in front of the Borgholzhausen, Dissen and Hesseln pass may be explained by more sheet-like flows in front of this 12 km wide overspill zone, although the calculated peak discharge is much higher than that calculated for the Bielefeld pass. The channel depth of the Borgholzhausen pass may also be underestimated, because fine-grained flood deposits are missing and it is not clear, if the lower channel was filled with coarser-grained sediments that cannot be distinguished in borehole logs from older meltwater deposits. The upper channel fill is characterized by diamicton, interpreted as flood-related hyperconcentrated flow deposits (diamicton II; Fig. 11). It is assumed that these hyperconcentrated flow sediments were deposited as the Borgholzhausen overspill was partly blocked by ice masses during the flood or as the discharge decreased at the final stage of the flood. These deposits were partly incised at the end of the flood, leading to the formation of large and incomplete streamlined hills directly in front of the Borgholzhausen pass (Fig. 8B).

Downslope from the Bielefeld and Borgholzhausen overspills the main channel systems pass into 0.1–2 km wide and 2–22 m deep valleys that merge into the River Ems. The Ems valley in the SW of the passes is uncommonly broad (up to 12 km wide). The basal valley-fill deposits of River Ems, Lippe and Lutter consist of 8 m thick interbedded pebbly sand and gravel, referred to as “Knochenkies” (“bone gravel”; Fig. 22A). These deposits, exposed further to the south of the study area, only occur in channels. It contains remains of mammals, molluscs, plants and wood. Components consist of clasts with a Scandinavian/Baltic provenance, Palaeozoic sandstone clasts derived from the Rhenish Massif and a high amount of Cretaceous limestone clasts (“Plänerkalkstein”) derived from the Teutoburger Wald Mountains and reworked pre-glacial fluvial deposits (Skupin, 2002, 2004; Lenz, 2003). The large amount of Cretaceous “Plänerkalkstein” clearly points to flows from the northeast and the erosion of basement rocks.

The age of the “Knochenkies” deposits is poorly constraint. Radiocarbon-dating of mammoth (Mammuthus antiquitatis) and rhinoceros (Coelodonta primigenius) bones from the Lippe valley, point to an early Weichselian age, pollen analysis to an Eemian age (Skupin, 1983). However, the results are critically discussed by other workers, who assume a late Saalian (Warthian) age based on U/Th dating of mammoth teeth (Schmitz, 1990). It is assumed that the “Knochenkies” represents an outburst flood-related channel-lag deposit, indicating strong erosion in the pass areas of the Teutoburger Wald Mountains and reworking of Pleistocene sediments in the Münsterland Embayment. The “Knochenkies” is overlain by up to 5 m thick interbedded sand, silt and gravel, which contain organic material (plants and molluscs remains). Sparse components have a Scandinavian/Baltic provenance. This succession is overlain by ~6 m thick fine- and medium-grained sand thought to be of Weichselian age (Lenz, 1997, 2003; Dölling, 2005).

Probably 55–65 km3 of flood water drained southwestward through an overspill channel, located in the south of the Münsterland Embayment at an altitude of ~115 m a.s.l. This spillway is indicated by the occurrence of “Knochenkies” (Fig. 22A).

Further ice-lobe retreat led to the opening of a 95 m a.s.l. overspill channel in the Teutoburger Wald Mountains (Fig. 22B) and a second outburst of Lake Weser during which ~20 km3 of water was released into the Münsterland Embayment (Winsemann et al., 2011, in press) This outburst event probably destabilized the ice-lobe in the Münsterland Embayment and triggered the opening of the 60 m a.s.l. overspill channel located at the margin of the retreating Drenthe ice sheet in the central Münsterland Embayment (Fig. 22B). As a consequence Lake Münsterland completely drained through the Lippe and Emscher valley (Fig. 22B) toward the west, releasing up to 65 km3 of water.
The “Delbrücker Rücken” might represent a large horse-shoe-shaped streamlined hill that formed during outburst of the Münsterland Lake and indicates the drainage pathway towards the southwest.

6.3.1. Influence of outburst floods on ice sheet dynamics

It is assumed that the formation and catastrophic drainage of proglacial lakes in the study area considerably influenced the stability of the Drentse ice sheet and may have initiated the Hondsrug ice stream. The progressively deepening lakes in the Münsterland Embayment and Weser Valley probably led to an increased removal of ice through calving, a rapid retreat of the western ice-lobes and opening of the 135 m a.s.l. and 95 m a.s.l. overspill channels in the Teutoburger Wald Mountains (Winsemann et al., 2011, in press). During the subsequent Weser Lake outburst floods 110 km$^3$ of water was released into the Münsterland Embayment and the lake level of the Weser Lake dropped by as much as 100 m. These two outburst floods must have led to an increase in ice temperature due to frictional heating and enhanced melting and rapid destabilization of the Münsterland ice lobe (cf. Stokes and Clark, 2004). Subsequently an ice re-advance occurred, leading to the renewed closure of the 95 m a.s.l. overspill channel and

---

**Fig. 24.** Supposed drainage pathway of lake outburst floods at the end of the late Saalian glaciation. A) Drainage through the Münsterland Embayment and the Central Netherlands into the North Sea. The distribution of Unit S5 is drawn after Busschers et al. (2008). The figure is based on SRTM data. B) Speculated further drainage pathway into the English Channel. The figure is based on GTOP030 (USGS).
a related lake-level rise of glacial Lake Weser (Winsemann et al., 2011). This ice-advance is related to the Hondsrug ice stream (cf. van den Berg and Beets, 1987; Passchier et al., 2010), which is the last ice-advance recorded from the Münsterland Embayment (Skupin et al., 1993). It is speculated that the Hondsrug ice stream may have been enhanced or even triggered by the combination of glacial lake formation in the Münsterland Embayment and outburst floods of glacial Lake Weser. The associated removal of ice may have led to a rapid draw-down of ice triggering fast ice flow (Winsborrow et al., 2010).

After the drainage of glacial Lake Weser and glacial Lake Münsterland the Hondsrug ice stream advanced into the Münsterland Embayment, probably considerably thinning the ice sheet profile in this region. The splayed, lobate pattern of the Hondsrug ice stream (Van den Berg and Beets, 1987; Skupin et al., 1993) indicates that it probably terminated on dry land or discharged into very shallow water. Stokes and Clark (2004) pointed out that once achieved the calving processes and losses might play a secondary role in the functioning of an ice stream and once rapid basal sliding is established thermomechanical feedback mechanism may sustain fast ice flow. Subsequently the thinned Drentse ice sheet deglaciated rapidly (Van den Berg and Beets, 1987; Passchier et al., 2010).

6.4. Pathway of the outburst floods into the North Sea and Strait of Dover?

The further pathway of the lake outburst floods has not been studied in detail yet. It is hypothesized that the main flood pathways in the Münsterland Embayment are indicated by the broad Emscher and Lippe Valley.

The Lippe valley is uncommonly broad with a width of up to 7 km and a depth of up to 14 m and it is most unlikely that the modern Lippe eroded such a broad valley. Therefore, Herget (1997) and Skupin (2002) suggested that the Lippe valley formed by meltwater erosion during the Saalian glaciation. The basal sediment infill of the valley consists of “Knochenkies” interpreted as an outburst related channel-lag deposit (Fig. 22B), which subsequently might have been reworked by younger fluvial processes. A further subordinate drainage route may be indicated by a 40 m wide and 2 m deep channel connecting the River Lippe and the River Emscher in the south (Fig. 22B). This “hanging” channel lies above the recent Lippe and Emscher valley at an altitude of 67 m a.s.l. and is filled with cross-stratified sand and gravel that contains components with a Scandinavian/Baltic provenance (Herget, 1998, 2002). The till cover between the Lippe and Emscher valley is strongly dissected (Braun and Thiermann, 1975). The till reliefs partly display streamlined shapes, pointing to southwestward directed flows, which are attributed to the outburst floods.

The outburst floods than probably crossed the Rhine valley, the area of the Central Netherlands and flowed into the Strait of Dover. The flow pathway may be indicated by the deep truncation of push moraine ridges at the western margin of the Rhine valley (Fig. 24A; Klostermann et al., 1984) and an up to 10 m deep and 5–7 km wide valley in the Central Netherlands (unit S5; cf. Busschers et al., 2008), formerly interpreted as a valley of the River Meuse that might have formed during the late Saalian in response to the opening of the Dover Strait. However, most other authors assume an Elsterian origin of the Strait of Dover by a significant flood event related to a megaflood from a proglacial lake in the southern North Sea Basin (Smith, 1985; Gibbard, 1995; Toucanne et al., 2009; Westaway and Bridgland, 2010).

OSL-dates of the valley-fill sediments (unit S5) range from 147 to 211 ka and point to a strong reworking of older fluvial deposits and poor bleaching (Busschers et al., 2008). The heavy mineral association differs from the underlying fluvial deposits and is characterized by a higher content of stable minerals, metamorphic minerals and tourmaline, thought to reflect most likely the Meuse catchment area (Busschers et al., 2008). However, this heavy mineral association may also be explained by reworking of fluvial and glacialic deposits in the Münsterland Embayment. The occurrence of augite indicates reworked Rhine sediments, related to the Eifel volcanism (Hennenings, 2006). It is therefore speculated that the valley formation in the Central Netherlands (unit S5, cf. Busschers et al., 2008) is related to outburst flood erosion. Subsequent the valley became filled with eroded material from the Münsterland Embayment during the waning flood. The floods probably drained into the Dover Strait and Bay of Biscay and may have caused the second erosional event in the English Channel recognized by Gupta et al. (2007). In core data from the Bay of Biscay Penaud et al. (2009) identified three main sedimentary events during MIS 6 (~150 ka) characterized by a high amount of reworked late Jurassic to Miocene dinocysts, attributed to an increased erosion in the English Channel region. These sediments with reworked dinocysts may have partly also been derived from the Münsterland Embayment and may represent the three outburst floods of glacial Lake Weser.

7. Conclusion

During deglaciation glacial Lake Weser catastrophically drained into the Münsterland Embayment. The drainage routes are characterized by streamlined hills, deep plunge pools, oversteepened channels and deep valleys, cut into bedrock and older Pleistocene deposits.

Probably up to 90 km³ of water drained catastrophically into the Münsterland Embayment during the first lake outburst event. The water drained via four major outlet channels (Dörenschlucht, Bielefeld, Borgholzhausen and Iburg pass), when the western ice dam failed.

Calculated peak discharges are 2.4 × 10⁶ m³/s to 1.3 × 10⁶ m³/s. Associated erosional features are:

- Up to 35 m deep, 780 m long and 400 m wide plunge pools cut into the Mesozoic basement rocks in front of the main outlet channels.
- Up to 4 km wide and 30 m deep channels in front of the Borgholzhausen and Bielefeld overspill channels.
- Streamlined hills that are 2–11 m high, 600–4300 m long and 170–1200 m wide and are arranged into three fields in front of the main lake outlet channels, covering an area of 60–130 km², each. The hills have quadrilateral to elongate, partly V-shaped chevron-like forms with an average aspect ratio of 1:3.3. The hill shapes are similar to hills in the Channeled Scablands (Patton and Baker, 1978). The arrangement of hills in the marginal channel zone resembles outburst related regions in the Northern Great Plains (Kelew and Lord, 1986; Kelew et al., 2009). The hills formed under submerged to partly submerged flow conditions.

- Anabranching channels separating the streamlined hills are 50–700 wide and 2–13 m deep. The channels were mainly filled with fine-grained flood sediments during the waning flood, deposited in a lake that formed when the outburst flood was dammed in the Münsterland Embayment.

- Opening of the 95 m overspill channel led to a second lake outburst of glacial Lake Weser, releasing up to 20 km³ of water. This outburst event probably destabilized the ice-lobe in the Münsterland Embayment and triggered the outburst of glacial Lake Münsterland, catastrophically draining south-westwards.

- The drainage pathway to the west is probably indicated by the uncommonly wide and deep Emscher and Lippe valley and the
deposition of a gravel lag (“Knochenkies”) at the base of the fluvial valleys.

- The further drainage pathway of the outburst floods may be indicated by highly dissected push moraines at the western margin of the Rhine Valley and a 10 m deep and 5–7 km wide valley in the Central Netherlands (Unit S5, cf. Busschers et al., 2008). It is speculated that the flood eventually flowed into the Dover Strait and Bay of Biscay and caused the second erosional event recognized in the English Channel by Gupta et al. (2007).

Acknowledgments

We thank reviewers Philip L. Gibbard and Freek S. Busschers for careful and constructive reviews, which helped to improve the manuscript. Borehole data were generously provided by the Geologischer Dienst NRW (Krefeld), the LBEG (Hannover) and the Umweltamt Bielefeld. The manuscript benefited from the discussion with many colleagues. In particular we would like to thank V. Baker, F.S. Busschers, P.A. Carling, J. Carrivick, K.M. Cohen, T. Fisher, R.L. Hooke, A. Kehew, D. Kellett, J. Klostermann, P. Komar, S. Morris, H.J. Pierik, J.A. Piotrowski, J. Shaw and K. Skupin. Many thanks are also due to Ariana Osman for improving the English.

References


Baker, V.R., 2002a. The study of super S. Busschers for carefl and constructiv reviews, which helped to improve the manucript. Borehole data were generously provided by the Geologischer Dienst NRW (Krefeld), the LBEG (Hannover) and the Umweltamt Bielefeld. The manuscript benefited from the discussion with many colleagues. In particular we would like to thank V. Baker, F.S. Busschers, P.A. Carling, J. Carrivick, K.M. Cohen, T. Fisher, R.L. Hooke, A. Kehew, D. Kellett, J. Klostermann, P. Komar, S. Morris, H.J. Pierik, J.A. Piotrowski, J. Shaw and K. Skupin. Many thanks are also due to Ariana Osman for improving the English.

References


Baker et al., 2008. It is speculated that the flood eventually flowed into the Dover Strait and Bay of Biscay and caused the second erosional event recognized in the English Channel by Gupta et al. (2007).

Acknowledgments

We thank reviewers Philip L. Gibbard and Freek S. Busschers for careful and constructive reviews, which helped to improve the manuscript. Borehole data were generously provided by the Geologischer Dienst NRW (Krefeld), the LBEG (Hannover) and the Umweltamt Bielefeld. The manuscript benefited from the discussion with many colleagues. In particular we would like to thank V. Baker, F.S. Busschers, P.A. Carling, J. Carrivick, K.M. Cohen, T. Fisher, R.L. Hooke, A. Kehew, D. Kellett, J. Klostermann, P. Komar, S. Morris, H.J. Pierik, J.A. Piotrowski, J. Shaw and K. Skupin. Many thanks are also due to Ariana Osman for improving the English.

References


Baker et al., 2008. It is speculated that the flood eventually flowed into the Dover Strait and Bay of Biscay and caused the second erosional event recognized in the English Channel by Gupta et al. (2007).

Acknowledgments

We thank reviewers Philip L. Gibbard and Freek S. Busschers for careful and constructive reviews, which helped to improve the manuscript. Borehole data were generously provided by the Geologischer Dienst NRW (Krefeld), the LBEG (Hannover) and the Umweltamt Bielefeld. The manuscript benefited from the discussion with many colleagues. In particular we would like to thank V. Baker, F.S. Busschers, P.A. Carling, J. Carrivick, K.M. Cohen, T. Fisher, R.L. Hooke, A. Kehew, D. Kellett, J. Klostermann, P. Komar, S. Morris, H.J. Pierik, J.A. Piotrowski, J. Shaw and K. Skupin. Many thanks are also due to Ariana Osman for improving the English.

References


