Distal delta-plain successions

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Distal delta-plain successions
Architecture and lithofacies of organics and lake fills in the Holocene Rhine-Meuse delta plain, The Netherlands

Distale deltavlakte-afzettingen
Architectuur en lithofacies van organische accumulaties en meeropvullingen in de Holocene Rijn-Maasdelta, Nederland

(met een samenvatting in het Nederlands)

Proefschrift

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Preface

What a wealth to have the opportunity to conduct a PhD-research project in fluvial sedimentology! The project turned out to be very dynamic, including field campaigns, conference trips, teaching, laboratory and office work. It comprised an excellent mix between autonomous and team work. I regard the contribution of the many people that assisted and supported me during the last few years as an essential architectural element to the success of this project and I acknowledge all of them with gratitude.

Ward, despite the fact that you retired soon after my appointment, you appeared to be the most constant factor in the supervision group. Thank you for the great effort and the valuable comments on the texts and the open-minded supervision. I'd like to thank the late Henk Berendsen for the time he put into this project, despite or possibly due to his retirement at the day I started this project. I deeply regret that he has not been able to join the final stages of this project. I thank him for the confidence, the insights in fluvial systems and physical-geographic research and the very fast turnover of manuscript commenting, often within 1 (one!) day. Besides, I am delighted he never used his notorious "bull shit" stamp for my manuscripts; I would suggest, nevertheless, that he was inclined to do so now and then. Esther, I very much appreciate your dedication and contributions to the project. You also gave me the fantastic opportunity to visit the Cumberland Marshes. The front cover is the only direct input from that trip to this thesis, but indirectly, that unforgettable trip has meant much more, as you know. I acknowledge Henk Weerts for his effort during the first years to make a success of this project. It was always helpful to discuss the many facets of the research, in particular sedimentology and palaeogeography. Jeroen, I'd like to thank you for the concise manuscript reviewing and clear insights. Especially your contribution to the palaeogeography paper, and scientific writing in general, is invaluable. I gratefully acknowledge Ward, Esther and Jeroen for providing the conditions to change plans halfway the project and to conduct the research that resulted in chapter 4 and 5.

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The very good atmosphere in the Physical Geography department—within the Gruppo Fluvial, among PhD-students and among the first-floor alley inhabitants—undoubtedly contributed to this project. Marc, I very much appreciate your company as roommate! I value the many good laughs, the fruitless attempts to neglect our giggling neighbours, and the flexibility to change personal 'heroes' on daily basis. I appraise your effort to hop on the plane with your family for assisting me during the defence, thanks! I'd also like to thank Sanneke, for all the good times before and during the PhD project. I am happy that you are one of the most constant factors for me in Physical Geography! Sibren, after having studied together in Amsterdam, it was a great pleasure to start on the same day in Utrecht and to share rooms during the first six months. Furthermore, I'd like to thank Juul Beltman, Frans Buschman, Roy Frings, Arien Lam, Wiebe Nijland, Jan Peters, Chris Roosendaal, Loes van Schaik, Hanneke Schuurmans and Willem Toonen. I would like to thank colleagues from other institutes as well: Bas van Geel, Freek Busschers, Peter Vos, Meindert van den Berg, Wim Booltink, Hennie van der Putten, Hennie Mensink, Geert-Jan Vis and Henk Kombrink.

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1 General introduction

1.1 CONTEXT

Delta plains, which can be defined as 'extensive areas of low slope that are traversed by distributary channels, having sub-aerial parts (e.g., floodplains with levees and crevasse splays) and sub-aqueous parts (e.g., marshes, lakes, interdistributary bays, and tidal flats)' (cf. Bridge, 2003, p. 307), are of significant societal and economic value. Modern delta plains are attractive for e.g., human habitation because they provide good conditions for agriculture, settlement, transport, etcetera and, hence, are amongst the most densely populated geomorphic features worldwide. Ancient fluvio-deltaic deposits, additionally, are reservoir rocks for hydrocarbons and water. The resulting need to increase both our understanding of delta formation and knowledge of the composition of delta plains stimulated a wide range of research activities in delta regions.

The current understanding of delta formation is largely based on field investigations, laboratory analogue experiments and model studies (Bridge, 2003). One of the key principles within these studies concerns alluvial architecture, which has been defined as 'three-dimensional geometry and configuration of various types of fluvial deposits in alluvial successions' (Allen, 1965; Leeder, 1978). Of special interest are the relations between observed architecture and the effect thereon of external controls, such as rate of base-level rise (e.g., Stanley and Warne, 1994) and sediment delivery (e.g., Erkens, 2009), and internal controls, such as avulsion (Slingerland and Smith, 2003) and compaction (Van Asselen et al., 2009) that govern delta-plain formation (e.g., Blum and Törnqvist, 2000, and references therein). Downstream changes of delta-plain architecture have been reported for various rivers, such as the Lower Mississippi Valley (LMV), USA (Schumm et al., 1994), the Brahmaputra, Bangladesh (Bridge, 2003, p. 150) and the Rhine-Meuse delta, The Netherlands (Wolfert, 2001; Gouw and Berendsen, 2007). Based on observed architectural and morphological changes, delta plains can be subdivided. In the Rhine-

Meuse, for example, two fluvial zones (proximal and distal) and an estuarine zone can be defined (Fig. 1.1). Similar subdivisions could also apply for other delta regions, such as the LMV (Gouw and Autin, 2008). Commonly, proximal delta plains are characterized by e.g., amalgamated channel-belt deposits, which usually form >50% of the succession, relatively high downstream gradient, single-thread meandering rivers, low average Holocene aggradation rates, low proportion of organics and relatively thin overbank successions. Distal delta plains, in contrast, are characterized by e.g., isolated channel-belt deposits, which typically form <30% of the succession, relatively low downstream gradient, straight-anastomosing channel systems, high average Holocene aggradation rates, high proportion of organics and relatively thick overbank successions. The estuarine zone is characterized by tidal deposits (e.g., tidal flat and tidal channel deposits).

Previous research in delta plains focused on the proximal zone and channel-belt deposits, whereas distal delta plains and overbank deposits received much less attention. For this reason, knowledge of architectural elements that are restricted to distal delta plains is much more limited. Prominent examples are organic accumulations and lake fills (organic and clastic) that form in delta-plain lakes, which both relate to the widespread occurrence of peat-forming wetlands, including lakes (Smith and Pérez-Arlucea, 1994). Other examples are the delta-wide spatial and temporal distribution and lithofacies composition of coarse-grained overbank deposits, which include crevasse-splay deposits, clastic lake fills, and bay-head delta deposits. Many questions regarding their architecture, lithofacies and formation remain to be answered, as will be discussed in this thesis.

The Holocene Rhine-Meuse delta (Fig. 1.2) is an excellent area to address these issues primarily owing to the presence of an extensive distal delta plain including and organics. Furthermore, the stratigraphic position of the delta plain (i.e., at or near the surface) implies relatively good access to the study object and morphological expression of the most recent deposits, which enables the use of high-resolution Digital Elevation Models. Additional conditions that support this choice of study area, are the availability of extensive geological databases, including time control, and the presence of a robust architectural and palaeogeographical framework (Berendsen and Stouthamer, 2001), for which the interplay of external and internal controls is comparatively well understood. Extending our knowledge of organics and lake fills in the Rhine-Meuse delta thus enables coupling of observed architectural-element properties with factors controlling delta formation in general, such as relative base-level rise and upstream sediment delivery. Therefore, the results of this study may contribute to understanding of the evolution of delta-plain successions.

In this chapter, I present an overview of the history of delta-plain mapping, both worldwide and in the Rhine-Meuse delta. Further, I present a literature review of the architecture and facies distribution of coarse-grained overbank deposits. The knowledge hiatuses emerging from this review ultimately lead to the formulation of a problem definition and the thesis research objectives. A description of the approach to assess these research objectives and how this thesis is organized conclude this chapter.

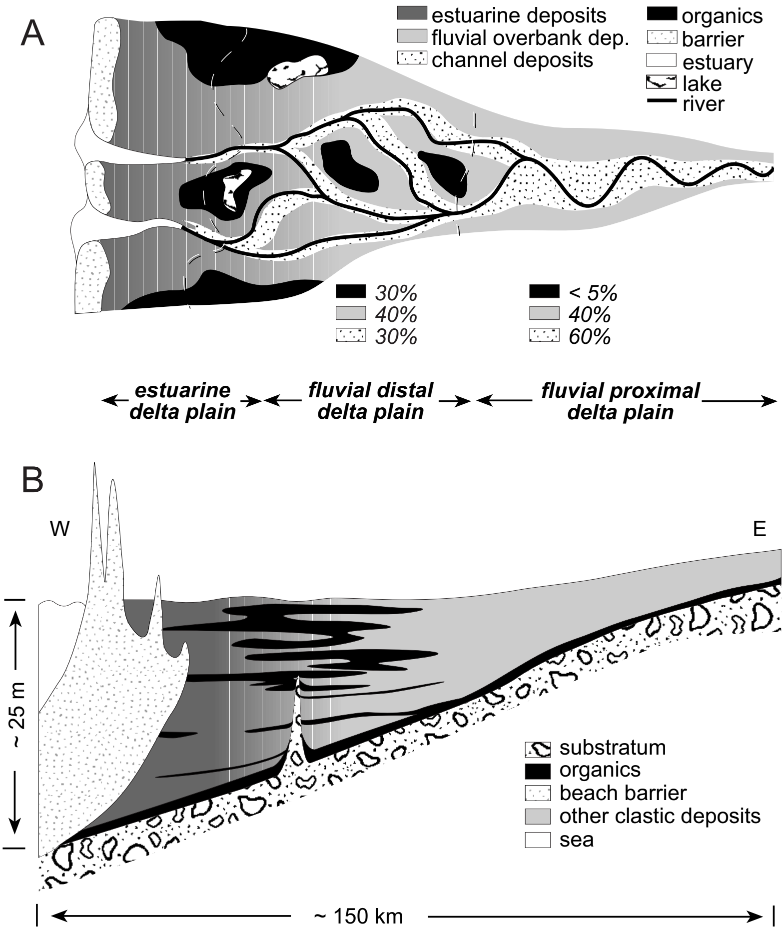

Figure 1.1. Schematic overview of the Rhine-Meuse delta A) in planview and B) in longitudinal cross section. Indicated is the subdivision in proximal and distal fluvial zone and the estuarine zone. The cross section is adapted from Hijma and Cohen (2010).

1.2 HISTORICAL PERSPECTIVE

Despite the relevance of delta plains, the fundamental tenets on their formation and composition only emerged during the 20th century, probably owing to their complex nature and difficult, i.e., expensive, data gathering. The seminal work of Harold Fisk (1944, 1947) in the LMV formed the starting point for an increasing number of studies on fluvio-deltaic successions, ultimately leading to the conception of basic principles on the formation and sedimentary characteristics of delta-plain successions (e.g., Allen, 1965; Schumm, 1968; Allen, 1978; Leeder, 1978; Miall, 1985; Stanley and Warne, 1994; Miall, 1996; Blum and Törnqvist, 2000; Bridge, 2003; Davies and Gibling, 2010).

Knowledge of the architecture and composition of fluvio-deltaic deposits significantly contributed to the understanding of the development of fluvio-deltaic successions. Previous research in the fluvio-deltaic realm (e.g., Rust, 1972; Miall, 1977; Bridge, 1993) primarily focused on coarse-grained successions, deposited by braiding rivers, which is why the architecture and facies composition of these deposits are relatively well understood. Fine-grained fluvial successions in fluvio-deltaic wedges (coastal prisms cf. Posamentier et al., 1992) deposited by, e.g., meandering and straight anastomosing rivers, in contrast, received less attention (e.g., Kraus, 1987; Miall, 1996). Furthermore, the available studies on fluvio-deltaic deposits are strongly biased towards channel deposits (e.g., Farrell, 1987; Miall, 1996; Gouw, 2008). Yet, according to Gouw (2008), overbank deposits and organics comprise large proportions of fluvial successions in delta plains. He concludes, based on several valley-wide cross sections in the Upper Rhine-Meuse delta, The Netherlands, and the LMV, that overbank deposits comprise almost half of the deposits in these delta plains. Further, organics comprise up to 30% of the fluvio-deltaic succession in the Rhine-Meuse delta plain (Gouw, 2008). Nevertheless, only a limited number of studies addresses attention to organics in delta plains (Van der Woude, 1981; Styan and Bustin, 1983a). Knowledge of the development, architecture and facies composition of overbank deposits and organics is essential to define the composition and the architecture of delta plains.

Research in the Holocene Rhine-Meuse delta plain (Fig. 1.2) commenced during the 20th century and is marked by the publications of Teunis Vink (1926; 1954) in which he concisely presented the results of extensive field investigations. In the '50s, investigations by workers from Wageningen University, led by Edelman, formed the basis for the present knowledge of the delta-plain composition (e.g., Pons and Wiggers, 1959, 1960). From the 60's onward, the subsoil of the delta was systematically mapped by various institutes and from different perspectives. The 'Stichting voor Bodemkartering' (Stiboka) constructed soil maps (e.g., Stiboka, 1965; Pons and Van Oosten, 1974), the Rijks Geologische Dienst (RGD) constructed geological maps (e.g., Van de Meene et al., 1988; Bosch and Kok, 1994) and both institutes jointly produced geomorphological maps (e.g., Van den Berg and Kluiving, 1992). From the '80s onward the large data archive that had been collected systematically by Henk Berendsen c.s. from Utrecht University, yielded and supported a probably unforeseen number of studies on fluvio-deltaic deposits in the Rhine-Meuse delta plain (Berendsen, 1982; Törnqvist, 1993a; Weerts, 1996; Makaske, 1998; Berendsen and Stouthamer, 2001, and references therein; Stouthamer, 2001a; Cohen, 2003; Gouw, 2007; Erkens, 2009; Hijma, 2009; Van Asselen, 2010). The mapping history of the Rhine-Meuse delta has been discussed in detail by Berendsen (2007).

Owing to increased data availability and querying techniques along with improved understanding of delta-plain composition and formation, the research focus in the Rhine-Meuse delta has widened from individual channel systems to the entire or larger part of the delta-plain. Examples of the latter are publications on palaeogeographic

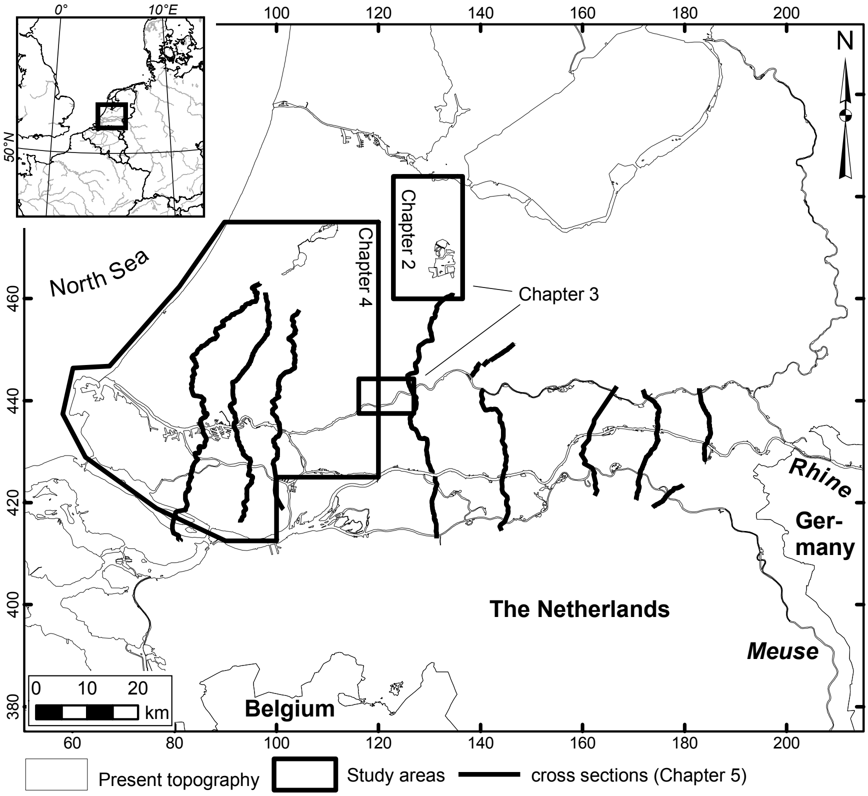

Figure 1.2. Topographical map of the Rhine-Meuse delta, The Netherlands, showing the study areas and location of the cross sections used. Coordinates are according to Rijksdriehoekstelsel.

development (Berendsen and Stouthamer, 2000), avulsions (Stouthamer and Berendsen, 2000), architecture (Gouw, 2008) and sediment budgets (Erkens, 2009).

1.3 ARCHITECTURE AND FACIES DISTRIBUTION OF COARSE-GRAINED OVERBANK DEPOSITS AND ORGANICS: A LITERATURE REVIEW

1.3.1 The architectural-elements concept

Miall (1985; 1996) stated that a limited number of units ('architectural elements', Tab. 1.1) would be sufficient to describe the architecture of fluvial deposits. For rivers in delta plains, for example, he defined six architectural elements: channels, sandy bedforms, downstream accretion deposits, lateral accretion deposits, scour hollows and overbank deposits. Many workers adopted Miall's concept (amongst others Törnqvist, 1993a; Jorgensen and Fielding, 1996; Hornung and Aigner, 1999), although the number of elements proved to be insufficient, especially concerning the overbank environment (e.g., Farrell, 1987; Kraus, 1987) as further discussed below. Nevertheless, architectural elements

Table 1.1. Definitions of sedimentological and morphological terms as used in this study.

Term	*Definition*	*Reference*
Alluvial architecture	Three-dimensional geometry and configuration of various types of fluvial deposits in alluvial successions	Allen (1965), Leeder (1978)
Architectural element	A sediment body that is characterized by a distinctive assemblage of lithofacies and geometry	After Miall (1996, p 91)
Facies	Reflects the character or nature of a deposit and reflects the environmental conditions of deposition	Salvador (1994, p16)
Meandering river	A river that has a sinuous channel that shifts its course by lateral erosion and accretion	Allen (1965), Brice (1984), Miall (1985), Collinson (1996)
Anastomosing river	A river that is composed of two or more interconnected channels that enclose floodbasins	Makaske (2001)

in fluvio-deltaic wedges can be arranged in three clusters: channel deposits, overbank deposits and organics (Fisk, 1944; Allen, 1964). Overbank deposits may be subdivided in, for example, natural-levee deposits, floodbasin deposits and coarse-grained overbank deposits.

1.3.2 General architecture of distal delta-plain successions

The architecture of fluvial deposits in delta plains is related to channel planform as has been found, for example, in the Rhine-Meuse delta (Weerts and Bierkens, 1993) and in the LMV (Aslan and Autin, 1999). Most delta plains predominantly comprise deposits of river channels that have a single-thread meandering or a straight-anastomosing channel

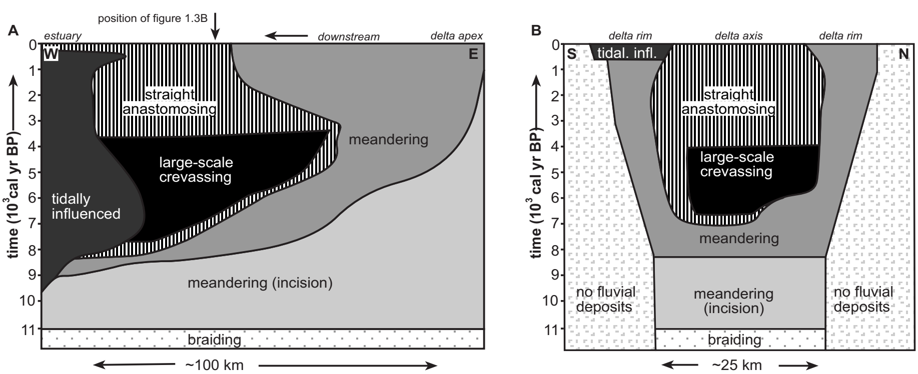

Figure 1.3. Diagrams showing (A) cross-valley (B) along-stream variation of channel planform during the Holocene in the Rhine-Meuse delta (after Törnqvist, 1993b; Berendsen and Stouthamer, 2001).

planform (Fisk, 1944; Törnqvist, 1993b; Saucier, 1994; Berendsen and Stouthamer, 2001, Tab. 1.1). Meandering river deposits have relatively wide channel-sediment bodies and the associated overbank deposits have a relatively simple geometry (Weerts and Bierkens, 1993; Aslan and Autin, 1999). Straight-anastomosing river deposits, in contrast, have comparatively narrow channel-sand bodies and the architecture of the overbank deposits tends to be very complex (Weerts and Bierkens, 1993). In the Rhine-Meuse delta, straight-anastomosing river systems occur in the distal delta plain (Van der Woude, 1981, 1984), mainly during the middle Holocene (Fig. 1.3) (Törnqvist, 1993b; Makaske, 1998; Makaske et al., 2007). Single-thread meandering-river systems dominated in the distal delta plain during the early and late Holocene and in the proximal delta plain throughout the Holocene (e.g., Törnqvist, 1993b).

The geometry of overbank deposits in meandering river systems generally is relatively simple and the thickness gradually decreases away from the channel (Fisk, 1944; Allen, 1964; Weerts and Bierkens, 1993; Aslan and Autin, 1999). Large spatial variations in the facies distribution and thicknesses, in contrast, have been reported for overbank deposits of straight anastomosing river systems (Törnqvist, 1993b; Weerts and Bierkens, 1993; Aslan and Autin, 1999).

1.3.3 Coarse-grained overbank deposits

Coarse-grained overbank deposits include crevasse-splay deposits and lacustrine deltas. Crevasse splays form when water from the main channel runs into a floodbasin via a crevasse-splay channel. Most often, water breaches through a natural levee in the outer bend of a river channel (e.g., Smith, 1986; Farrell, 1987) and upon flow velocity reduction sediment is deposited. Lacustrine deltas form when channel sediment is diverted to a floodplain lake (Tye and Coleman, 1989b), although crevasse-splay deposits may also form in lakes, such as those in the Cumberland Marshes, Canada (e.g., Smith et al,. 1989). Terminology, apparently, is not unambiguous, which illustrates the need to further investigate these architectural elements. It seems reasonable to consider two types of elements, being those that form in periodically inundated floodbasins (crevasse-splay deposits) and those that form in permanent lakes. I propose the latter to be termed organic-clastic lake fills, instead of lacustrine deltas, for reasons that will be outlined below. The differences in depositional setting between both architectural elements tentatively suggests the existence of fundamental architectural and facies differences.

Crevasse-splay deposits

Smith et al. (1989) proposed a 3-stage conceptual model to describe the development and facies distribution of crevasse splays (Fig. 1.4). During stage 1, the splay is relatively wide and thin, and flow is unconfined resulting in a fan-shaped sediment body. As the splay grows (stage 2), confined flow becomes apparent in a diverging channel network in the upstream part of the splay. During stage 3, the splay is stabilized and only a few active channels are left, which form an anastomosing channel pattern. This model, although based on organic-lacustrine lake fills, has been proven to be of value also for 'true' crevasse-splay deposits in other settings. Crevasse-splay deposits have been reported from a wide range of fluvial settings (see extensive overview provided by Miall, 1996, and references therein), including distal delta-plain successions. It remains unknown, however, how crevasse-splay deposits are spatially and temporally distributed in delta-plain successions and which factors govern that distribution (Stouthamer, 2001b).

The plan-view shape of crevasse-splay deposits is mostly lobate, showing (terminal) channels that, as fingers, extent farther into the floodbasin. Crevasse-splay deposits in the Rhine-Meuse delta have maximum thicknesses of 1-2 m for the splay and 8 m for the channel deposits (Weerts, 1996; Makaske, 1998; Makaske et al., 2007). In the LMV, the maximum thickness of splay deposits is 5 m (Farrell, 1987).

The lithofacies distribution of crevasse-splay deposits tends to be very complex (e.g., Smith, 1986; Makaske, 1998; Hornung and Aigner, 1999; Makaske, 2001). Crevasse splays may comprise sand (mainly crevasse-channel deposits), silt (crevasse-channel and crevasse-levee deposits) and clay (crevasse-levee and abandoned crevasse-channel deposits). Farrell (2001) estimated that fine-grained deposits in crevasse splays may make up 70% of the total volume of the crevasse splay.

Organic-clastic lake fills

Deposits in delta plains do not often include significant portions of both clastic deposits and organics. Lakes, however, may be filled by a combination of organic lithofacies and clastic sediment (e.g., by lacustrine-delta deposits), which commonly were regarded as two separate architectural elements: organics and, e.g., crevasse-splay deposits respectively. To

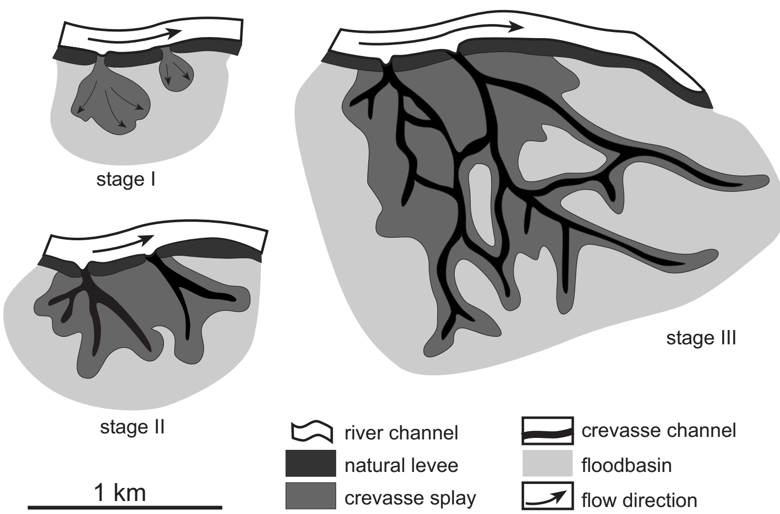

Figure 1.4. Conceptual model for the development of crevasse-splay deposits (Smith et al., 1989). See text for explanation.

account for the genetic congruence, these are combined in this thesis in organic-clastic lake fills. These deposits have sharp lateral boundaries, as has been observed in both in the Cumberland Marshes (Smith and Pérez-Arlucea, 1994) and in the Rhine-Meuse delta (Van de Meene et al., 1988; Weerts et al., 2002). An organic-clastic lake fill, herewith, is defined as a laterally sharply-bounded sediment body that is deposited in a delta-plain lake, which comprises gyttja and overlying clastic deposits–lacustrine deltas–that exhibit a coarsening-upward succession. The clastic part of organic-clastic lake fills has been termed crevasse-splay deposits (e.g., Smith et al., 1989), crevasse-complex (e.g., Weerts et al., 2002) and lacustrine deltas (e.g., Tye and Coleman, 1989a).

Despite the relatively low number of studies on organic-clastic lake fills, (re) interpretation of published examples show that they occur in both modern (Smith et al., 1989; Smith and Pérez-Arlucea, 1994; Pérez-Arlucea and Smith, 1999), (sub)recent (Van de Meene et al., 1988; Tye and Coleman, 1989b, a; Weerts et al., 2002) and ancient settings (Gersib and McCabe, 1981; Jorgensen and Fielding, 1996; Hornung and Aigner, 1999; Rajchl and Uličný, 2005). Publications on organic-clastic lake fills suggest that they preferably occur in low-gradient, organic-rich settings. This has been found, for example, in the Cumberland Marshes (Smith et al., 1989; Smith and Pérez-Arlucea, 1994; Pérez-Arlucea and Smith, 1999), in the Rhine-Meuse delta (Van de Meene et al., 1988) and in the Atchafalaya Basin (LMV) (Tye and Coleman, 1989b, a). Currently, no studies exist that have investigated systematically the delta-scale spatial and temporal distribution of organic-clastic lake fills, nor the factors that controlled this distribution.

The development of organic-clastic lake fills partly follows the principles outlined by the aforementioned conceptual model of Smith et al., (1989), which indeed has been based on organic-lacustrine lake fills. Notwithstanding its contribution to the present understanding of coarse-grained overbank deposits, it does not explicitly account for the lacustrine origin of the deposits, as it ignores the impact of lake geometry on the architectural-element geometry and the presence of organic lake deposits underlying the clastic succession.

Thicknesses of organic-clastic lake fills in modern and (sub)recent fluvio-deltaic successions of 1-5 m have been reported (Van de Meene et al., 1988; Smith and Pérez-Arlucea, 1994). Reported surface areas typically range from 7,2 to 29 km^2 (Tye and Coleman, 1989a, b; Smith and Pérez-Arlucea, 1994; Pérez-Arlucea and Smith, 1999). Thicknesses of organic-clastic lake fills in ancient settings range from <2 m (Hornung and Aigner, 1999) to 5 m (Rajchl and Uličný, 2005).

Commonly observed vertical facies successions in organic-clastic lake fills comprise organic-lacustrine deposits overlain by a prodelta (clay), which is overlain by a delta front (silty and sandy clay) and distributary mouth bar (Tye and Coleman, 1989a; Smith and Pérez-Arlucea, 1994). A much thinner clastic fining-up succession, being levee deposits, usually overlies these deposits and forms the upper part of organic-clastic lake fills.

1.3.4 Organics in delta-plain successions

Organics accumulate in case of shortage of deltaic sediment relative to the available accommodation space. They include in situ peat and organic lacustrine deposits (gyttja cf. Van Post, 1860; Hansen, 1959; 'fluvio-lagoonal deposits' cf. Van der Woude, 1981; Weerts, 1996).

The proportion of organics in delta plains tends to increase from the apex to the coast; organics predominantly occur in the distal zone. This has been found for instance in the Po delta, Italy (Amorosi et al., 1999), in the Fraser delta (Styan and Bustin, 1983b) and in the Cumberland Marshes (Morozova and Smith, 2003). In the Rhine-Meuse delta, the proportion of organics increases in a downstream direction, being 30% halfway between the apex of the delta and the coast (Gouw, 2007).

Organics in general are sensitive to compaction, especially when buried by clastic deposits. Measurements in Holocene deltaic successions revealed that peat layer thickness may be reduced by more than 50% (e.g., Van Asselen et al., 2009) whereas coal layers in ancient deposits probably represent only 10% of the initial peat-bed thickness (Nadon, 1998).

Peat

Subdivision of peat is primarily based on nutrient availability and botanical content (e.g., Wassen et al., 1989; Wheeler and Proctor, 2000; Succow and Joosten, 2001). A limited number of studies in delta plains subdivide organics. Few distinguish in situ peat from organic lacustrine deposits (e.g., Van der Woude, 1981; Morozova and Smith, 2003) and even less subdivide peat and/or gyttja types (e.g., Styan and Bustin, 1983a; Van de Meene et al., 1988; Bosch and Kok, 1994). Commonly observed peat types are Sphagnum, Carex, Phragmites and wood peat (e.g., Styan and Bustin, 1983a; Van de Meene et al., 1988). Furthermore, Styan and Bustin (1983a, b) illustrate how organic successions relate to the position in the Fraser River delta plain, Canada.

Maximum thicknesses of organic beds of 6 m are reported from the Rhine-Meuse delta (Bosch and Kok, 1994; Gouw and Erkens, 2007). In the Cumberland Marshes, the thickness of in situ peat beds does usually not exceed 1 m although thicknesses of up to 3.2 m occasionally were found (Morozova and Smith, 2003). Peat beds may cover areas well over 10 km^2 (Törnqvist, 1993b; Bosch and Kok, 1994; Morozova and Smith, 2003). Often, however, organic accumulations incorporate (thin) clastic beds that represent a period of increased fluvial activity (Jorgensen and Fielding, 1996; Gouw and Erkens, 2007).

Gyttja

The most frequently reported types of organic lake deposits are algae gyttja and detrital gyttja (e.g., Pons and Wiggers, 1960; Smith et al., 1989). Other gyttja-types, though less often documented, are calcareous gyttja and siderite gyttja, which are often found in abandoned channels fills (e.g., Tebbens et al., 2000).

Gyttja layers in the Holocene Rhine-Meuse delta are usually less than 1 m thick (Van der Woude, 1981; Bosch and Kok, 1994; Weerts, 1996). The areal extent may reach several square kilometres (Bosch and Kok, 1994). In the Cumberland Marshes gyttja layers are generally not over 0.5 m thick although at one location a thickness of 1.5 m was measured (Morozova and Smith, 2003). Currently, there are no published area measurements of gyttja layers.

1.4 PROBLEM DEFINITION AND OBJECTIVES

In this thesis, two related issues are addressed. First, the occurrence of organic-clastic lake fills is potentially underestimated in delta-plain successions owing to their classification as crevasse-splay deposits. Second, at present a key to classify organics in delta plains lacks, which hampers systematic analyses of the spatial distribution of organic lithofacies and botanical facies.

The few publications that describe the geometry, facies composition and formation of organic-clastic lake fills commonly consider only the clastic part of the lacustrine succession, and ignore implications of a lacustrine setting on geometrical and sedimentary properties. Furthermore, the delta-wide distribution of coarse-grained overbank deposits, including crevasse-splay deposits and organic-clastic lake fills, and their effect on channel processes are presently not known. Natural delta-plain lakes and their fills, which are known to exist in other distal delta plains worldwide, are presently not incorporated in the geological framework of the Rhine-Meuse delta, notwithstanding tentative suggestions for their existence (Pons and Wiggers, 1960; Van der Woude, 1981, 1984; Berendsen and Stouthamer, 2001; Weerts et al., 2002).

Organics are often not differentiated in the field, despite their potential as proxy for a wide range of environmental conditions, such as hydrological state (e.g., water depth), nutrient availability and salinity. The botanical composition of peat has recently been identified to be an important factor in compaction susceptibility of organic intervals, which subsequently effects delta-plain architecture in a complex fashion (Van Asselen et al., 2010). Our understanding of the architecture and the evolution of the distal delta-plain successions, including organic-clastic lake fills, would greatly benefit from better knowledge of organic facies and their spatial and temporal distribution. Investigating organic facies in the entire Rhine-Meuse delta plain, however, seems not feasible as differential compaction obstructs straightforward time-depth correlations. Nevertheless, within the scope of this research it seems feasible to undertake the first step towards delta-plain organic-facies classification and mapping based on archived data. Analyses of basal peat provide an opportunity because this peat directly covers uncompactable substratum, which strongly limits differential subsidence. It, therefore, is well-suited to test classification procedures of organics and to determine the palaeo-environmental conditions that controlled its formation. Further, the notion that organic-clastic lake fills occur predominantly in bottom parts of Holocene successions, suggests that knowledge of the facies composition of basal peat (both botanical and lithological) will contribute to understanding distributions of organic-clastic lake fills.

The general objective of this thesis is to analyze and explain the architecture, facies distribution and formation of coarse-grained overbank deposits, with special attention for organic-clastic lake fills, and organics in the distal Holocene Rhine-Meuse delta plain. This objective has been specified by the formulation of research goals, which are:

1. To describe the geometrical properties and depositional facies of organic-clastic lake fills, to understand how they form and to determine their influence on delta-plain architecture and facies composition.
2. To investigate the sedimentary and botanical variability of organics in the basal peat layer of the distal delta plain and to assess the factors that controlled their spatial distribution patterns.
3. To analyze and explain spatial and temporal distribution patterns of coarse-grained overbank deposits and their lithofacies composition at delta scale.

1.5 APPROACH AND ORGANISATION

The main part of this thesis comprises four research chapters (Chap. 2-5), which are essentially all individual studies and outlined as scientific papers. This general introduction (Chap. 1) and a synthesis (Chap. 6) discuss the scientific contribution of these four studies together.

The first two research chapters (Chap. 2, 3) address the first research goal. Chapter 2 describes the Holocene palaeogeographical development of the Angstel-Vecht area (Fig. 1.2), which illustrates the history of a fluvial system that invades a peat area including lakes. This study is based on cross sections and a geomorphogenetic map (using archived and newly drilled core data and a high-resolution Digital Elevation Model), dating evidence (^{14}C, OSL, pollen, archaeological artefacts, stratigraphical relations, historical sources) and palaeo-environmental analyses (diatom and pollen analyses).

Chapter 3 zooms in on organic-clastic lake fills specifically by investigating their geometry, depositional-facies composition, sedimentation rates and formative duration, i.e., how much time of bankfull discharge was needed to fill the lake. This chapter ultimately presents a conceptual model for the development of organic-clastic lake fills. This chapter concentrates on the Angstel-Vecht area, where it profits from the results presented in chapter 2, and the Schoonhoven area (Fig. 1.2), where buried middle Holocene organic-clastic lake fills occur. The sedimentary characterization is partly based on earlier presented cross sections (Chap. 2), complemented with sedimentary logs of mechanical cores. The calculation of sedimentation rates and formative duration benefits from the palaeogeographic reconstruction presented in chapter 2.

Chapter 4 presents a classification key for organics in delta-plain successions, based on macroscopic core descriptions. This classification is the basis for an automated procedure with which organic facies are identified in archived core descriptions. Interpolation of the resulting subset of classified samples yield a distribution map of organic facies in the basal peat layer for the distal Rhine-Meuse delta (Fig. 1.2), which is subsequently interpreted in terms of dominant formative hydrological regime.

Chapter 5 presents the delta-scale spatial and temporal distribution of coarse-grained overbank deposits and their lithofacies composition, based on examination of eight existing valley-wide cross sections. By enlarging the scope to delta scale, the significance of coarse-grained overbank deposits for delta-plain architecture and facies composition can be addressed.

Chapter 6 synthesises the main conclusions of the preceding chapters, points out potential and realised practical applications, and signals opportunities for future research.

2 Influence of organics and clastic lake fills on distributary channel processes in the distal Rhine-Meuse delta (The Netherlands)

With contributions of Rik Feiken, Frans Bunnik and Jeroen Schokker

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ABSTRACT

Organic-clastic lake fills, which include lacustrine deltas, are prominent architectural elements in the distal zone of delta plains. We hypothesize that lakes and fossil lake fills affect the development of a fluvial distributary channel. To test this hypothesis, a Holocene palaeogeographic reconstruction is presented of the Angstel-Vecht area, The Netherlands based on a geomorphogenetic map, lithogenetic cross sections, microfossil analyses, and various dating methods (i.e. ^{14}C-dating, OSL-dating, pollen analyses and archaeological evidence). We found that peat accumulation on the Pleistocene substratum started ~8000 cal yr BP in response to base-level driven groundwater-level rise. Peat-bounded lakes existed in which gyttja was formed. Fluvial sedimentation in the study area commenced when a new Rhine distributary – the Angstel-Vecht – formed ~2970 cal yr BP. As a result, the lakes were filled with fluvially supplied clastic material. Fluvial activity diminished after ~2300 cal yr BP although the system continued to function as a high-discharge spillway for Rhine distributaries. This study illustrates the influence of lakes and lake fills on fluvial architecture and sedimentology. Downstream of a lake, levee deposits are thin compared to upstream reaches and the channel-belt sediment is clayey. Furthermore, channels maintain a meandering pattern when organic-clastic lake fills form the substrate because of the relative low bank stability. This is in contrast to straight channels in the distal delta plains that are influenced by erosion-resistant peat substrate.

2.1 INTRODUCTION

Delta plains are extensive low gradient depositional features traversed by distributary channels, and are comprised of clastic channel and overbank sediments and organics. Previous studies concerning delta plains have primarily focused on channel deposits whereas overbank deposits as well as organics have received much less attention. Overbank deposits, however, comprise approximately 40 % of the Holocene Rhine-Meuse and

Mississippi delta plains (Gouw, 2008; Gouw and Autin, 2008). This suggests an opportunity to extend the knowledge of fluvial architecture and delta plain development from the perspective of understanding the importance of overbank deposits. Such investigations also link to a comprehensive range of interdisciplinary topics, such as ground subsidence and flooding, the properties of reservoir rocks, ecosystem and environmental habitat characterization, and the geomechanical characterization of delta-plain deposits in relation to building activities.

Commonly recognized overbank fluvial architectural elements are natural levee, floodbasin, and crevasse-splay deposits. Lacustrine deltas or crevasse splays formed in peat bounded lakes are included in organic-clastic lake fills, and have been recognized in Holocene settings (e.g., Tye and Coleman, 1989b; Tye and Coleman, 1989a; Smith and Pérez-Arlucea, 1994; Pérez-Arlucea and Smith, 1999; Weerts et al., 2002; Davies-Vollum and Smith, 2008) and ancient environments (e.g., Jorgensen and Fielding, 1996; Hornung and Aigner, 1999). The mechanism of lacustrine-delta formation as well as their facies distribution is relatively well understood (Smith et al., 1989). The influence of lakes and organic-clastic lake fills on the pattern and dynamics of associated distributary channels, however, is not well understood. This stands in strong contrast to our understanding of the interrelations between channel processes and development in alluvial valleys and proximal delta plains (Ferguson, 1987; Van den Berg, 1995). Differences between channels within alluvial valleys and deltaic settings, however, represent fundamental differences in processes and constraints. For example, in deltaic settings lakes are a common component of the overbank environment and represent a substantial sediment sink. This fundamental tenet of deltaic fluvial systems elucidates two hypotheses examined in this paper. First, the presence of lakes reduces downstream sediment loads and consequently alters overbank sedimentation. Secondly, the presence of an organic-clastic lake fill, in an otherwise organic-dominated setting, changes bank stability and results in differences in distributary channel pattern. Although the reasoning behind the two hypotheses are straightforward, actual formal investigations of these hypotheses are lacking.

The floodplain of the Angstel and Vecht Rivers (the Angstel-Vecht area), part of the Holocene Rhine-Meuse delta (Fig. 2.1), is an ideal setting to test these hypotheses. The relatively uncomplicated fluvial stratigraphy enables the key stratigraphic units that influence channel processes to be identified. Further, this study benefits from regional scale palaeogeographic reconstructions of the relevant settings (Berendsen, 1982; Berendsen and Stouthamer, 2001), providing an opportunity to link distributary channel changes to the development of fluvial and coastal-deltaic environments.

The aim of this study is to determine how the presence of lakes and organic-clastic lake fills in the Holocene Angstel-Vecht delta plain affect the architecture and grain-size characteristics of fluvial deposits, and the development of distributary channels. We approach this by reconstructing the palaeogeography of the Angstel-Vecht delta plain using a variety of geomorphic, sedimentological, and palaeoecological data.

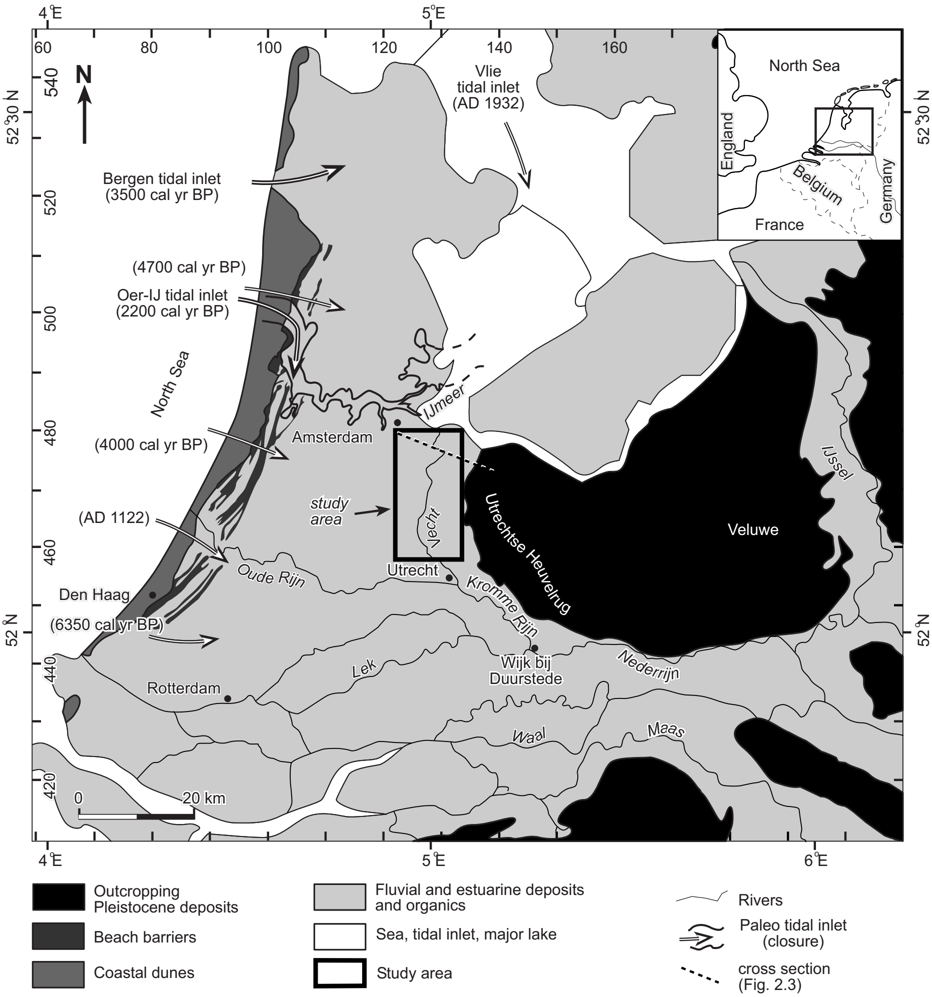


Figure 2.1. Overview of the Rhine-Meuse delta, the Netherlands. The palaeotidal inlets in the Holland coast and the extent of deposits associated with the Oer-IJ tidal inlet are indicated (compilation of maps from Beets et al., 1992; Jelgersma et al., 1970; Pons and Van Oosten, 1974; Pons and Wiggers, 1959; Van der Valk, 1996; Westerhoff et al., 1987). Note that there are only two Holocene distributaries that drain in northern direction: the IJssel and the Vecht-system. The box that indicates the location of the study area corresponds to the extent of Fig. 2.2, 2.5 and 2.9. Latitude-longitude coordinates are shown (outside frame), as well as Dutch coordinates (inside frame, Rijksdriehoekstelsel).

2.2 THE STUDY AREA

The study area is within the Angstel-Vecht delta plain, located between the city of Utrecht and the present IJmeer (Fig. 2.1). Three middle to late Holocene distributaries of the Rhine are present in the study area: the Vecht, the Oud Aa and the Angstel Rivers

(Fig. 2.2). These rivers flow from south to north and are currently flanked by cultivated peat areas and lakes, such as the Vinkeveense Plassen and the Loosdrechtse Plassen that resulted from peat digging (Fig. 2.2).

2.2.1 Geological setting – Pleistocene

East of the study area, a north-south oriented Saalian ice-pushed ridge (Utrechtse

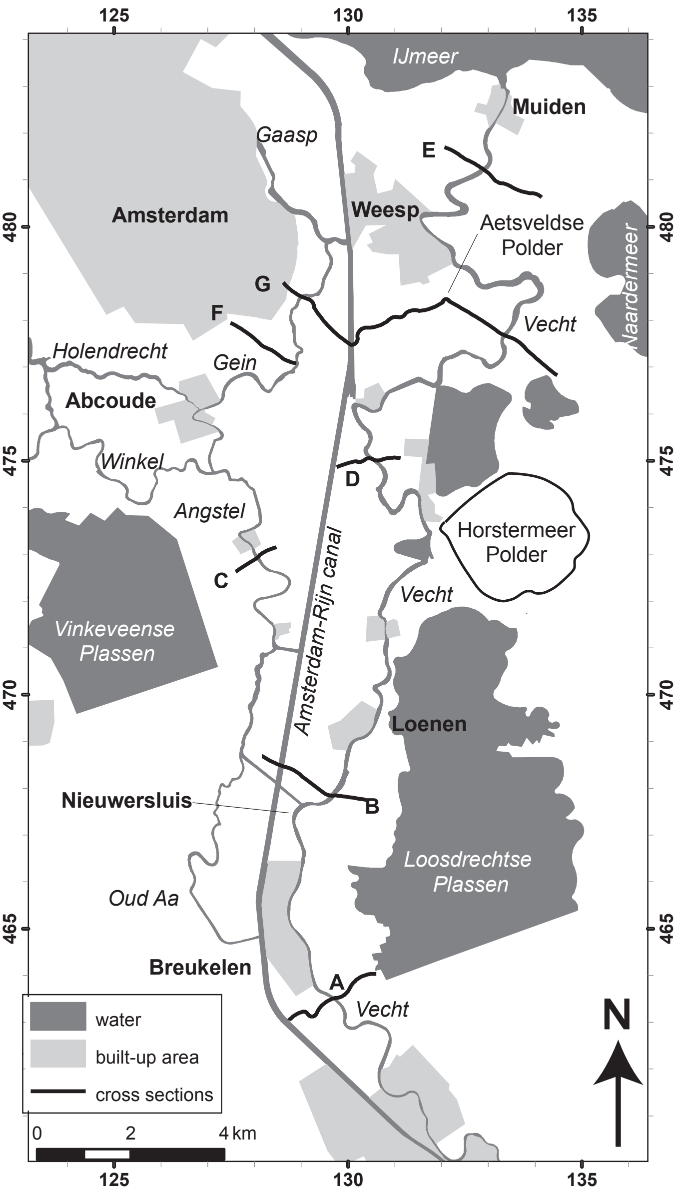


Figure 2.2. Planimetric map of drainage and built-up areas. The cross sections shown in Fig. 2.6 are indicated.

Heuvelrug) is present. Westward-dipping fluvio-glacial sandur deposits (Schaarsbergen Member, Drente Formation, Fig. 2.3, 2.4) are present adjacent to the ice-pushed ridge and continue in the subsurface of the study area. In the southern portion of the study area the Saalian ice-marginal Rhine and Meuse Rivers deposited sand and gravel (Krefteneye Formation, Unit S4, cf. Busschers et al., 2008). Sandy deposits (Boxtel Formation) overlie the sandur deposits and Saalian Rhine deposits. These deposits include a widespread 1 to 3 m thick layer of aeolian sand (so called 'coversand', Wierden Member, Boxtel Formation, Fig. 2.3), consisting of $CaCO_3$-poor, well-sorted quartz sand. The undulating topography of the Late Weichselian coversand is superimposed on the westward slope of the sandur and is visible in the surface topography in the eastern study area (Stiboka, 1970; Van de Meene et al., 1988).

2.2.2 Geological setting – Holocene

Holocene aggradation in the western Netherlands has largely been governed by sea-level rise and associated groundwater-level rise. In the study area, this resulted in extensive organic accumulation (mainly peat), which in places was interrupted by periods of marine or fluvial sedimentation. The organic succession (Nieuwkoop Formation, Fig. 2.4) in the northwestern study area, north of Abcoude, commonly consists of a thin bed of mesotrophic *Carex* peat covered by eutrophic *Phragmites* peat, which, in turn, is overlain either by eutrophic wood peat or ombrotrophic *Sphagnum* peat (Stiboka, 1965; Poelman, 1966). East of the Vecht the organic succession is thinner (Fig. 2.3) and comprises mesotrophic *Carex* peat overlain by oligotrophic *Sphagnum* peat (Stiboka, 1965).

During the Atlanticum (~8700-5700 cal yr BP) (Fig. 2.3), as the rate of base-level rise gradually decreased, eastward-migrating beach barriers formed (Fig. 2.2) (e.g., Pons et al., 1963; Van der Valk, 1996). Initially, the beach barriers were segmented by a number of tidal inlets (Fig. 2.1), which enabled marine invasion of back-barrier areas and subsequent formation of estuarine and lagoonal deposits (Wormer Member, Naaldwijk Formation). Most of the tidal channels silted in during the Subboreal (~5700-2600 cal yr BP) (Fig. 2.1). After the closure of the Bergen tidal inlet ~3500 cal yr BP (calibrated radiocarbon dates from Westerhoff et al., 1987; Roep and Van Regteren Altena, 1988), the Oer-IJ remained

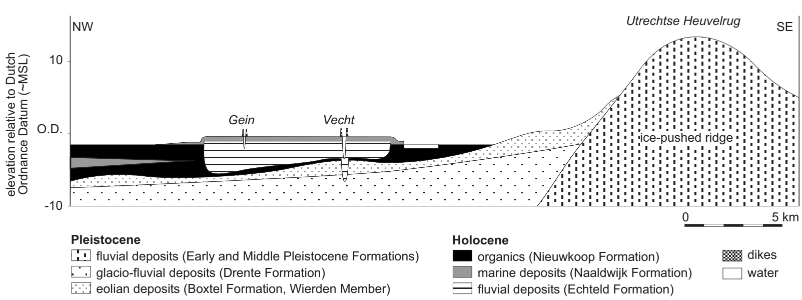


Figure 2.3. Schematic geologic cross section (mainly based on Stiboka, 1965; Van de Meene et al., 1988).

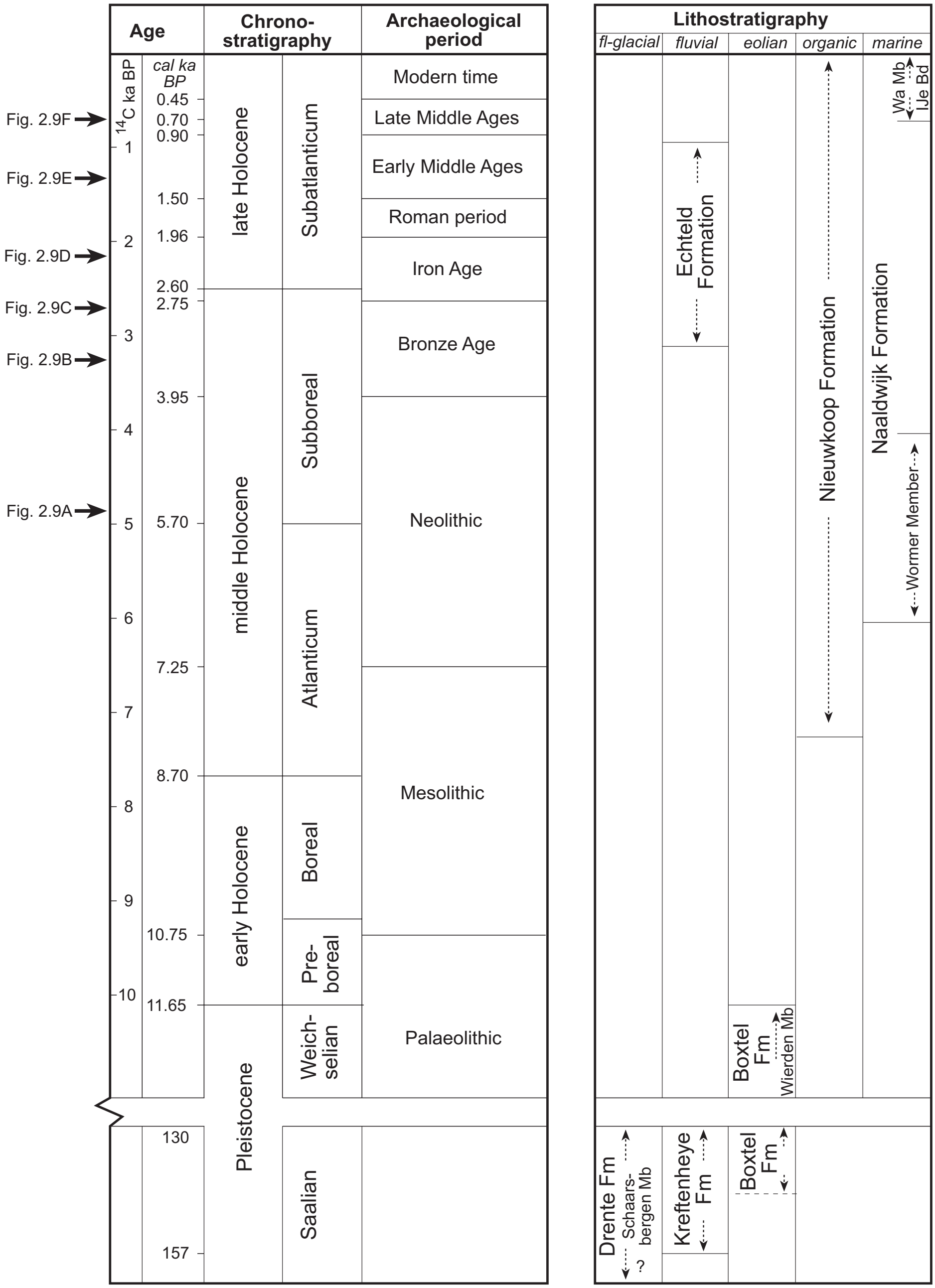


Figure 2.4. Chronostratigraphy, archaeological periods and lithostratigraphy valid for the study area. Chronostratigraphy following Van Geel et al. (1981) and Walker et al. (2009). Lithostratigraphy following Westerhoff et al. (2003b). Archaeological periods partly following Louwe Kooijmans (2005).

the single active inlet. Presumably after the closure of the Oer-IJ tidal inlet, 2270 ± 90 cal yr BP (Vos, 2008), a marine connection was established in the north, the so-called Vlie tidal inlet (Fig. 2.1) (Westerhoff et al., 1987; Vos, 2008). Between the 12th and 14th century AD, storm surges resulted in the enlargement of the Vlie and the formation of a lagoon, the so- called Zuiderzee. Under brackish conditions, a lagoonal clay bed of ~0.5 m (IJe Bed, Walcheren Member, Naaldwijk Formation) was deposited in the northern study area (Pons and Wiggers, 1960). Following damming in AD 1932 and subsequent land reclamations, the Zuiderzee was transformed into the freshwater lakes IJsselmeer, Markermeer and IJmeer.

Fluvial activity south of the study area started 6370 ± 70 cal yr BP (calibrated radiocarbon dates from Berendsen, 1982) when the Oude Rijn was formed (Fig. 2.1). The Oude Rijn was fed by a number of successive channels. The youngest channel, the Kromme Rijn, existed from 3200 ± 130 cal yr BP (calibrated radiocarbon dates from Berendsen, 1982) onward and connected to the Oude Rijn at Utrecht. An avulsion at that location ~2700 cal yr BP (Törnqvist, 1993a) resulted in the formation of the Vecht and Angstel Rivers. Weerts et al. (2002) suggest that natural sedimentation in the study area ceased between 1690 and 1540 cal yr BP.

The fluvial deposits in the Angstel-Vecht area (Echteld Formation, Fig. 2.4) comprise channel deposits (sand), natural levee deposits (sandy clay to silty clay), floodbasin deposits (clay) and organic-clastic lake fills (gyttja, clay, sandy/silty clay and sand). The channel sediment bodies of the Angstel and Vecht are partly embedded in Pleistocene deposits. In addition, the study area includes five organic-clastic lake fill sediment bodies.

2.2.3 Human activity

The first indications of structural landscape modification by humans within the Angstel-Vecht delta plain date to 11th century when increased population density forced agricultural exploitation of peat areas that fringed the fluvial system (Buitelaar, 1993, p. 152). In order to improve growing conditions drainage capacity was enlarged by a regular ditch network, which regulated groundwater levels. By the end of the 16th century most of the Angstel-Vecht delta plain was cultivated (Gottschalk, 1956).

The first recorded human regulation of the Vecht River dates from the first half of the 12th century (Manten, 2001, p. 261) an attempt was made to prevent flooding of the city of Utrecht by means of a dam near Breukelen. During the following centuries, the river courses in the study area became successively constrained by dikes. The ultimate protection against flooding from the Zuiderzee was realized with the building of a dam near Muiden in AD 1673/'74. The construction of dams and dikes had a large influence on sediment deposition in the area, as they prevented frequent and large scale flooding from the Zuiderzee during storm surges. Instead, erosion and sedimentation occurred during dike breach events at a more local scale.

Increasing demand for peat fuel after the 15th century resulted in significant landscape change (Gottschalk, 1956). Mining of the extensive ombrogenic peat, and to a lesser extent the mesotrophic *Carex* peat, resulted in the formation of numerous geometric lakes (e.g., the Vinkeveense Plassen and the Loosdrechtse Plassen) (Fig. 2.2). Subsequent groundwater lowering resulted in substantial subsidence of peat areas. The surface is presently > 1 m lower than the surface where peat is absent, including, for example, channel belt deposits (e.g., Fig. 2.5).

2.3 METHODS

We reconstructed the palaeogeographic development of the study area for six periods during the Holocene to assess the role of lakes and lake fills on the subsequent development of fluvial systems. The palaeogeographic reconstruction included three aspects. First, the sedimentary architecture, i.e. the composition and geometry of the deposits, was reconstructed and depicted in maps and cross sections. Second, at specific localities depositional environments within former lakes were reconstructed on the basis of microfossil analyses. Third, age control was obtained to reconstruct the period of existence of these environments or the period of sedimentation.

2.3.1 Architecture

To examine the sedimentary architecture of the Holocene deposits, we constructed a digital elevation model (DEM) of the underlying Pleistocene surface, a geomorphogenetic map (Fig. 2.5) and seven lithogenetic cross sections (Fig. 2.6). We used a high resolution DEM of the present land surface (Rijkswaterstaat-AGI, 2005) as well as borehole data. The latter comprised lithological information of ~3300 manual cores that are archived in the databases of Utrecht University (Berendsen, 2005) and TNO Geological Survey of the Netherlands (TNO, 2009). The DEM of the Pleistocene surface was constructed with simple kriging interpolation between interpreted borehole descriptions.

To compare levee development along channel sections upstream and downstream of lake fills, we measured the maximum thickness of natural levee deposits at nine cross sections (see Fig. 2.5 for locations). Cross sections were oriented perpendicular to the channel belt axis, with a maximum core spacing of 100 m. Interpretation of lithological cross sections resulted in the identification of, among others, natural levee deposits (consisting of silty to sandy clay), of which the maximum thickness could be measured. Further, we analyzed the lithological composition of 24 cores that penetrated the complete channel sediment body (see Fig. 2.5 for locations). In addition, to examine the influence of sedimentology (organics or organic-clastic lake fill) on the channel planform geometry, we calculated the sinuosity index of channel sections as a function of subsoil type. The sinuosity index (P_{ind}) is defined as the ratio between channel distance and channel belt axis distance (Brice, 1964). The sinuosity index is commonly utilized to distinguish meandering ($P_{ind} \geq 1.3$) or straight channels ($P_{ind} < 1.3$) (Makaske, 2001). Channel-belt length, channel length and subsoil type were identified on the geomorphogenetic map (Fig. 2.5). Where former channel positions could not be identified we utilized the minimum (along the present channel) and maximum (along meander bends) values for the channel length.

2.3.2 Environmental conditions

To determine local environmental conditions in the former lakes two sites were selected for pollen and diatom analyses, core Abcoude (B25G1057, pd1 in Fig. 2.5) and core Weesp (B25H0739, pd2 in Fig. 2.5). From core Abcoude, 41 pollen samples and 29 diatom samples were analysed (Fig. 2.7), and from core Weesp 33 pollen and 20 diatom samples were analysed (Fig. 2.8). Pollen samples were prepared largely following the procedure of Fægri and Iversen (1975). Diatoms samples were prepared following Cremer et al. (2001).

Sample intervals were chosen according to lithological variation and ranged between 1 and 51 cm for the pollen samples, and between 3 and 60 cm for the diatom samples. The sample depths are indicated in Fig. 2.7 and 2.8.

2.3.3 Age control

Age control was provided by radiocarbon dating, optically stimulated luminescence (OSL) dating, pollen analyses, and archaeological materials.

Radiocarbon dates were used to estimate the period of activity for fluvial distributaries following the principles outlined by Törnqvist and Van Dijk (1993). Preferentially, we based our results on AMS radiocarbon dates. The AMS method is considered more accurate than conventional bulk-sample analyses, and requires small samples of organic material. This enables the use of botanical macrofossils, thereby avoiding contamination of the sample by roots of younger vegetation (Törnqvist et al., 1992). For this study, 11 ^{14}C samples were collected for AMS dating (Fig. 2.5, Tab. 2.1). From these, terrestrial botanical macrofossils were selected (mainly Alnus seeds and catkins) to avoid ageing due to the hard water effect (e.g., Shotton, 1972). The ^{14}C-calibration curve (Reimer et al., 2004) and the OxCal v4.0.5 program (Bronk Ramsey, 1995, 2001) were used to calibrate ^{14}C ages to calendar ages and to calculate a combined value of ^{14}C ages that determine the same event. Further, we assumed that sedimentation of clastic material downstream of a lake is restricted as long as the lake interrupts the pathway of the fluvial channel as follows. Deltaic lakes trap sediment thereby preventing the formation of levee along downstream channels. The age of radiocarbon samples underlying levee deposits along such downstream channels thus indicates the moment that upstream lakes were filled.

OSL dating (e.g., Wallinga, 2002) was used to obtain time control on the various stages of organic-clastic lake fill formation. Six samples from three sites in the Aetsveldse organic-clastic lake fill (Fig. 2.5, Tab. 2.3), were analyzed in the Netherlands Centre for Luminescence dating (NCL). Here we provide only a short summary of the methods used; for additional information we refer to Wallinga and Johns (Wallinga and Johns, 2008). The sampled sediment was composed of intercalated sand and clay with varying laminae thicknesses. Dose rates were obtained from radionuclide concentrations determined by laboratory gamma ray spectrometry. In calculating the dose rates experienced by the grains we took into account the heterogeneity of the laminated deposits, attenuation by water and organic material, and a contribution from cosmic rays. We assumed water

Table 2.1. AMS ^{14}C dates in the study area. See text for explanation.

Laboratory nr[a]	*Nr*[b]	*^{14}C-age*	*Calendar age (cal yr BP)*	*River*	*DT*[c]	*Coordinates*[d] *(x/y) (m)*	*Surface elevation (m ± O.D.)*	*Depth below surface (cm)*	*Sample name*	*Dated and/or parent material*
UtC-01900[e]	1	2650 ± 80	2730 ± 230	Vecht	b	133405 / 460280	-0,10	43-46	Oud Zuilen I-1a	Alnus glutinosa fragments
UtC-01901[e]	2	2620 ± 50	2690 ± 170	Vecht	b	133405 / 460280	-0,10	43-46	Oud Zuilen I-1b	Alnus fragments
UtC-14573	13	2367 ± 44	2510 ± 190	Crevasse	er	130221 / 478555	-1,44	121-122	Weesp 1	tm[f]
UtC-14574	7	2920 ± 70	3100 ± 220	Spengen	b	123751 / 464139	-1,67	177-181	Spengen 2	tm[f]
UtC-14575	8	1577 ± 43	1460 ± 90	Angstel	er	129542 / 470254	-0,72	194-195	Loenen 1	tm[f]
UtC-14576	6	2414 ± 47	2530 ± 180	Spengen	er	124780 / 465251	-1,57	134-135	Noordeinde 1	tm[f]
UtC-14577	12	2420 ± 50	2530 ± 180	Crevasse	er	128141 / 477478	-1,50	133-134	Gein 1	tm[f]
UtC-14578	9	2853 ± 46	3000 ± 150	Angstel	b	128508 / 473158	-1,55	146-147	Angstel 1	tm[f]
UtC-14580	11	1694 ± 45	1580 ± 130	Vecht	er	124678 / 474147	-2,05	213-214	Winkel 5	tm[f]
UtC-14581	3	2280 ± 50	2260 ± 100	Vecht	b	129342 / 466569	-1,15	72-73	Weeresteijn 1	tm[f]
UtC-14582	4	2810 ± 50	2930 ± 140	Oud Aa	b	129342 / 466569	-1,15	217-218	Weeresteijn 2	tm[f]
UtC-14584	5	2870 ± 46	3010 ± 150	Oud Aa	b	125711 / 465662	-1,63	202-203	Portengen 3	tm[f]
UtC-14585	10	2352 ± 45	2440 ± 250	Vecht	b	130257 / 474979	-1,57	65-69	Vecht 1	tm[f]

a Bold numbers are new dates.
b In Fig. 2.5.
c DT = Date type: b = beginning of sedimentation; c = continuing sedimentation; e = end of sedimentation; er = end of sedimentation (sample from the base of abandoned channel fill).
d In Dutch coordinate grid (Rijksdriehoekstelsel).
e Published by Törnqvist (1993a).
f tm = terrestrial macrofossils

Table 2.2. Combined ages for the onset of clastic sedimentation by the Angstel-Vecht river system based on data from Törnqvist (1993a) (C1) and this study (C2).

nr	*Laboratory nr*	*Combined ^{14}C-age*	*Calendar age (cal yr BP)*	*Reference*
C1	UtC-14582 UtC-14584 UtC-14574	2857 ± 31	2970 ± 100	This study
C2	UtC-01900 UtC-01901	2630 ± 43	2730 ± 110	Törnqvist (1993)

saturation for the samples since the time of burial. This yields ~20% water by weight for sandy intervals; for finer-grained intervals we used water contents as measured in the laboratory. The equivalent dose was obtained from OSL measurements on small aliquots (2 mm) of sand-sized quartz (180-212μm) using the single-aliquot regenerative-dose (SAR) procedure (Murray and Wintle, 2003). A preheat of 220°C for 10s was selected, based on a three-sample preheat-plateau test. Data was accepted if the recycling ratio was within 20% of unity. Using this procedure a laboratory dose could be accurately retrieved (average dose recovery ratio 1.05 ± 0.02) and sensitivity changes were corrected for satisfactorily (average recycling ratio 1.00 ± 0.01). For each sample equivalent dose estimates were made until at least 22 aliquots passed the acceptance criteria. No evidence was found for incomplete resetting of the OSL signal. Optical ages were calculated by dividing the measured equivalent doses by the calculated dose rates. Results are presented with a one sigma uncertainty (standard error).

Pollen-based age control on fluvial deposits is much less accurate than ^{14}C dating. In some instances, however, pollen of a specific species can act as a regional marker. This is the case with Juglans (walnut). This species was introduced in southern and central Europe during the Late Iron Age (150-12 BC; 2200-1962 yr BP) but became widespread in Europe during the Roman period in association with the expansion of the Roman Empire (Zohary and Hopf, 2000). It is assumed that Juglans did not exist in the Rhine catchment before the Roman period (i.e before 1962 yr BP) and therefore, deposits in the Rhine catchment that contain Juglans pollen are considered to have been formed since the Roman Period.

Prehistoric human habitation atop natural levees is considered contemporaneous with fluvial activity or later (Louwe Kooijmans, 1974; Berendsen, 1982; Berendsen and Stouthamer, 2001). More specifically, post-neolithic settlement atop natural levees is considered to begin after the period of maximum activity of the associated fluvial distributary (L.P. Louwe Kooijmans and W. Van Zijverden, personal communication). In this study we estimated the period of prehistoric habitation on various fluvial channels based on the Dutch Rijksdienst voor Archeologie, Cultuurlandschap en Monumenten (RACM, the present RCE). This database contains information on archaeological artefacts (e.g., age and position) in The Netherlands (Tab. 2.4).

Table 2.3. Results of 6 OSL-dated samples. Note that sample 'I' is somewhat older than a ^{14}C sample (UtC-14577) that has been taken from approximately the same location. The same holds for sample VI that is slightly older than a ^{14}C sample (UtC-14573).

Nr[a]	*NCL lab-nr*[b]	*OSL age ± 1σ (ka)*[c]	*Dose rate (Gy/ka)*	*Equivalent dose (Gy)*	*Coordinates*[d] *(x/y) (m)*	*Surface elevation (m±O.D.)*	*Depth below surface (cm)*	*Relevance*
I	3206026	2.87 ± 0.16	1.73 ± 0.07	4.91 ± 0.15	128144/ 477486	-1.10	228-240	Last active phase of Gd1
II	3206022	3.49 ± 0.21	1.64 ± 0.07	5.22 ± 0.14	128526/ 477358	-1.30	406-446	Suspended load input in lake
III	3206023	3.16 ± 0.17	1.76 ± 0.07	5.51 ± 0.16	128526/ 477358	-1.30	324-357	Suspended load input in lake
IV	3206024	2.80 ± 0.13	1.69 ± 0.06	4.74 ± 0.12	128526/ 477358	-1.30	234-239	Active phase of Gd1
V	3206025	2.83 ± 0.14	1.68 ± 0.06	4.76 ± 0.15	128526/ 477358	-1.30	163-176	Active phase of Gd1
VI	3206027	2.45 ± 0.12	1.70 ± 0.06	4.13 ± 0.15	130220/ 478551	-1.35	157-170	Last active phase of Gd2

a In Fig. 2.5.
b Laboratory number of the Netherlands Centre for Luminescence Dating (NCL), Delft University of Technology.
c Analyses carried out in 2006.
d In Dutch coordinate grid (Rijksdriehoekstelsel).

The onset of organic accumulation on the underlying Pleistocene deposits is approximated by the moment of drowning. It is assumed that this moment depends only on the elevation of the Pleistocene surface. Holocene palaeogroundwater levels (Cohen, 2005), in combination with the DEM of the Pleistocene surface, were used to estimate the moment of drowning. Palaeogroundwater levels are only available for the southern study area, but because of the similarity in hydrological setting, are assumed similar for the northern portion of the study area.

Table 2.4. Selection of archaeological artefacts that have been found in the study area (from the RCE Archis-database).

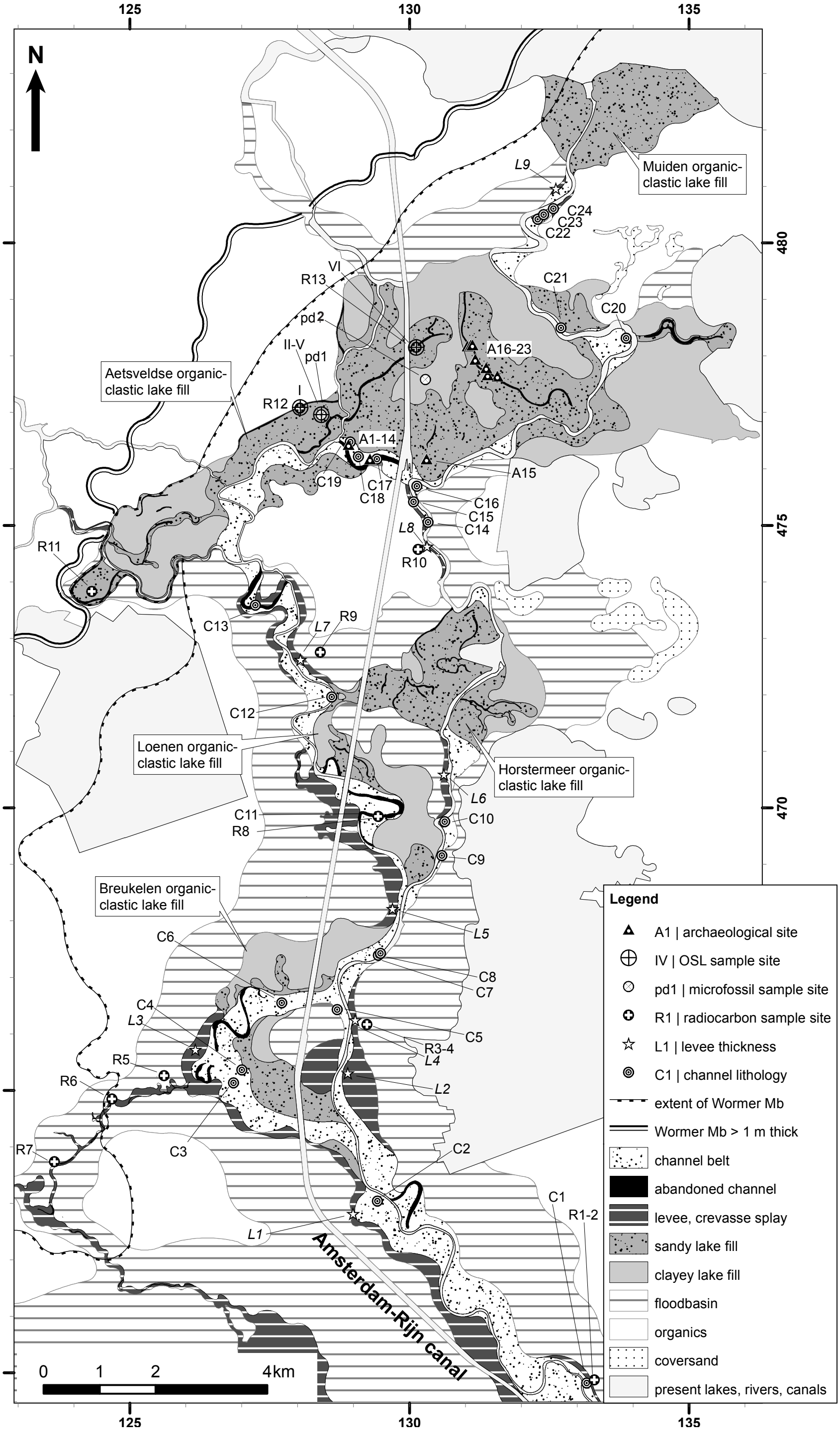
Nr[a]	*ROB-nr*[b]	*Coordinates*[c] *(x/y) (m)*	*Depth below surface (cm)*	*Archaeological period*	*Age (yr BP)*
A1	w-104644	129217/476419		Late Iron Age	2200-1962
A2	w-104700	129455/476542		Late Iron Age	2200-1962
A3	w-104679	129396/476581		Late Iron Age	2200-1962
A4	w-104661	129316/476608		Late Iron Age	2200-1962
A5	w-104677	129388/476609		Late Iron Age	2200-1962
A6	w-104674	129371/476635		Late Iron Age	2200-1962
A7	w-104673	129360/476642		Late Iron Age	2200-1962
A8	w-104692	129438/476656		Late Iron Age	2200-1962
A9	m-1913	129380/476661	10-100	Late Iron Age	2200-1962
A10	w-104666	129335/476674		Late Iron Age	2200-1962
A11	w-104668	129342/476769		Late Iron Age	2200-1962
A12	w-48510	129100/476800	0-100	Late Iron Age	2200-1962
A13	w-104613	129031/476836		Late Iron Age	2200-1962
A14	w-104611	129012/476964		Late Iron Age	2200-1962
A15	m-11575	130409/476581		Early Iron Age	2650
A16	w-138317	131675/478050	0-50	Middle Iron Age	2450-2200
A17	m-10892	131500/478071		Late Iron Age	2200-1962
A18	m-1290	131475/478199		Middle-Late Iron Age	2450-1962
A19	m-1908	131280/478345		Late Iron Age	2200-1962
A20	m-10891	131138/478579		Late Iron Age	2200-1962
A21	w-15207	131200/478600		Middle Iron Age	2450-2200
A22	w-33987	131200/478600	25-55	Middle Iron Age	2250-2150
A23	m-1289	131229/478602		Middle Iron Age	2250-2150

a In Fig. 2.5.
b ROB nrs refer to the database of the Rijksdienst voor het Oudheidkundig Bodemonderzoek (ROB), the present Rijksdienst voor het Cultureel Erfgoed (RCE).
c In Dutch coordinate grid (Rijksdriehoekstelsel).

2.4 PALAEOGEOGRAPHIC DEVELOPMENT

2.4.1 ~5500 cal yr BP (3550 BC)

Approximately 5500 cal yr BP the study area was characterised by surficial Pleistocene deposits in the east *(1, italic* numbers refer to locations in Fig. 2.9) and wetland *(2)* and lagoonal environments *(3)* in the west (Fig. 2.9A). Fluvial activity was absent during this period.


N
125
130
135
480
475
470
465
460
Muiden organic-clastic lake fill
L9
C24
C23
C22
C21
C20
VI
R13
pd2
II-V
pd1
A16-23
Aetsveldse organic-clastic lake fill
I
R12
A1-14
A15
C19
C17
C18
C16
C15
C14
L8
R10
R11
C13
L7
R9
C12
Loenen organic-clastic lake fill
Horstermeer organic-clastic lake fill
L6
C11
R8
C10
C9
Breukelen organic-clastic lake fill
L5
C6
C8
C7
C4
L3
C5
R3-4
L4
R5
L2
R6
C3
R7
C2
L1
Amsterdam-Rijn canal
C1
R1-2
0 1 2 4km
Legend
A1 | archaeological site
IV | OSL sample site
pd1 | microfossil sample site
R1 | radiocarbon sample site
L1 | levee thickness
C1 | channel lithology
extent of Wormer Mb
Wormer Mb > 1 m thick
channel belt
abandoned channel
levee, crevasse splay
sandy lake fill
clayey lake fill
floodbasin
organics
coversand
present lakes, rivers, canals

Soil development occurred atop sandy Pleistocene deposits, until drowned by rising groundwater. Rising groundwater that was controlled by rising relative sea-level resulted in the accumulation of organics (mainly peat) on the Pleistocene surface. Peat accumulation was initially restricted to the western part of the study area where the groundwater level was elevated above the Pleistocene substratum ~8000 cal yr BP. As the groundwater level rose, the eastern limit of organic accumulation steadily moved to the east, and by ~5500 cal yr BP peat already covered over half of the study area (Fig. 2.9A). For most of the study area peat accumulation continued until groundwater levels were artificially lowered, beginning in the Middle Ages. In general the peat-forming vegetation can be categorized in north-south oriented zones that shifted eastwards following sea-level rise. The eastern zone was a mesotrophic *Carex* wetland (Stiboka, 1970; Bakker, 1976), whereas in the western part of the area extensive reed marshes developed under the influence of marine inundations.

At ~6500 cal yr BP inland-moving beach barriers became stabilized near the present shoreline (Fig. 2.1) (Beets et al., 2003). Through tidal inlets, marine sediments were deposited in the back-barrier area. Between ~6900 and ~4700 cal yr BP (Pons and Wiggers, 1959; Van de Meene et al., 1988), marine deposits (Naaldwijk Formation, Wormer Member, Fig. 2.4) formed in the northwestern study area *(3)*, decreasing in thickness towards the southeast (Fig. 2.5). The marine deposits are key stratigraphic markers that form a clastic wedge intercalated with the organic succession, indicating alternating periods of increased and reduced marine influence.

2.4.2 ~3500 cal yr BP (1550 BC)

Approximately 3500 cal yr BP the study area was comprised of surficial Pleistocene deposits *(4)*, extensive wetlands with active peat development *(5)*, and lakes *(6)* (Fig.

← Figure 2.5. Geomorphogenetic characterization of the study area. The map location is indicated in Figure 2.1. The locations of ^{14}C-samples, OSL-samples, pollen and diatom sample cores and the positions of archaeological finds. Numbers on the map correspond with numbers in Tab. 2.1 (^{14}C-dates, nrs R1-13), Tab. 2.3 (OSL-dates, nrs I-VI) and Tab. 2.4 (archaeological artefacts, nrs A1-23). The pollen and diatom sample locations pd1 and pd2 correspond to the cores Abcoude and Weesp respectively. Note that a few cores were sampled for multiple analyses (e.g., samples for both OSL and radiocarbon dating in an abandoned channel in the Aetsveldse organic-clastic lake fill) and that those locations are indicated with overlapping sample symbols. Furthermore, the locations where the thickness of natural levees has been determined are shown (labelled in italics and with "L" as prefix) as well as the locations where the channel sediment lithology (labelled with "C" as prefix) has been determined.
The top of the marine deposits (Wormer Member, Naaldwijk Formation) in the north-west of the map present at ~6 m –O.D., and in general a few meters below the surface. Note that the channel planform of fluvial channels in the peat area, for instance north of the Horstermeer organic-clastic lake fill, are predominantly straight, whereas they are meandering where organic-clastic lake fills are present. See Addendum 1 for a full-colour version.

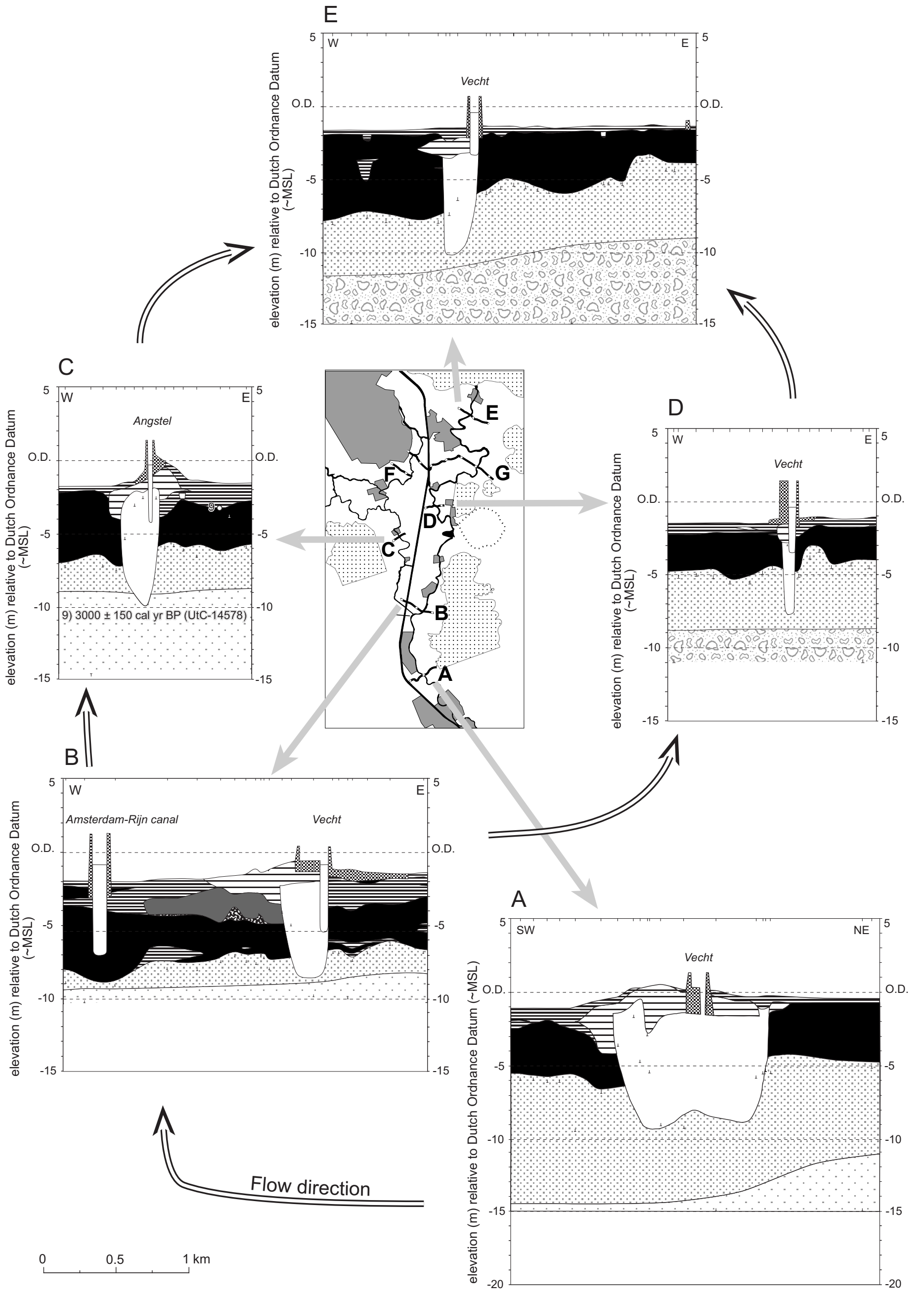


Figure 2.6. Cross sections A-G. The position of the cross sections is indicated in Fig. 2.2. The cross sections associated with the oldest river course are shown, in downstream direction, in A, B, C and E. Cross section D shows a channel belt that formed later, shortly before fluvial activity decreased.

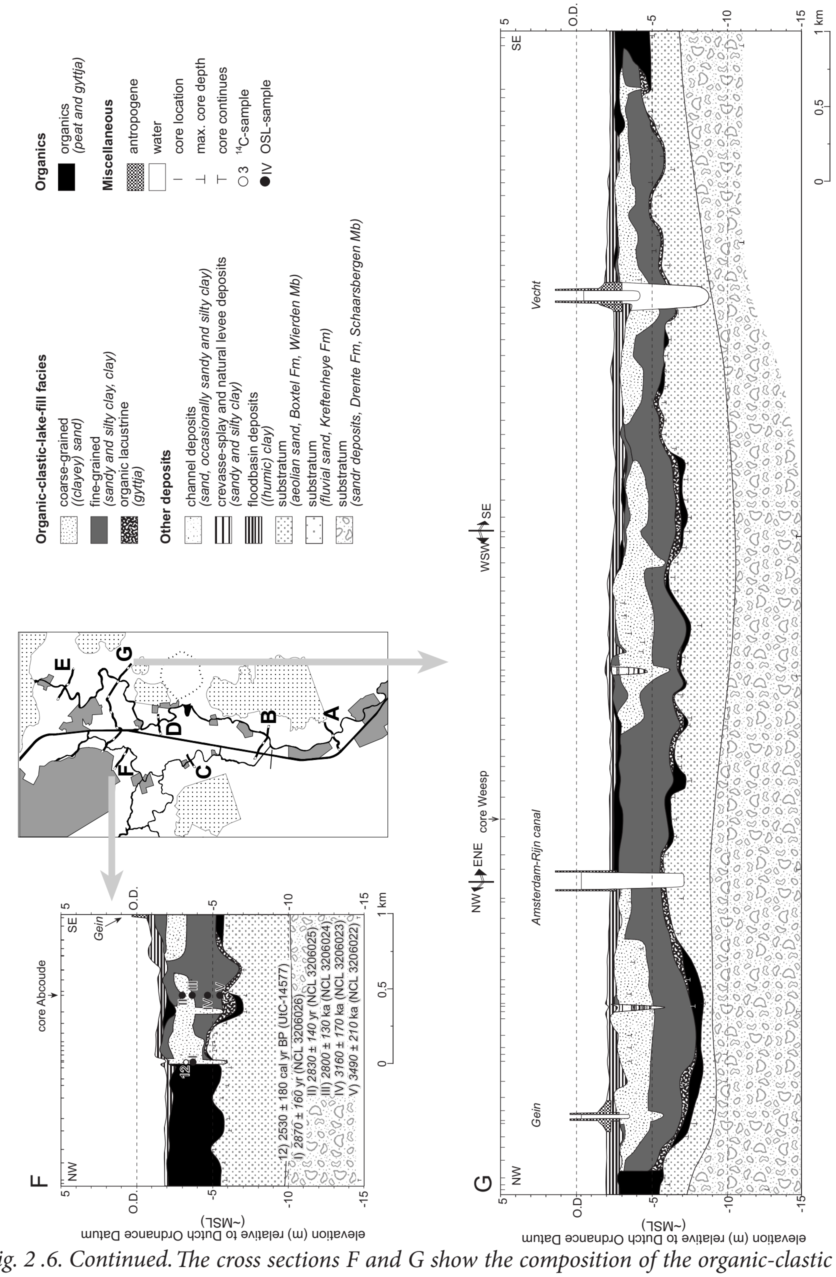


Fig. 2 .6. Continued. The cross sections F and G show the composition of the organic-clastic lake fills. The dimensions of the channel-belt sediment body of the oldest course changes downstream. The channel belt is relatively wide in cross section A, whereas it is narrow in cross section E. The younger channel belt that is shown in cross section D, is even narrower. See Addendum 2 for a full-colour version.

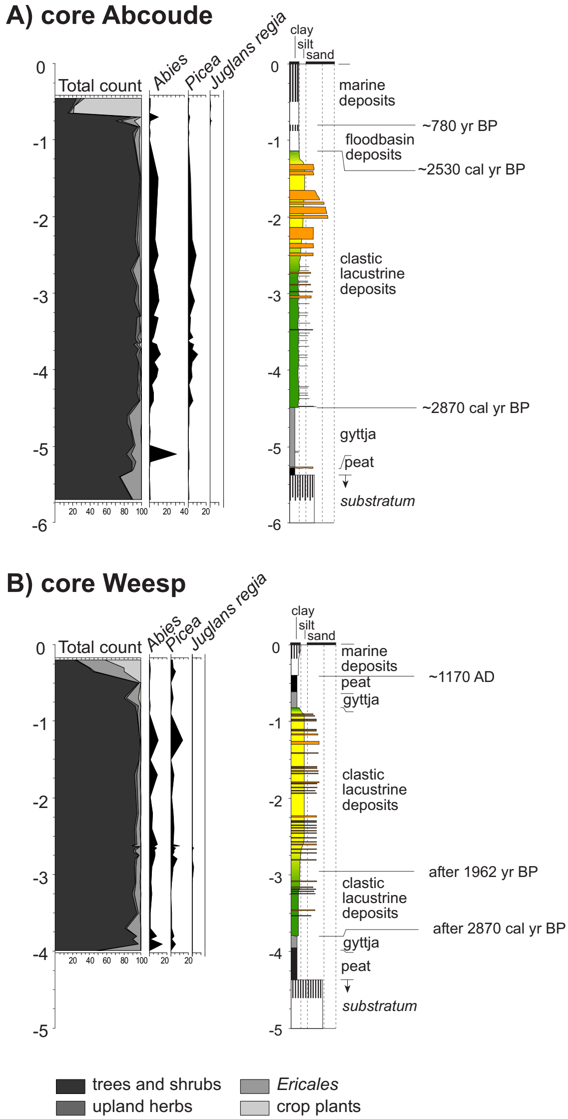


Figure 2.7. Pollen diagrams of selected species for A) core Abcoude and B) core Weesp (locations are indicated in Fig. 2.5). The first clastic input corresponds to a strong increase of fluvial transported pollen such as Picea and Abies from the hinterland, e.g., Schwarzwald, Germany, and Vosges, France.

2.9B). Further, a small or intermittent connection to the Rhine became established *(7)*.

The area of peat accumulation enlarged after 5500 cal yr BP in the west where marine regression provided conditions for peat accumulation and in the east where the limit of peat accumulation shifted landward *(8)* in response to raised groundwater levels. Bog peat (*Sphagnum*) formed in a zone parallel to and eastward of the present Vecht River (Bennema, 1951; Poelman, 1966) (Fig. 2.2), succeeding *Carex* peat. Bog peat also accumulated west of the present Angstel (Fig. 2.2). Between these two areas, a ~6 km

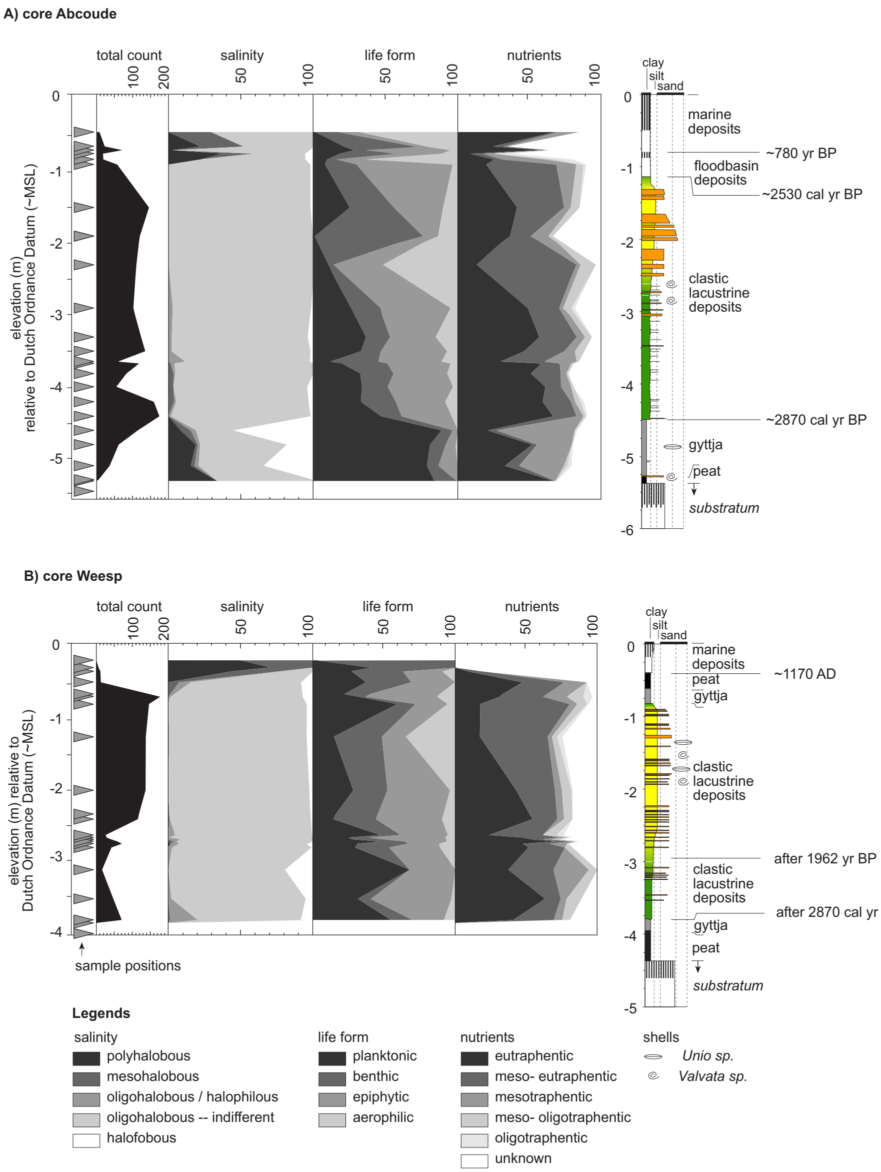


Figure 2.8. Diagram that shows the relative abundance of groups of diatoms with similar salt-tolerance ranges, life form, and nutrient-tolerance ranges. The position of Unio sp. and Valvata sp. is also indicated.

wide corridor of fen peat accumulated (Bennema, 1951; Poelman, 1966). The latter is dominated by wood peat and reed peat (Poelman, 1966) and influenced by nutrient-rich fluvial-derived water (Pons and Van Oosten, 1974).

Eutrophic freshwater lakes (Fig. 2.8, 2.9B) within the wetlands existed prior to the onset of fluvial sedimentation, as inferred from the presence of organic lake sediment (gyttja and detritus) underlying fluvial deposits (Fig. 2.6F, G). The freshwater lake conditions are indicated by the diatom association, and especially by the presence of shells (*Valvata sp.* and *Unio sp.*) that require a freshwater environment. Although the moment of lake formation is not dated directly, it is likely that a period of peat accumulation existed prior to the formation of the lakes. This assertion is supported by the presence of peat underlying gyttja that formed in the lakes (for example see Fig. 2.6F, G), as noted in the western Aetsveldse organic-clastic lake fill. In this lake basin, the eastern limit of marine deposits (Wormer Member, Fig. 2.5) is truncated by organic-clastic lake fills, indicating formation after cessation of marine sedimentation at ~4700 cal yr BP. If the lake had previously existed marine diatoms would have been more prominent in the gyttja.

The relatively minor Winkel, Holendrecht and Gaasp Rivers are possible remainders of a local drainage network that would have been required to drain seepage and precipitation-derived waters from the Angstel-Vecht delta plain. These local channels also drained intermittent fluxes of Rhine water. This is indicated by the presence of *Picea* and *Abies* pollen in the gyttja (Fig. 2.7) for which the only available source at that time was the upper reaches of the Rhine catchment, i.e. Schwarzwald and Vosges (Lang, 1994). A marine signal is recorded by the diatom assemblage in the gyttja deposits ('salinity' panels in Fig. 2.8A-B), which suggests the presence of a drainage pathway to a tidal inlet. Along this 55 km long channel (measured from core Abcoude to the beach barriers along the nearest tidal inlet, the Oer-IJ, as mapped by Pons and Wiggers, 1959) large lakes were probably absent, as key marine derived evidence, such as marine derived diatom skeletons, are not recorded.

2.4.3 ~2800 cal yr BP (850 BC)

By ~2800 BP the Vecht-Angstel system functioned as a Rhine branch, resulting in the formation of fluvial deposits and organic-clastic lake fills (Fig. 2.9C).

AMS radiocarbon dates at three locations indicate that fluvial sedimentation in the upstream study area initiated in 2970 ± 100 cal yr BP (C1 in Tab. 2.2). This differs from two AMS ^{14}C dates published by Törnqvist (1993a), based on *Alnus*-wood remains from a single location, which set the beginning of fluvial sedimentation somewhat later: 2730 ± 110 cal yr BP (C2 in Tab. 2.2). Age differences could be explained by downstream prograding sedimentation, which implies that upstream samples yield older dates than downstream sites. Although sampled from upstream locations, we find the dates reported by Törnqvist (1993a) to be younger. Alternatively, the age difference can be explained by the nature of the dated material. In palaeorecords it is nearly impossible to discern Alnus-wood remains from *Alnus*-root remains (Schweingruber, 1978), and it thus cannot be excluded that *Alnus* root is dated, which obviously would return a younger date than the beginning of fluvial activity. We therefore regard the ages for the onset of fluvial sedimentation obtained by Törnqvist (1993a) as too young. OSL ages of the fine-grained clastic sediments in the Aetsveldse organic-clastic lake fill, 3490 ± 210 and 3160 ± 170 (NCL-3206022, -3206023, with 1-σ, Fig. 2.6F, Tab. 2.3) are too old, although the latter overlaps the radiocarbon age for beginning of fluvial sedimentation. Indications for incomplete resetting of the OSL-signal are lacking. Thus, we explain the overestimation of the OSL age by lithological heterogeneity of the samples which introduces uncertainties concerning β- and γ-radiation and water-content estimations (Wallinga, personal communication). Overestimation of the water content by 1%, for instance, may result in overestimation of the OSL age by 1% (Wallinga, 2002).

The initial river course followed parts of the present Vecht *(10, 11)*, the Oud Aa *(12)* and the Angstel *(13)* and presumably drained towards Muiden *(14)*. The channel-belt deposits associated with this initial river course are the widest (Fig. 2.6A-C) and most coarse-grained (Fig. 2.10B). In general, the planform pattern of the fluvial channels changes from upstream meandering (see, for example, the section between *10* and *11* in Fig. 2.9C) to essentially a straight channel in the downstream segment (the channels downstream of *13* in Fig. 2.9C). Fluvial channels that traverse organic-clastic lake fills, however, tend to exhibit a meandering planform *(15)*.

The lakes were successively filled with clastic sediment supplied by the fluvial system (Fig. 2.5, 2.9C). The Breukelen and Loenen organic-clastic lake fills presumably formed within several decades, as can be inferred from similar upstream dates for the onset of sedimentation (2970 ± 100, C1 in Tab. 2.2) and downstream of these former lakes (3000 ± 150 cal yr BP, UtC-14578, Tab. 2.1).

After the Breukelen and Loenen Lakes had been filled with clastic material, the Angstel channel transported sediment further downstream, to the Aetsveldse Lake *(16)*. From the point where the river debouched into the Aetsveldse Lake, initially located near Abcoude, lake infilling prograded in a north-eastern direction, towards the primary exit that was located near Weesp (Fig. 2.2). The progradation direction is indicated by the position and orientation of a few small distributary channels (*16*, Fig. 2.9C). Although the exact timing of the initiation of these distributary channels is not known, their formation has been diachronous and related to the progradation stage of the organic-clastic lake fill. As a new river course enters a lake it deposits most of its sediment close to the river mouth. In response, water depth decreases and channelized flow transports clastic material through a network of small channels further into the lake. These channel sections close to the initial river mouth therefore are oldest whereas the downstream part formed at a later stage. The upstream segments of the distributary channels in the Aetsveldse Lake are therefore assumed to have formed shortly after the onset of sedimentation in the Aetsveldse Lake, whereas the downstream segments were likely formed shortly before they were abandoned.

2.4.4 ~2200 cal yr BP (250 BC)

By ~2200 cal yr BP, two avulsions had occurred, fluvial sedimentation had ceased and the lakes were largely filled (Fig. 2.9D).

Approximately 2440 ± 250 cal yr BP (UtC-14585, Tab. 2.1) an avulsion *(17)* reduced discharge through the Angstel in favour of a more eastward new channel. This channel supplied sediment to the Horstermeer Lake *(18)* and continued towards the remainders of the Aetsveldse Lake *(25)*. The levee deposits downstream of the Horstermeer Lake *(19)* are thin relative to those upstream (Fig. 2.10). This implies that the lake was not completely filled before sediment supply to the Horstermeer Lake diminished. Approximately 2260 ± 100 cal yr BP (UtC-14581, Tab. 2.1), an avulsion resulted in another channel *(20)* formed parallel to the Oud Aa (Fig. 2.2). The natural levees along this channel are also relatively thin (L2 and L4 in Fig. 2.10A). The immaturity of the new channels *(17, 19* and *20)* is

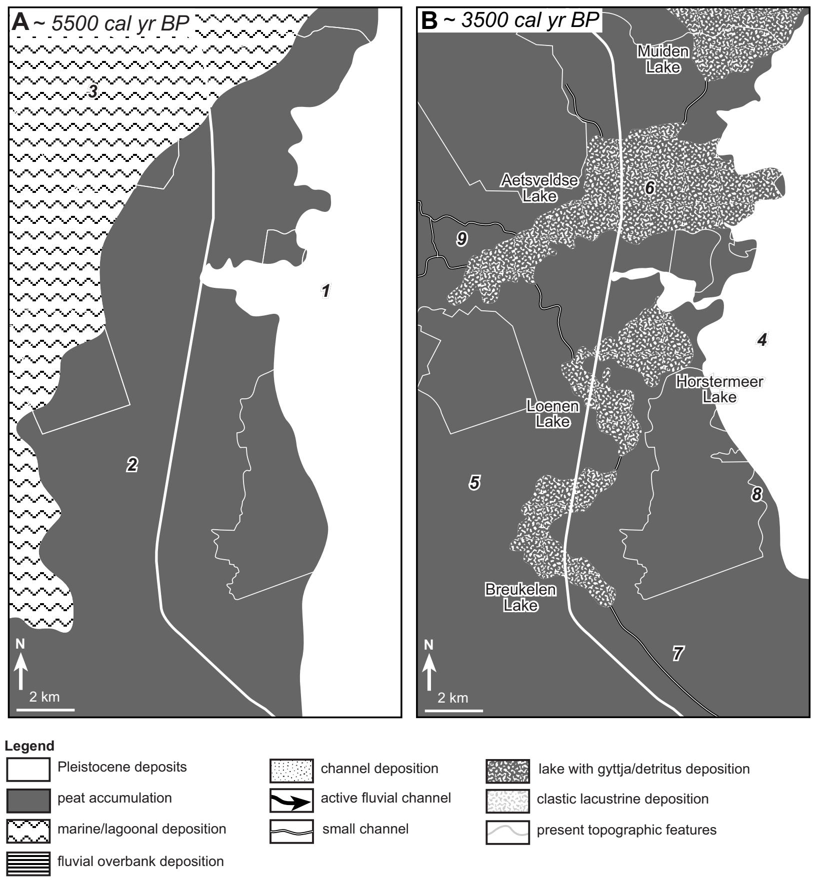
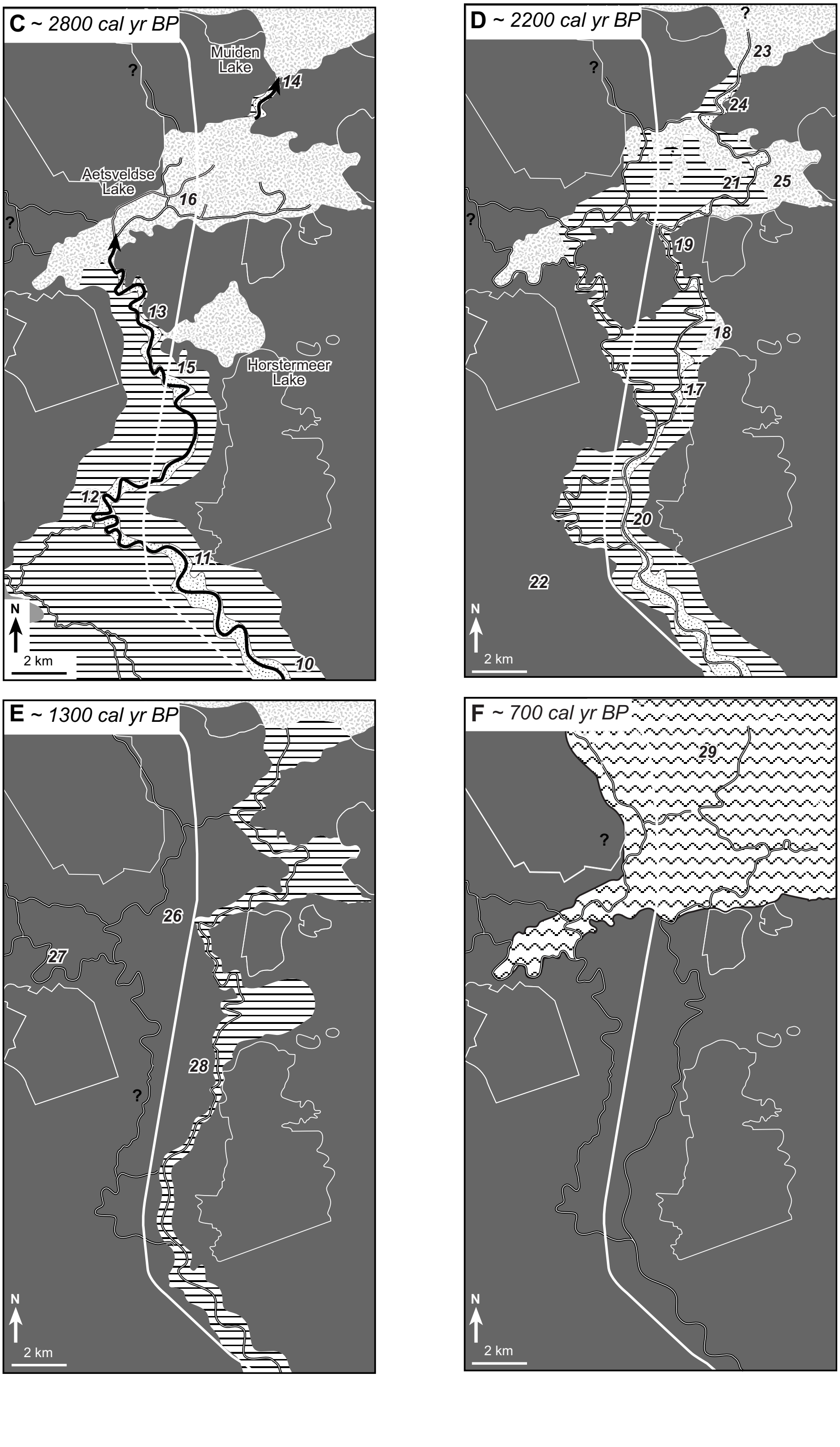


Figure 2.9. (C-E on next page) Holocene palaeogeographic maps of the Angstel-Vecht area. Italic numbers are referred to in the text. See text for further explanation. A. Palaeogeography of ~5500 cal yr BP. B. Palaeogeography of ~3500 cal yr BP. C. Palaeogeography of ~2800 cal yr BP. D. Palaeogeography of ~2200 cal yr BP. E. Palaeogeography of ~1300 cal yr BP. F. Palaeogeography of ~700 yr BP. See Appendix 8 for a full-colour version.


C ~ 2800 cal yr BP
Muiden Lake
Aetsveldse Lake
Horstermeer Lake
?
14
16
13
15
12
11
10
N
2 km
D ~ 2200 cal yr BP
?
23
24
21
25
19
18
17
20
22
N
2 km
E ~ 1300 cal yr BP
26
27
28
?
N
2 km
F ~ 700 cal yr BP
29
?
N
2 km

illustrated by the thin natural levees and the fine-grained composition of the channel sediment (Fig. 2.10B), as well as the narrow channel sediment body (Fig. 2.5, 2.6D). We, therefore, contend that shortly after formation of the new channels, fluvial activity in these channels was strongly reduced.

The reduction of discharge and sediment through the Angstel is inferred from the formation of a new channel *(17)* and is also recorded in the fluvial record downstream of the Angstel. The distributary channels in the Aetsveldse Lake *(16)*, for example, were abandoned between 2530 ± 180 cal yr BP (UtC-14576, Tab. 2.1) and 2200 yr BP (A21, Tab. 2.4) in favour of a single-channel course traversing the eastern Aetsveldse organic-clastic lake fill *(21)* (Fig. 2.9D). The OSL ages that indicate the last active period of distributary channel activity (2870-2800, NCL-3206024-3206026, Tab. 2.3) are older than the abandonment (2530 ± 180 cal yr BP, UtC-14577, Tab. 2.1). The abandonment of the distributary channels between 2650 cal yr BP and 2200 BP is probably the result of strongly reduced fluvial activity of the Angstel. Concurrently, fluvial activity ceased in the southwestern portion of the study area *(22*, compare Fig. 2.9C, D) (2530 ± 180 cal yr BP, UtC-14576, Tab. 2.1).

The presence and sandy character of the Muiden organic-clastic lake fill *(23)* seems to conflict with the relatively thin levees (L9 in Fig. 2.10A) and fine-grained character of the upstream channel sediment (C22-24 in Fig. 2.10B) *(24)*. The prominence of sandy facies in the Muiden organic-clastic lake fill points to a significant influx of fluvial bed-load sediment. The thin levees along this channel, in contrast, indicate limited sediment transport to the Muiden Lake. Another upstream source or even a downstream source for the sediment in the Muiden organic-clastic lake fill, however, is not available. We contend, therefore, that sediment was supplied to the channel for a very short period, i.e. few decades. This period was sufficient to produce the Muiden organic-clastic lake fill, but was insufficient to produce thicker levees.

Because fluvial activity was strongly reduced prior to 2200 yr BP, both in the parallel-running Angstel and new channel *(17, 19),* the Rhine effectively had abandoned the study area prior to the system shut down. Fluvial sedimentation continued at low rates and predominantly occurred in the remainders of the Aetsveldse Lake. The presence of *Juglans regia* pollen (Walnut) in the remainder of the Aetsveldse Lake deposits (core Weesp, Fig. 2.5, 2.8B) implies that these sediments were deposited during or after Roman occupation (Zohary and Hopf, 2000), i.e. after 1962 yr BP.

2.4.5 ~1300 cal yr BP (650 AD)

By about 1300 cal yr BP, the Rhine influence had substantially diminished, as organics covered the downstream fluvial deposits (Fig. 2.9E).

Prior to the Roman period, discharge and sediment supply to the main downstream portion of the Angstel *(26)* was presumably further reduced. This is illustrated by the presence of a Late Iron Age (2200-1962 yr BP) settlement on natural levees east of Abcoude *(26)* (A1-14 in Tab. 2.4), which indicates that channel activity decreased prior to the Roman Period (i.e. before 1962 yr BP). Despite reduced discharge and sediment transport, the fluvial system did not completely silt up. River-transported pollen (e.g., *Abies* and *Picea*) is present in the upper Aetsveldse organic-clastic lake fill sequence (Fig. 2.7A, B), which suggests a connection with the Rhine system during the Roman period. The Angstel and the parallel Vecht channel coexisted until 1460 ± 90 cal yr BP (UtC-14575, Tab. 2.1) when the Angstel was finally abandoned. The activity of the Angstel before that time is evidenced by continuing clastic deposition in the Aetsveld Lake (at location *27*), close to the downstream end of the Angstel, until at least 1590 ± 140 cal yr BP (UtC-14580, Tab. 2.1). Around 1300 yr BP (Fig. 2.9E), however, this portion of the lake no longer received clastic sediment, as organics started to accumulate.

At approximately 1300 yr BP the Vecht River remained open and connected to the Rhine River (Fig. 2.9E) as induced from two observations. First, a palaeohydrological modelling study indicates that between AD 1 and 1350, seepage produced average discharge volumes of ~2 m^3s^{-1} (Van Loon et al., 2009), suggesting the presence of a small channel. From the 6th century onwards, trade became increasingly important between cities along the Rhine (e.g., Dorestad, presently: Wijk bij Duurstede) and northern regions (e.g., Scandinavia), with the Vecht representing an important link in the trade route (Besteman et al., 1993). The reconstruction of a drainage course north of Muiden is hampered by later erosion of the deposits. The channel must have drained to the Vlie tidal inlet, however, because by this time the Oer-IJ inlet was closed. The absence of marine diatoms from this period indicates that there was no marine influence in the study area.

After the Angstel was abandoned at ~1460 cal yr BP, the Vecht *(28)* was a single channel. As a result of reduced sediment input, organics started to accumulate over the organic-clastic lake fill deposits in the west (Fig. 2.6F, G).

2.4.6 ~700 yr BP (1250 AD)

Rhine influence definitely ceased AD 1122, after the Kromme Rijn was dammed at Wijk bij Duurstede (Dekker, 1980) (Fig. 2.1). The channels in the study area, however, continued to function as local drainage systems (Fig. 2.9F). After AD 1170 (Gottschalk, 1971), the northern part of the study area was periodically inundated by Zuiderzee storm surge events (29). This is evidenced by, amongst others, the catch of a marine pout (Dutch: bolk) in the canals of Utrecht during a severe storm event in AD 1170 (based on historic sources examined by Gottschalk, 1971, p. 80-94). As a result of the inundations, clastic deposits (IJe Bed, Fig. 2.4) formed in the northern part of the study area under brackish conditions (Fig. 2.9F). A brackish diatom association in the topsoil at two locations in the Aetsveldse Polder ('salinity' panels in Fig. 2.8) indicates the influence of a marine environment during deposition. The only source for brackish conditions at that time is the Zuiderzee (Fig. 2.9F). Further, a brackish diatom association in the channel fill of the Aetsveldse distributary channel (Stiboka, 1966) points to channel reactivation during marine inundations. Other distributary channels in the Aetsveldse Polder, however, were abandoned much earlier (2530 ± 180, UtC-14576 and 2510 ± 190, –14573, Tab. 2.1) and reveal that not all distributary channels were reactivated during marine inundations. It remains difficult, however, to validate the supposed marine nature of the topsoil deposits because pedogenic processes have obscured macroscopic characteristics. Exceptions occur where organic-clastic lake fills in the subsoil are separated from the marine clay layer by an organic soil horizon.

2.5 DISCUSSION

2.5.1 The origin and position of the lakes in the Angstel-Vecht area

Lakes and ponds are a common element in modern mires (e.g., Fairchild and Eidt, 1993; Heino, 2000). They may be a remnant of an older and larger lake that was reduced in size by peat accumulation. Alternatively, they may form when small water-logged depressions amalgamate into larger water bodies (Succow and Joosten, 2001, p. 196). Eventually, such a water body may reach an area and depth such that wind-driven waves become effective agents of erosion, further increasing the size of lakes. Knowledge of the mechanisms that control the formation of lakes, however, remains limited.

The palaeolakes of the Angstel-Vecht area developed in a mesotrophic environment. The lakes were erosional elements, and evidence for terrestrialization of the lacustrine environment is lacking. This implies that after the lakes formed, erosion was dominant over organic accumulation, probably as the result of wind-driven waves. In prior times, for instance, the Naardermeer extended northeastwards (Heydanus, 1877), which is in accordance to the dominant southwestern wind direction. This wind direction previously has been inferred from inland aeolian dune morphology that developed from the beginning of the Neolithic to the present (Koster, 1982; Castel et al., 1989).

The formation of the lakes is likely controlled by the groundwater discharge from the ice –pushed ridge. The lakes in the Angstel-Vecht area are not randomly distributed, but positioned at approximately equal distances from the ice-pushed ridge. This suggests a causal relation between lake formation and the ice-pushed ridge, which is elucidated as follows. Seepage from the ice-pushed ridge increased the quantity of surface water (Van Loon et al., 2009) and greatly changed the water quality (Wassen et al., 1989). Both factors influenced the conditions for plant growth and, subsequently, botanical composition of the peat. Seepage-dominated mesotrophic *Carex* peat typifies the organic bed in a zone that is positioned largely east of the present Vecht and parallel to the ice-pushed ridge (Poelman, 1966). In places, this peat is overlain by ombrogenic peat. In the subsurface directly overlying the substratum the *Carex* peat extends 4-6 km further west (Stiboka, 1970). The infilled lakes are mostly located within this extended zone. The position of this *Carex* peat and the position of the former lakes overlap, and the periods of subsurface *Carex*-peat accumulation coincides with lake formation. It is therefore inferred that the formation and position of lakes in the study area were at least partly controlled by seepage from the ice pushed ridge.

It is possible that peat type has also influenced the geometry of the lakes. The composition of peat, for instance, may have affected the erodibility of lake banks and therefore the size of lakes. This facet of lake evolution, however, remains the subject for future investigation.

2.5.2 Regional implications of the palaeogeography of the Angstel-Vecht area

Upstream connection of the Angstel-Vecht River system

The Angstel-Vecht system is the first Holocene Rhine-distributary system that drained to the northern part of the Holocene delta plain. Until then the channels had a general E-W orientation governed by the position of Saalian ice-pushed ridges (Fig. 2.1). By 2970 cal yr BP the conditions were favourable for a fluvial channel to adopt a SE-NW drainage pathway (from Utrecht to the Oer-IJ tidal inlet), which is parallel to the longitudinal axis of the North-Sea Basin.

The virtual closure of the Angstel and Vecht channels 2650-2200 cal yr BP is in accordance with the tendency of the Rhine-Meuse system to shift its drainage direction southward during the Subatlanticum (after 2600 cal yr BP). Berendsen and Stouthamer (2001) attribute this both to a gradient advantage resulting from a long sedimentation record in the northern delta related to the Oude Rijn, as well as enlargement of the Maas River estuary. From ~2250 cal yr BP (calibrated radiocarbon dates from Berendsen, 1982) onward, discharge of the main Rhine branch, the Oude Rijn, was captured by channels that drained to a tidal inlet near Rotterdam (Fig. 2.1) (Berendsen and Stouthamer, 2001, p. 75).

Downstream connection of the Angstel-Vecht River system

Before the Angstel and Vecht Rivers came into existence, ~2970 cal yr BP, the locally generated discharge of the Angstel-Vecht area drained to the Oer-IJ tidal inlet. A marine signal in the diatom record from gyttja in the Aetsveldse organic-clastic lake fill (Fig. 2.8A) indicates that the sea occasionally had access to the study area.

The connection between the Angstel-Vecht area and the Oer-IJ became more important after initiation of the Angstel and Vecht Rivers. Previously, various authors (Güray, 1952; Pons and Wiggers, 1960; Zagwijn, 1971; Beets et al., 2003) postulated the presence of a connection between the Oer-IJ tidal inlet and the Angstel-Vecht. However, partly because of insufficient age control, satisfactory evidence of the inferred connection was lacking. Nevertheless, palaeogeographic reconstructions of the Oer-IJ tidal inlet (Vos, 2008) and the Angstel-Vecht area (this study) show that both systems coexisted and that the silting-up of the Oer-IJ 2270 ± 90 cal yr BP (Vos, 2008) coincided with a strong decrease in fluvial activity of the Angstel and Vecht Rivers. Moreover, environmental changes in both areas seem to be related. In addition, the relatively short episode of fluvial activity in the Angstel-Vecht area coincided with a prevailing freshwater-brackish environment in the otherwise salt-brackish Oer-IJ (Vos, 2008). This, as well as the absence of an alternative tidal channel, strongly suggests a connection. From this perspective the presence of the Angstel and Vecht Rivers may also have contributed to the prolonged existence of the Oer-IJ tidal inlet relative to other tidal inlets of the Holland coast (Fig. 2.1). Later lake erosion and expansion of the city of Amsterdam, however, restrict mapping and determination of the exact position of the connection between the Angstel and Vecht Rivers and the Oer-IJ.

The Oude-Rijn and the Oer-IJ are the only two tidal inlets that existed along the northern Holland coast after ~3500 cal yr BP, and both functioned as fluvial outlets. It therefore seems likely that the activity of the Angstel and Vecht Rivers prolonged the existence of the Oer-IJ after ~3000 cal yr BP and that the decreased discharge of the Angstel-Vecht system 2650-2200 cal yr BP contributed to the silting up of the Oer-IJ ~2270 cal yr BP (Vos, 2008). The closure of the Oer-IJ forced drainage of the Angstel-Vecht and Flevomeer areas to the north, probably to the Vlie tidal inlet (Westerhoff et al., 2003a).

2.5.3 Effect of lakes and organic-clastic lake fills on the development of fluvial systems

The palaeogeographic reconstruction of the Angstel-Vecht area illustrates that the presence of lakes and organic-clastic lake fills in peat areas of delta plains can affect the development of a fluvial system. Three aspects are discussed below.

Natural levees

Natural levee size of delta channels tends to decrease in downstream direction (Kolb, 1963; Smith and Pérez-Arlucea, 1994; Hudson and Heitmuller, 2003) because of a reduction in the amount of coarse sediment required to construct natural levees, caused by downstream fining (Pirmez and Imran, 2003). The same trend has also been observed in the Angstel-Vecht area. At this location, however, the thickness of natural levees abruptly decreases downstream of where a lake is present in the drainage course (Fig. 2.10A). The relatively smooth downstream decrease shown in the left panel of Figure 2.10A, for example, is disrupted by the thin levee downstream of the Aetsveldse organic-clastic lake fill. The thickness of levees seems to be largely controlled by the presence or absence of a palaeolake in the upstream channel during levee building. For example, the levees upstream of the Horstermeer organic-clastic lake fill are 0.7 m thicker than the levees downstream of the former lake. We attribute this difference to a depletion of sediment in the channel downstream of the lake. Alternative controls on downstream thinning of natural levees involve downstream loss of discharge, as discussed by Makaske et al. (2007), downstream decrease of river stage, and downstream fining of sediment (Kolb, 1963). In the Angstel-Vecht area, for example, loss of discharge can be inferred from the presence of Winkel and Holendrecht Rivers that partly drain the Aetsveldse Lake (Fig. 2.2). Otherwise, the sharp drop in levee thickness is not accompanied with loss of discharge. Further, downstream decrease of river stage generally results in a gradual downstream decrease in levee thickness whereas the observed reduction of levee thickness in the Angstel-Vecht area is abrupt. We therefore associate the drop in levee thickness with the presence of palaeolakes. The preservation of this situation is explained by a strong decrease in upstream sediment supply before the Horstermeer Lake was completely filled. The levees upstream and downstream of the Loenen organic-clastic lake

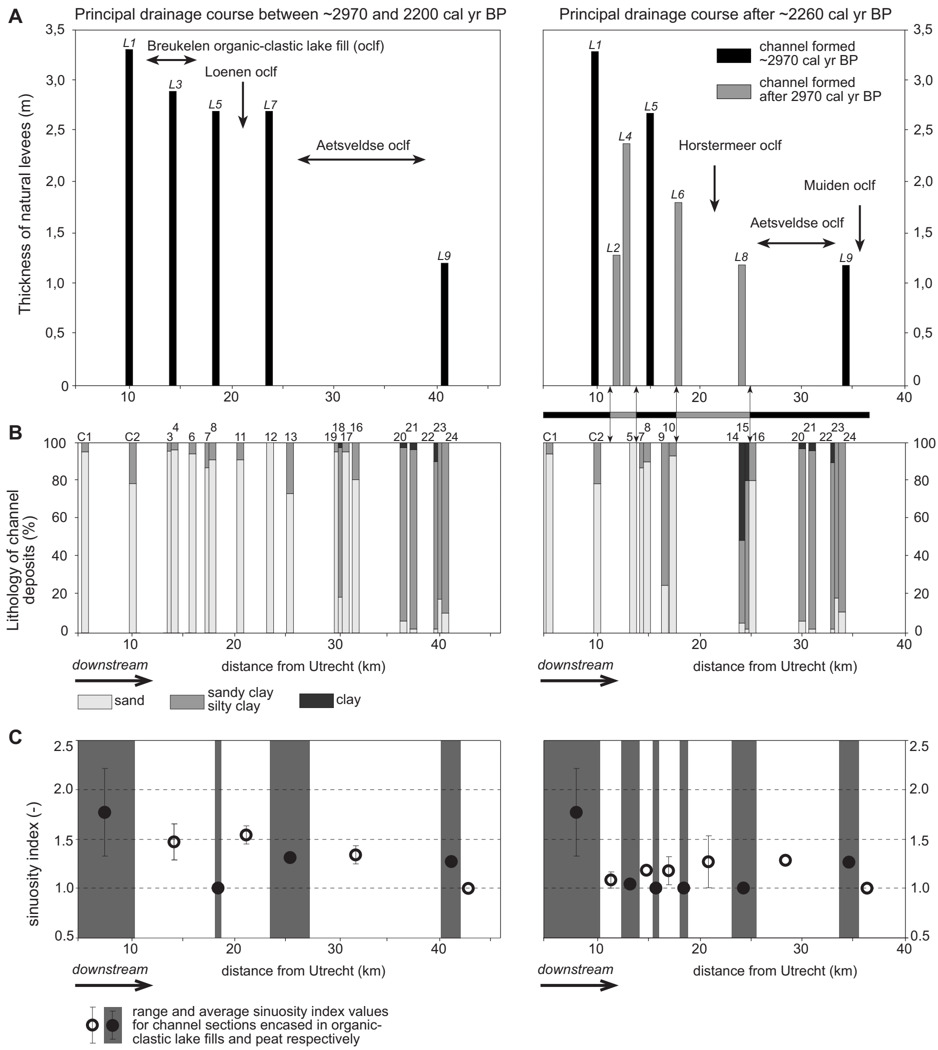


Figure 2.10.

A *Graph showing the thickness of natural levees in downstream direction following the initial river course (outlined in Fig. 2.9C) and the river course that includes the channels that formed at a later stage and follows the present River Vecht. The position of the organic-clastic lake fills relative to the measurement locations is also indicated. The locations of the measurements are shown in Fig. 2.5.*

B *The lithological composition of channel deposits for the same river courses as outlined in graph A.*

C *The sinuosity index values of channel sections in the Angstel-Vecht area. The channel sections are either encased in organics or in organic-clastic lake fills. Sinuosity values indicate whether fluvial channels are straight (<1.3) or meandering. The channel sections encased in organic-clastic lake fills have higher sinuosity index values compared with those that are encased in organics (predominantly peat). See Appendix 8 for a full-colour version.*

fill, in contrast, have comparable thicknesses resulting from continuing fluvial activity after formation of the organic-clastic lake fill.

Channel-belt sediment

The majority of fluvial channel deposits on deltaic plains is composed of sand and may include minor portions of silty and clayey material (Allen, 1965). Under low-energy conditions, however, which usually prevail in the distal (near coastal, cf. Makaske and Weerts, 2005) zone of deltaic plains, muddy lateral accretion deposits may be a significant component of channel deposits (Nanson and Page, 1983; Makaske and Weerts, 2005). Makaske and Weerts (2005) largely attributed this to low stream power and relative high bank stability, which subsequently promotes downstream rather than lateral migration of point bars and limits the formation of wide channel belts. At some locations in the Angstel-Vecht area, channel belt sediment is not entirely composed of sand (Fig. 2.10B). These locations mainly are positioned in fluvial systems downstream lakes. The channel belt sediment downstream of the Horstermeer organic-clastic lake fill, for example, is finer grained than the upstream channel-belt sediment (right panel in Fig. 2.10B). The same applies for the channel belt sediment downstream of the Aetsveldse organic-clastic lake fill (left panel in Fig. 2.10B). Because lakes trap sediment, coarse sediment is limited downstream of lakes. Channel-belt sediments, therefore, abruptly fine downstream of lake basins.

Channel planform

The planform geometry of distributary channels in the Angstel-Vecht area is largely influenced by the nature of the encasing substrate. Channel planform is controlled by, amongst other factors, discharge, amount and grain-size distribution of the supplied sediment, and bank stability (Bridge, 2003, and references therein). Makaske (2001) illustrated that the transition between straight and meandering channel planforms in distal parts of delta plains is largely governed by stream power and bank stability. When bank stability (controlled by the characteristics of the subsoil such as cohesiveness and the presence of roots) is high, channel migration is restricted (Smith, 1976). Peat accumulations, for example, promote high bank stability, resulting in straight channels where peat is prominent in the subsoil (Gradziński et al., 2003; Makaske et al., 2007). Clastic deposits, especially if composed of sand, are associated with low bank stability relative to peat banks. This implies that channels are able to laterally migrate and form meanders where adequate sand exists (Törnqvist, 1993b; Berendsen and Stouthamer, 2000). Downstream trends from meandering to straight channel planform patterns are common for lowland rivers (e.g., in the Rhine-Meuse delta (Törnqvist, 1993b; Makaske et al., 2007) and in the LMV (Gouw and Autin, 2008), where increasing bank stability and decreasing stream power occur because of a loss of discharge through crevasse channels.

The sedimentology of the Angstel-Vecht area that influences bank stability primarily consists of organics and organic-clastic lake fills. The channels that dissect organics show a downstream change from meandering to straight (Fig. 2.10C). We contend that as an effect of progradation the downstream channel reaches are younger and therefore have not had time to develop a typical meandering pattern. The peat-encased reaches upstream of the Aetsveldse Lake (Fig. 2.9C) started virtually simultaneous (within a few centuries) and yet show a remarkable reduction in sinuosity, which suggests a downstream loss of stream power (i.e. because of a lower gradient). In the Angstel-Vecht area, this trend is disturbed. There, the contiguous channels in downstream order display alternating modest Pind values (1.2-1.6) indicative of lateral channel migration and low P_{ind} values (1.0-1.2) indicative of very limited lateral channel migration (Fig. 2.10C). Wider channel belts with higher sinuosity values occur where organic-clastic lake fills are present (Fig. 2.10C). The alternation of channel reaches with comparably high and low P_{ind} values, reflects the spatial differences in sedimentology. Thus, the data illustrates that over time, where a fluvial distributary debouches into lake basins, a period of coarse clastic lake infilling will eventually result in the development of meandering channel patterns and wider channel belts as the coarse sediment is reworked and redistributed within the fluvial system.

2.6 CONCLUSIONS

Fluvial deposition in the Angstel-Vecht area predominantly occurred between 2970 and 2300 cal yr BP. Before and after that period a very small river system existed that occasionally connected to the Rhine river system. The Angstel and Vecht Rivers drained to the North Sea via the Oer-IJ tidal inlet. The Oer-IJ silted up ~2300 cal yr BP in response to reduced Rhine discharge through the Angstel-Vecht system. From that time onward, the remaining discharge through the Angstel-Vecht system drained to the north, towards the Vlie tidal inlet.

Lakes existed in the Angstel-Vecht area prior to fluvial sedimentation. The origin of the lakes is probably controlled by seepage related to the presence of an ice-pushed ridge.

The presence of lakes and organic-clastic lake fills in peat areas affects subsequent fluvial development in the following ways:

- Levees are thinner along the channel downstream of lakes compared to those along the upstream channel.
- Channel sediment bodies include more clay downstream of lakes than upstream of lakes or along rivers without lakes.
- Channel planform is affected by the presence of organic-clastic lake fills. Organic-clastic lake fills develop in peat-bounded lakes, which implies that bank stability in areas that surround organic-clastic lake fills is comparably high. In such settings lateral migration is largely restricted, which advocates the development of straight channels. In contrast, channels that traverse organic-clastic lake fills are able to migrate laterally (due to the sandy subsoil in which they are embedded) and thereby produce meander bends.

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3 Architecture and facies distribution of organic-clastic lake fills in the fluvio-deltaic Rhine-Meuse system, The Netherlands

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ABSTRACT

Lakes are prominent features in distal zones of delta plains. Organic-clastic lake fills may form in these lakes, which include gyttja and lacustrine deltas. Understanding of the geometrical properties, the facies distribution, and the origin of organic-clastic lake fills is limited. Knowledge of their development and composition, however, will improve our fundamental understanding of both sediment distribution in and evolution of distal delta plains. Furthermore, organic-clastic lake fills potentially form connectors in hydrocarbon reservoirs. This study aims to describe and explain the characteristics of organic-clastic lake fills and thereby to contribute to the understanding of delta evolution, especially of distal delta plains.

A field study conducted in the Schoonhoven and Angstel-Vecht areas, located in the distal part of the Rhine-Meuse delta, the Netherlands indicates that organic-clastic lake fills developed in peat-bounded lakes after an avulsion initiated fluvial sedimentation. The characteristics of organic-clastic lake fills, such as geometry, sedimentary facies, and sedimentation rates, were both qualitatively and quantitatively determined by cross sections, geomorphogenetic maps, and logs. The results show that in the Angstel-Vecht area as much as 30% of the fluvial-sand volume is stored in organic-clastic lake fills. This contrasts to the general notion that overbank deposits do not contain significant portions of sand. Further, the lacustrine setting often results in vertically bounded sediment bodies that include gyttja and underlies clastic depositional facies, which show a prominent coarsening-upward succession. It is concluded that the presence of organic-clastic lake fills enlarges the reservoir capacity of ancient fluvio-deltaic successions as they contribute to the reservoir volume and also can form connectors between isolated channel-belt sand bodies.

3.1 INTRODUCTION

Lakes are common features in distal delta plains. They are widespread, for example, in the distal delta plains of the modern Columbia River (Makaske et al., 2002), in the

Cumberland Marshes (Smith and Pérez-Arlucea, 1994)(, both situated in Canada, and in the Laitaure delta, Sweden (Andren, 1994). When fluvial sediment is transported to these lakes, it is trapped and subsequently, a clastic sediment body is formed. These clastic sediment bodies have been termed lacustrine deltas (e.g., Tye and Coleman, 1989a) or crevasse-splay deposits (Smith et al., 1989). Studies in delta plains, however, are strongly biased to proximal portions of delta plains. Hence, depositional facies that are restricted to distal delta plains, such as lacustrine deltas, received relatively little attention. Nevertheless, they have been recognized in ancient settings (e.g.,Fielding, 1984; Hornung and Aigner, 1999), (sub)recent settings (e.g., Smith, 1996; Pérez-Arlucea and Smith, 1999; Bos et al., 2009; Chap. 2) and modern settings (e.g., Smith and Pérez-Arlucea, 1994; Pérez-Arlucea and Smith, 1999). They form potential reservoirs for hydrocarbons and improve reservoir connectivity. Further, they affect geomechanical properties of distal delta-plain successions, which is of importance for building activities.

Based on relatively well-studied examples in the Lower Mississippi Valley, USA (Tye and Coleman, 1989a, b; Smith, 1996) and the Cumberland Marshes, Canada (Smith et al., 1989; Smith and Pérez-Arlucea, 1994; Pérez-Arlucea and Smith, 1999) the facies composition of these sediment bodies is well understood. In addition, Smith et al. (1989) proposed a conceptual model for the development of these deposits. Little attention, however, has been given to the lacustrine depositional setting in which the deposits form. In distal portions of lakes and prior to clastic input, for example, gyttja is deposited. Gyttja is an organic lacustrine deposit, composed of a mixture of dead-organism remains, chemical precipitation, and minerogene matter (Von Post, 1860; Hansen, 1959). Further, the geometry of lakes likely affects the geometry of lacustrine sediment bodies. The composition of lake fills, including gyttja that formed prior to clastic input, and the geometry of lake fills, therefore, needs further investigation. The aim of this study, therefore, is to determine the geometry, sediment distribution, and origin of organic-clastic lake fills in the Holocene Rhine-Meuse delta, The Netherlands.

In this study we focus on the complete lake fill, including organics and lacustrine deltas. To emphasize the unity of this system, the term organic-clastic lake fill is introduced, which encompasses the complete lacustrine succession. An organic-clastic lake fill is herewith defined as a sharply bounded sediment body that has been deposited in a delta-plain lake, which comprises gyttja and overlying clastic deposits – lacustrine deltas – that exhibit a coarsening-upward succession.

The Holocene Rhine-Meuse delta, The Netherlands, offers a unique opportunity to study organic-clastic lake fills because it has been studied extensively the alluvial architecture and the palaeogeography (Berendsen and Stouthamer, 2001, and references therein; Gouw, 2008). Alluvial architecture has been defined as the geometry, proportion, and spatial distribution of different types of fluvial deposits in the alluvial succession (e.g., Leeder, 1978). In the fluvio-delta plain of the Holocene Rhine-Meuse system, organic-clastic lake fills of various ages are present but a detailed description of their geometry and depositional facies is lacking. A study on organic-clastic lake fills in the Rhine-Meuse delta, therefore, can be implemented in a well defined architectural framework, which will contribute to a better understanding of delta evolution.

3.2 STUDY AREA AND GEOLOGIC SETTING

The Holocene Rhine-Meuse delta is located at the southeastern margin of the subsiding North Sea Basin, roughly between the German-Dutch border and the Dutch coast (Stouthamer and Berendsen, 2000) (Fig. 3.1). As a result of rising base level, the apex of the Holocene delta shifted from offshore Rotterdam towards the east, and simultaneously the delta plain expanded laterally by the formation of peat areas fringing the delta plain. The evolution of the Holocene Rhine-Meuse delta is influenced by, amongst others, eustatic sea-level rise (dominant before 5000 cal yr BP), basin subsidence (dominant between 5000 and 3000 cal yr BP), and increased discharge and sediment supply (dominant after 3000 cal yr BP) (Gouw and Erkens, 2007). The Holocene deposits are ~1 m thick in the eastern part of the Netherlands and thicken westwards, being ~0 m thick near Rotterdam. Natural sedimentation in the Rhine-Meuse delta ended between AD 1100 and 1350 when the embankment of the Dutch rivers was completed (e.g., Lambert, 1985).

Within the Rhine-Meuse delta, two sites were selected to study organic-clastic lake fills: (1) the Schoonhoven area, where organic-clastic lake fills of middle Holocene age are covered by approximately 5 m of younger sediments and (2) the Angstel-Vecht area, where organic-clastic lake fills of middle to late Holocene are present near the surface (Fig. 3.1).

3.2.1 Schoonhoven area

The substratum in the Schoonhoven area (Fig. 3.1) is present at approximately -10 m relative to Dutch O.D. (Ordnance Datum, ~mean sea level). It consists of fluvial coarse-grained sand (Kreftenheye Formation, Fig. 3.2), which is overlain by a distinct bed that is ~0.5 to 1.0 m thick and comprises loamy overbank deposits (Wijchen Member, Kreftenheye Formation) that has often been affected by soil formation. Eolian sand dunes (Boxtel Formation, Delwijnen Member), such as the Zevender (Fig. 3.3A) are locally present on top of the Kreftenheye Formation, although they are currently largely buried by fluvio-deltaic deposits.

Holocene aggradation in this area started ~8500 cal yr BP and resulted in the accumulation of organics (peat and gyttja) and Rhine deposits (channel-belt, natural-levee, floodbasin, crevasse-splay deposits and organic-clastic lake fills). Organic-clastic lake fills in this area intercalate with and underlie deposits from a middle Holocene channel belt that has been active between 6800 and 6125 cal yr BP (Cabauw channel belt, cf. Berendsen and Stouthamer, 2001).

3.2.2 Angstel-Vecht area

The substratum in the Angstel-Vecht area is of Pleistocene age and comprises fluvial coarse-grained sand (Kreftenheye Formation, Fig. 3.2) and fluvio-glacial sandur deposits

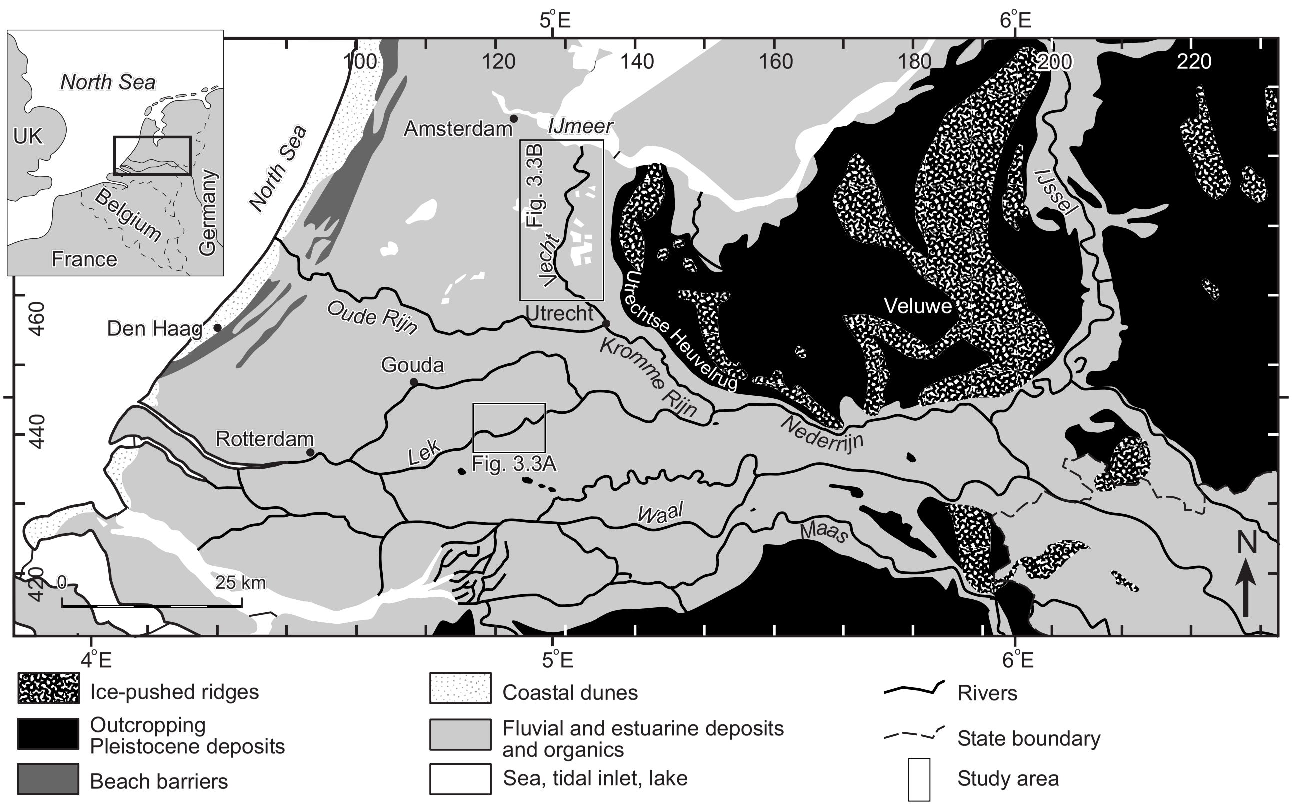


Figure 3.1. Overview of the central part of the Netherlands, covering the Rhine-Meuse delta. Indicated are the positions of the two study sites: the Schoonhoven area (Fig. 3.3A) and the Angstel-Vecht area (Fig. 3.3B). Dutch coordinates (Rijksdriehoekstelsel, inner tick marks) as well as latitude-longitude coordinates (outer tick marks) are indicated.

(Schaarsbergen Member, Drente Formation) that are overlain by Late Pleistocene eolian sand (so-called "coversand", Wierden Member, Boxtel Formation). Coversand is 1-2 m thick and exhibits an undulating topography. Furthermore, coversand consists of homogeneous, $CaCO_3$-poor, fine-grained quartz sand, which contrasts with the vast majority of the fluvial deposits that contain calcite silt grains.

The thickness of the Holocene deposits in the eastern part of the study area is ~1-2 m and increases towards the west to up to 8 m. The Holocene sequence comprises extensive organic accumulations (mainly peat, Nieuwkoop Formation), fluvial deposits of the Angstel and Vecht Rivers (channel-belt, natural-levee and floodbasin deposits and organic-clastic lake fills, Echteld Formation), and marine deposits (Naaldwijk Formation) (Van de Meene et al., 1988; Weerts et al., 2002). The fluvial deposits are formed after 2970 cal yr BP (Bos et al., 2009; Chap. 2). A marine bed (IJe Bed, Walcheren Member, Naaldwijk Formation; Fig. 3.2) up to 0.5 m thick and composed of silty clay, forms the top of the Holocene sequence (Stiboka, 1965).

Human activity has not been of any influence on the formation of organic-clastic lake fills. The earliest human intervention in the Angstel-Vecht drainage system occurred perhaps during the Roman period, from 12 BC onwards (Weerts et al., 2002), which is a few hundred years after the main period of organic-clastic lake-fill formation.

3.3 METHODS

The approach of this study consisted of three steps, which are listed here and outlined in the following paragraphs. First, the geometry and architectural setting of the organic-clastic lake fills was measured on newly constructed geomorphogenetic maps and cross sections (Fig. 3.3, 3.4). These cross sections and maps were mainly based on lithological borehole descriptions of manual cores, and on Digital Elevation Model (DEM) data. Second, depositional facies interpretations were based on lithological cross sections, concisely recorded sedimentary characteristics of five cores (Fig. 3.5, 3.6), and microfossil analyses (i.e., diatom counts) (Fig. 3.7). Third, quantitative analyses for determining sedimentation rates and clastic sedimentation duration in the Aetsveld Lake were carried out.

3.3.1 Determining architecture

Lithological cross sections and geomorphogenetic maps provided information on the position of organic-clastic lake fills within the fluvio-deltaic succession as well as on their geometry. Furthermore, the cross sections and maps showed the morphology of the boundaries between the organic-clastic lake fill and the surrounding deposits. For this study we used ~5500 archived core data and additional manually drilled cores. Archived core descriptions were stored in the geological databases of Utrecht University (Berendsen, 2005) and TNO Built Environment and Geosciences, Geological Survey of the Netherlands (TNO, 2009) and include, amongst others, lithological (e.g., texture, organic content, colour, $CaCO_3$ content) and sedimentological (e.g., bedding and

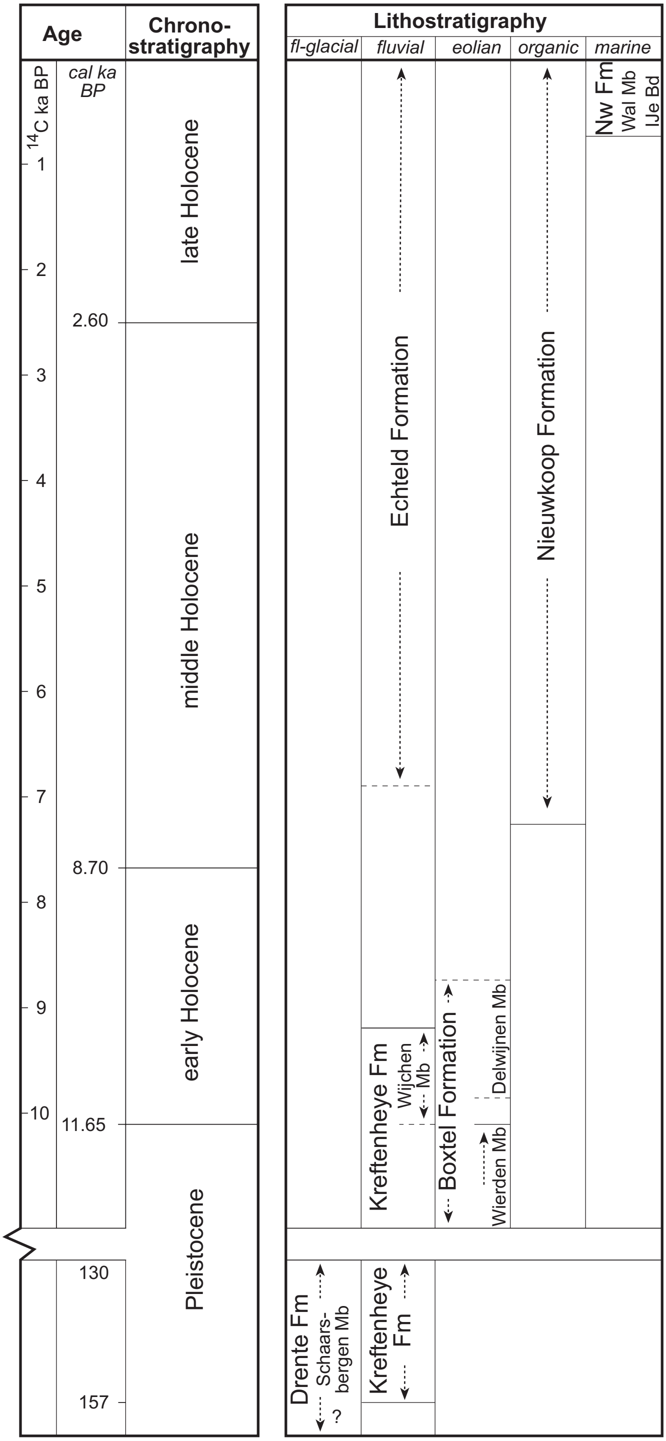


Figure 3.2. Chronostratigraphy and lithostratigraphy of the study area. Nw Fm = Naaldwijk Formation, Wal Mb = Walcheren Member. Radiocarbon ages follow Van Geel et al. (1981) and Walker et al. (2009). Lithostratigraphy follows Westerhoff et al. (2003b) and Busschers et al. (2008).

boundary characteristics) information. Core descriptions in the Utrecht University database contain information at 10 cm intervals (outlined by Berendsen and Stouthamer, 2001) whereas these in the TNO database are organized in homogeneous layers of variable thickness (following the method outlined by Nederlands Normalisatie Instituut, 1989). In the Schoonhoven area, core density was ~36 cores/km^2; 34% of these penetrated into the substratum. In the Angstel-Vecht area, core density was ~24 cores/km^2; 64% of these penetrated into the substratum.

In addition to the lithological borehole data a high-resolution (1 measurement/m^2; 1 cm vertical resolution) laser altimetry DEM (Rijkswaterstaat-AGI, 2005) was used in the Angstel-Vecht area to construct the map and cross sections. The deposits of interest in that area are present near the surface and therefore can be adequately mapped on the basis of their morphological expression. Human-induced groundwater-level lowering, in addition, resulted in relatively strong surface lowering in peat areas due to oxidation of organic constituents. The surface level of peat-underlying areas, therefore, is much lower than areas where clastic deposits such as organic-clastic lake fills are present, which further enhances geomorphogenetic mapping of the Angstel-Vecht area.

3.3.2 Depositional facies descriptions

Sedimentary characteristics of the deposits were obtained by macroscopic analyses of five cores (codes and locations indicated in Fig. 3.3) with 10 cm diameter, drilled with a mechanical bailer-drilling device (Oele et al., 1983). The results were compiled in sedimentary logs (Fig. 3.5) that include texture, organic content, admixtures (e.g., shells, peat clasts, clay pebbles), sedimentary structures, bounding-surface descriptions, soil features, and bioturbation characteristics.

Diatom analysis was used to reconstruct the environmental conditions (e.g., salinity and trophy) during deposition, based on ecological indicator values proposed by Van Dam et al. (1994). The relative quantity of epiphytic diatoms – attached to plants – was used to roughly indicate the density of submerged vegetation. It was expected that in dynamic environments – such as mouth bars – the relative abundance of epiphytic species would be less than in quiet environments. For diatom counting, 44 samples (24 from core B25G1057 and 20 from core B25H0739) were prepared following the procedure of Cremer et al. (2001). The sampling strategy was to cover the main lithological layers (sampling positions are indicated in Fig. 3.7).

3.3.3 Quantitative analyses

Quantification of sand volume in organic-clastic lake fills

The total volume of sand that is contained in the organic-clastic lake fills in the Angstel-Vecht area was calculated by multiplying the surface area by the average thickness. The surface areas of the individual organic-clastic lake fill sediment bodies were calculated in a geographic information system (GIS) and based on the geomorphogenetic map (Fig. 3.3B). The average thickness of the organic-clastic lake fill and mouth-bar deposits was

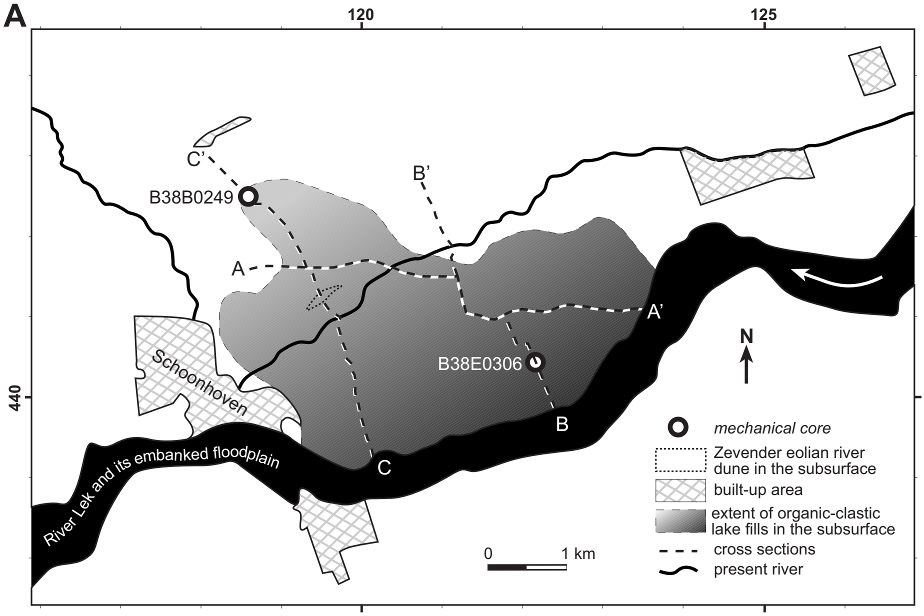


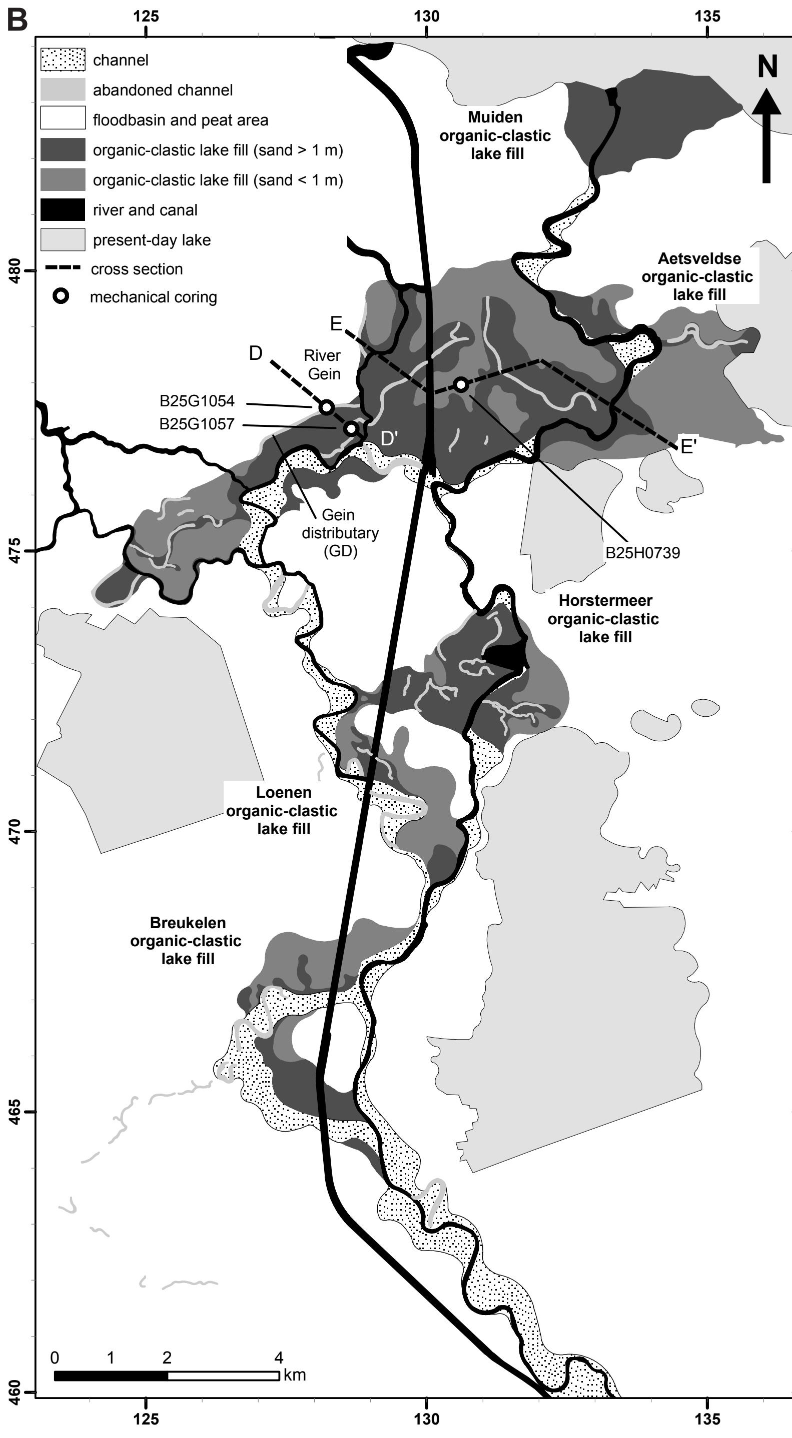
Figure 3.3. A) Spatial extent of organic-clastic lake fills in the Schoonhoven study area. These deposits include sandy mouth-bar deposits that are ~2.5 m thick in the southeast (dark grey) and in general taper off towards the northwest (light gray), where they eventually cease. The position of the cross sections (Fig. 3.4A-C) and sedimentary logs (Fig. 3.5A, B) are shown. B) (on facing page) Spatial extent of organic-clastic lake fills in the Angstel-Vecht area. Five organic-clastic lake fill bodies are found.

estimated, based on the cross sections (Fig. 3.4D, E).

Duration of clastic lacustrine sedimentation

A simple, physics-based model (based on Kleinhans, 2005) was used for calculation of both flow parameters and sediment transport capacity, based on the architecture of the feeding channel. The resulting prediction of supplied sediment was compared with the volume of the sand in the Aetsveldse organic-clastic lake fill and the period of time as defined by independent age constraints.

Discharge and sediment transport through the Angstel to the Aetsveldse Lake were calculated using channel-belt characteristics (e.g., grain size, dimensions) as input parameters (as outlined by Kleinhans, 2005, and summarized in Appendix 1). In order to determine the dominant sediment-transport mechanism, both bed-load sediment transport (based on Meyer-Peter and Mueller, 1948) and suspended-load transport (based on Engelund and Hansen, 1967) were calculated. Characteristics of palaeochannels, such as width, depth, and gradient, are reflected by the channel-belt deposits, especially for straight river channels where the limited lateral migration minimizes strong discrepancies


B
125
130
135
480
475
470
465
460
channel
abandoned channel
floodbasin and peat area
organic-clastic lake fill (sand > 1 m)
organic-clastic lake fill (sand < 1 m)
river and canal
present-day lake
cross section
mechanical coring
N
Muiden
organic-clastic
lake fill
Aetsveldse
organic-clastic
lake fill
E
D
River
Gein
B25G1054
B25G1057
D'
E'
Gein
distributary
(GD)
B25H0739
Horstermeer
organic-clastic
lake fill
Loenen
organic-clastic
lake fill
Breukelen
organic-clastic
lake fill
0
1
2
4
km

between channel and channel-belt width. In this study, the characteristics of the Angstel palaeochannel were measured on maps (maximum width), individual cores (wherein thickness of channel-belt deposits ≈ maximum channel depth), and gradient of top of sand (GTS) lines (≈ gradient of paleochannel). Reduction of the channel-belt deposits thickness as the result of compaction due to loading is probably minimal because overlying sediments are thin, typically < 1 m. Furthermore, compaction of deposits underlying channel-belt deposits, which would affect the GTS lines, is considered to be absent because the studied channel sediment bodies are encased in sandy substratum. Channel-belt dimensions are maximum measures of the palaeochannel dimensions, and the resulting sediment transport rate therefore is a maximum value. The calculated sediment transport rate, in combination with the known sand volume in the Aetsveldse organic-clastic lake fill, was used to calculate a minimum time scale for the formation of this sediment body. It is assumed that the Aetsveldse Lake functioned as a perfect sediment trap for sand.

Given that both the duration of bankfull discharge and the total period of clastic deposition, obtained by radiocarbon dating by Bos et al. (2009; Chap. 2), were determined, the hydrological regime could be estimated. It was feasible to calculate the intermittency (I_f) of the Angstel palaeochannel, where intermittency is defined as the fraction of time that a river has or exceeds bankfull discharge (Paola et al., 1992).

Sedimentation rates

The period of sedimentation within the Aetsveldse palaeolake is constrained by the beginning of clastic sedimentation by the feeding channel – the Angstel River as determined by ^{14}C analyses (Tab. 3.3) and the moment that marks the final abandonment of the distributary channels in the Aetsveldse organic-clastic lake fill, which also has been determined by ^{14}C analyses (Bos et al., 2009; Chap. 2). Furthermore, four optical stimulated luminescence (OSL) samples (NCL-3206022 -25, Tab. 3.4) were taken in a vertical sequence from one core. Two samples were taken from prodelta facies; the other two were taken from distributary-mouth-bar facies. Analyses of OSL samples were carried out following the procedure outlined by Wallinga (2002) and specified by Wallinga and Johns (2008). Radiocarbon ages are presented with 1σ uncertainty range to allow for comparison with OSL-ages.

3.4 RESULTS

3.4.1 Architecture of organic-clastic lake fills

Clastic lake fills in the study areas (Fig. 3.5) occur directly on top of the substratum or overlie a thin peat bed that covers the substratum (Fig. 3.4). In the Schoonhoven study area, two successions of stacked organic-clastic lake fills are present, of which the lower one is the most extensive (Fig. 3.4A, C).

The shape of the studied organic-clastic lake fills in cross-sectional view is often rectangular with sharp vertical edges (see, for example, the western limits of the organic-

Table 3.1. Surface areas (km²) of organic-clastic lake fill deposits. The location of the organic-clastic lake fills is indicated in Fig. 3.3. The reconstructed areas of the palaeolakes potentially were larger still because only the area that received clastic sediment is included. The preservation potential of organic-clastic lake fills increases in downstream direction. This is probably due to higher stream power in the upstream part of the study area, which advocates lateral migration and subsequently erosion of the organic-clastic lake fill. Besides, the lakes in the upstream part were filled earlier than those in the downstream part, which means that the period available for lateral migration – i.e. erosion – is longer in the upstream part of the study area.

	Organic-clastic lake fill	*Reconstructed palaeolake[a] (km²)*	*Organic-clastic lake fill (km²)*	*Preserved organic-clastic lake fill[b] (%)*	
	Schoonhoven	-	14.9	-	
Angstel-Vecht	Breukelen	6.0	3.8	64	downstream ↓
	Loenen	3.4	2.6	76	
	Horstermeer	5.1	4.7	91	
	Aetsveld	25.6	23.8	93	
	Muiden	3.7	3.7	100	

a Area of the palaeolake = organic-clastic lake fill + area of channel deposits in that domain.
b Preserved organic-clastic lake fill = (organic-clastic lake fill / palaeolake) * 100 %

clastic lake fills in Fig. 3.4D-E) or wedge-like with edges that taper out (see, for example, the eastern limit of the organic-clastic lake fill in Fig. 3.4E). These examples also show that lateral boundaries of organic-clastic lake fills are often defined by organics. The lower boundary of these deposits is often sharp (erosional), and its morphology generally is defined by the relief of the substratum. The upper boundary, typically defined by organics or floodbasin deposits (Fig. 3.4), has a gradual character and undulates on a (sub)horizontal level. The plan-view shape of the studied organic-clastic lake fills is highly variable and ranges from hook-shaped (Breukelen organic-clastic lake fill, Fig. 3.5B) to rectangular or circular-shaped (Schoonhoven, Horstermeer and Muiden organic-clastic lake fills, Fig. 3.3A-B). Regularly, younger channel-belt deposits dissect organic-clastic lake fills in the Schoonhoven area (Fig. 3.4A-C). It should be noted that deposits that cover peat may have subsided due to compaction of the underlying peat. Therefore, the deposits that currently laterally bound the organic-clastic lake fills in the Schoonhoven area cannot straightforwardly be regarded as the former lake banks because these deposits overlie peat. Before this peat was buried by clastic deposits it must have been much thicker. This means that the top of the peat layer originally was present at a higher level, and, therefore, the peat probably represents a former lake bank. The thin and compacted peat bed north of the organic-clastic lake fill that covers the substratum in cross section B-B' (Fig. 3.4B), for example, may have bounded the lake in which the organic-clastic lake fill developed.

The surface area of the studied organic-clastic lake fills ranges from 2.6 km² to 23.8 km²

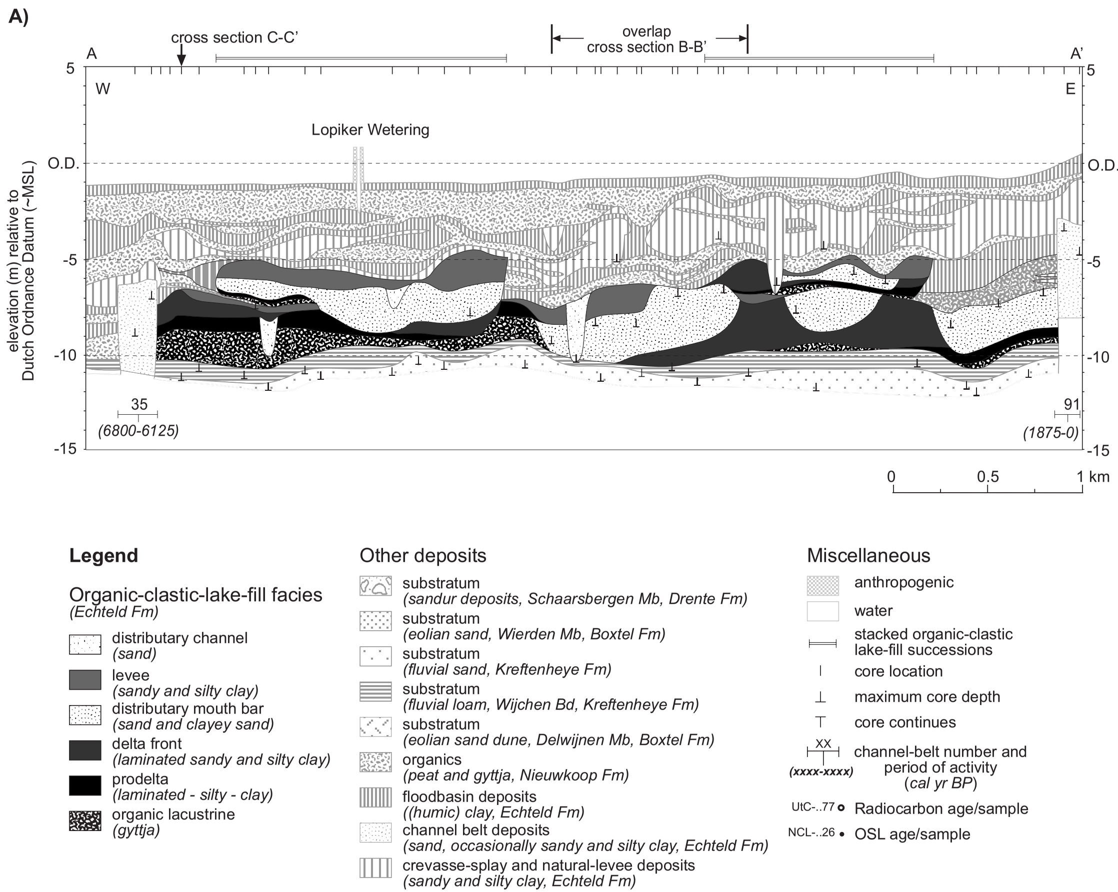


Figure 3.4. Cross sections in the Schoonhoven study area (A-C) and in the Angstel-Vecht study area (D, E). The location of the cross sections is indicated in Figure 3.3.
Cross section A-A' is oriented parallel to the progradation direction of the organic-clastic lake fills. Note that two generations of organic-clastic lake fills are present. Moreover, the overbank deposits of the Cabauw channel belt intercalate with the younger-generation organic-clastic lake fills, just east of the intersection with cross section C-C'.

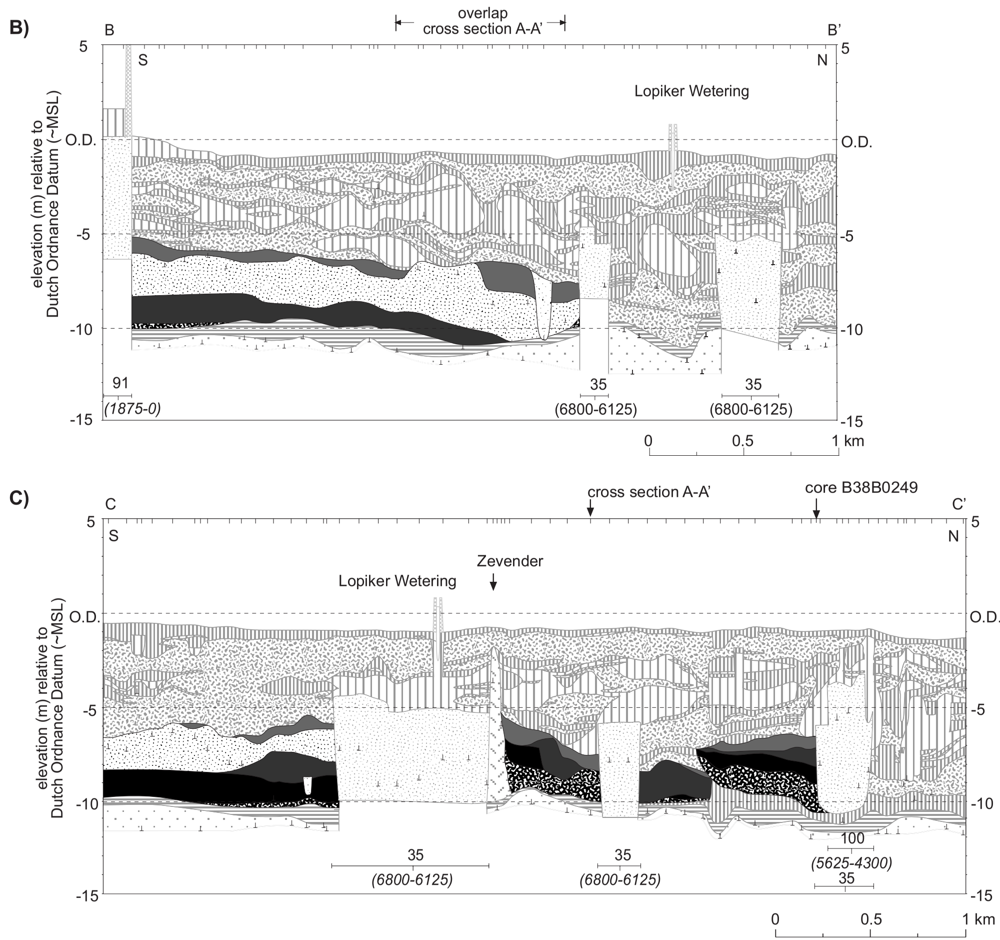


Figure 3.4. Continued. Cross section B-B' shows the facies distribution in the upstream part of the organic-clastic lake fill. In this part the prodelta facies is absent and the organic lacustrine and the delta-front facies are relatively thin whereas the distributary-mouth-bar facies is relatively thick. Cross section C-C' is positioned in the downstream part of the organic-clastic lake fill where the organic and fine-grained clastic (prodelta) facies dominate, especially north of the Zevender.

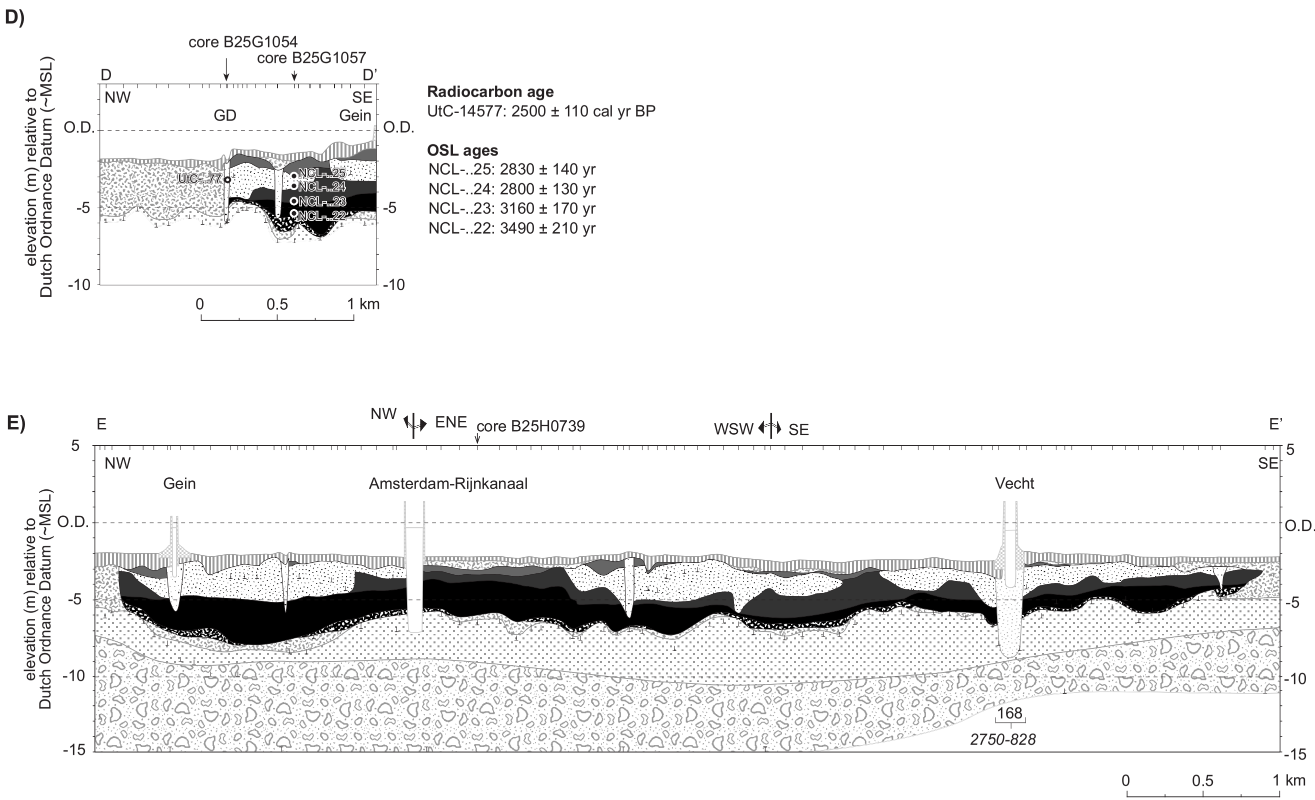

D)
core B25G1054
core B25G1057
D
D'
NW
SE
GD
Gein
O.D.
-5
-10
elevation (m) relative to Dutch Ordnance Datum (~MSL)
UtC-..77
NCL-..25
NCL-..24
NCL-..23
NCL-..22
0
0.5
1 km
Radiocarbon age
UtC-14577: 2500 ± 110 cal yr BP
OSL ages
NCL-..25: 2830 ± 140 yr
NCL-..24: 2800 ± 130 yr
NCL-..23: 3160 ± 170 yr
NCL-..22: 3490 ± 210 yr
E)
E
E'
NW
ENE
core B25H0739
WSW
SE
Gein
Amsterdam-Rijnkanaal
Vecht
5
O.D.
-5
-10
-15
elevation (m) relative to Dutch Ordnance Datum (~MSL)
168
2750-828
0
0.5
1 km

Table 3.2. Organic-clastic lake-fill volume in the Angstel-Vecht area and the proportion of sandy facies therein.

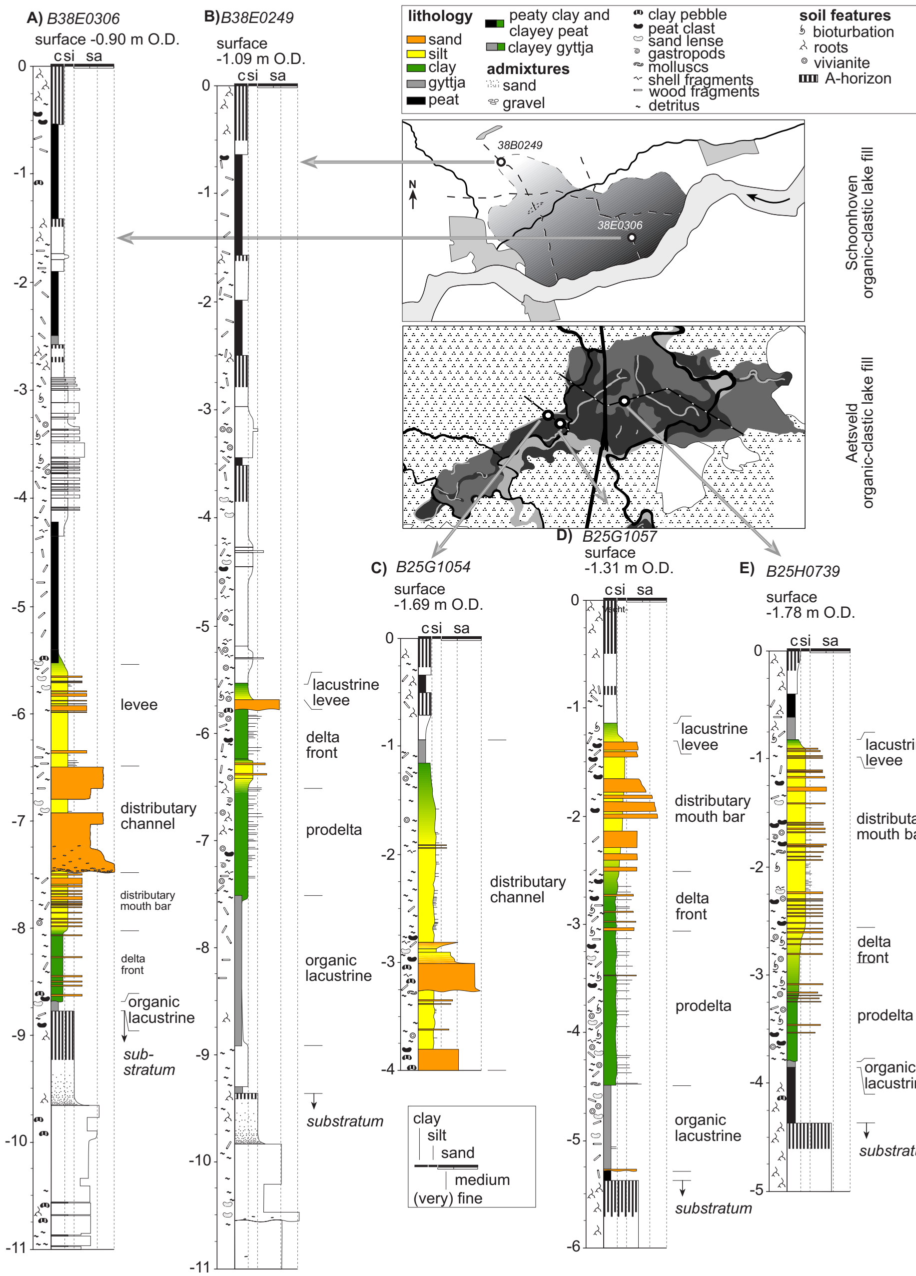
Organic-clastic lake fill		*Average thickness (m)*	*Area (km^2)*	*Volume (km^3)*	*Sandy lithofacies volume (km^3)*	*Sandy lithofacies in organic-clastic lake fill (%)*
Breukelen		3.7	3.8	0.0139	0.0042	30
Loenen	downstream	3.5	2.6	0.0091	0.0021	23
Horstermeer		3.9	4.7	0.0181	0.0064	35
Aetsveld		4.0	23.8	0.0951	0.0259	27
Muiden	↓	4.3	3.7	0.0159	0.0070	44
total		3.9	38.6	0.1522	0.0457	30

(Tab. 3.1). The thickness of individual organic-clastic lake fills is fairly constant or decreases slightly away from the river mouth, typically being 3-4 m, both in the Schoonhoven and the Angstel-Vecht study area (Fig. 3.4). In Figure 3.4E, for example, the thickness of the organic-clastic lake fills decreases in the progradation direction from ~5 m in the west to ~3 m in the east. The morphology of the substratum causes local variations in the thickness of less than ~1.5 m. In the Angstel-Vecht area (Fig. 3.4D-E) the substratum as well as the lower boundary of the organic-clastic lake fill undulates whereas the upper limit of the organic-clastic lake fill represents a horizontal level causing the thickness to vary according to the topography of substratum. Compaction of the clayey lithofacies contained in the organic-clastic lake fills and especially of the organics underlying the clastic succession likely occurred synchronously with the formation of the organic-clastic lake fills in the Angstel-Vecht area. Postdepositional compaction due to additional loading by younger deposits is limited because the organic-clastic lake fills in that area (Fig. 3.4D-E) are covered only by a thin (~0.3 m) marine clay layer (IJe Bed, Naaldwijk Formation). The volume of individual organic-clastic lake fills range from 0.01 to 0.10 km^3 (Tab. 3.2).

← Figure 3.4. Continued. Cross section D-D' shows a relatively "complete" succession of organic-clastic lake fill facies including organic lacustrine, prodelta, delta-front, distributary-mouth-bar and levee deposits. Note that the lower boundary of the organic-clastic lake fills runs parallel to the relief of the substratum and that the boundary to the adjacent peat is very abruptly as defined by the GD distributary channel. Note also that the two upper OSL samples in core B25G1057 are from mouth-bar deposits that are associated with the GD channel (Fig. 3.3B). Cross section E-E' covers the central part of the Aetsveldse organic-clastic lake fill. Note that the organic-clastic lake fill in the northwest has a very steep edge and that it displays a wedge shape towards the southeast limit. The distributary-channel facies are associated with distributary-mouth-bar deposits, as is clearly visible, for example, between Gein and the Amsterdam-Rijn canal. See Addenda 2 and 3 for full-colour versions.

3.4.2 Facies description of organic-clastic lake fills

Six depositional facies have been discerned in the studied organic-clastic lake fills: organic lacustrine, prodelta, delta front, distributary mouth bar, levee, and distributary



channel depositional facies (terminology largely according to Tye and Coleman, 1989a). Frequently, fluvial channel-belt deposits are encased in organic-clastic lake fills. Because these deposits form mainly after the formation of organic-clastic lake fills, their deposits are not regarded as part of organic-clastic lake fills.

Organic-lacustrine facies

Organic-lacustrine facies comprise gyttja. Most frequently observed is the very fine-textured (particle diameter < ~10 μm) algae gyttja, although, in places, gyttja composed of reworked plant remains (often including twigs) has been found. Organic lacustrine deposits form the lower part of organic-clastic lake fills. In some places though, these are absent, which is probably due to wave action causing turbulent water and thereby preventing sedimentation or eroding the top part of the lake bottom. This can be shown by calculating the wave base (D_w) for waves that potentially form in these lakes during regular storms. The wave base determines the depth at which waves can exert erosional force. The wave base can be approximated by half the wave length (L_w), which relates to wave period (T_w) by $L_w = 1.56\ T_w^{\ 2}$. The wave base depends predominantly on the fetch (lake length in downwind direction) and wind velocity. Groen and Dorrestein (1976), in their diagram I, presented the quantitative relation between wind velocity, fetch and wave period. From this, it could be deduced that, during gales (wind velocity ≈ 20 m/s), the wave base reaches a depth of ~3 and ~5 m below the water surface for a fetch of 1 and 2 km respectively. This illustrates that wave action is of prominence in shallow lakes and affects lake-bottom sediments.

Organic lacustrine deposits generally are less than 0.5 m thick (e.g., Fig. 3.4B, E), although, in exceptional cases, thicknesses of up to 2 m have been found (see the western part of the organic-clastic lake fill at ~10 m depth in Fig. 3.4A). The sediments often display blue vivianite dots in association with plant remains, which generally indicate a fresh-water depositional environment (e.g., Fagel et al., 2005). Regularly, very thin (< 2 mm thick) sand and silt lenses and laminae are observed. In the Angstel-Vecht study area these lenses and laminae are $CaCO_3$-poor and hence originate from the coversand substratum. Further, these lenses and laminae become thicker towards the eastern limit of the organic-clastic lake fill (for example, compare Fig. 3.4D and 3.4E) where the substratum has a higher elevation and the organic-clastic lake fill is thinner. This suggests that wave action eroded the

← Figure 3.5. Sedimentary logs of five cores in organic-clastic lake fills. The organic-clastic lake fills are indicated with a color scheme whereas the other deposits are shown in black and white. In the sedimentary logs, the median grain size is indicated for analyzed samples. One radiocarbon age and four OSL ages are indicated in the logs; their codes refer to Table 3.3 and Table 3.4 respectively. The prodelta and delta-front facies in Part D exhibit primarily a coarsening-upward succession and include evidence of strong bioturbation. The thin sand bed at -5.27 m O.D. in Part D comprises reworked substratum sediment that is present at the base of the organic lacustrine facies. See Addendum 4 for a full-colour version.

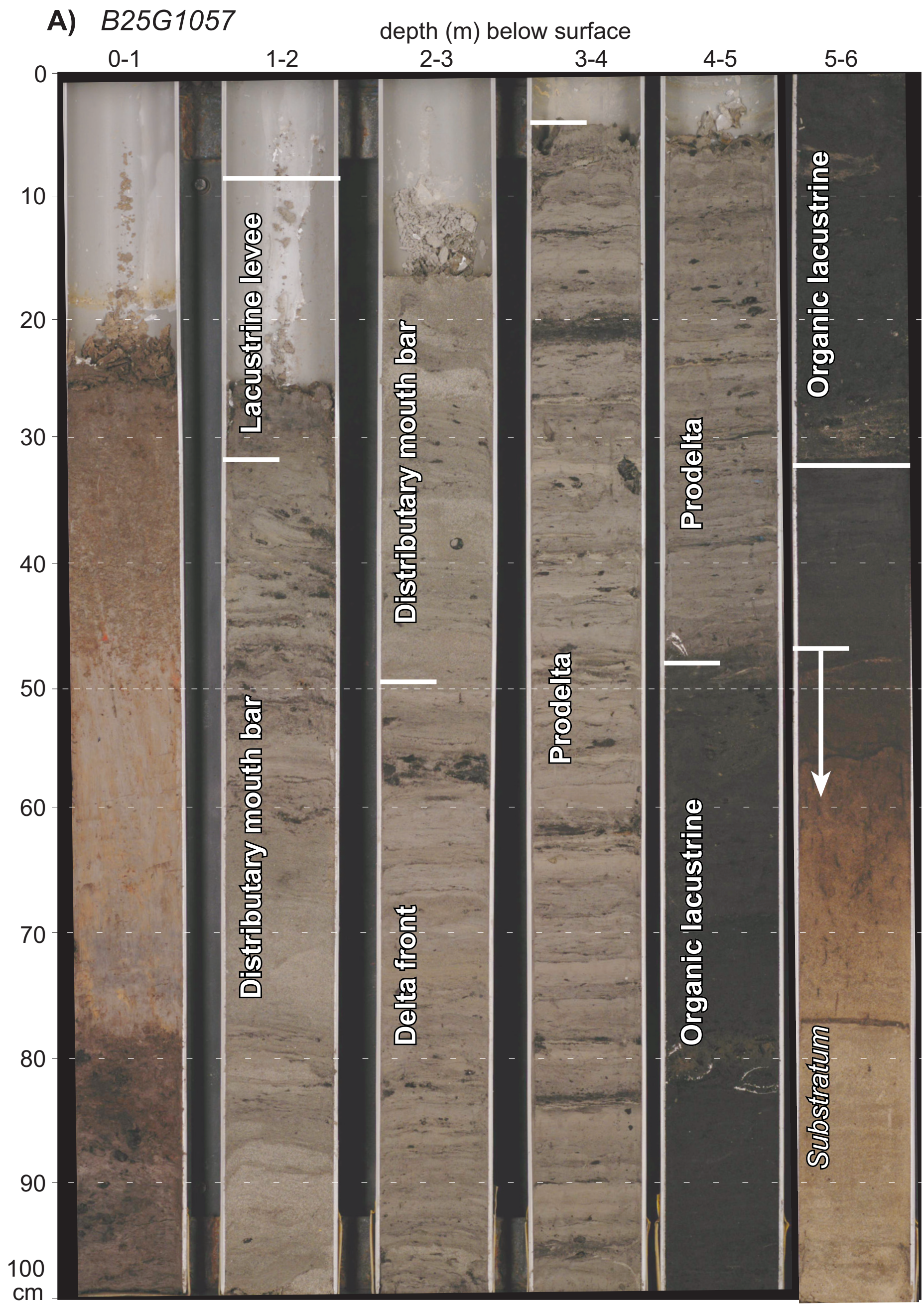
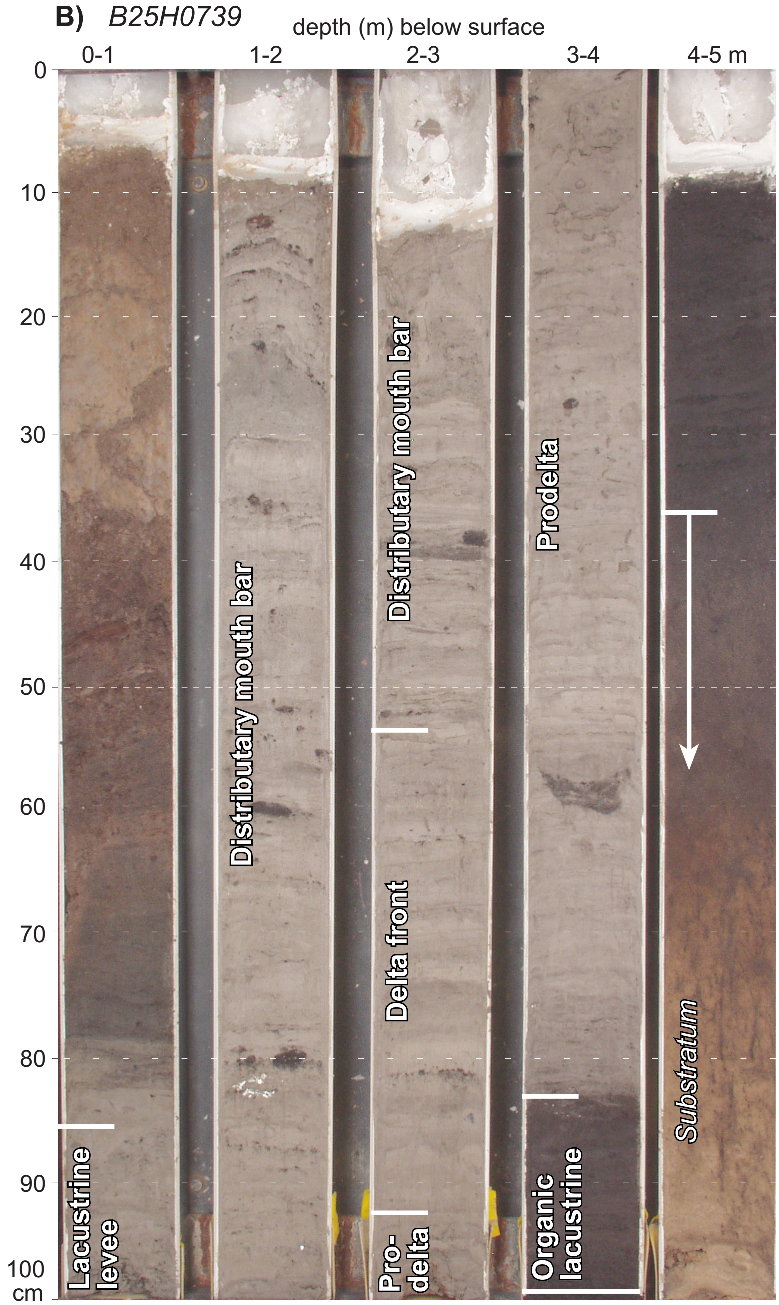


Figure 3.6. Pictures of mechanical cores in A) sandy organic-clastic lake fills and B) clayey organic-clastic lake fills. See Fig. 3.3A-B for the location and sedimentary logs. Surface elevation of the cores is -1.09 O.D. (A) and -1.78 O.D. (B). The interpretation of the sections that were not recovered – empty



parts of the tubes – is based on manual cores from the same location. See Appendix 8 for a full-colour version

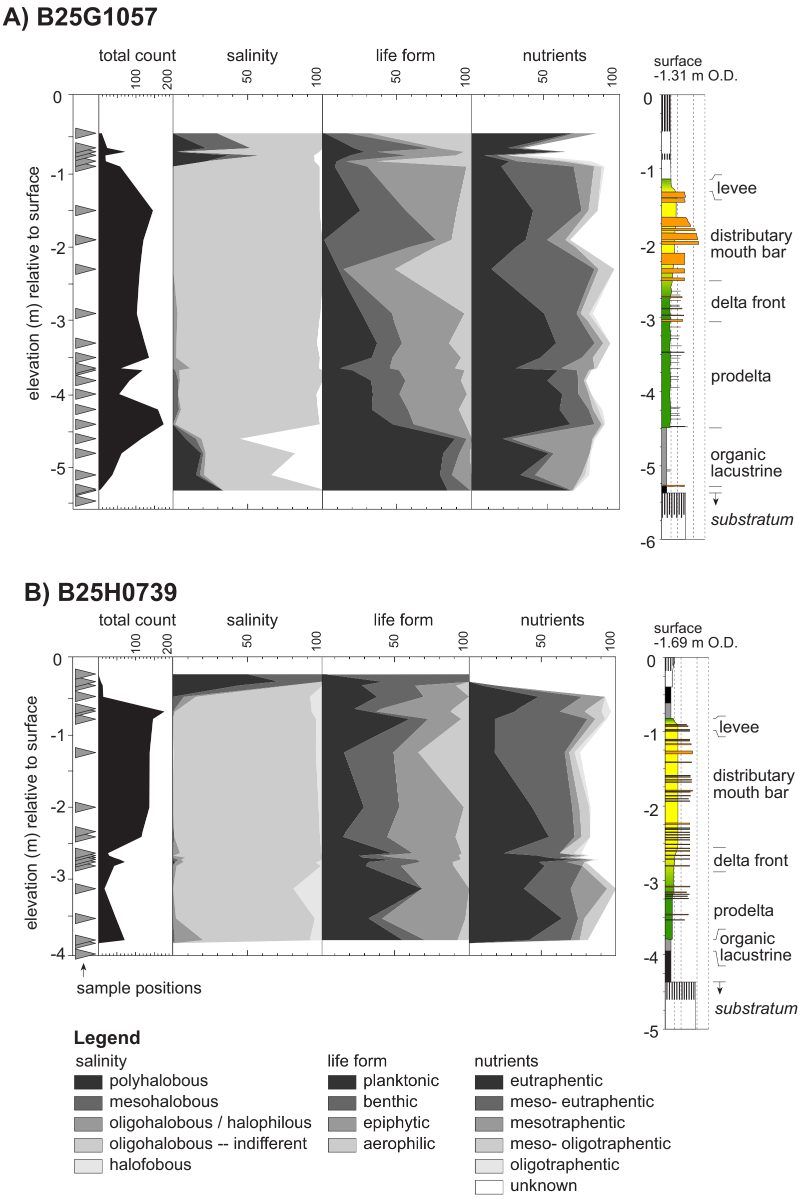


Figure 3.7. Diagrams showing the abundance of diatom assemblages in two cores penetrating organic-clastic lake fills in the Angstel-Vecht study area (modified from Bos et al., 2009; Chap. 2). The diatoms are grouped, from the left to the right diagram, in salt-tolerance classes, life-form classes, and nutrient-requirement classes. The position of the cores is indicated in Figure 3.5B. A) Core B25G1057 is positioned near the location where initially the river debouched into the lake. B) Core B25H0739 is positioned in the center of the lake, a location that received relatively little sand.

Table 3.3. AMS radiocarbon ages used in this study based on analyses of terrestrial macrofossils selected from organics (Bos et al., 2009; Chap. 2). Ages are given with 1σ range to allow comparison with OSL dates. A) ^{14}C ages for the end and beginning of sedimentation in the Aetsveldse organic-clastic lake fill. B) Combined ages for the onset of clastic sedimentation by the Angstel-Vecht river system.

A

Laboratory no.	*^{14}C age*	*Age intervals[a]*	*Mean age (cal yr BP)[a]*	*Coordinates[b] (x/y) (m)*	*Surface elevation (m ± O.D.)*	*Depth below surface (cm)*	*Sample name*	*Relevance*
UtC-14574	2920 ± 70	3205 - 3188 3164 - 2965	3080 ± 110	123751 / 464139	-1,67	177-181	Spengen 2	Onset Angstel-Vecht
UtC-14575	1577 ± 43	1415 - 1473 1478 - 1517	1470 ± 48	129542 / 470254	-0,72	194-195	Loenen 1	End Angstel
UtC-14577	2420 ± 50	2680 - 2642 2609 - 2602 2493 - 2355	2500 ± 110	128141 / 477478	-1,50	133-134	Gein 1	End Gein distributary
UtC-14582	2810 ± 50	2976 - 2851	2920 ± 70	129342 / 466569	-1,15	217-218	Weeresteijn 2	Onset Angstel-Vecht
UtC-14584	2870 ± 46	3072 – 2928	3000 ± 80	125711 / 465662	-1,63	202-203	Portengen 3	Onset Angstel-Vecht

B

nr	*Laboratory no.*	*Combined ^{14}C age*	*Age intervals[a]*	*Mean age (cal yr BP)*
C1	UtC-14574, ..82, ..84	2857 ± 31	3060 - 3053 3026 - 3014 3006 - 2924 2906 - 2891	2970 ± 50

a Applying 1σ range
b In Dutch coordinate grid (Rijksdriehoekstelsel).

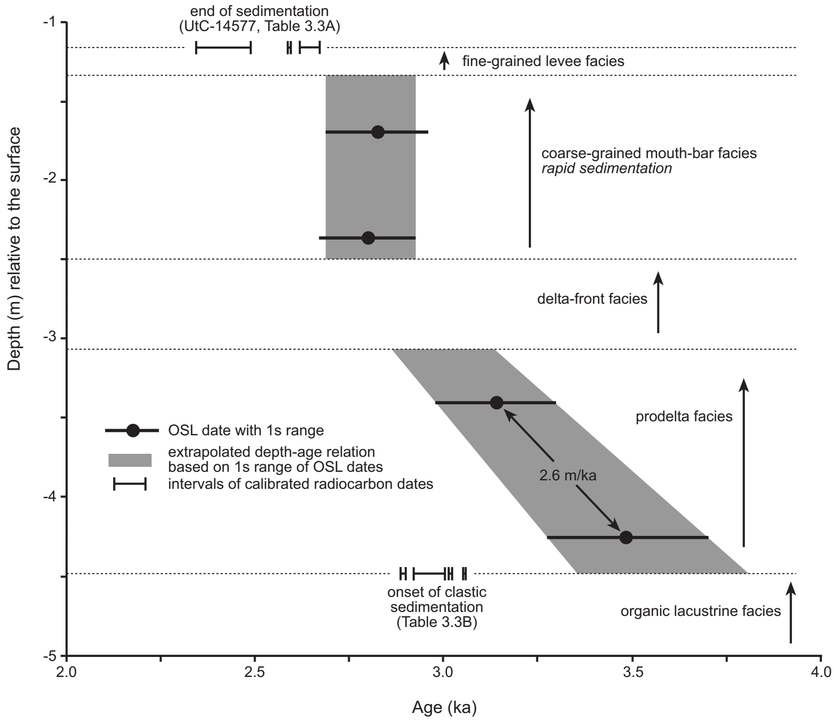


Figure 3.8. Age-depth diagram of relevant ^{14}C and OSL dates for sedimentation-rate calculations in the Aetsveld Lake (modified from Wallinga and Bos, 2010). The ranges correspond with 1σ uncertainties.

substratum more effectively in the eastern part, where the lake was shallower. The environment is mesotrophic to eutrophic, as is indicated by the diatom assemblage ("nutrients" graphs in Fig. 3.7) wherein the planktonic *Melosira italica*, *M. ambigua*, and *M. granulata* are key species. Growing conditions for benthic diatoms (low numbers of e.g., *Navicula porifera*) are not favorable ("life form" graphs in Fig. 3.7), which probably indicates that aquatic plants were abundantly present, thereby preventing sunlight to reach the lake bottom.

Prodelta facies

Prodelta facies form the bottom part of the clastic succession in organic-clastic lake fills. The lower boundary with organic-lacustrine deposits is generally gradual and non-erosional. Prodelta deposits are 0.5 to 1.5 m thick. They consist of thinly laminated (silty) clay and contain numerous very small plant remains that are organized in laminae (< 0.05 cm thick) and distributed throughout the sediment (e.g., Figs. 3.5B, 3.6). Sand laminae (< 0.1 cm thick) are not very abundant, although an increase of their frequency towards the top of this facies is often observed (for example, see the succession between -3.5 and

-3.0 m O.D. in Fig. 3.5E). The sediment contains shells – or shell fragments – of *Valvata sp.* and *Unio sp.*, indicating very low flow velocities in a freshwater environment (Dillon, 2000). This is supported by the abundance of diatoms found in this facies (e.g., *Melosira ambigua*) that require a freshwater environment (see the succession between -3.5 and 2.8 m O.D. in the "salinity" graph of Fig. 3.7B). Furthermore, (strongly) eutrophic conditions prevail, as is indicated by the diatom assemblage ("nutrients" graphs in Fig. 3.7) wherein *Stephanodiscus hantzschii* and *Cocconeis placentula* are key species.

Delta-front facies

Delta-front facies overlie prodelta deposits and underlie distributary-mouth-bar deposits (e.g., Fig. 3.4E and 3.5D, E). The delta front is differentiated from the underlying prodelta depositional facies by the abundance and thickness of sand layers. In the delta-front facies, sand laminae occur frequently and their thickness is regularly (approximately every 10 cm) over 0.5 cm. Delta-front deposits, which are 0.5-1 m thick, record a strong coarsening-upward succession: silt content of the clayey matrix increases upwards as well as the abundance and thickness of sand and silt lenses and laminae (see Fig. 3.5D, E). In the bottom part, for example, the sandy beds are thin (< 0.2 cm) and occur only sporadically. In the top part, in contrast, their thickness is up to 1 cm and the sandy beds may comprise almost half of the sediment volume in places where the delta front underlies distributary-mouth-bar deposits. The diatom assemblage includes both benthic (e.g., *Navicula porifera*) and planktonic (e.g., *Fragilaria inflata*) diatom species, which probably reflect variable growing conditions, which may correspond to alternating periods with and without sediment input.

Distributary-mouth-bar facies

Distributary-mouth-bar facies overlie delta-front facies. Distributary-mouth-bar deposits are generally 1-2 m thick and consist predominantly of fine- to coarse-grained sand intercalated with finer-grained laminae (loam). The thickness of this facies is ~1 m on average (e.g., Fig. 3.4B), but thicknesses of up to 3 m have been found (see, for example, the eastern part of cross section A-A' between -10 and -7 m O.D., Fig. 3.5A). The decrease of planktonic (e.g., *Melosira ambigua*) and epiphytic (e.g., *Achnanthes lanceolata*) diatom species (e.g., the section between -2.5 and -1.3 m O.D. in the "life form" graph in Fig. 3.7A), especially towards the top of this facies, points to less favorable conditions for plant growth. The benthic species (e.g., *Amphora perpusilla, Navicula porifera, N. costulata*) instead, which prefer sandy lake bottoms in clear water, are relatively abundant. The shift from planktonic to benthic species indicates that higher sedimentation rates and flow velocities (compared to the other organic-clastic lake-fill depositional facies) prevented plants to grow abundantly.

Levee facies

Levee facies most frequently overlie the distributary-mouth-bar facies and in most places form the top of the organic-clastic lake fill sequence. Levees are approximately 0.5 m thick, but thicknesses may be up to 2 m. Levee deposits include texture-class mixtures of sand, silt, and clay, ranging from clayey sand to sandy/silty clay. Occasionally, thin laminae (< 2 mm) consisting of pure sand occur in the lower portion of levee deposits. This depositional facies exhibits a fining-upward succession from very fine, fine, or medium coarse, clayey sand to silty clay whereas its laminated character is progressively obscured towards the top as a result of bioturbation. In contrast to the underlying facies, levee deposits lack the presence of shells (or shell remains) and peat clasts. The number of diatom species present in this facies that reside on sandy lake bottoms (e.g., *Opephora martyi, Navicula porifera*) decrease rapidly in favor of species that prefer vegetated environments (e.g., *Gomphonema angustatum var. producta,* and "life-form" graphs in Fig. 3.7). Furthermore, planktonic species (e.g., *Stephanodiscus hantzschii, Melosira italica, Melosira granulata*) are absent, which points to strongly decreased water depths.

Distributary-channel facies

Distributary-channel facies form elongated sediment bodies and occur in association with distributary mouth bars (Fig. 3.3B). It also includes the fine-grained and organic – often gyttja – channel-fill deposits that form after abandonment of the channel. Distributary-channel deposits may form the top of the sequence, and are then associated with the last stage of clastic lake-fill development, or may be overlain by levee deposits. Distributary channel deposits are up to 4 m thick and up to 0.1 km wide (Fig. 3.4), with erosional boundaries. Distributary-channel deposits are composed predominantly of fine to medium coarse sand and minor portions of clay and organics (Fig. 3.5D). Sandy lithofacies are characterized by a fining-upward succession; towards the top, very fine silty-clay lamination is regularly observed (Fig. 3.5C). The bottom portion of this depositional facies and may comprise a gravel-bearing lag (Fig. 3.5A), which probably contains reworked gravel from a local source, i.e. the sandur deposits (Angstel-Vecht area) or braided-channel deposits (Schoonhoven area). The sandy lithofacies commonly exhibits cross-bedding structures, although interpretation thereof is somewhat speculative due to core-sediment disturbance. Admixtures of peat clasts (diameter < 4 cm) and clay pebbles (diameter < 2 cm) are frequently observed (for example see Fig. 3.5C).

3.4.3 Facies distribution in organic-clastic lake fills

A typical vertical depositional-facies succession in organic-clastic lake fills comprises, from bottom to top, organic-lacustrine, prodelta, delta-front, distributary-mouth-bar, and levee facies (see for instance Fig. 3.4D), although slight deviations are also present. For example, the prodelta (Fig. 3.5A) or the distributary-mouth-bar (Fig. 3.5B), in places, may not be present. Furthermore, the succession may be truncated at any depth by a distributary channel (Fig. 3.5A). Away from the river mouth the thickness of the fine-grained depositional facies (organic lacustrine, prodelta, and delta front) increases whereas distributary-mouth-bar deposits become much thinner, less wide and finer grained. This is shown in Figure 3.5B, where sandy organic-clastic lake fills (thick distributary-mouth-bar deposits) prevail near the river mouth (near the location of cross section D-D') and become less dominant towards distal parts of the lake (for instance at the location of cross section E-E'), where prodelta and delta-front deposits are more frequently observed. Most of the distributary channels (e.g., in the Aetsveldse organic-clastic lake fill, Fig. 3.3B) terminate in the former lake and do not have a downstream connection to the main fluvial channel.

In the Angstel-Vecht area, 30% of the organic-clastic lake fill volume is composed of mouth-bar deposits, which chiefly contain sand (Tab. 3.2). The percentage of sand in individual organic-clastic lake fills ranges from 23% (Loenen) to 44% (Muiden) (Tab. 3.2).

3.4.4 Sediment supply and sedimentation rate

Sediment supply

Clastic sedimentation in the Angstel-Vecht area started after 2970 cal yr BP (Tab. 3.3B) and strongly decreased prior to 2200 yr BP (Bos et al., 2009; Chap. 2). However, low-rate sedimentation continued in remote parts of the lakes until 1470 cal yr BP (UtC-14575, Tab. 3.3A) (Bos et al., 2009; Chap. 2). The period of clastic deposition (henceforth referred to as T_c) in the Angstel-Vecht area thus is 1500 yr, although the greater part of the deposits formed during the first 770 years of that episode. The fluvial deposits in Angstel-Vecht study area contain a total of ~0.32 km^3 clastic sediment (Appendix 2), of which 0.15 km^3 (46%) is included in organic-clastic lake fills. This means that average sediment supply by the Angstel-Vecht system was at least 220,000 m^3/yr (when T_c = 1340 yr) and

Table 3.4. OSL ages in core B25G1057. The position of the core is indicated in Figure 3.3B.

NCL lab no[a]	*OSL age ± 1σ (ka)*[b]	*Dose rate (Gy/ka)*	*Equivalent dose (Gy)*	*Coordinates*[c] *(x / y) (m)*	*surface elevation (m±O.D.)*	*Depth below surface (cm)*	*Facies*
3206022	3.49 ± 0.21	1.64 ± 0.07	5.22 ± 0.14	128526 / 477358	-1.30	406-446	Prodelta
3206023	3.16 ± 0.17	1.76 ± 0.07	5.51 ± 0.16	128526 / 477358	-1.30	324-357	Prodelta
3206024	2.80 ± 0.13	1.69 ± 0.06	4.74 ± 0.12	128526 / 477358	-1.30	234-239	Mouth bar
3206025	2.83 ± 0.14	1.68 ± 0.06	4.76 ± 0.15	128526 / 477358	-1.30	163-176	Mouth bar

a Laboratory number of the Netherlands Centre for Luminescence Dating (NCL), Delft University of Technology.

b Analyses carried out in 2006.

c In Dutch coordinate grid (Rijksdriehoekstelsel)

	W^a (m)	h^a (m)	S^a (cm/km)	W/h	Fr	Q_w (m^3/s)	Bedload transport (m^3/yr)	Total-load transport (m^3/yr)	Total load / bedload	Formative duration (yr)	I_f
preferred	135	6.0	3,6	23	0,10	628	51292	262101	5.1	99	0.13
sc 1	90	4.0	3,6	23	0,12	261	30413	82899	2.7	312	0.41
sc 2	90	5.0	3,6	18	0,11	334	31116	124252	4.0	208	0.27
sc 3	90	6.0	3,6	15	0,10	408	31404	172451	5.5	150	0.20
sc 4	90	7.0	3,6	13	0,09	481	31424	227076	7.2	114	0.15
sc 5	90	8.0	3,6	11	0,09	555	31262	287774	9.2	90	0.12
sc 6	135	4.0	3,6	34	0,12	398	48432	125555	2.6	206	0.27
sc 7	135	5.0	3,6	27	0,11	512	50193	188525	3.8	137	0.18
sc 8	135	6.0	2,7	23	0,09	544	29580	127680	4.3	203	0.26
sc 9	135	6.0	4,5	23	0,11	702	76657	457871	6.0	57	0.07
sc 10	135	7.0	3,6	19	0,10	744	51947	345677	6.7	75	0.10
sc 11	135	8.0	3,6	17	0,09	861	52287	438748	8.4	59	0.08
sc 12	180	4.0	3,6	45	0,12	536	66565	168231	2.5	154	0.20
sc 13	180	5.0	3,6	36	0,11	691	69446	252842	3.6	102	0.13
sc 14	180	6.0	3,6	30	0,10	849	71429	351832	4.9	74	0.10
sc 15	180	7.0	3,6	26	0,10	1008	72800	464417	6.4	56	0.07
sc 16	180	8.0	3,6	22	0,09	1169	73731	589942	8.0	44	0.06

a input parameters

sc = scenario
W = Channel width
h = Channel depth
S = Slope of upstream channel

Fr = Froude Number
Re = Reynolds Number
Q_w = Discharge

I_f = Intermittency (effective formative duration divided by ^{14}C-based duration of activity of 770 years) probably 431,000 m^3/yr (when T_c = 770 yr). These are minimum values because sediment loss downstream of the Muiden organic-clastic lake fill (Fig. 3.3B), which probably has only been marginal (Bos et al., 2009; Chap. 2), has not been taken into account. Besides, the sediment load of the Angstel-Vecht has probably not been uniform throughout its period of clastic sedimentation.

The formative duration of the Aetsveldse organic-clastic lake fill, as calculated from sediment transport prediction, ranges between 44 yr (sc 16 in Tab. 3.5, Appendix 3) and 312 yr (sc 1 in Tab. 3.5, Appendix 3). Kleinhans (2005) showed that the uncertainty for these calculations is a factor of 3-4. The total-load/bed-load ratio for the Angstel paleochannel is 2.5 to 9.2 (Tab. 3.5), which means that sediment transport in the Angstel paleochannel has been suspended-load dominated. The width-thickness ratio (W/h) of the channel, which ranges between 17 and 45 (Tab. 3.5), is comparable to what is generally found for straight channels. The intermittency of the Angstel river (when T_c = 770) ranges between 0.06 and 0.41. Channel dimensions are probably overestimated, which means that the results of scenarios 5 and 11-16 (Tab. 3.5) have to be considered as non-optimal. Furthermore, Bos et al. (2009; Chap. 2) showed that the activity (i.e., discharge and sediment transport) of the Angstel River decreased over its full period of activity. Therefore, it is possible that during the period of full activity (770 yr) discharge and sediment transport already slightly decreased. This would mean that the channel-belt dimensions are representative only of the initial period of fluvial activity and overestimate the channel dimensions afterwards. The preferred scenario (Tab. 3.5) accounts for overestimation of the channel dimensions and yields a formative duration of 99 years and an intermittency value for the Angstel channel of 0.13. This is similar to the intermittency of the modern Rhine, which is ~0.16 and can be approximated as follows. Based on measurements carried out in 1792 in the upstream part of the Rhine, which was not embanked at that time, it was calculated that bankfull discharge was ~3400 m^3/s (Hesselink et al., 2006). Discharge measurements covering the 20th century from the same channel (Rijkswaterstaat, 2009) showed that this volume was exceeded 16% of the time, which means that the intermittency of the modern Rhine is 0.16, which can be considered as relatively high (Parker et al., 2008).

Sedimentation rate

Age constraints for clastic sedimentation in the Aetsveld Lake are provided by ^{14}C ages (Tab. 3.3). In the Aetsveldse Lake, clastic deposition started after 2970 cal yr BP (Tab.

← Table 3.5. Formative time scale of Aetsveldse Lake fed by the Angstel channel based on flow and sediment transport-rate calculations. Bold numbers indicate changed parameters with respect to measured geometry on the channel-belt deposits. Details on the applied methodology are outlined in Apendix 1. A complete overview of the calculations is given in Appendix 3. Median grain size (D_{50}) of the transported sediment was 300 μm; the sand volume in the clastic lake fill is 0.026 km^3. Scenario 5 and 11 show that overestimation of the channel depth has significant influence on the intermittency, which is also true for overestimation of the channel width (scenario 12-15) and errors on the channel gradient (scenario 8-9)

3.3B), the moment that the Angstel-Vecht river system came into existence. Sedimentation associated with the activity of the Gein distributary channel ended 2500 cal yr BP (UtC-14577, Tab. 3.3A and Fig. 3.3B). The average sedimentation rate at the location of core B25G1057, where the clastic lacustrine succession is 3.35 m thick, was 7.13 mm/yr.

The resulting OSL ages (Tab. 3.4) are in correct stratigraphical order (within uncertainties) (Fig. 3.8). The upper two samples have the same age, which is in agreement with anticipated rapid sedimentation rates in coarse-grained mouth-bar facies. The OSL ages of the two oldest samples (NCL-3206022, -23) overestimate the period of deposition as obtained with ^{14}C dating, which probably results from incorrect assumptions on the water content (Wallinga and Johns, 2008). These samples are both from prodelta facies, which means that erroneous assumptions on water content would yield similar age deviations. Therefore, it is feasible to use their relative ages for sedimentation-rate calculations. The sedimentation rate for prodelta facies, based on these OSL dates, is 2.6 mm/yr (modified after Wallinga and Bos, 2010).

3.5 DISCUSSION

3.5.1 Spatial distribution of organic-clastic lake fills in delta plains

Organic-clastic lake fills of Holocene age occur in distal zones of delta plains. Modern and recent examples of organic-clastic lake fills in the Cumberland Marshes (Smith and Pérez-Arlucea, 1994), sub-recent examples in the Atchafalaya Basin (Tye and Coleman, 1989a, b) as well as (sub-)recent examples in the Rhine-Meuse delta (Schoonhoven and Angstel-Vecht area in this study) developed in peat-bounded lakes in distal zones of these deltas. Although many questions regarding the formation of lakes remain to be answered, it can tentatively be concluded that lakes develop only when there is sediment shortage relative to the available accommodation space. Delta-plain lakes are usually incorporated in peat-forming environments (e.g., fens and bogs), which occur in places with a relative shortage of clastic deposits (Gouw, 2008). Two factors, delta-plain morphology and relative base-level rise, control the formation of accommodation space, whereas a third factor, sediment supply, controls the reduction of accommodation space. First, delta-plain widths in general increase in downstream direction. Fluvial systems therefore have more lateral space for sedimentation in distal delta plains than in their proximal counterparts. Hence conditions for the formation of excess accommodation space, and thus peat and associated lakes, are favorable in distal delta plains relative to proximal delta plains. Second, the rate of relative base-level rise is positively related to the formation of accommodation space. Commonly, therefore, early and middle Holocene conditions (high rate of relative sea-level rise) were much more favorable for the formation of accommodation space than late Holocene conditions (low rate of relative sea-level rise). I therefore contend that organic-clastic lake fills potentially are common both in lower portions of Holocene fluvio-deltaic successions that are formed during periods of high relative base-level rise, and in distal delta plains where the delta plain is generally wider. Besides, this tentatively suggests that organic-clastic lake fills are indicative for transgressive (TST) or highstand systems tracts (HST) according to sequence stratigraphic nomenclature.

The results presented in this study show that organic-clastic lake fills that develop in permanent standing water bodies have a geometry and facies distribution different than crevasse-splay deposits that form in periodically inundated floodbasins. Organic-clastic lake fills, for example, consist of a coarsening-upward sequence overlying gyttja. They do not necessarily become thinner with increasing distance from the trunk channel (Fig. 3.4E and examples in Smith et al., 1989; Tye and Coleman, 1989a; Weerts et al., 2002). Moreover, the sediment body may be sharply bounded by peat (Fig. 4E and examples in Van de Meene et al., 1988; Smith and Pérez-Arlucea, 1994) or by levees (see examples in Tye and Coleman, 1989a, b). Crevasse-splay deposits that develop in a floodbasin, in contrast, tend to have a less heterogeneous sediment body than organic-clastic lake fills. Furthermore, the sediment body of crevasse splays, excluding the crevasse-channel deposits, becomes thinner with increasing distance from the trunk channel (e.g., Smith, 1983; Farrell, 1987; Makaske et al., 2007) and includes relatively small portions of sandy lithofacies (Farrell, 2001; Stouthamer, 2001a).

3.5.2 Factors controlling the geometry and composition of organic-clastic lake fills

Clastic sedimentation in a lake will be as fast as possible, because lakes function as relative efficient sediment traps. This process has also been recognized in the tidally influenced part of the Rhine-Meuse delta, where rapid fluvial sedimentation followed marine erosion of a local basin (Kleinhans et al., 2010). The trapping efficiency, however, relates to the flow diversion at the entrance location of the river and the subsequent drop of flow velocity and transport capacity. When a lake is not sufficiently deep, the flow is not able to diverge completely, which means that discharge and sediment transport – especially of sand – is largely concentrated in a channel. Sand is not deposited in the lake, but instead is transported downstream, which prevents the formation of, for example, mouth-bar deposits. A shallow lake in which a fluvial channel debouches, therefore, does not result in a typical organic-clastic lake fill. Organic-clastic lake fills, so far, have been found only in distal zones of delta plains. This is probably because organic beds that can accommodate lakes of sufficient depth (> 2 m) are generally restricted to those portions of delta plains.

Organic-clastic lake fills potentially contain large proportions of the sand volume of a fluvial system. Almost half of the organic-clastic lake fill volume in the Angstel-Vecht area includes sand, which is 30% of the total volume of fluvial sand of the Angstel-Vecht distributary system. Sandy deposits are also prominent in the Windy Lake splay in the Cumberland Marshes, where they constitute 10-15% of the organic-clastic lake-fill volume (Smith and Pérez-Arlucea, 1994). A generic geometrical property of organic-clastic lake fills is their thickness, which typically is 2-4 m. This has been found in the Rhine-Meuse delta (this study), the Cumberland Marhses (Smith and Pérez-Arlucea, 1994; Pérez-Arlucea and Smith, 1999), and the Atchafalaya Basin (Tye and Coleman, 1989b, a). The thickness is probably determined by the depth of the lake in which they

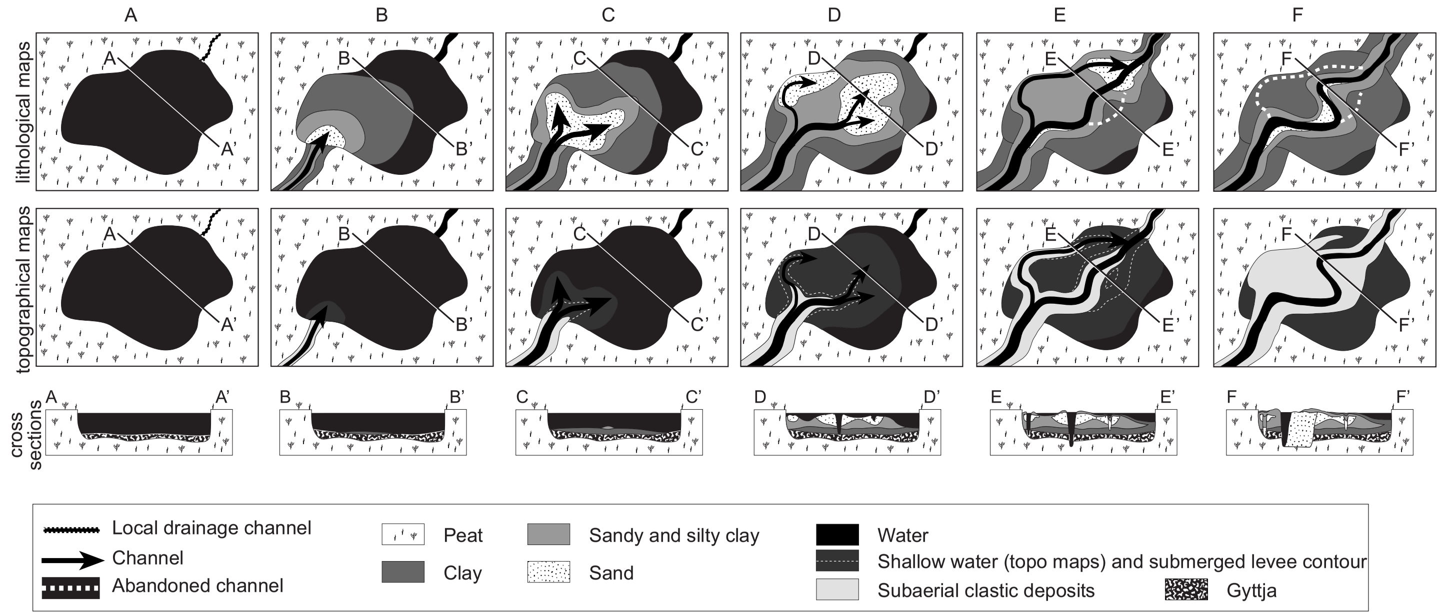


Figure 3.9. Development of organic-clastic lake fills in six time steps (A-F). The top row (lithological map) shows the distribution of the lithology present at the top of the succession. The middle row (topographical map) shows the distribution of the subaerial and subaqueous parts and the contours of submerged levees. The bottom row (cross sections) shows the lithological cross sections indicated in the maps (lithological and topographical). The unit "shallow water" roughly indicates relative shallow parts of the lake, where clay particles are still in suspension. See text for further explanation.

form. This suggests that organic-clastic lake fills do not form in lakes that are shallower than 2 m and that 4 m is the maximum depth of lakes on delta plains. This tentatively suggests that autogenic processes prevent lakes to become deeper than ~4 m. Although speculative, this may be related to intrinsic characteristics of delta-plain lakes such as bank stability. Deeper lakes may give rise to more wave-induced lake-bank erosion, thereby enlarging the lake and, by redistribution of eroded bank material, heighten the lake bottom. Consequently, water depth is reduced. Evidence that either approves or rejects the efficiency of this process, however, is not available yet. Other geometrical properties (e.g., surface area and shape) as well as the sedimentation rate are determined by site-specific factors such as size, shape, and configuration of the lake within the fluvial network as well as the ratio between the sediment transport rate and the lake volume and the period of activity of the feeding channel.

3.5.3 Development of organic-clastic lake fills: a conceptual model

On the basis of the results presented in this study (e.g., cross sections, maps, sedimentary logs) a conceptual model (Fig. 3.9) was developed that describes the facies distribution and morphology of organic-clastic lake fills. It confirms an earlier published model based on studies in the Cumberland Marshes (Smith et al., 1989) on the planform evolution of crevasse splays and organic-clastic lake fills. However, the model presented here sheds light on two implications of a lacustrine setting for the architecture and facies distribution of organic-clastic lake fills. Firstly, prior to clastic deposition, organic lacustrine deposits (gyttja) are formed. Secondly, shorelines often are relatively resistant to erosion, which therefore define the geometry of the architectural element. Furthermore, this model includes the development of the fluvial channel after organic-clastic lake-fill formation and the effects on the preservation of the clastic lake fills.

Organic-clastic lake fills form in lakes that exist prior to the input of clastic sediment (Fig. 3.9A), which is evidenced by algae gyttja (algae remains) or coarse-detrital gyttja that underlies the clastic succession. Due to wave action, the top of the substratum may be partly eroded and redeposited on the lake floor, thus becoming incorporated in the organic lacustrine facies. The initial lake probably has a small outflow channel and may also receive water from a local drainage system. The shorelines of lakes in which organic-clastic lake fills form are not strictly fixed, but may be subject to change, which has also an effect on the geometry of the organic-clastic lake fill. This change may be due to both erosion by wave action and erosion by distributary channels. Nevertheless, peat banks are relatively erosion resistant (e.g., Makaske et al., 2007; Bos et al., 2009; Chap. 2), and therefore rapid and radical changes of the lake morphology is unlikely.

At a given moment, an avulsion upstream of the lake results in the supply of clastic sediments to the lake (Fig. 3.9B). The deposited sediment close to the river mouth is coarse – mainly sand – and becomes finer-grained downstream with increasing distance to the river mouth. Levees are formed along the channel upstream of the lake. As sediment is trapped in the lakes, there is a shortage of sediment in channels downstream of lakes, which strongly limits the formation of levee and floodbasin deposits downstream of lakes. Simultaneously with avulsion, the initially small channel network has to accommodate larger discharge volumes. This results in deeper and wider channels, both upstream and downstream of the lake (Fig. 3.9B).

A – sandy – mouth bar is formed where the river debouches into the lake, which leads to the formation of a bifurcation (Edmonds and Slingerland, 2007) (Fig. 3.9C). When lake filling has prograded, each distributary may bifurcate again farther downstream (Fig. 3.9D). This may ultimately result in an irregular channel network such as shown in the Angstel-Vecht area (this study), the Windy Lake Splay (Smith and Pérez-Arlucea, 1994), and the Laitaure delta, Sweden (Andren, 1994). Occasionally, however, a symmetric dendritic distributary system develops, as is exemplified in the Mossy River delta, Cumberland Marshes (Edmonds and Slingerland, 2007).

The mouth bar and levees near the initial river mouth are fully developed and are raised above the lake level (Fig. 3.9D). During this stage and in this part of the lake, one of the initial small channels may become the dominant one. Smith et al. (1989) conclude that one channel becomes dominant when the lake is more or less completely filled. In the Aetsveldse organic-clastic lake fill a dominant channel already is established, at least in the upstream part of the organic-clastic lake fill before the lake is filled.

Once a particular part of the lake is filled, the crevasse channels are abandoned in favor of a single fluvial channel, which marks a change from a divergent to a convergent channel network. Bifurcations apparently are not stable anymore, and one downstream distributary closes. Kleinhans et al. (2008) relate this to the backwater effect, which directs flow to the outlet due to gradient advantage. Associated with lake filling, the lacustrine environment is progressively substituted by a fluvial environment including a single channel with levees and floodbasins. The efficiency of sediment trapping in the lake is negatively related to the progression of the filling of the lake. As the lake is being filled, an increasing volume of sediment is transported through the lake and enters the channel downstream of the lake, where it becomes available for levee and floodbasin sedimentation along the downstream channel reach (Fig. 3.9E).

When the lake is almost completely filled in (Fig. 3.9F), a fluvial channel crosses the organic-clastic lake fill and erodes part of it. This subsequently leads to a change in fluvial style from straight to meandering and thus in geometry of the fluvial deposits, as has been shown by Bos et al. (2009; Chap. 2). The remainder of the lake is filled in by accumulation of organics and overbank deposition. The fluvial channel is able to migrate laterally and therefore erodes the organic-clastic lake fill as long as it remains active.

3.6 CONCLUSIONS

- Organic-clastic lake fills are sediment bodies that encompass lacustrine successions in delta plains. Six depositional facies have been discerned and described in organic-clastic lake fills in the Holocene Rhine-Meuse delta: organic-lacustrine, prodelta, delta-front, distributary-mouth-bar, levee and distributary-channel depositional facies. Quantitative analyses show that up to 35% (volume) of organic-clastic lake fills is composed of distributary-mouth-bar deposits, which predominantly contain sand.

- A conceptual model is presented that illustrates the formation of organic-clastic lake fills in six sequential steps. It demonstrates that the lacustrine setting affects both the facies composition by the accumulation of gyttja and the geometry by the presence of relatively fixed lake banks.
- Variations in the geometry and size of organic-clastic lake fills are determined largely by site-specific factors. The most dominant of these are: shape, size (up to 25 km^2) and depth (2-4 m) of the lake, the orientation of the lake relative to the general flow direction, the sediment supply and the sediment transport rate, and the period of activity of the feeding channel relative to the volume of the lake.
- This study, for the first time, calculates the formative duration of a deltaic sediment body (organic-clastic lake fill) by application of a simple physics-based model and geological interpretations of both the delta sediment body and the upstream channel belt. The physics-based model along with the reconstructed channel dimensions enabled quantification of flow parameters and sediment-transport characteristics of the upstream channel. It is therefore concluded that calculation of formative duration of lacustrine deltas as carried out in this research is a powerful tool for reconstructing former flow and sediment-transport conditions such as discharge, intermittency, and sediment transport capacity.
- Lakes in which organic-clastic lake fills develop are incorporated in wetlands where peat accumulates. Favorable conditions for these wetlands and associated lakes to form are characterized by a shortage of supplied clastic sediments compared to accommodation space. These conditions commonly exist both during periods when the rate of relative base-level rise is high and in distal delta plains where the floodplain generally is relatively wide. Organic-clastic lake fills, therefore, are probably limited to transgressive (TST) or highstand systems tracts (HST).
- Organic-clastic lake fills accumulate in delta-plain lakes. Lakes are relatively good sediment traps due to diverging of the channel flow and decrease in gradient of the water surface, which lead to a strong decrease in flow velocity and thereby in sediment transport capacity. The sediment characteristics of the feeding channel are therefore strongly reflected in the composition of organic-clastic lake fills, which explains their relatively large proportion of sand, up to 35%. Because of the high sand content and its influence on channel-belt geometry in distal deltas, ignoring organic-clastic lake fills in reservoir studies on ancient fluvio-deltaic successions leads to significant underestimation of hydrocarbon reservoir volumes.

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4 Organic-facies determination: a key for understanding facies distribution in the basal peat layer of the Holocene Rhine-Meuse delta, The Netherlands

With contributions of Freek Busschers and Wim Hoek

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ABSTRACT

Peat and gyttja (organic lake sediment) are important indicators for past environmental conditions. They form in areas where the supply of clastic sediment is insufficient to fill the accommodation space. Previous studies of delta sediments, however, have focused mainly on clastic deposits whereas organics have received only minimal attention. The focus in this study is on the composition and distribution of organic facies in the Rhine-Meuse delta plain, The Netherlands. For this, we developed a field method for identifying organic facies in delta plains and we concisely described organic facies in selected cores from the Rhine-Meuse delta. Furthermore, we determined the spatial distribution of organic facies in the diachronous basal peat layer of the distal Holocene Rhine-Meuse delta by means of a semi-automated procedure to select and classify samples from a database. The resulting dataset was used for mapping organic facies in the basal peat layer using indicator kriging and inverse distance weighting. Here we show that the formation of the basal peat, which marks the onset of Holocene aggradation, has been controlled by marine-dominated, fluvial-dominated and seepage-dominated environments. Before 9000 cal yr BP, marine processes influenced the initial stage of organic accumulation in the western part of The Netherlands. After 9000 cal yr BP, on the flanks of topographical higher regions, seepage-dominated mesotrophic organic facies characterized the onset of accumulation. Simultaneously, nutrient-rich organic facies could develop in the Rhine-Meuse valley: gyttja formed in the sediment-limited Meuse realm whereas reed peat accumulated in the Rhine realm, indicating lower water depths and thus a higher elevated surface level. Reconstruction of a large set of environmental conditions is feasible by means of organic-facies determination. Organics host an archive, for example for water depth, with which accommodation space can partly be reconstructed. Identification of organic facies and determination of their spatial distribution at delta scale, as is shown in this study, is valuable for understanding delta evolution.

4.1 INTRODUCTION

Deltas are complex sedimentary systems composed of clastic and intercalated organic deposits. In most sedimentological delta studies, emphasis is on clastic sedimentary facies whereas organic deposits are usually considered as uniform units and facies descriptions are often lacking. From ecological studies in present delta plains it is known, however, that there is a large variety of plant communities colonizing these wetlands (e.g., Dirschl, 1972; Dirschl and Coupland, 1972; Kosters et al., 1987; Baldina et al., 1999). Also studies on recent and subrecent organics from a large variety of environments show a wide range of organic facies (Succow and Joosten, 2001). Here, organic facies refers to environmental conditions that prevailed during their formation. Currently, it remains poorly known which organic facies are present in fluvio-deltaic successions and how these are distributed at delta scale. Knowledge thereof can contribute in various ways to understand delta formation. Organic facies have been formed under different palaeoenvironmental conditions and hence can be used for palaeoclimatological, palaeogeographical and palaeohydrological reconstructions. Also, different geomechanical and geochemical properties may have consequences for both subsidence susceptibility and the suitability of organics as source rock for fossil fuels.

Only few studies in recent and subrecent delta-plain settings differentiate between organic facies including both organic lacustrine deposits – gyttja – as well as various types of in situ accumulated peat. Most of these studies are clustered in four regions, being the Fraser delta, Canada (Styan and Bustin, 1983a, b), the alluvial plain of the continentally positioned Cumberland Marshes, Canada (Dirschl and Coupland, 1972; Davies-Vollum and Smith, 2008), the Mississippi delta plain, USA (e.g., Frazier, 1967; Kosters et al., 1987) and the Rhine-Meuse delta, The Netherlands (e.g., Pons and Van Oosten, 1974; Van der Woude, 1981). Documented facies descriptions, however, either have a general character or strongly rely on palynological data. This limits the feasibility of organic-facies identification, especially in the field, in other regions by a wider scientific community. Besides, a classification key for identifying organic facies, based on field characteristics, lacks. Earlier proposed description and classification schemes for organics (e.g., Dirschl, 1972; Hobbs, 1986; Myślińska, 2003) usually discriminate between various levels of decomposition resulting in amorphous peat and very fibrous peat at the extremes and various stages in between (Hobbs, 1986). Alternatively, classifications discriminate between peat and gyttja on various trophic levels, partly also on the basis of organic content (Succow and Joosten, 2001; Myślińska, 2003). Although the latter scheme yields a distinction of few organic facies, a descriptive framework for organic facies in delta plains is lacking yet. Hence, there is a need to develop a classification procedure and a field key related to concise field descriptions for organic facies in delta plains.

The wealth of subsurface data for the Holocene Rhine-Meuse delta and the availability of a 3D subsurface model (TNO, 2009) for the distal delta plain offer a unique opportunity to undertake the first crucial steps towards understanding the spatial and temporal distribution of organic facies in delta plains. Of special interest is the so-called basal peat layer that occurs at the base of the Holocene coastal prism. Despite its time-diachronic nature, this layer is regionally widespread and can easily be recognised in cores and archived core descriptions. Examination of the spatial distribution of the organic facies in this layer, therefore, provides a unique framework for exploring 3D regional-scale peat-type variation and factors that controlled the onset of organic aggradation.

The aims of this study are 1) to develop a field method for classifying organics in combination with concise facies descriptions and 2) to determine factors that controlled both the spatial and temporal distribution of organic facies in the diachronous basal peat layer in the distal Holocene Rhine-Meuse delta plain.

4.2 GEOLOGICAL SETTING

4.2.1 Substratum of the Holocene Rhine-Meuse delta plain

The substratum of the Holocene Rhine-Meuse delta comprises late Pleistocene and early Holocene fluvial and aeolian sediments (Kreftenheye Fm and Boxtel Fm, cf. Westerhoff et al., 2003b; Fig. 4.1-4.3). In the central part of the area (further referred to as 'palaeovalley', Fig. 4.1), the top of the substratum is several meters lower than to the north and south due to fluvial incision during the late glacial and early phase of the Holocene. The palaeovalley separates two areas where a thin layer (1-2 m) of aeolian deposits (so-called 'coversand', Boxtel Fm) forms the top part of the substratum. Further, north of the Holocene delta, Saalian-age (MIS-6) ice-pushed ridges occur, of which the highest points reach over 50 m above Dutch Ordnance Datum (O.D., ~mean sea level). The deposits in the palaeovalley consist of fluvial sands and gravels that are frequently covered by a distinct loam bed (Wychen Bd, Kreftenheye Fm) including an often prominent palaeosol (Törnqvist et al., 1994; Autin, 2008). Abundantly present in the southern part of the palaeovalley are aeolian river dunes of Late Weichselian to early Holocene age (Delwijnen Mb, Boxtel Fm, Fig. 4.2). The coversand deposits, present outside the palaeovalley, consist of carbonate-poor homogeneous fine quartz sands (Wierden Mb, Boxtel Fm).

4.2.2 Holocene Rhine-Meuse delta plain

The Holocene Rhine-Meuse delta formed in response to post-glacial relative sea-level rise and associated rise in groundwater level. Initially, the thus formed accommodation space was filled by organics (the basal peat layer, Nieuwkoop Fm), except for areas close to river channels where clastic deposition prevailed. Basal peat, in this study, is defined as the organic bed that directly overlies the substratum (i.e., Kreftenheye Fm (often Wijchen Mb) or Wierden Mb, Boxtel Fm) and is an easily identifiable unit that marks the onset of Holocene aggradation (Vermeer-Louman, 1934; Bennema, 1954; Jelgersma, 1961). Basal peat does not include only peat sensu stricto, as the term suggests, but comprises organics in general, including organic lacustrine deposits, i.e., gyttja.

During the Holocene, the fluvio-deltaic wedge expanded both upstream (eastwards) and lateral (invading the coversand area), thereby covering the basal peat layer. Lateral expansion occurred earlier in the western part of the area because the inherited Pleistocene

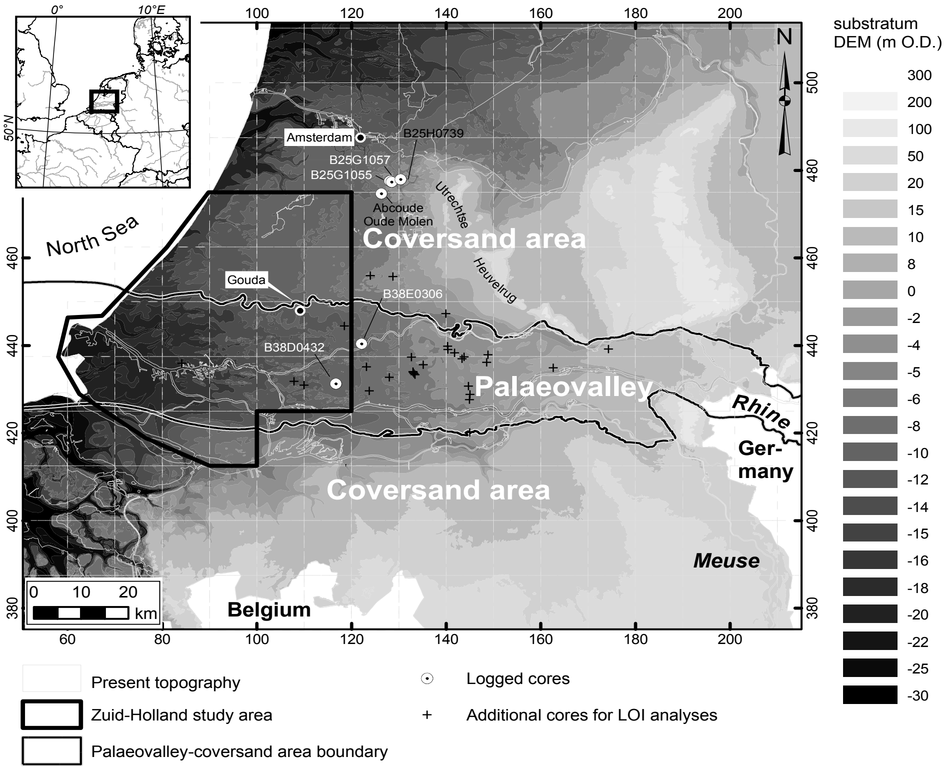
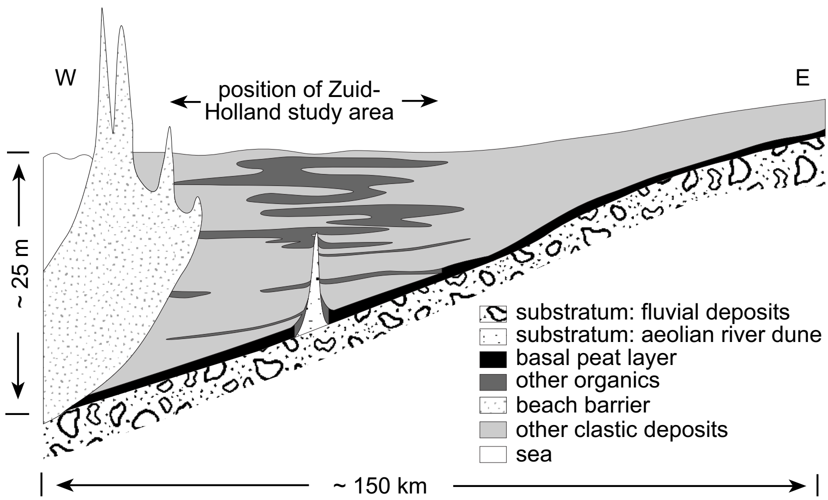


Figure 4.1. Digital Elevation Model of the substratum in the central part of the Netherlands, including the Rhine-Meuse delta. The positions of logged cores (Fig. 4.5) are indicated as well as the area for which we determined the spatial distribution of organic facies (Fig. 4.7B). The palaeovalley, which underlies in the centre of the present delta, is constrained by coversand deposits to the north and south. Coordinates are according to the Rijksdriehoekstelsel (Dutch national grid system).

and early Holocene substratum dips in western direction (~0.2 m/km; Stouthamer and Berendsen, 2000). Owing to the presence of an extensive area with low-lying substratum, especially to the north (roughly between Gouda and Amsterdam, Fig. 4.1), lateral expansion of the distal delta plain continued.

The fluvio-deltaic wedge of the Holocene Rhine-Meuse delta includes intercalated marine, fluvial and organic beds (Naaldwijk, Echteld and Nieuwkoop Formations, respectively; Westerhoff et al., 2003; Fig. 4.2), which are dissected by channel sand bodies of various sizes (Berendsen and Stouthamer, 2000; Berendsen and Stouthamer, 2001). Gouw and Erkens (2007) found that during the Holocene, three factors successively controlled the formation of the delta: eustatic sea-level rise prior to 5000 cal yr BP, basin subsidence between 5000 and 3000 cal yr BP and sediment supply after 3000 cal yr BP. Accumulation of organics was most prominent between 5000 and 3000 cal yr BP and



← *Fig. 4.2. Schematic longitudinal cross section in the palaeovalley of the Holocene Rhine-Meuse delta, indicating the main depositional units.*

mainly occurred in the distal part of the delta (Berendsen and Stouthamer, 2001; Gouw and Erkens, 2007) when the existence of beach barriers limited marine influence in the inland area. The volumetric proportion of organics in that part of the delta is expected to be more than 30% (Gouw, 2008). Various workers distinguished organic facies in the (distal) Rhine-Meuse delta plain (Vermeer-Louman, 1934; Visscher, 1949; Bennema, 1951; Poelman, 1966; Pons and Van Oosten, 1974; Van der Woude, 1981; Van de Meene et al., 1988; Bosch and Kok, 1994; Van Asselen and Bos, 2009). Commonly, categories like wood peat, reed peat, sedge peat and *Sphagnum* peat were described. Additionally, some studies (Pons and Van Oosten, 1974; Van der Woude, 1981; Van de Meene et al., 1988; Bosch and Kok, 1994) reported the presence of other facies such as reed-sedge peat and gyttja. Further, Bosch and Kok (1994) mentioned the presence of *Sphagnum* peat in the western portion of the central delta (west of Gouda, Fig. 4.1). These and other oligotrophic peat accumulations that once were prominent in the delta plain outside the fluvial realm, however, have largely been extracted for fuel from medieval times onwards.

Natural sedimentation in the delta largely ended between 1150 and 1350 AD due to the construction of dikes and dams (Lambert, 1985).

4.3 METHODS

4.3.1 Classification and description of organics

Six continuous mechanical cores (Abcoude Oude Molen (AOM), B25G1055, B25G1057, B25H0739, B38D0432 and B38E0306, positions given in Fig. 4.1), obtained with a bailer drilling device, were examined macroscopically and documented in logs. Site selection aimed to include all relevant organic facies and, therefore, cores were positioned both inside

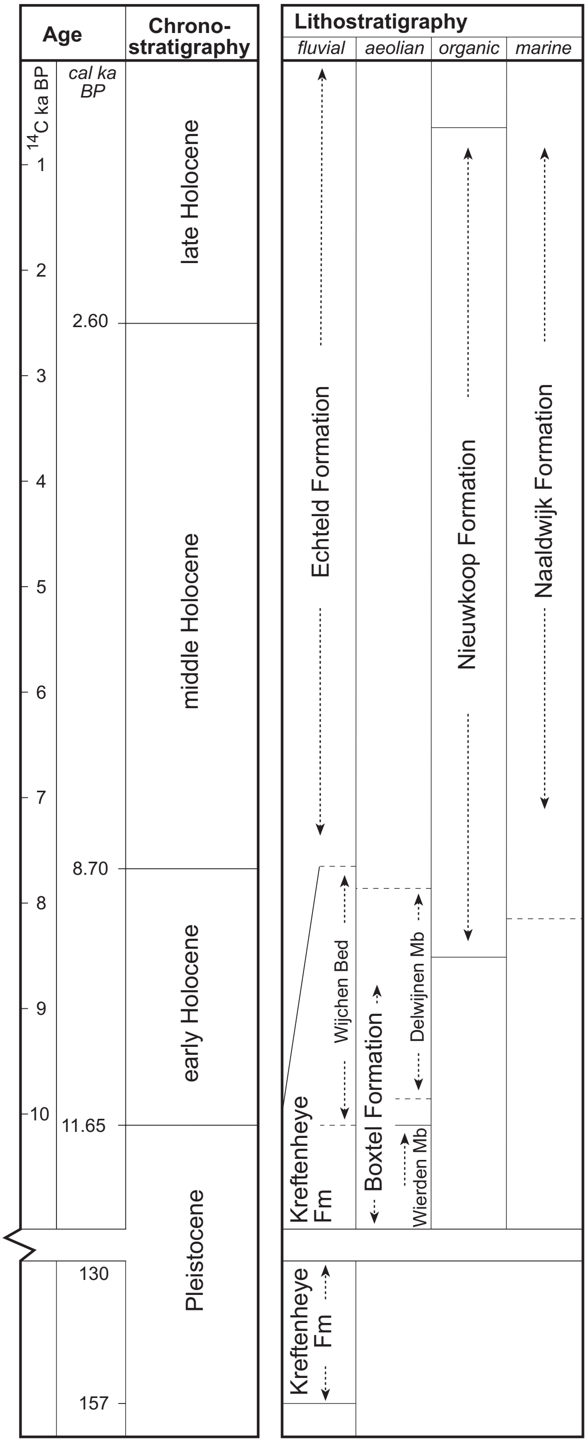


Figure 4.3. Overview of chronostratigraphic (Walker et al., 2009) and lithostratigraphic (Westerhoff et al., 2003b; Hijma et al., 2009) units for the Rhine-Meuse delta.

Table 4.1. Classification key for organics based both on literature cited in the text and on field experience. Although the key is designed for organics in the Rhine-Meuse delta, it could be used in other delta plains in the temperate climate zone, probably with slight modifications.

applicable for freshly-cut material	
The colour of freshly cut material is grey, brown-grey, yellow-grey	CLASTIC FACIES
The colour of freshly cut material is brown, green-brown, grey-brown, yellow-brown, white-brown, red-brown, or black	ORGANIC FACIES
For identification of organic facies, the material should be pulled loose	
1. The material lacks sedimentary structures and recognizable plant remains, is often coloured black	→ amorphous organics
2. The material is (finely) laminated, contains abundant plant remains and may strongly react to 5% HCl, and may contain shells, shell fragments, ostracods and other small aquatic animal remains and clastic laminae/lenses (sand/silt/clay), breaks open following the laminations	→ gyttja
2.1. The deposits consist of predominantly coarse plant remains (e.g., wood pieces, twigs, leaves)	→ coarse detrital gyttja
2.2. The deposits consist of a fine-textured mass, generally green-brown, and may contain abundant very small plant remains (<1 mm), may have a gummy-like appearance	→ fine detrital gyttja
2.3. The deposits predominantly consist of white-brown material, which may appear as brittle concretions or as a very fine textured mass, abundant shell fragments, opercula or Chara stems and strongly reacts to HCl	→ calcareous gyttja
2.4. The deposits predominantly consist of yellow-brown material, usually represented as stains, which may appear as brittle concretions or as a very fine (Ø <1 mm) textured mass, and reacts to HCl, though only at higher temperatures (>30°C)	→ siderite gyttja
3. The material is composed of plant remains and is strongly rooted	→ peat
3.1. The material consists of wood fragments, which may be embedded in a fine-textured matrix	→ wood peat
3.2. The dominant components are 5-10 mm wide and elongated, curly, yellowish plant remains (reed culms, rhizomes and roots), which are generally accompanied by fine roots (Ø <1 mm)	→ reed peat
3.3. The dominant components are flattened, 0.5 - 2.0 mm wide, elongated, light-brown, curly plant remains (sedge roots), often accompanied by 1.0-2.0 mm round orange-brown (or black when oxidized) seeds of *Menyanthes*	→ sedge peat
3.4. The dominant components are either (1) fine moss leaves and stems, small moss layers can be peeled, (2) highly fibrous, rope-like (*Eriophorum*), (3) 1 cm long, lancet shaped soft plant remains (*Sphagnum*) or (4) red curly *Ericales* roots	→ oligotrophic peat

(n=2) and outside (n=4) the palaeovalley. Classification of organics was based on a key (Tab. 4.1) that was designed for this study, based on field experience and earlier published descriptions of organics in the Rhine-Meuse delta (for references see section Holocene Rhine-Meuse delta plain). Furthermore, lithological and sedimentary characteristics were documented, such as colour and texture, sedimentary structures, rooting and botanical and clastic content. Certain intervals with organics lacked both sedimentary structures and macroscopically identifiable botanical components. These samples were classified as amorphous organics. The amorphous appearance of the material may be due to the nature of the peat-forming vegetation. Plant remains of common rush (*Juncus effusus*), for example, are not preserved and hence, the associated peat type is amorphous and can only be identified by analyzing botanical macro remains microscopically (see Appendix 7). Other reasons for organics to become amorphous are pedogenic processes (e.g., bio-activity, oxidation), which potentially completely obscure both sedimentary structures and botanical components and always result in freshly cut material to appear blackish. Compaction of organics as a result of loading, in addition, may limit the identification of organic facies in the field as it hampers both parting of the material and picking of individual (botanical) components. Amorphous and unidentified organics have not been analysed further.

Additional information on palaeovegetation and palaeoenvironment was obtained from the analyses of macroremains and diatoms, respectively. Macroremain analyses were conducted on the transition trajectories from substratum to basal peat in two cores (10 samples in B25G1057 and 12 samples in B25H0739) in the coversand area. We obtained quick-scan diatom counts from 18 samples from core B38D0432 (Fig. 4.1), which was positioned at the same location as an earlier core that was analysed palynologically (core Molenaarsgraaf H1110 in Van der Woude, 1981). Quick-scan analyses involve species counting and only crude estimations of their abundance. The results give only rough indications of environmental conditions, but may be helpful to distinguish between, for example, marine and fluvial conditions.

Organic content measurements on organic sections in 30 cores (see Fig. 4.1 for locations and Appendix 4 for sample details) were used to investigate the relation between clastic input (which can indicate fluvial, marine or aeolian input) and organic facies (partly unpublished data of Van Asselen). 'Loss on ignition' (LOI, weight %), as outlined by Heiri et al. (2001), was calculated as

$$LOI_{550} = \frac{DW_{105} - DW_{550}}{DW_{105}} * 100 \qquad \text{[Eq. 4.1]}$$

where DW_{105} = dry weight bulk sample and DW_{550} = weight of the mineral components. LOI was also measured on the organic sequence of core B25G1057 and on the total recovered Holocene succession cores AOM, B25G1055, B25H0739, B38D0432 and B38E0306. Samples covered intervals of 5 cm and were taken contiguously, such that continuous LOI profiles were obtained, which were only interrupted by unrecovered intervals. The volume of each individual sample was 5 cm^3. Dry weight was measured after heating the sample in a crucible for 24 hours at 105 °C. The weight of the mineral components was measured after heating the sample in a crucible for 4 hours at 550 °C.

Calcium carbonate content was measured using the Scheibler method as described by

Table 4.2. *Literature overview of identified organic facies in delta plains in areas with a temperate climate. (A)* In situ *peat and (B) gyttja types.*

A

	trophy facies	*OLIGOTROPHIC PEAT*					*MESOTROPHIC PEAT*			*EUTROPHIC PEAT*		
	botanical facies						*(REED-)SEDGE PEAT*			*WOOD PEAT*		*REED PEAT*
Reference	↓ *region* / *botanical facies (subdivided)*	*Nuphar peat*	*Erica/Calluna peat (oligotrophic)*	*Erica-Sphagnum peat*	*Sphagnum peat*	*Sedge-Sphagnum peat*	*sedge peat*	*reed-sedge peat*	*Sedge-wood peat*	*White spruce peat*	*Willow-alder peat*	*reed peat*
This study	Rhine-Meuse delta, The Netherlands		*		*		*	*			*	*
Gradziński et al., 2003	Narew River, Poland							*			*	*
Badina et al., 1999[1]	Lower Volga delta, Russia	?						*			*	*
Styan & Bustin (1989ab)	Fraser delta, Canada	*		*	*	*	*	?	*			
Dirschl, 1972[1]	Cumberland Marshes, Canada	?					*			*	*	*

B

Reference	↓ *region* / *sedimentary facies*	*fine detrital gyttja*	*coarse detrital gyttja*	*calcareous gyttja*	*siderite gyttja*
This study	Rhine-Meuse delta, The Netherlands	*	*	*	*
Gradziński et al., 2003	Narew River, Poland				
Badina et al., 1999[1]	Lower Volga delta, Russia	?			
Styan & Bustin (1989ab)	Fraser delta, Canada	*			
Dirschl, 1972[1]	Cumberland Marshes, Canada	?			

1 study in modern environment: predominantly peat-forming vegetation has been reported, not botanical peat content.

Hoek and Bohncke (2001). Samples with a volume of 1 cm^3 were dried and crushed. Five percent HCl was added to 250 or 500 mg of material (according to anticipated $CaCO_3^-$ content) in a closed cylinder to separate $CaCO_3$ into $CaCO_3^{2-}$ and HCO_3^- ; the escaping CO_2 gas was measured volumetrically.

4.3.2 Spatial distribution of organic facies in the basal peat layer

We limited the spatial analysis of the organic components of the basal peat to the Zuid-Holland area, which is positioned in the distal part of the delta, roughly the area of the Dutch province of Zuid-Holland (Fig. 4.1). For that area, a 3D subsurface model has recently been developed, which enabled automated selection of cores that include the basal peat layer (TNO, 2009).

To obtain the spatial distribution of organic facies in the basal peat, we used digitally archived core descriptions from the geological database of TNO Built Environment and Geosciences, Geological Survey of The Netherlands (TNO, 2009). Borehole descriptions include, amongst many other variables, texture, colour(s), quantitative and qualitative records for organic content, botanical content, admixtures (clay pebbles, peat clasts, etc.), sedimentary structures, carbonate content, shell content and occasionally also indications of the depositional facies. Analyses of this database involved six sequential steps (Fig. 4.4) that resulted in a facies map. These steps were (1) selection of all cores that include the basal peat layer, (2) selection of the lowermost organic-rich sample in each core, (3) primary and (4) secondary classification of the organic sample, (5) exclusion of conflicting cores and (6) interpolation of all remaining samples that were classified in step 3. Details on these procedures are outlined below.

In the Zuid-Holland area (Fig. 4.1), the DINO database provides us with over 50,000 borehole descriptions, which were queried in a number of sequential query steps using Python-based software and a spatial grid. A first selection included all cores that reached the substratum utilizing an existing DEM of the top of the substratum (Vos and Kiden, 2005). By 1) examination of an interval (+/- 2.5 m) surrounding this depth and 2) by analysing each individual core in upward direction, we were able to identify the basal peat as the first peat or gyttja layer that was encountered (step 1 in Fig. 4.4). This automated procedure, in which all 50,000 cores were evaluated, resulted in approximately 8000 boreholes containing basal peat. This selection also excludes basal peat samples on top of aeolian river dunes (Delwijnen Member, Boxtel Formation). This selection yielded a relatively high data density in the eastern part of the study area, being ~4.4 cores/km^2, which decreased in westerly direction to 0.2 cores/km^2 near the present coast (see also Appendix 5). The basal peat on dune slopes is dominated by wood peat due to dryer conditions (e.g., Van der Woude, 1981). This may result in short-distance gradients in depositional conditions, which would obscure facies differences that potentially are present in areas between the dunes.

For each core where the basal peat consisted of two or more superimposed layers, we selected the lowermost layer that contained organics without or with only minor

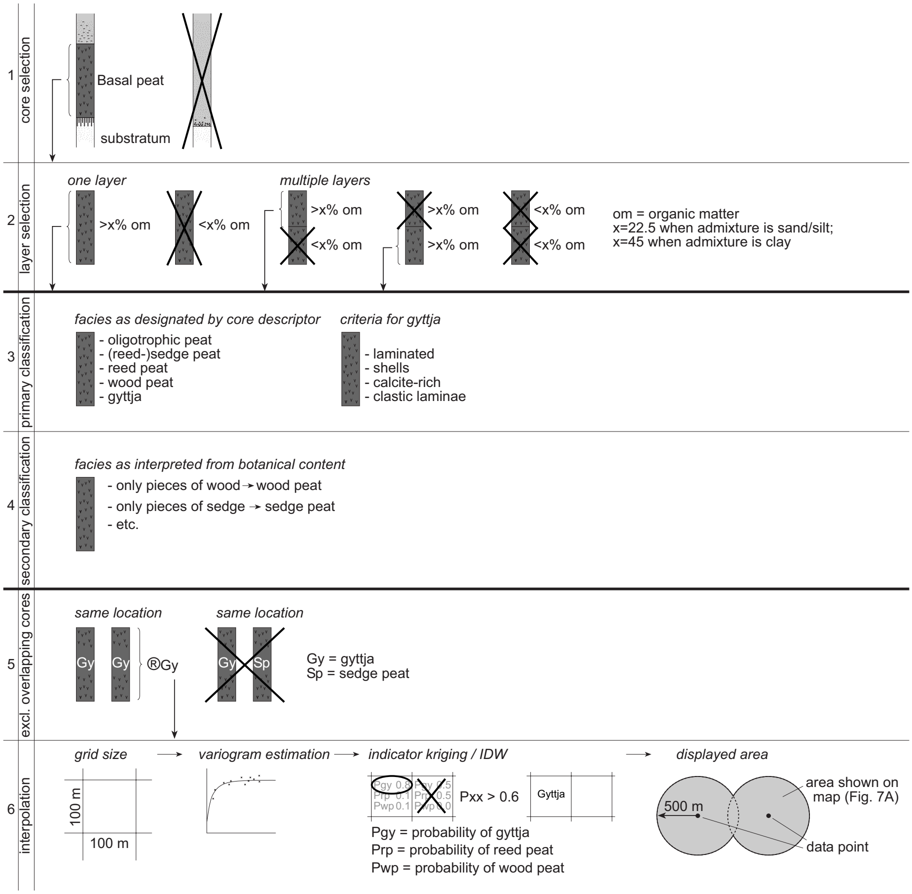


Figure 4.4. Overview of selection, classification and interpolation procedures on archived cores for identification and mapping of basal peat organic facies.

quantities of clastic admixtures (step 2 in Fig. 4.3). By selecting the lowermost sample (0.62 m thick on average), we ensured that the selected layer represents the first period of Holocene aggradation at that location. Owing to the presence of a palaeosol in the top of the substratum and the presence of organic matter therein, occasionally the palaeosol was included in the basal peat in the DINO database. The selection of samples without or with little clastic admixture ensured that the samples originated from a position above the substratum.

The classification of organics was based on the classification key developed for this study (Tab. 4.1) and on documented facies characteristics (see section Results and Fig. 4.5). The primary classification (step 3 in Fig. 4.4), in which 44% of the samples were classified, differentiated between gyttja and in situ peat and, subsequently, classified samples that were assigned to a specific organic facies by the core describer. Based on documented sedimentary sample characteristics, irrespective of facies identification by the core descriptor, 2% of the samples that were recorded as peat in situ in the database were reclassified as gyttja. The other criteria were used for the secondary classification (step 4 in Fig. 4.4), which was conducted on the remaining set of unclassified cores. Samples that contained large quantities of reed without having significant amounts of wood, sedge and oligotrophic plant remains, for example, were classified as reed peat. Next, overlapping cores (tolerance range of 1 m) where excluded when conflicting (step 5 in Fig. 4.4). From overlapping cores with identical organic facies, only one sample was used for the analyses. Finally, primary classified samples (n=3411) were interpolated (step 6 in Fig. 4.4) to produce a basal peat organic-facies distribution map. Secondary classified samples were not included for interpolation because they, in places, yielded facies distribution patterns that correspond with database heterogeneity (see sections Results for further details).

Interpolation was carried out separately for the palaeovalley and the coversand area (Fig. 4.1) because of anticipated differences between both regions, e.g., in correlation length. Furthermore, we only included reed peat, wood peat and gyttja in the interpolation. It was not feasible to differentiate gyttja types by database queries, possibly because of the scarcity of gyttja types other than algae gyttja. Gyttja, therefore, was included as single unit in the interpolation. Subsets of cores that were classified as oligotrophic peat or (reed-) sedge peat were not included in the interpolation procedure. They did not contain the minimum of 30-50 disjoint pairs for variogram determination as proposed by Journel and Huijbregts (1978). We determined the variograms of the organic facies to characterize the spatial correlation (see Appendix 6 for details). We used indicator kriging if the samples of an organic facies class were spatially correlated and inverse distance weighting (IDW) for those facies that were not spatially correlated. Grid-cell size was set at 100 m (Fig. 4.4). The probabilities for each cell were then normalised. When one of the facies in a grid cell had a probability of over 60%, that facies was assigned to that cell under the restriction that it was less than 500 m away from the closest classified sample. Otherwise the grid cell was left blank.

4.4 RESULTS

4.4.1 Organic-facies characteristics

Division of organics is based on sedimentary characteristics and distinguishes peat in situ and gyttja. Further division of peat is based on commonly used trophic levels (Succow and Joosten, 2001), being oligotrophic, mesotrophic and eutrophic peat. Mesotrophic and eutrophic peat are subdivided according to botanical content. Gyttja types, additionally, are differentiated based on sedimentary (mainly lithological) characteristics. We could identify six botanical peat facies (*Erica-Calluna* peat, *Sphagnum* peat, sedge peat, reed-sedge peat, reed peat and wood peat), and four sedimentary gyttja facies (fine-detrital gyttja, coarse-detrital gyttja, calcareous gyttja and siderite gyttja). The field characteristics and environmental conditions in which these facies form are outlined below in six sections: oligotrophic peat undifferentiated, mesotrophic peat, reed peat, wood peat, detrital gyttja and carbonate dominated gyttja. We combined various facies for practical reasons. Oligotrophic peat is a large and heterogeneous group of peat types that accumulates under nutrient-poor conditions. Strictly, oligotrophic peat includes a number of different organic facies. However, most of these facies are rare and of limited areal extent, which is why we treat these facies as a single unit. Gyttja is described in two sections, one that includes detrital gyttja, both fine and coarse, and one that includes gyttja consisting of chemical precipitates, calcareous and siderite gyttja.

Oligotrophic peat: undifferentiated

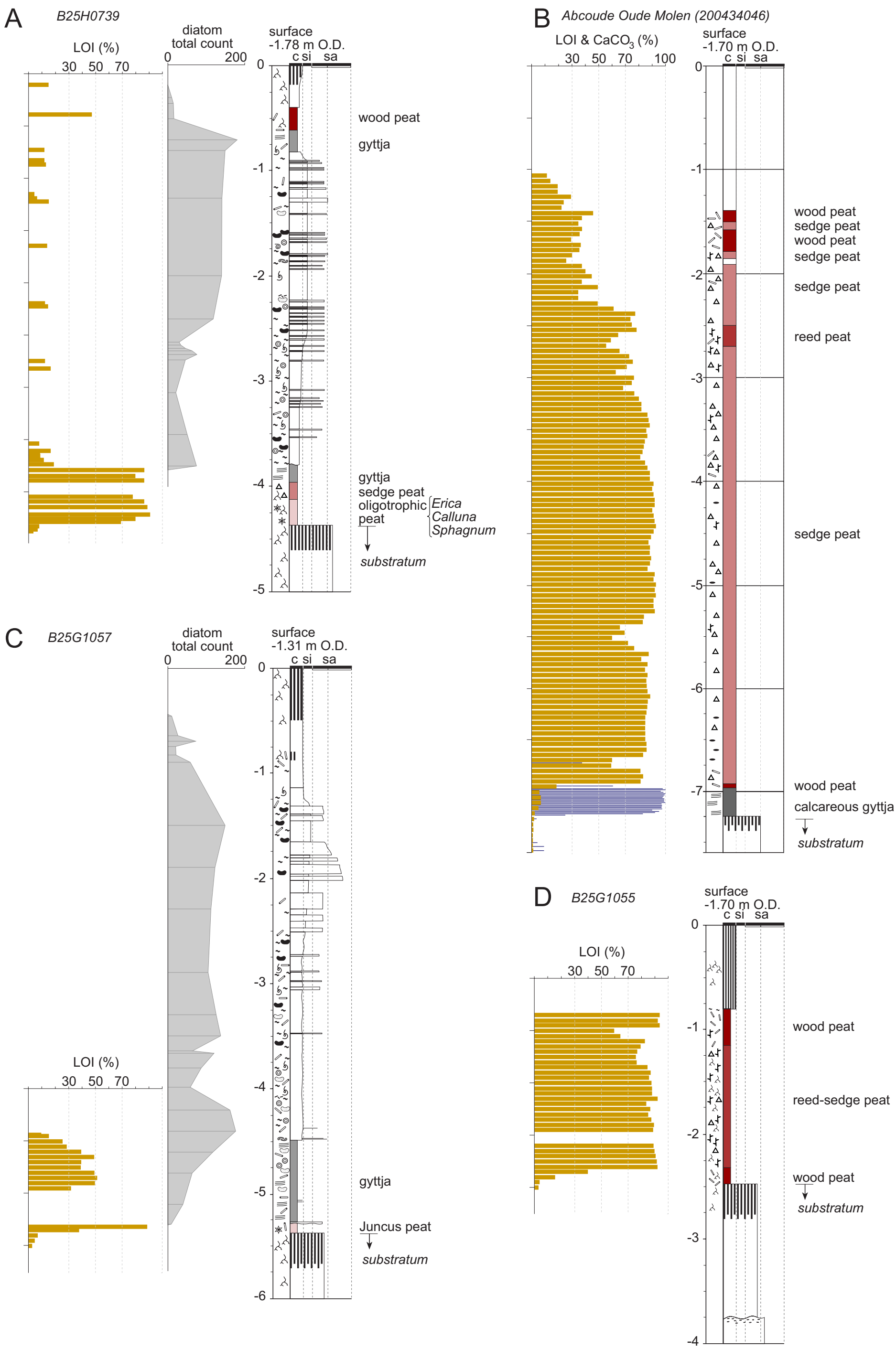
Common oligotrophic peat types are composed of peat moss (*Sphagnum spp.*), cottongrass (*Eriophorum sp.*) or heather, e.g., common heather (*Calluna vulgaris*) and cross-leaved heath (*Erica tetralix*) (Fig. 4.5A). Most of these species can fairly easily be identified based on their macro remains. Peat moss has tightly arranged clusters of branches and leaves. It is usually somewhat spongy and contains significant proportions of water, which can be squeezed out by hand. Cotton grass can readily be recognized by its well-preserved hairy, fibrous inflorescences (flower clusters). The LOI value of oligotrophic peat in our cores ranges from 69% to 91% (Fig. 4.6), which is low compared to LOI values (> 95%) that are common in bog peat (Unkel et al., 2008). The relatively low LOI values of the samples in our study can be explained by their position relative to the substratum: these samples were collected from organics that directly overlie sandy Pleistocene deposits, which consist almost completely of quartz grains. As a result of bioturbation and aeolian transport, sand grains probably became incorporated in the peat thus yielding relatively low LOI values.

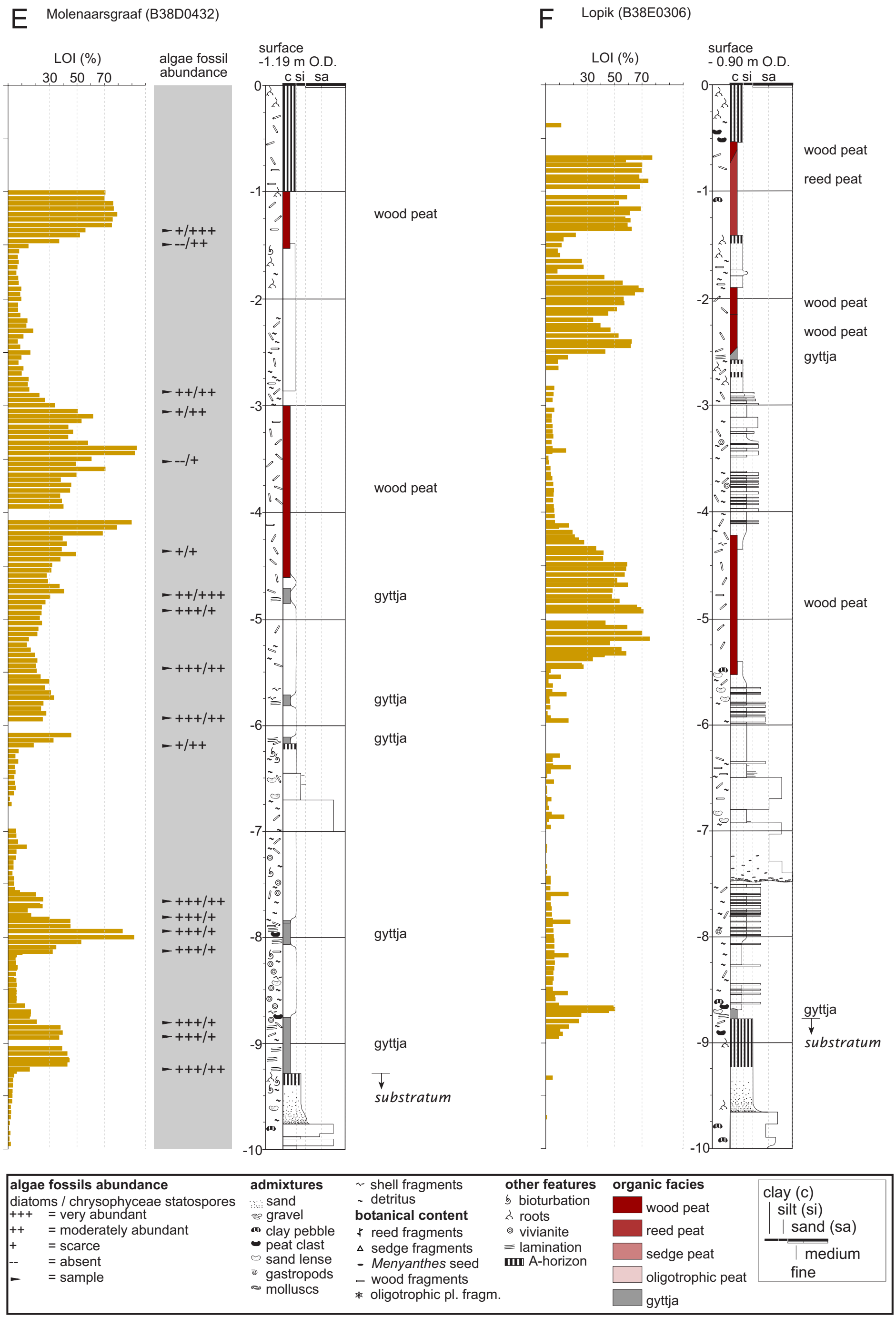
Oligotrophic peat is composed of plants that grow under nutrient-poor conditions associated with bogs in places that were isolated from regional groundwater. The most extensive bog-peat accumulations in the delta plain consisted of peat moss and were positioned in the upper part of the Holocene sequence (e.g., Visscher, 1949). Other types of oligotrophic peat, e.g., *Ericales* peat, have a rather scattered distribution and are generally found directly on top of coversand (Fig. 4.5A).

Mesotrophic peat: sedge peat and reed-sedge peat

Sedge peat and reed-sedge peat – in short: (reed-)sedge peat – consist partly of remains of sedges (*Carex spp.*). Within the latter, remains of reed (*Phragmites australis*) accompany sedge remains. Sedge remains in peat accumulations most often consist of roots, which

→ Next page. Figure 4.5. Sedimentary logs and LOI characteristics of mechanical cores that are positioned in the coversand area (B25H0739 (A), Abcoude Oude Molen (B) , B25G1057 (C), B25G1055 (D)) and in the palaeovalley (B38D0432 (E), B38E0306 (F)) (partly after Bos, 2010; Chap. 3). LOI values are given in brown horizontal bars; $CaCO_3$ values are given in blue horizontal bars (only in 4.5A). Diatom occurrence shown in A and C is based on data published by Bos et al. (2009; Chap. 2). See Addenda 5 and 6 for full-colour versions.


A
B25H0739
LOI (%)
30 50 70
diatom
total count
0
200
surface
-1.78 m O.D.
c si sa
0
-1
-2
-3
-4
-5
wood peat
gyttja
gyttja
sedge peat
oligotrophic
peat
Erica
Calluna
Sphagnum
substratum
B
Abcoude Oude Molen (200434046)
LOI & CaCO3 (%)
30 50 70 100
surface
-1.70 m O.D.
c si sa
-6
-7
wood peat
sedge peat
wood peat
sedge peat
sedge peat
reed peat
sedge peat
wood peat
calcareous gyttja
substratum
C
B25G1057
diatom
total count
0
200
surface
-1.31 m O.D.
c si sa
LOI (%)
30 50 70
gyttja
Juncus peat
substratum
D
B25G1055
surface
-1.70 m O.D.
c si sa
LOI (%)
30 50 70
wood peat
reed-sedge peat
wood peat
substratum


E
Molenaarsgraaf (B38D0432)
LOI (%)
30 50 70
algae fossil abundance
surface -1.19 m O.D.
c si sa
+/+++
--/++
++/++
+/++
--/+
+/+
++/+++
+++/+
+++/++
+++/++
+/++
+++/++
+++/+
+++/+
+++/+
+++/+
+++/+
+++/++
0
-1
-2
-3
-4
-5
-6
-7
-8
-9
-10
wood peat
wood peat
gyttja
gyttja
gyttja
gyttja
gyttja
substratum
F
Lopik (B38E0306)
LOI (%)
30 50 70
surface - 0.90 m O.D.
c si sa
wood peat
reed peat
wood peat
wood peat
gyttja
wood peat
gyttja
substratum
algae fossils abundance
diatoms / chrysophyceae statospores
+++ = very abundant
++ = moderately abundant
+ = scarce
-- = absent
= sample
admixtures
sand
gravel
clay pebble
peat clast
sand lense
gastropods
molluscs
shell fragments
detritus
botanical content
reed fragments
sedge fragments
Menyanthes seed
wood fragments
oligotrophic pl. fragm.
other features
bioturbation
roots
vivianite
lamination
A-horizon
organic facies
wood peat
reed peat
sedge peat
oligotrophic peat
gyttja
clay (c)
silt (si)
sand (sa)
medium
fine

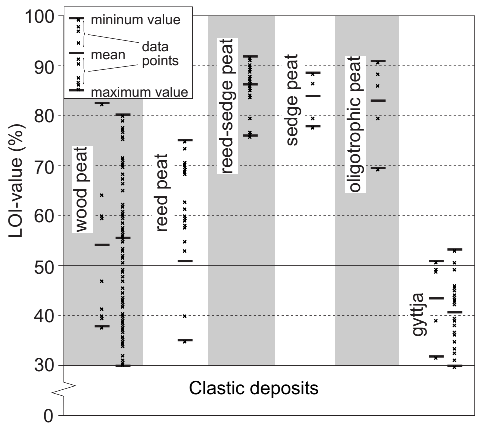


Figure 4.6. LOI values for organic facies in the Rhine-Meuse delta. For these analyses we used data from the cores presented in Figure 4.5 and from additional cores (locations are shown in Figure 4.1), (partly unpublished data of Van Asselen). For wood peat and gyttja, the data has been differentiated between the palaeovalley (shown at the right) and the coversand area (shown at the left). Data points may overlap.

are flat and 1 – 3 mm wide and they have a rolling tobacco-like appearance. Often present in (reed-)sedge peat are 2 – 3 mm large, disc-shaped, dark red to black coloured seeds of bogbean (*Menyanthes trifoliata*) (Fig. 4.5B). Especially when reed remains are ubiquitous, very concise visual inspection is needed to distinguish this facies from reed peat. Wood remains of birch (*Betula spp.*) may be found in (reed-)sedge peat, indicating relatively low (ground)water levels during accumulation. We found that the LOI value of sedge and (reed-)sedge peat is on average 85% and ranges between 76 and 92% (Fig. 4.6).

Sedge peat and reed-sedge peat preferentially accumulate under mesotrophic, freshwater conditions with water depths of 0.1 – 0.3 m (Den Held et al., 1992). Sedge and bogbean are indicators for mesotrophic conditions that, in the Rhine-Meuse delta plain, are often associated with seepage-affected positions along the margins of ice-pushed ridges (e.g., Pons and Van Oosten, 1974).

Eutrophic peat: reed peat

Reed peat is composed of plant remains of common reed (*Phragmites australis*) that are embedded in a fine-textured gyttja-like matrix. This matrix is partly composed of yellowish reed leaves of which often the veins can be recognized. Observed reed culms, roots and rhizomes had diameters of up to 1 cm. This size is large relative to remains of other non-wood plants. Macro remains of reed plants can therefore easily be identified. The LOI-value of this facies is on average 51% and ranges between 35% and 75% (Fig. 4.6); the minerogenic component predominantly consists of clay.

Reed peat accumulates in eutrophic lacustrine or marshy environments, e.g., lake edges. Reed is tolerant for a wide range of conditions that occur in delta plains: mesotrophic to eutrophic and freshwater to (slightly) brackish water. Reed preferentially accumulates in water with depths ranging from 0.2 to 0.5 m (Den Held et al., 1992). The combination of reed plant remains and gyttja-like material relates to the environment in which it accumulated: in shallow flooded parts of the floodplain like marshes, lake edges, inland parts of lagoons.

Eutrophic peat: wood peat

Wood peat is composed of woody material that is embedded in a soft matrix. Wood peat includes remains of trees (e.g., wood, roots, twigs, bark, leaves, fruits and seeds) as well as the remains of other plants that formed the undergrowth of the forest vegetation, and which in general cannot be identified by macroscopic inspection. However, wood peat generally includes significant portions of amorphous, gyttja-like material that accumulated in pools and ponds on the forest floor. Commonly found in wood peat are the remains of black alder (*Alnus glutinosa*) and willow (*Salix spp.*). Less often, poplar wood (*Populus spp.*) and oak wood (*Quercus spp.*) can be identified in the peat, which indicates drier conditions, possibly associated with higher stands on fluvial levees or aeolian river dunes. The LOI value for wood peat is on average 54%, ranging between 30 and 83% (Fig. 4.6).

Wood peat accumulated in eutrophic, freshwater environments, which are dry relative to other organic facies. The groundwater level on average approximates the surface level (-0.1 – 0.1m) (Den Held et al., 1992). The presence of alder and willow indicates eutrophic conditions whereas only rarely recognized birch and oak generally indicate mesotrophic stands. The latter in general occurs in combination with sedge undergrowth. Furthermore, the proportion of gyttja-like material can tentatively be related to the wetness of the environment, i.e. average water depth and part of the year that the area is flooded.

Gyttja: detrital gyttja

Detrital gyttja is composed of three components, being organism remains, chemical precipitates and minerogenic matter (Von Post, 1860; Hansen, 1959), although the proportion of chemical precipitates in detrital gyttja is limited. Because of the lacustrine depositional environment, gyttja often exhibits a very fine horizontally laminated structure, which may be emphasized by the presence of sand, silt or clay lenses/laminae and commonly observed plant remains. Furthermore, beetle elytra, i.e. wing covers, (fragments of) shells and peat clasts (up to 2 cm diameter) are regularly observed. Gyttja usually contains abundant fossilised diatoms, which contrasts to in situ peat, where they are absent or very scarcely present (as reported for cores B25G1057 and B25H0739 by Bos, 2010; Chap. 3, and shown in Fig. 4.5A, C). The diatoms in these gyttja deposits are indicative for standing water of ~2 m water depth (Bos, 2010; Chap. 3). Gyttja ranges from fine-detrital algae gyttja to coarse-detrital gyttja. Fine-detrital algae gyttja often has a greenish-brown colour. Coarse-detrital gyttja is composed of reworked plant material such as leaves and twigs. LOI values in the palaeovalley and coversand area on average are 41% and 44% respectively. Furthermore, the LOI values do not exceed 53% (palaeovalley) and 51% (coversand area) (Fig. 4.6). Part of the gyttja sequence in the coversand area thus accumulated while supply of clastic sediment was very limited.

Detrital gyttja is a deposit that forms in standing water bodies, which is in contrast to in situ peat that is composed of organic material from plants that grew at that location. It preferentially forms inland in floodbasin lakes. Less commonly, it forms in sediment starved lagoons where marine-derived shells and diatoms (i.e., species with tolerance for salt indicate marine influence (Bos et al., 2009; Chap. 2).

Gyttja: carbonate-dominated gyttja

A gradational scale exists between detrital gyttja and carbonate-dominated gyttja. Both gyttja types may alternate in vertical successions. When volumetrically more than 50% of a succession consists of carbonates, it classifies as carbonate gyttja (for example, see bottom part of organic succession in Fig. 4.5B). The most frequently found precipitates are calcite ($CaCO_3$) and siderite ($FeCO_3$). Calcareous gyttja is a soft white mass that strongly reacts with 5% HCl. Siderite gyttja is a light-brown to dark-yellow soft mass that reacts with 5% HCl, though only weakly and notably stronger with higher temperatures (> 30˚C). Lamination within carbonate intervals usually is absent although strong laminations may appear if alternation between organic and carbonate precipitates exists.

At least two conditions must be met for gyttja to form carbonates: the hydrology is seepage-dominated and chemical disequilibria exist between conditions in the surface water and the groundwater. When groundwater is able to dissolve more calcite or siderite (or any other potential precipitate) than surface water, carbonates will precipitate once seepage forces groundwater to surface. Seepage occurs in relation to elevated infiltration regions. In The Netherlands, therefore, large-scale seepage is confined the lower flanks of substrate plateaus or ice-pushed ridges (e.g., the Utrechtse Heuvelrug, Fig. 4.1). This accounts, for example, for the AOM site (Figs. 4.1, 4.5B), where the lowermost part of the gyttja sequence is almost entirely composed of calcium carbonate.

4.4.2 Organic facies in the basal peat layer

Spatial distribution

We found that reed peat is the most abundant organic facies in the basal peat layer (Fig. 4.7B), occupying 63% of the area. Gyttja is also frequently found (29%) whereas wood peat (8%) is least common of the three organic facies in the basal peat. In general, reed peat dominates vast, continuous areas of up to 200 km^2 (*1, italic* numbers refer to locations in Fig. 4.7B). Areas where continuous gyttja is expected are less extensive and measure up to 40 km^2 in the coversand area *(2)* and up to ~10 km^2 in the palaeovalley (e.g., *3, 4)*. Wood peat occurs in even smaller patches, with maximum areas of ~10 km^2 *(5)*, but frequently much smaller, being < 1 km^2 (e.g., *6)*.

In the coversand area, the dominance of reed peat is prominent, especially in the northern part *(1)*. There, only relatively small areas of wood peat (7) interrupt the vastness of reed peat. The spatial distribution of wood peat has a patchy character although larger contiguous areas (5-10 km^2) also occur *(5)*. The majority of cores that include indicators for sedge peat or oligotrophic peat are also located in the coversand area *(1, 8, 9)*. Frequently their occurrence coincides with the occurrence of reed peat facies. Gyttja is restricted to the western part of the northern coversand area, which relate to deeper positions (-12 to -15 m O.D.). Furthermore, rarely found calcareous gyttja, which indicates carbonate-rich, mesotrophic seepage conditions, is also restricted to basal peat in the coversand area in the vicinity of ice-pushed ridges (see, for example, Fig. 4.5B).

In the palaeovalley, reed peat dominates the northern part *(10)*. The central and southern parts, are characterized by buried Late Weichselian and early Holocene aeolian river dunes, and are dominated by gyttja *(3, 4)* and wood peat *(11, 12)* respectively. The elevation of wood peat samples *(11, 12)* on average is 0.8 m above those of the gyttja samples *(3, 4)*, which may explain the hydrological differences between both areas. Further, wood peat is restricted to small and scattered patches (e.g., *6)*.

Marine shells that are included in the gyttja deposits in the western part of the investigated area *(2, 13, 14)* indicate that this facies partly accumulated in a brackish or salt-water environment, such as a lagoon or an estuary.

Temporal distribution

Prior to 10,000 cal yr BP, reed peat *(15)* and marine-shells containing gyttja *(2)* accumulated both in the coversand area and in the palaeovalley, although only limited data is available. After that, between 10,000 and 9000 cal yr BP, reed peat accumulated at large scale, both in the coversand area *(1)* and in the palaeovalley (progressively from *16* to *17)*. Between 9000 and 8000 cal yr BP, gyttja was prominent both in the palaeovalley *(3, 4, 18)* and in the western part of the coversand area *(13)*, although in the latter, marine shells were frequently encountered. Further, reed-peat accumulation continued at large scale in the eastern part of the coversand area *(19)*. Between 8000 and 7000 cal yr BP, gyttja formation in the northern coversand area ceased and simultaneously, wood peat accumulation became more abundant *(5, 20)*. It was not only limited to small patches, but occurred also in larger areas *(5)*. Further, reed peat accumulation continued in the eastern portion of the northern coversand area *(7)*. By this time, the basal peat layer in the studied part of the palaeovalley had almost completely formed.

4.5 DISCUSSION

4.5.1 Identifying and mapping organic facies in delta plains

We designed a classification key, which enabled identification of organic facies on the basis of field characteristics. Concise core descriptions as well as palynological, diatom, LOI and $CaCO_3$-content analyses provide a comprehensive descriptive framework that forms the basis for the key. Although the classification key and facies descriptions are based on data from the Rhine-Meuse delta record only, we stress that this tool is applicable also in other delta regions with a temperate climate. First, differentiation of in situ accumulated peat and organic sediment (gyttja), and subdivision of the latter, is based on fundamental sedimentological characteristics. Second, the main peat types present in the Holocene Rhine-Meuse delta are likely to have large similarities to main facies groups in other deltas with temperate climates although at species level, differences probably exist. In the Fraser delta, for example, mesotrophic (*Carex*) peat, oligotrophic peat, wood peat and gyttja were found (Styan and Bustin, 1983b) (Tab. 4.2). A major difference between the Eurasian

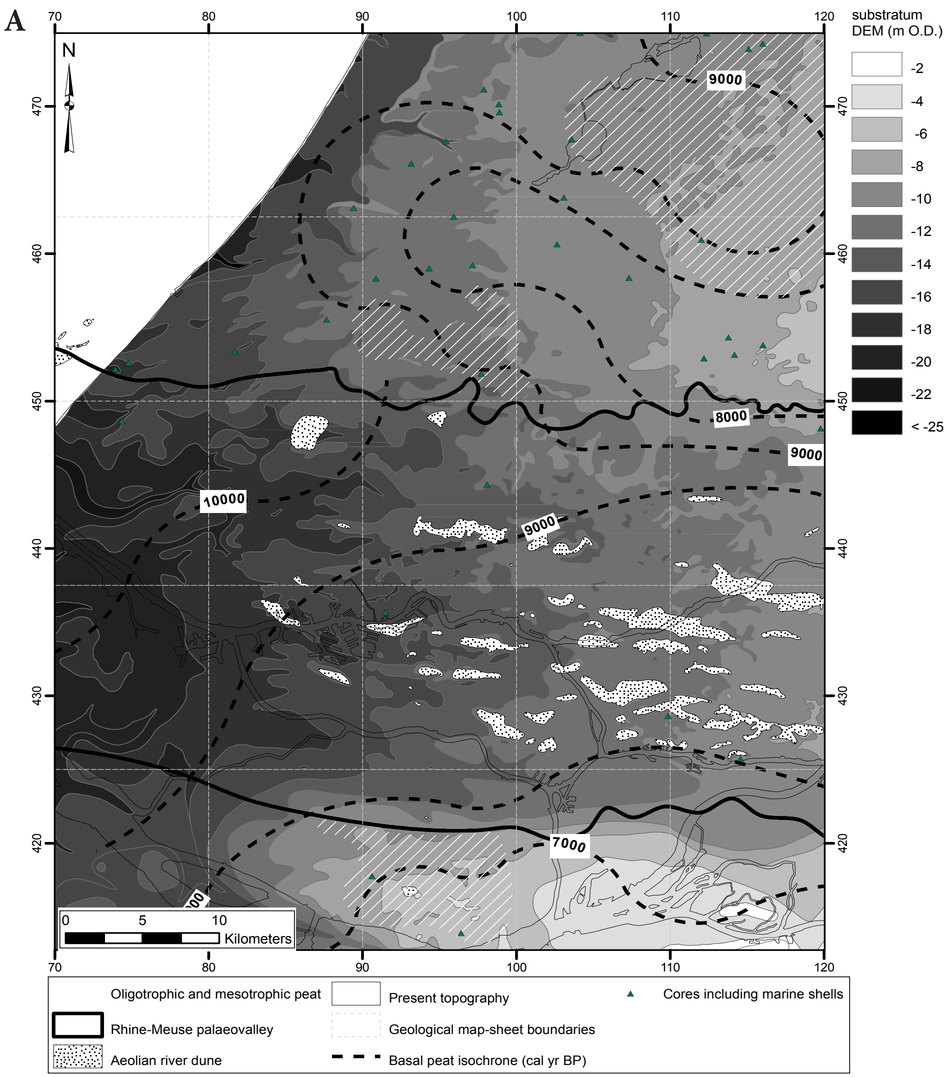
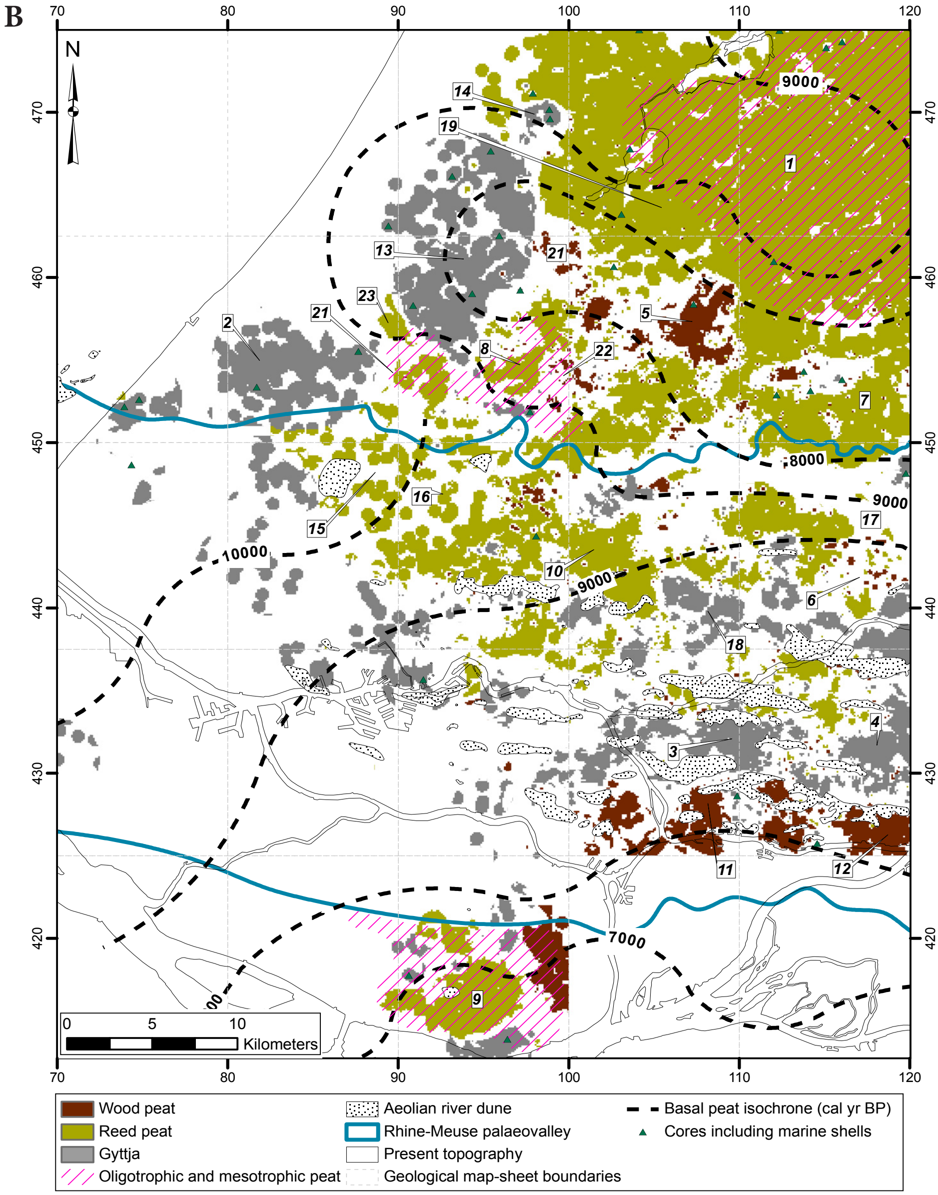


Figure 4.7. A. DEM of the substratum. B. (next page) Spatial distribution of organic facies in the (diachronous) basal peat layer in the distal part of the Holocene Rhine-Meuse delta. Isochrones (corresponding to palaeogroundwater levels of Cohen, 2005) indicate the onset of Holocene aggradation and hence the maximum age of the basal peat. Facies are displayed for locations where the distance to the nearest core is less than 500 m, which corresponds to the maximum range of spatial correlation. The area shown corresponds to Fig. 4.7A, which enables comparison of the spatial distribution of organic facies in the basal peat layer and the topography of the substratum. Italic numbers are referred to in the text. See Appendix 8 for a full-colour version.



and American continent, however, is the presence of reed. While reed is omnipresent on the Eurasian continent, it formed only minor portions of the North American vegetation until ~200 years ago when it dramatically expanded (Chambers et al., 1999). This means that on the North American continent, reed peat is not present at large scale in Holocene delta plains. In the modern Cumberland Marshes, however, reed forms a substantial part of both the vegetation and recent peat accumulations (Dirschl, 1972). Further, organic facies there (Dirschl and Coupland, 1972) correspond very well with those found in the Rhine-Meuse delta (Tab. 4.2). In the Biebrza National Park, Poland, the dominant organics are reed peat, sedge peat, wood peat and gyttja (Gradziński et al., 2003; Van Asselen and Roosendaal, 2009). Also in the Lower Volga delta, Russia (Baldina et al., 1999) these facies groups are likely dominant as can be inferred from the wetland plant communities that are present in these areas today (Tab. 4.2). Remarkable is that only very few studies in delta plains report the presence of gyttja (Tab. 4.2B). The presence of lakes in modern delta plains, such as the Cumberland Marhses (Smith and Pérez-Arlucea, 1994) and the Laitaure delta, Sweden (Andren, 1994), though, suggests that they were commonly ignored in delta-plain studies, especially when filled by organics. We contend that our understanding of delta-plain formation would greatly benefit from identification of gyttja because it is associated with different environmental conditions (i.e., the formation of gyttja requires a certain water depth, > ~0.5 m) compared to in situ peat.

Additional analyses can be useful both to support visual facies identification and to further subdivide the organic facies groups, thereby improving the quality of interpretations and increasing the opportunities and the level of detail for various applications such as palaeogeographic reconstructions. LOI analyses, for example, may support visual differentiation between reed peat (LOI values < 75%) and reed-sedge peat (LOI values > 75%). Another method by which organic facies can be further subdivided is macro-remains analysis, as is shown by e.g., Biserni and Van Geel (2005). Palynology and micropalaeontology (e.g., diatom analysis) may be a helpful tool for reconstructing the environment in which the organics accumulated. Due to significant influx of regional species, however, pollen and diatom counts alone cannot be used for organic facies identification.

Identification of organic facies not only serves reconstruction of palaeoenvironmental conditions in (sub)recent delta plains. Earlier studies on deltaic coals in ancient rock records point to the value of distinguishing facies in coals and of understanding organic-facies distributions in delta plains (e.g., Scott, 1989; Robinson Roberts and McCabe, 1992). In the Fraser delta, Canada, for example, the relationships between organic facies that were identified in modern settings and maceral composition in ancient coal layers were investigated (Styan and Bustin, 1983a). In the Rhine-Meuse delta, however, this issue awaits further investigation.

We contend that the procedure of database querying and geostatistical analyses can also be applied on other stratigraphical levels (older deposits) and in other regions. When borehole density is > 0.2 cores/km^2, interpolation of interpreted organic samples may yield results relevant for organic-facies distribution at delta scale.

4.5.2 Organic-facies map of the basal peat layer: robustness

The applied method for reconstructing organic-facies distribution in the basal peat layer enables us to see trends in the data. The spatial distribution of organic facies in the basal peat layer, nevertheless, corresponds very well to what has been observed in the logged cores presented in this study (Fig. 4.5) and in earlier, mostly local, studies. Hijma et al. (2009), for example, presented three valley-wide north-south oriented cross sections that are positioned in our study area and wherein three organic units were identified (wood peat, gyttja and undifferentiated). Cross sections in the eastern part of the palaeovalley (between x=100 and x=120 in Fig. 4.7B), presented by Bosch and Kok (1994), seem to correspond with our results, although they did only differentiate reed peat and wood peat and incorporated gyttja deposits in the floodbasin clay unit, which hampers comparison. Furthermore, Van der Woude (1981) found eutrophic organic facies to be dominant in the palaeovalley. From a location in the central part of the palaeovalley (core Molenaarsgraaf H1110, positioned at the same location as core B38D0432, Fig. 4.1), he reported significant proportions of gyttja (cf. Figs. 4.5E, 4.7B). Bennema (1951) and Pons and Van Oosten (1974), additionally, studied parts of the northern coversand area. They observed, apart from eutrophic peat types, large areas where mesotrophic peat and to a lesser extent oligotrophic peat, formed large parts of the basal peat layer. The same was found in logged cores from the northern coversand area (Fig. 4.5A-D). Comparison between our observations and the observations from other studies supports the robustness of the resulting spatial distribution of basal peat organic facies.

In the field, organics can be classified by application of the key presented in this study (Fig. 4.4). Nonetheless, factors that complicate classification of organics by means of visual inspection also exist, two of which will be discussed below. First, disproportionate identification of organic facies composed of easily recognized and well-preserved plant species, yields biased results. Remains of reed and *Sphagnum*, for example, are easily identified in organic successions, whereas remains of sedge plants are more difficult to recognize. Macroscopically identifiable remains of *Juncus spp*. are not preserved at all. Second, roots that penetrated into earlier accumulated material 'polluted' and partly obscured the characteristics of the underlying succession. Especially deep rooting and easily identifiable plant species may limit facies identification. Reed roots, for example, can penetrate as deep as 2.5 m (Kohzu et al., 2003) into the soil. This means that the amount of reed peat is overestimated in borehole descriptions, and consequently also in results presented in this study. This may, for example, apply to reed-peat dominated areas where occasionally oligotrophic peat and sedge peat has been identified (e.g., *1, 8, 9)*.

The database used for this study is heterogeneous, which is caused, for example, by input of a large number of core describers and different sample quality related to different core techniques. Because map sheet boundaries often dictated planning of mapping activities, different core describers were often separated by these map boundaries. Hence, it was expected that spatial distribution patterns of organic facies would partly correspond to these map sheets. The results (Fig. 4.7B), however, indicate that this expectation was too pessimistic. We only could find one map sheet boundary that determined the spatial distribution of organic-facies samples. Within one of the map sheets in the northern coversand area (between *21* and *22*), the number of cores with sedge and oligotrophic plant remains increases in eastward direction. This trend is not continued across the map-sheet boundary in eastward direction, where sedge and oligotrophic plant remains were only observed in a few cores. In other parts of the study area, database heterogeneity across map sheet boundaries does not affect the spatial distribution of organic facies in the basal peat layer.

The average thickness of the basal peat samples that were classified in the primary classification is 0.62 m. Compaction of the analysed organics can be as much as 60% (personal communication Van Asselen), which means that a 0.62 m thick layer originally (i.e., in uncompacted state) could have been 1.2 – 1.5 m thick. The rate of groundwater-level rise, which controlled accumulation of basal peat, has been estimated at ~1 mm/yr for the lowest part of the Holocene sequence in the Zuid-Holland study area (Cohen, 2005). Thus, basal peat samples used in this study, on average formed in 1200 to 1500 yr. The analyzed samples, therefore, do not only represent initial basal onlap but also include the first successive stages of Holocene aggradation. This means that our results are likely biased towards more nutrient-rich and probably also wetter conditions in relation to initial aggradation conditions, especially in the coversand area. The lower part of the basal peat layer there consists of oligotrophic or mesotrophic peat (e.g., Fig. 4.5A-C; Bennema, 1951), whereas eutrophic peat is less common (e.g., Fig. 4.5D).

4.5.3 Controls on the organic facies distribution in the basal peat layer

In this study, we distinguished four basal peat organic-facies domains: precipitation-dominated, seepage-dominated, fluvial-dominated or marine-dominated deposition (Fig. 4.8). The spatial pattern of these domains suggests a tight relationship of organic deposition to large-scale hydrological conditions, which are partly governed by the topography of the substratum. We found that basal peat formation was governed by marine processes in the westernmost part of the area. Basal peat in the eastern portion of the palaeovalley (e.g., *3, 4, 6, 10-12, 17, 18)* accumulated in the fluvial realm under eutrophic conditions while basal peat in the eastern portion of the coversand area (e.g., *1, 8, 21, 22)* has developed predominantly under seepage-affected mesotrophic conditions. Seepage is strongly related to the topography of the substratum and, therefore, mesotrophic peat is often situated on the flank or at the foot of topographical higher regions. Mesotrophic peat in the southern coversand area *(9)*, for example, is probably associated with elevated substratum that exists south-eastwards thereof (compare Fig. 4.7A and 4.7B). Another example occurs in the northern coversand area where mesotrophic peat (1) formed westwards of an ice-pushed ridge (Fig. 4.1, 4.7B). A potential other domain is the ombrogenic hydrological regime, which probably is contained within the seepage realm. Oligotrophic peat, however, could not be mapped due to the limited number of identified samples.

In the westernmost part of the study area, the basal peat predominantly consists of gyttja. The presence of marine shells shows that these gyttja deposits have at least partly been formed in a brackish-lagoonal setting. Not all gyttja deposits represent the same palaeogeographical setting. There are gyttja deposits *(2)* that formed at 18.8 m -O.D. on

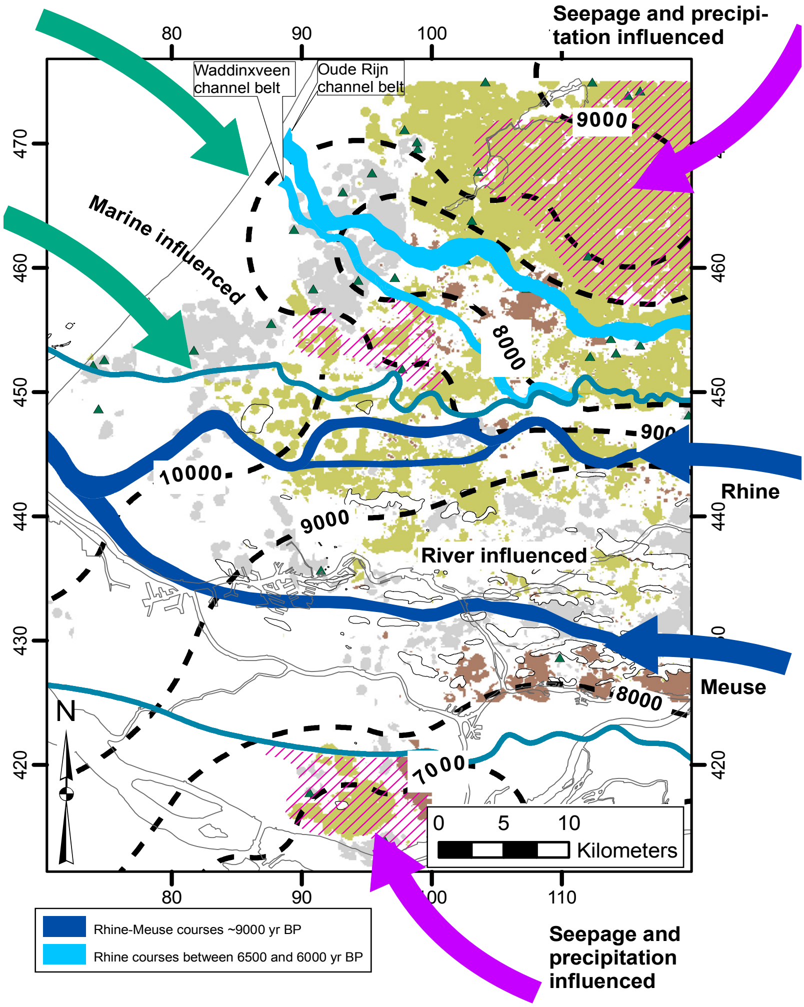


Figure 4.8. Dominant hydrological regimes controlling the onset of Holocene organic aggradation as inferred from the facies composition of the basal peat. Only legend elements that differ from Fig. 4.7B are explained. The position of fluvial channels is after Hijma (2009). Between 9000 and 6500 cal yr BP the Rhine is positioned in the palaeovalley, corresponding to the situation at 9000 cal yr BP. See Appendix 8 for a full-colour version.

average whereas other gyttja deposits *(13)* formed at 13.4 m -O.D. on average. Both areas are separated in space (by reed peat *(23)*) and time (note the position of the 9000 cal yr BP isochrone in Fig. 4.7B). This indicates that gyttja in the northern coversand area formed in two phases. We tentatively suggest that these phases are controlled by marine processes. Hijma et al. (2009) indicate that in the northern gyttja area *(13)* basal peat (gyttja) is largely covered by a thin (< 0.5 m) fluvial clay bed that is overlain by intertidal deposits, confirming the existence of near-coastal environmental conditions during the early Holocene. We contend that the existence of marine-influenced gyttja is related to the occurrence of rapid sea-level rise during the early and middle Holocene transgression. Possibly, there is a relation with accelerated relative sea-level rise around 8200 cal yr BP (Von Grafenstein et al., 1998; Hijma and Cohen, 2010), triggering large scale drowning of the area and development of a brackish lagoon. Marine-influenced gyttja also occurs in basal peat samples from locations that correspond to the later course of the Oude Rijn (compare Figs. 4.7B and 4.8). This suggests that marine invasions have persistently or repeatedly affected that area already when basal peat accumulated.

We found a nutrient gradient in the basal peat layer from the palaeovalley (eutrophic) to the coversand area (mesotrophic) (Fig. 4.7B), which is controlled by the Pleistocene topography that prevented nutrient-rich river water to affect the first stages of Holocene aggradation outside the palaeovalley. The separation between both hydrological domains is observed from west to east along progressively younger accumulations. Hence, the palaeovalley-coversand boundary (i.e., the height difference between both) functioned as an effective barrier for fluvial water during the first stages of aggradation, even though the locus of initial aggradation (the apex) shifted upstream over time.

The facies difference between the northern (reed peat, *10, 16)* and central portion (gyttja, *3, 4, 18)* of the palaeovalley (Fig. 4.7B) indicates smaller water depths in the northern part. This probably reflects differences in sediment transport between the Rhine in the northern part and Meuse in the southern part (positions during the early Holocene are indicated in Fig. 4.8). The Rhine, which during the Holocene supplied much more sediment than the Meuse (81 vs. 19 weight %) (Erkens, 2009), apparently had a larger potential to fill accommodation space.

4.5.4 Accommodation space

Accommodation space can be considered as a key parameter for delta evolution and hence, alluvial architecture. Identification of organic facies in delta-plain successions can be used to reconstruct past accommodation space and thereby contributes to the understanding of delta evolution, as will be discussed below. Accommodation space is created in response to relative base-level rise (Blum and Törnqvist, 2000) and compaction due to loading (Van Asselen et al., 2009). The latter happens when compactable sediments (e.g., clay) or organics are buried by younger deposits. Owing to the increased vertical pressure, water is squeezed out of the buried succession and consequently, its volume (i.e., vertical thickness) decreases. Accommodation space can become filled with fluvial and/or estuarine sediment. A shortage of clastic sediment input relative to accommodation-space formation results in the creation of excess accommodation space. Under these conditions, the water level will rise above the surface, which enhances organic accumulation. Water depth, in combination with the potential amount of subsidence due to compaction of the underlying succession, determines the vertical space that is available for sedimentation

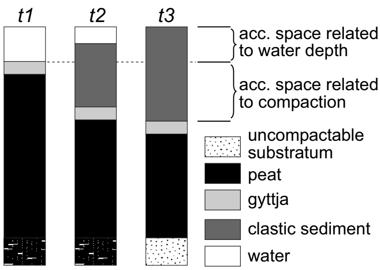


Figure 4.9. Three time steps that show the relation between a setting in which organics accumulate and the fossilised situation after burial by younger deposits. Accommodation space at a certain location and certain time can be reconstructed by identification of the organic facies, which is a proxy for water depth, and estimation of subsidence of the former surface due to compaction.

(Fig. 4.9). Organic facies can be used as proxy for palaeo water depth. reed peat, for example, forms in lakes or along lake shores where water depths are between 0.1 to 1.0 m, whereas wood peat forms when the water table is near the surface (Den Held et al., 1992). Hence, identification and mapping of organic facies in delta plains is useful for reconstruction of palaeo accommodation space.

4.6 CONCLUSIONS

This is one of the first studies in which a spatial and temporal distribution of organic facies in the basal peat layer of a delta plain is presented. An integral approach, which included concise facies descriptions and analyses of a lithological database, resulted in the identification of 12 organic facies. The spatial distribution of the three most frequently found facies (reed and wood peat and gyttja) have been obtained using geostatistical analyses. Mapping of these facies is important because they host an enormous archive for palaeoenvironmental reconstructions such as hydrological characteristics, nutrient availability and salinity level, which provide great potential to better understand delta formation. As such, this study may serve as a starting point for research on the effect of vegetation/peat type on fluvial processes and the environmental conditions that prevailed during the period of organic accumulation.

In this study, a field key for classifying organics in delta plains has been designed and tested in the field and against laboratory analyses (LOI, macroremains analyses etc.). It is shown that the most frequently found organic units (i.e., oligotrophic peat, (reed-) sedge peat, reed peat, wood peat and gyttja) are likely to be dominant in other delta plains with temperate climates as well. We conclude that identification of organic facies in delta plains, especially gyttja, would be very feasible and enhance reconstruction of past environmental conditions.

We show that semi-automated database analyses are a powerful tool for organic facies identification in archived core descriptions and that these are an excellent starting point for organic facies mapping at delta scale, especially when large geological databases are available.

Identification of organic facies is a powerful tool for reconstructing accommodation space. Accommodation space is partly determined by water depth, for which organic facies are an easily obtainable proxy.

ACKNOWLEDGEMENTS

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5 Spatial and temporal distribution of coarse-grained overbank deposits in the Holocene Rhine-Meuse delta, The Netherlands

With contributions of Esther Stouthamer

ABSTRACT

It has been generally recognized that coarse-grained deposits, including crevasse-splay deposits, organic-clastic lake fills, and bay-head delta deposits, are present in the overbank realm. Their quantitative significance, however, has never been studied at delta scale, neither at Holocene time scale. Knowledge thereof would be beneficial for understanding sediment distribution in delta plains, which subsequently contributes to explaining delta evolution. This study systematically addresses delta-scale distribution of coarse-grained overbank deposits and their sand-body proportion variability from 9000 cal yr BP onwards, based on eight valley-wide geological cross sections in the Holocene Rhine-Meuse delta, The Netherlands. We found that coarse-grained overbank deposits form 7.1% of the fluvio-deltaic wedge, of which crevasse-splay deposits are most frequently observed halfway between the delta apex and the coast. Organic-clastic lake fills and bay-head delta deposits, in contrast, are limited to the distal delta plain. Over successive periods, the largest proportions of crevasse-splay deposits remain at 50-100 km downstream of the upstream-shifting delta apex. It is shown that this proportion partly depends on inter-channel floodbasin width. Intermediate floodbasin widths (3.1-3.6 km) yield the highest proportions of crevasse-splay deposits. Narrow floodbasins limit the development of crevasse-splay deposits whereas wide floodbasins (> 3.6 km) do not necessarily result in bigger crevasse-splay deposits but instead facilitate the formation of floodbasin deposits and organics thereby reducing the proportion of crevasse-splay deposits. The formation of organic-clastic lake fills is promoted by high rates of base-level rise and wide floodplains, which both facilitate the creation of accommodation space. Subsequently, large volumes of accommodation space provide excellent conditions for extensive peat-forming wetlands wherein organic-clastic lake fills can develop.

The results from the sand-body proportion analyses show that sand forms 26-30% of the coarse-grained overbank deposits. The sand-body proportion of coarse-grained overbank deposits is shown to be controlled, amongst other variables, by channel planform, superelevation (= elevation difference between levee crest and adjacent floodbasin surface) and the composition of the substratum. It is concluded that large volumes of overbank sand bodies exist in distal delta plains. They form substantial portions of the reservoir volume (up to 39% in the distal Rhine-Meuse delta) and yield relatively high connectedness ratios.

5.1 INTRODUCTION

Enormous amounts of exploitable fossil-fuel reserves are stored in ancient delta-plain successions. Partly because of the commercial value of these natural resources, a long history of scientific research into delta-plain deposits exists. This research mainly focussed on channel deposits, whereas non-channel deposits, i.e., overbank deposits and organics, received much less attention. This is surprising because especially overbank deposits account for significant volumetric proportions of delta plains. They form as much as 40% of the delta-plain deposits in, for example, the Holocene Rhine-Meuse delta, The Netherlands (Gouw, 2008; Erkens, 2009) and in the Lower Mississippi Valley (LMV), USA (Gouw and Autin, 2008). Still, various studies have reported on the geometry, depositional-facies composition and formation of delta-plain overbank deposits, such as natural-levee, floodbasin and crevasse-splay deposits (see, for example, Miall, 1996, and references therein; Cazanacli and Smith, 1998; Hornung and Aigner, 1999; Farrell, 2001; Stouthamer, 2001b; Ghosh et al., 2006). These studies demonstrated that the depositional-facies and lithological heterogeneity of overbank deposits potentially is very high (e.g., Fielding, 1986; Farrell, 1987; Weerts and Bierkens, 1993; Willis and Behrensmeyer, 1994). The facies variability especially occurs in crevasse-splay deposits, organic-clastic lake fills and bay-head delta deposits (Farrell, 1987; Smith et al., 1989; Tye and Coleman, 1989b; Weerts and Bierkens, 1993; Smith and Pérez-Arlucea, 1994; Pérez-Arlucea and Smith, 1999; Stouthamer, 2001b; Hijma et al., 2009; Bos, 2010; Chap. 3), which we collectively term coarse-grained overbank deposits. Coarse-grained overbank deposits form in floodplain depressions and basins that become filled by sediment supplied by a fluvial channel, and contain relatively large amounts of sandy lithofacies. Knowledge of the delta-scale distribution of these deposits is limited, but would increase our understanding of delta-plain building. For example, sediment-distribution patterns in delta-plain successions are an important aspect of delta-plain architecture. Better understanding of the distribution of coarse-grained overbank deposits would contribute to the quality of deltaic-architecture models and hence the predictability of reservoir characteristics.

The aim of this study is to analyze and explain the spatial and temporal distribution and lithofacies variability of coarse-grained overbank deposits in the Holocene Rhine-Meuse delta and to assess the contingent controlling factors.

For a number of reasons the Rhine-Meuse delta is a very suitable area to study the spatial and temporal variability of coarse-grained overbank deposits. First, eight recently compiled valley-wide cross sections that cover the delta plain from apex to the coast offer a unique overview of the alluvial architecture of the Holocene delta-plain succession (Fig. 5.1). Second, there is good time control on the deposits, which support several palaeogeographic reconstructions (Berendsen and Stouthamer, 2001; Gouw and Erkens, 2007; Hijma et al., 2009). The architectural and palaeogeographic framework that exists enables determination of the distribution of coarse-grained overbank deposits. Third, the allogenic controlling factors on delta-plain formation, such as sea-level rise (Van de Plassche, 1982; Hijma and Cohen, 2010), associated groundwater-level rise (Cohen, 2005), and sediment supply (Erkens, 2009) are well known.

5.2 THE HOLOCENE RHINE-MEUSE DELTA

5.2.1 Geological setting

The Rhine-Meuse delta encompasses the combined delta plains of the River Rhine (mean annual discharge 2350 m^3/s) and the River Meuse (mean annual discharge 230 m^3/s) (Rijkswaterstaat, 2010) and is located at the south-eastern margin of the subsiding North Sea Basin. Locally, large differences in relative subsidence rates exist between fault-separated tectonic blocks. For example, Quaternary displacement rates along the Peel Boundary Fault (PBF, Fig. 5.1) separating the subsiding Roer Valley Graben from the relatively stable Peel Block, averaged 60 mm/ka (Van den Berg, 1994). Holocene displacements along the PBF have influenced fluvial processes in that area resulting in changing architecture across this fault (Cohen et al., 2002).

Holocene onlap, triggered by base-level rise, is often marked by the presence of organics at the base of the delta-plain succession. Delta formation has predominantly been controlled by base-level rise and upstream sediment delivery (Gouw and Erkens, 2007). Until 5000 cal yr BP eustatic sea-level rise was the principal component in base-level rise and thus the dominant controlling factor on the formation of accommodation space for delta-plain sedimentation. By 5000 cal yr BP, eustatic sea-level rise ceased and from that moment on base-level rise was principally governed by basin subsidence. After 3000 cal yr BP increased sediment delivery from the hinterland related to human activity, controlled sedimentation in the delta plain (Erkens, 2009).

Avulsion is one of the major processes that governs delta-plain architecture of the Rhine-Meuse delta (e.g., Stouthamer and Berendsen, 2000; Stouthamer and Berendsen, 2007). It can be regarded as a key process in the diversion of coarse-grained sediments to floodbasins. Avulsion often commences by breaching of natural levees along trunk channels, accompanied by the formation of a crevasse splay. Avulsion is controlled by both allogenic and autogenic factors. Avulsion location in the Rhine-Meuse delta, for example, is shown to be largely related to allogenic controls (rate of base-level rise, fault activity, sediment load, and discharge volume). The average period of activity of the newly formed channel belts was found to be controlled by autogenic processes (Stouthamer and Berendsen, 2007).

Natural sedimentation in the larger part of the delta plain ended ~800 cal yr BP when embankment of the river channels limited flooding of the delta plain. Deposits formed after 800 cal yr BP are considered to be associated with regulated river channels and therefore are not included in the study.

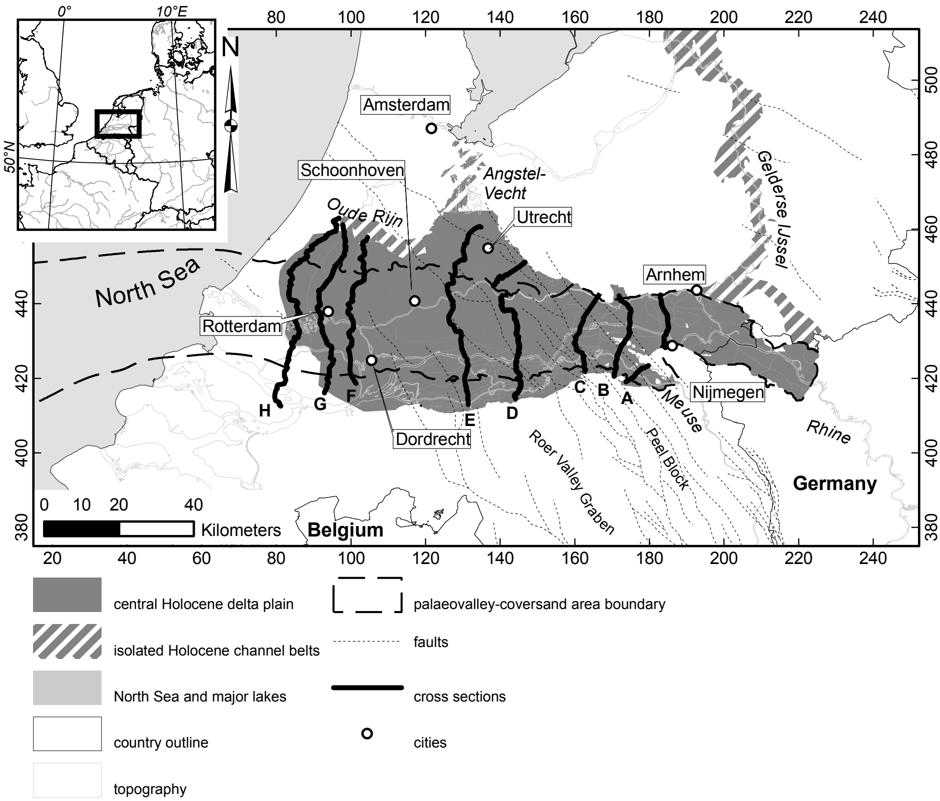


Figure 5.1. Overview of the Holocene Rhine-Meuse delta. Indicated are the positions of valley-wide cross sections. Cross sections A-E were constructed by Gouw and Erkens (2007); cross section C initially was published by Cohen (2003) and the larger part of cross section E was modified after Törnqvist (1993a). Cross sections F-H were all newly constructed by Hijma et al. (2009). The boundary between the palaeovalley and the coversand area is after Berendsen and Stouthamer (2001) and Hijma et. al (2009). Coordinates are conform the Dutch coordinate system, i.e., Rijksdriehoekstelsel.

5.2.2 Coarse-grained overbank deposits

The existing architectural framework of the Holocene Rhine-Meuse delta contains three coarse-grained overbank architectural elements: crevasse-splay deposits, organic-clastic lake fills and bay-head delta deposits (e.g., Hijma et al., 2009; Bos, 2010; Chap. 3). These elements, in general, are coarser-grained than other overbank deposits such as floodbasin deposits, but nonetheless also contain fine-grained deposits. In this study 'coarser-grained' and 'finer-grained' designate relative proportions of sand within architectural elements. The dominant coarse-grained lithofacies consists of sand, usually with diameters up to 300 μm; gravel, if present, is confined to lags in the distributary-channel deposits.

Crevasse-splay deposits form in periodically inundated floodbasins. They consist predominantly of sand-clay mixtures although sporadically sand layers occur (Stouthamer, 2001b). Crevasse-splay deposits most frequently overlie floodbasin deposits (silty clay) or in situ peat. Due to intense bioturbation and soil-formation processes during low-stage intervals (e.g., summer period) sedimentary structures in crevasse-splay deposits are often disturbed or obscured.

Organic-clastic lake fills form in peat-bounded lakes (i.e., permanently inundated) and include amalgamated mouth-bar deposits, composed of sand of up to 3 m thick (Bos, 2010; Chap. 3). These prominent sand bodies overlie finer-grained deposits (ranging from clay to clayey sand) that exhibit a coarsening-upward sequence, and organic lacustrine sediments (gyttja), which form the base of these lake fills. Their geometry is defined by the shape of the lake in which they form. The often near-vertical lateral boundaries give these deposits a rectangular shape when viewed in cross section (Bos, 2010; Chap. 3).

Bay-head delta deposits form in upper estuaries where fluvially supplied sediments are deposited under freshwater but tidal conditions (Dalrymple et al., 1992). In the western, downstream part of the delta plain, Hijma et al. (2009) found a large sand body composed of amalgamated mouth-bar deposits of a bay-head delta. Because of the narrow zone in which bay-head delta deposits occur and the few (i.e., one) examples present in the cross section, we assume that they are not very abundant in the delta. However, when present, they are a marked architectural element because they may contain a large volume of sand, which affects both the geomechanical properties and reservoir characteristics of the delta-plain deposits.

5.3 METHODS

5.3.1 Approach and input data

The spatial and temporal distribution of coarse-grained overbank deposits were identified in 8 valley-wide cross sections (A-H) that cover the Holocene Rhine-Meuse delta from the apex to the present coast (Fig. 5.1). Based on borehole descriptions (including lithological and sedimentary characteristics) and, to a lesser extent, cone penetration tests (CPT's, exclusively in cross sections F-H), a number of lithogenetic units in the fluvial domain were distinguished, e.g., channel-belt deposits, levee and crevasse-splay deposits and organics (see Table 5.1 for a complete list). Average spacing of data points (cores and CPT's) was ~100 m, though it varied considerably in cross sections F-H, being less in built-up areas and at the base of the deltaic wedge. A set of time lines divides the succession in temporal intervals, mostly spanning 1000 years (Tab. 5.2). Methodological details on the construction of the cross sections are provided by Gouw and Erkens (2007, cross sections A-E) and Hijma et al. (2009, cross sections F-H).

Based on the geological cross sections and time lines therein, we calculated the proportions of coarse-grained overbank deposits and their lithofacies composition relative to 1) cross sections in order to obtain spatial trends and 2) time intervals in order to obtain temporal trends.

Table 5.1. Identified lithogenetic units in valley-wide cross sections and those identified in this study.

Cross sections A-E	*Cross sections F-H*	*This study*
Channel-belt deposits Abandoned-channel fill	Channel-belt deposits Abandoned-channel fill	
Crevasse-splay and natural-levee deposits	Crevasse-splay and natural-levee deposits	Natural-levee deposits Crevasse-splay deposits Organic-clastic lake fills
	Bay-head delta deposits	Bay-head delta deposits
Floodbasin deposits	Humic clay (silty) Clay Silty clay with abundant tree debris	
Organics	Reed peat Wood peat Gyttja	Peat Peat gyttja

Table 5.2. Time intervals (cal yr BP) analyzed in this study and corresponding time-intervals as identified in existing cross sections of Gouw and Erkens (2007)(A-E) and Hijma et al. (2009)(F-H). The time-intervals in cross sections F-H slightly deviate from analyzed periods. This means that, for example, the volume of coarse-grained overbank deposits in cross sections F-H is slightly underestimated because deposits that formed between 7500 and 7000 are included in the subsequent time interval.

This study	*A-E*	*F-H*
<9000	<9000	<9000
9000-7000	9000-8000 8000-7000	9000-8500 8500-8000 8000-7500
7000-5000	7000-6000 6000-5000	7500-6500 6500-5000
5000-3000	5000-4000 4000-3000	5000-2500
2000-800	3000-2000 2500-800	2500-800

5.3.2 Identification of crevasse-splay deposits, organic-clastic lake fills and bay-head delta deposits

Despite the cross section's detail, they do not discriminate between crevasse-splay deposits and natural-levee deposits, which lithologically show strong similarities. Furthermore, organic-clastic lake fills, were interpreted as crevasse-splay deposits (Tab. 5.1). Information on grain size within individual elements, which is essential to assess lithofacies distribution, is only available for individual cores, archived in two large geological datasets. The borehole descriptions include information, among many other variables, on grain-size distribution, sedimentary characteristics, admixtures and rooting

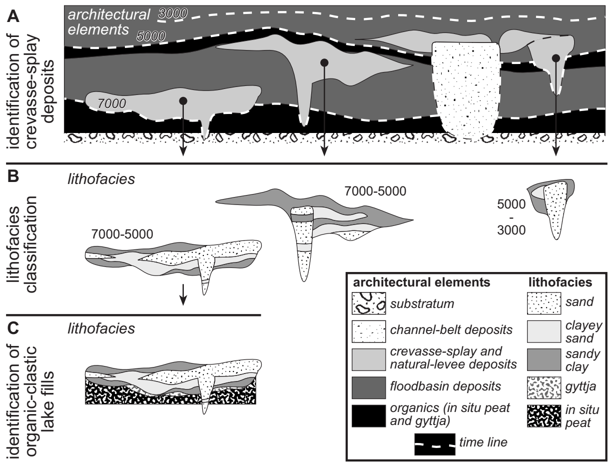


Figure 5.2. Procedure for identification of crevasse-splay deposits and organic-clastic lake fills and their lithofacies composition in the analysed cross sections. The age of the deposits is indicated in step B. See text for further explanation.

(Berendsen, 2005; TNO, 2010).

We conducted the following steps to identify crevasse-splay deposits, organic-clastic lake fills and bay-head delta deposits and the lithofacies proportions for individual sediment bodies in the cross sections (Fig. 5.2). First, we limited our analyses to those fluvio-deltaic units that contained coarse-grained overbank deposits, being 'crevasse-splay deposits and natural-levee deposits' and 'bay-head delta deposits' (Tab. 5.1, Fig. 5.2A). Second, we selected coarse-grained overbank deposits in the combined unit 'crevasse-splay deposits and natural-levee deposits' (Fig. 5.2A), which involved two consecutive procedures: 1) elements in the cross sections that occur isolated from channel-belt deposits were interpreted as coarse-grained overbank deposits, 2) elements that were connected to channel-belt deposits were further examined because they both could comprise natural-levee deposits and coarse-grained overbank depsits. To differentiate those, we applied the widely used conceptual model of meandering rivers (Miall, 1985) whereby levees become thinner and finer-grained away from the channel belt, making use of aforementioned geological databases in combination with the cross-sectional geometry. The third step in our analyses involved subdivision of coarse-grained overbank deposits into three texture classes: 'sandy clay', 'clayey sand' and 'sand' (Fig. 5.2B). The fourth step was identification of organic-clastic lake fills and bay-head delta deposits (Fig. 5.2C). Organic-clastic lake fills were designated using both geometrical and lithological properties, also of the underlying succession. The presence of gyttja underneath coarse-grained overbank deposits is a strong indication for the presence of organic-clastic lake fills (Fig. 5.2C). It recently has been suggested that the presence of gyttja in delta plains, also in the Rhine-Meuse delta, has been underestimated (Chap. 4). This means that the extent of organic-clastic lake fills, for which gyttja is an important indicator, represents minimum values. Lithologically, bay-head deltas are very similar to organic-clastic lake fills. Differentiation between both, therefore, was primarily based on micro- and macrofossil content (Hijma et al., 2009). For example, the estuarine environment of bay-head deltas deposits is recorded in the diatom association and shells. In organic-clastic lake fills, in contrast, marine signals in diatom associations and shells, if present, are only weak.

Crevasse-channel deposits were not uniformly classified across the delta plain. Hijma et al. (2009)(cross sections F-H) included crevasse-splay channels in the fluvial-channel architectural element whereas Gouw and Erkens (2007)(cross sections A-E) included those in the crevasse-splay and natural-levee architectural element. Gouw and Erkens (2007) used channel thickness (T_c) to distinguish crevasse-splay channels ($T_c < 5.2$ m) from mature fluvial channels ($T_c \geq 5.2$ m), following observations of Gouw and Berendsen (2007). Accordingly, we used 5.2 m as dividing value for fluvial and crevasse-splay channels in cross sections F-H. Gouw and Berendsen (2007) found that channel-deposit thickness increases in downstream direction. Hence, the use of 5.2 m as maximum thickness for crevasse-splay channel deposits means that cross-sectional areas and proportions of these deposits are minimum values.

5.3.3 Age estimation

The studied cross sections included time lines, which were drawn on the basis of radiocarbon dates and stratigraphic principles as outlined by Gouw and Erkens (2007). We analyzed periods of 2000 years, which were defined by the time lines 9000, 7000, 5000, 3000 and 800 cal yr BP. For cross sections A-E these time lines were already available (Gouw and Erkens, 2007). In cross section F-H some of these time lines were not determined and instead, we used time lines with slightly different ages (Tab. 5.2).

We chose 2000-year periods for the following reasons. Principal controlling factors on delta formation and the resulting architecture were shown to change at 5000 and 3000 cal yr BP (Gouw and Erkens, 2007). The period before 5000 cal yr BP was subdivided in (i) a period (9000-7000 cal yr BP) when the locus of fluvio-deltaic sedimentation was constrained by the palaeovalley (Fig. 5.1) and (ii) a period (7000-5000 cal yr BP) during which fluvio-deltaic sedimentation strongly extended upstream and, in addition, invaded areas outside the incised palaeovalley (Berendsen and Stouthamer, 2000).

5.3.4 Coarse-grained overbank-deposit proportion and lithofacies composition

The relative volumetric importance of the studied deposits, both spatially and temporally, can be quantified by the coarse-grained overbank-deposit proportion (CGODP), herewith defined as the volume of coarse-grained overbank deposits divided by the volume of all fluvial deposits and organics in the delta-plain succession. Similarly, we also obtained proportions of crevasse-splay deposits (CSDP), organic-clastic lake fills (OCLFP) and bay-head delta deposits (BHDDP), in the delta-plain succession. To obtain CGODP, CSDP, OCLFP and BHDDP, we calculated cross-sectional areas of architectural elements (crevasse-splay deposits, organic-clastic lake fill and bay-head delta deposits) and compared these with cross sectional areas of time slices and cross sections. The procedure we applied was modified from an existing method (Erkens, 2009, p. 138) and involved rasterizing the units of interest (lithofacies units per cross section, per period, per architectural element). Subsequently, the number of pixels per unit was multiplied by the cross-sectional area that one pixel represents. With photo-editing software (Adobe Photoshop 10.0) the number of pixels per unit was counted and the area represented by one individual pixel was calculated. The length of pixels was calculated by dividing the 'length represented by scale bars of the drawn cross sections' by the 'number of pixels that fitted within the length of these scale bars'.

To obtain alongstream CGODP, the cross sectional areas of the studied deposits were compared with the cross-sectional area of the Holocene fluvial deposits and organics per cross section. Cross-sectional areas for the total Holocene fluvial succession included only deposits above the reconstructed substratum level. The thus obtained cross sectional areas, therefore, do only account for Holocene upstream-delivered sediment and ignore reworked sediment of the palaeovalley. It should be noted that cross-sectional areas, notably in cross sections F-H, are minimum values because mining of peat from medieval time onwards lowered the surface and hence resulted in smaller cross sectional areas.

To obtain CGODP for time slices, the cross sectional areas of the deposits within a certain time slice must be compared with the summed cross-sectional areas of that time slice. However, reconstruction of cross-sectional areas of time slices is hampered due to erosion during successive periods. Alternatively, we used reconstructed delivered-sediment volumes per time slice (Erkens, 2009), which were based on extensive database analyses and GIS computations and whereby erosion was taken into account and corrected for. Comparing these volumes with cross sectional areas does not yield true proportions, but still enables comparison with other time slices and thus detection of temporal trends.

An important variable determining delta-plain architecture is floodplain or delta-plain width (e.g., Bridge and Leeder, 1979). We measured delta-plain widths at the locations of the cross sections for four moments during the Holocene corresponding to the four analysed periods using a map indicating Holocene delta-plain growth (Erkens, 2009, his Fig. 6.4).

5.4 RESULTS

5.4.1 Coarse-grained overbank-deposit proportion

In the Holocene Rhine-Meuse delta plain, CGODP is 0.07, which includes crevasse-splay deposits (0.046), organic-clastic lake fills (0.010) and bay-head delta deposits (0.015).

Spatial variability

In general, the volume of coarse-grained overbank deposits and CGODP increases in downstream direction (Fig. 5.3C).

Crevasse-splay deposits occur throughout the delta-plain (Fig. 5.3A). The CSDP in the cross sections, however, has an optimum in the proximal delta plain (cross sections B-E), being at its maximum in cross section D (0.075) (Fig. 5.3A). Upstream (cross section A) and especially downstream (cross sections F-H) from these cross sections, proportions

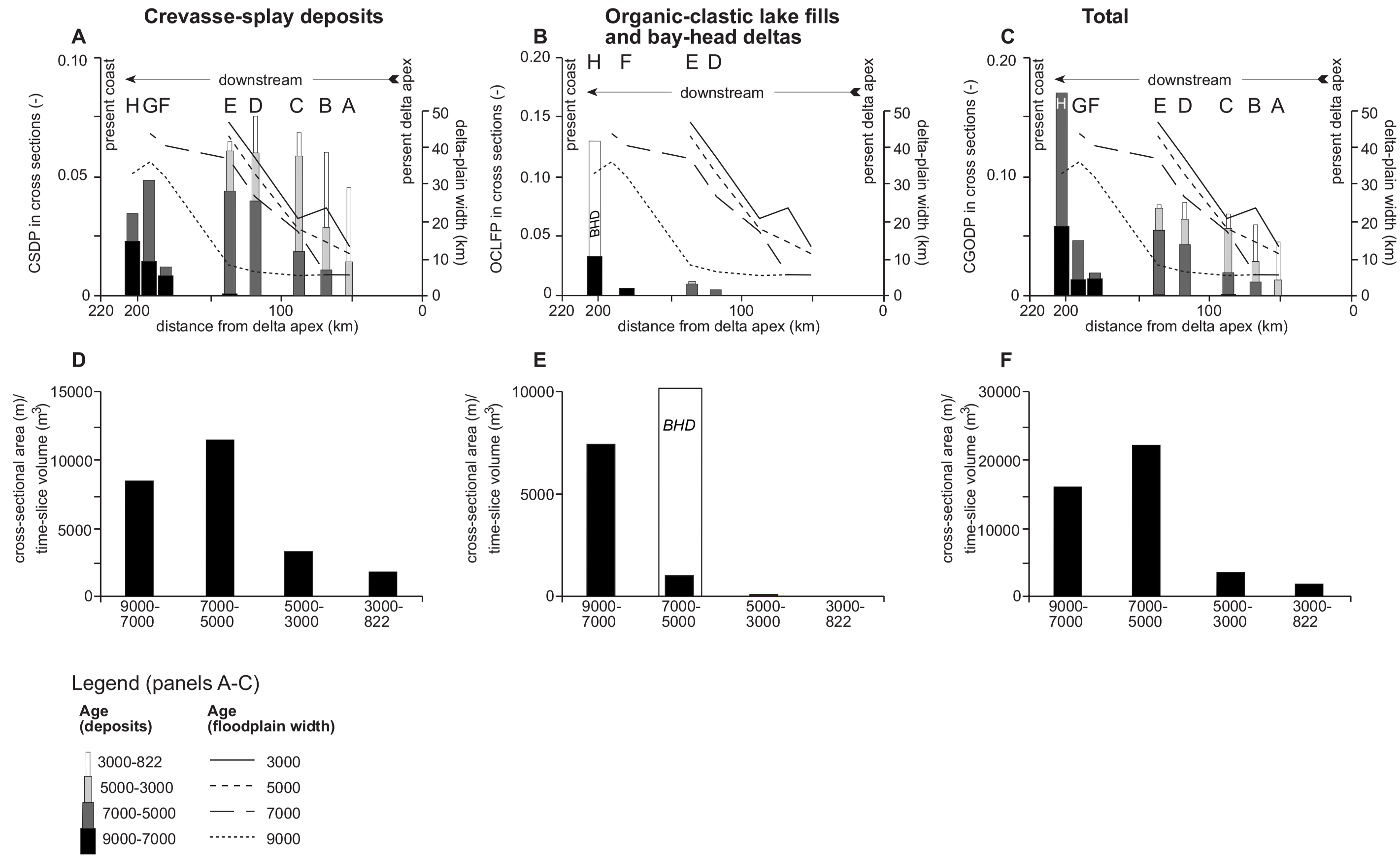


Figure 5.3. Spatial (A-C) and temporal (D-F) distribution of the proportions of crevasse-splay deposits (left-hand panels), and organic-clastic lake fills and bay-head delta deposits (middle panel) and their summed values. Floodplain width is given in diagrams A-C. See Appendix 8 for a full-colour version.

of crevasse-splay deposits are considerably lower. Also in individual cross sections, the CSDP has an optimum 50-100 km downstream of the corresponding delta apex, which shifted upstream, in response to base-level rise.

Organic-clastic lake fills exclusively occur in the distal part of the delta plain, in cross sections D-H (Fig. 5.3B). The OCLFP increases strongly in downstream direction, being 0.04 in the most downstream part of the delta plain (cross section H, Fig. 5.3B).

Bay-head delta deposits were confined to the most downstream part of the study area (cross section H), where BHDDP is 0.13 (Fig. 5.3B).

Temporal variability

Coarse-grained overbank deposits predominantly formed prior to 5000 cal yr BP (Fig. 5.3D-F). The majority of crevasse-splay deposits, also proportionally to time-slice volume, formed prior to 5000 cal yr BP, being at its maximum between 7000 and 5000 cal yr BP (Fig. 5.3D). Organic-clastic lake fills most extensively formed in the period 9000-7000 cal yr BP (Fig. 5.3E). The abundance of these deposits in the delta-plain succession strongly decreases thereafter (compare time slices 9000-7000 and 7000-5000 in Fig. 5.3E). After 5000 cal yr BP, extensive organic-clastic lake-fill formation did not occur despite the occurrence of a relatively wide floodplain (Fig. 5.3B). Bay-head delta deposits formed between 7000 and 5000 cal yr BP (Fig. 5.3E).

5.4.2 Sand body proportion in coarse-grained overbank deposits

The proportion of sand in coarse-grained overbank deposits varies considerably. The minimum, average and maximum proportions (per time slice, per cross section) of sand in crevasse-splay deposits are 0.00, 0.26 and 0.77 respectively. For organic-clastic lake fills, we found values of 0.02, 0.28 and 0.38 respectively. The bay-head delta deposits were entirely composed of sand (Hijma et al., 2009). The proportion of sand bodies contained in coarse-grained overbank deposits in the Holocene delta-plain is 0.030, which encompasses sand bodies in crevasse-splay deposits (0.012), organic-clastic lake fills (0.003) and bay-head delta deposits (0.015). The geometry of sand bodies in coarse-grained overbank deposits varies. Sand bodies in crevasse-splay deposits are chiefly channel deposits whereas sand bodies in organic-clastic lake fills predominantly are amalgamated, sheet-like mouth-bar deposits.

Spatial variability

Crevasse-splay deposits in general are coarser-grained in the distal delta plain. The proportion of sand, for example, increases from 0.10 (cross section B) to 0.62 (cross section F, Fig. 5.4A). The trend within individual time slices, however, is less well expressed or absent (Fig. 5.4C, G, and I) except for time slice 7000-5000 (Fig. 5.4E) where crevasse-splay deposits in the downstream part (cross sections F, G) are much sandier than those in the proximal part (cross sections B-E). The overall downstream coarsening largely reflects a temporal trend. First, through time, crevasse-splay deposits become finer-grained (Fig.

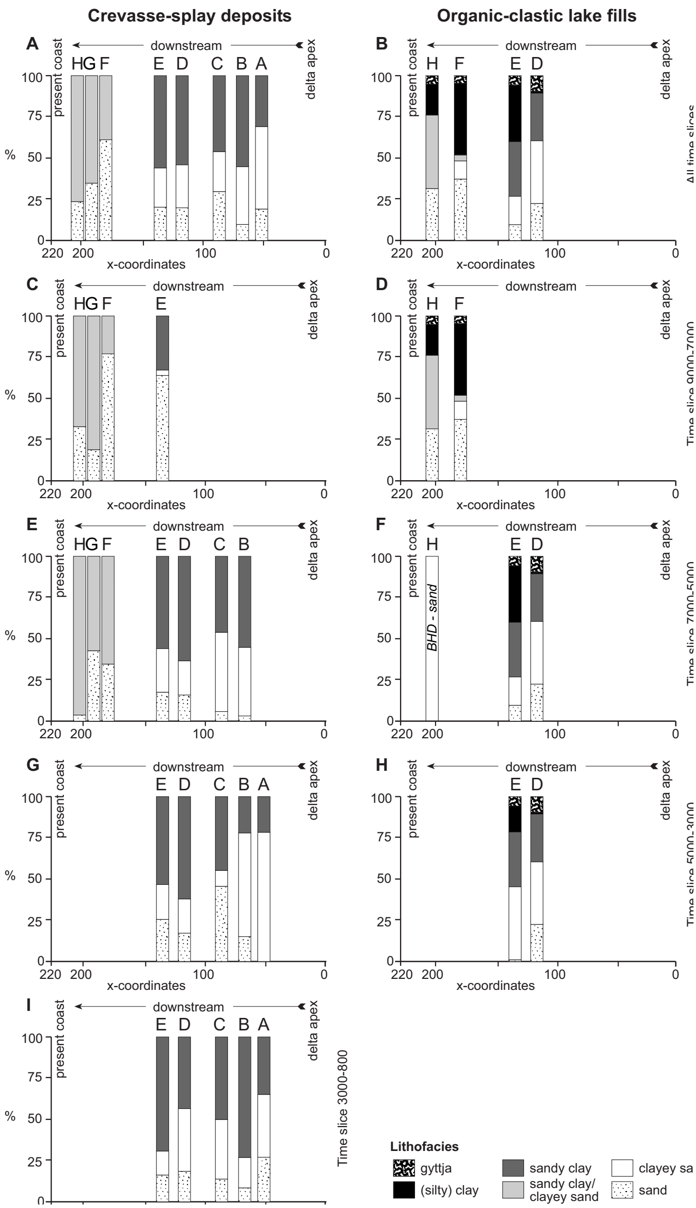


Figure 5.4. Lithofacies composition of crevasse-splay deposits and organic-clastic lake fills related to longitudinal position in the delta plain. Diagrams A and B indicate composition for the whole delta. The other diagrams give the composition for various periods, being 9000-7000 cal yr BP (C and D), 7000-5000 cal yr BP (E-F), 5000-3000 cal yr BP (G and H) and 3000-800 cal yr BP (I). Organic-clastic lake fills did not form after 3000 cal yr BP in the central delta plain. See Appendix 8 for a full-colour version.

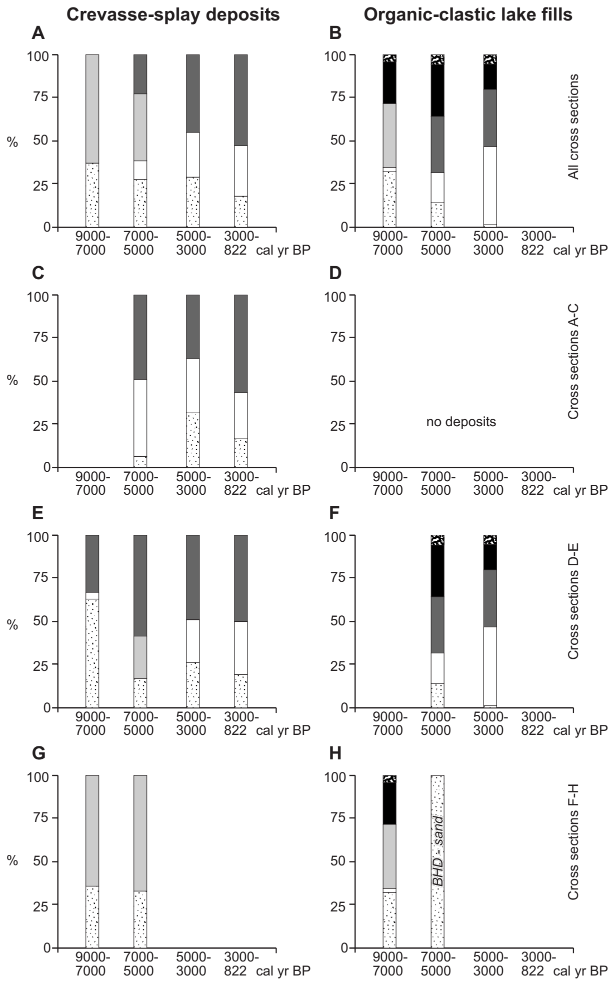


Figure 5.5. Lithofacies composition of crevasse-splay deposits and organic-clastic lake fills related to time slices. Diagrams A and B indicate composition for the whole delta. The other diagrams give the composition for three parts of the delta, being the proximal delta plain (summed values of cross sections A-C), the intermediate delta plain (summed values of cross sections D and E) and the distal delta plain (summed values of cross sections F-H). For legend, see figure 5.4. See Appendix 8 for a full-colour version.

5.5A, E). Second, the delta-plain area simultaneously expanded during the Holocene mainly by upstream extension, which means that overall proximal delta-plain successions are younger than their distal counterparts. The combination of both observations implies that coarse-grained overbank deposits in the proximal delta plain averagely contain less sand than those in the distal delta plain. The downstream coarsening of crevasse-splay deposits and organic-clastic lake fills, therefore, is believed to reflect fining through time in combination with an upstream shift of the delta plain (Fig. 5.6).

Also organic-clastic lake fills exhibit a downstream increase in the proportion of sand, increasing from 0.09 in cross section E to 0.88 in cross section H (Fig. 5.4B). Within time slices, there is a maximum of two cross sections wherein organic-clastic lake fills have been found (Fig. 5.3E), which limits detection of downstream trends in the composition of these deposits in time slices. The upstream shift of the occurrence of organic-clastic

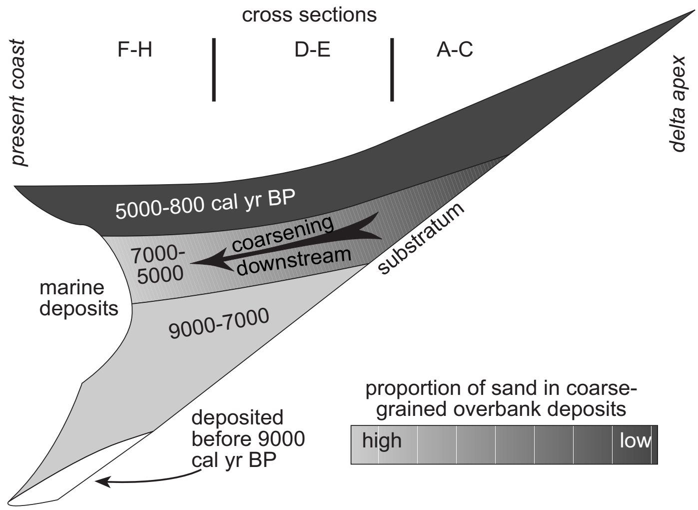


Figure 5.6. Schematic cross-section of observed sand proportions within crevasse-splay deposits throughout the Rhine-Meuse delta for the period after 9000 cal yr BP. This overview illustrates how a temporal trend affects a spatial trend. Crevasse-splay deposits formed between 9000 and 7000 cal yr BP, for example, are coarser-grained than younger crevasse-splay deposits. Deposition in the proximal delta plain (cross sections A-C) commenced after 7000, which means that crevasse-splay deposits in that part of the delta, on average, are finer-grained than those in the distal delta plain (cross sections F-H). As a result, a general coarsening-downstream trend is observed for crevasse-splay deposits, but this largely reflects a temporal fining trend (see text for explanation).

lake fills from cross sections F and H (Fig. 5.4D) to D and E (Fig. 5.4F, H) reflects the upstream shift of the entire delta-plain. The results, therefore, suggest that organic-clastic lake fills exclusively form in a narrow zone perpendicular to the general flow direction within the distal delta plain (see also the discussion section).

Temporal variability

The lithofacies composition of crevasse-splay deposits seems to be correlated with the period of formation: they become finer-grained with decreasing age (Fig. 5.5A). Crevasse-splay deposits, for example, that formed between 9000 and 7000 cal yr BP are coarser-grained than those formed between 3000 and 800 cal yr BP (Fig. 5.5A). This trend is well visible in the central part of the delta plain (cross sections D-E, Fig. 5.5E), where the proportion of sand in the oldest crevasse-splay deposits is 0.63 and decreases to 0.19 in the most recent time slice.

Organic-clastic lake fills also become finer-grained with decreasing age (Fig. 5.5B). This trend, however, cannot be further spatially resolved (Fig. 5.5D, F, and H) due to the limited temporal range of the deposits within individual cross sections.

5.5 DISCUSSION

5.5.1 Explanation of spatial and temporal distribution patterns of coarse-grained overbank deposits

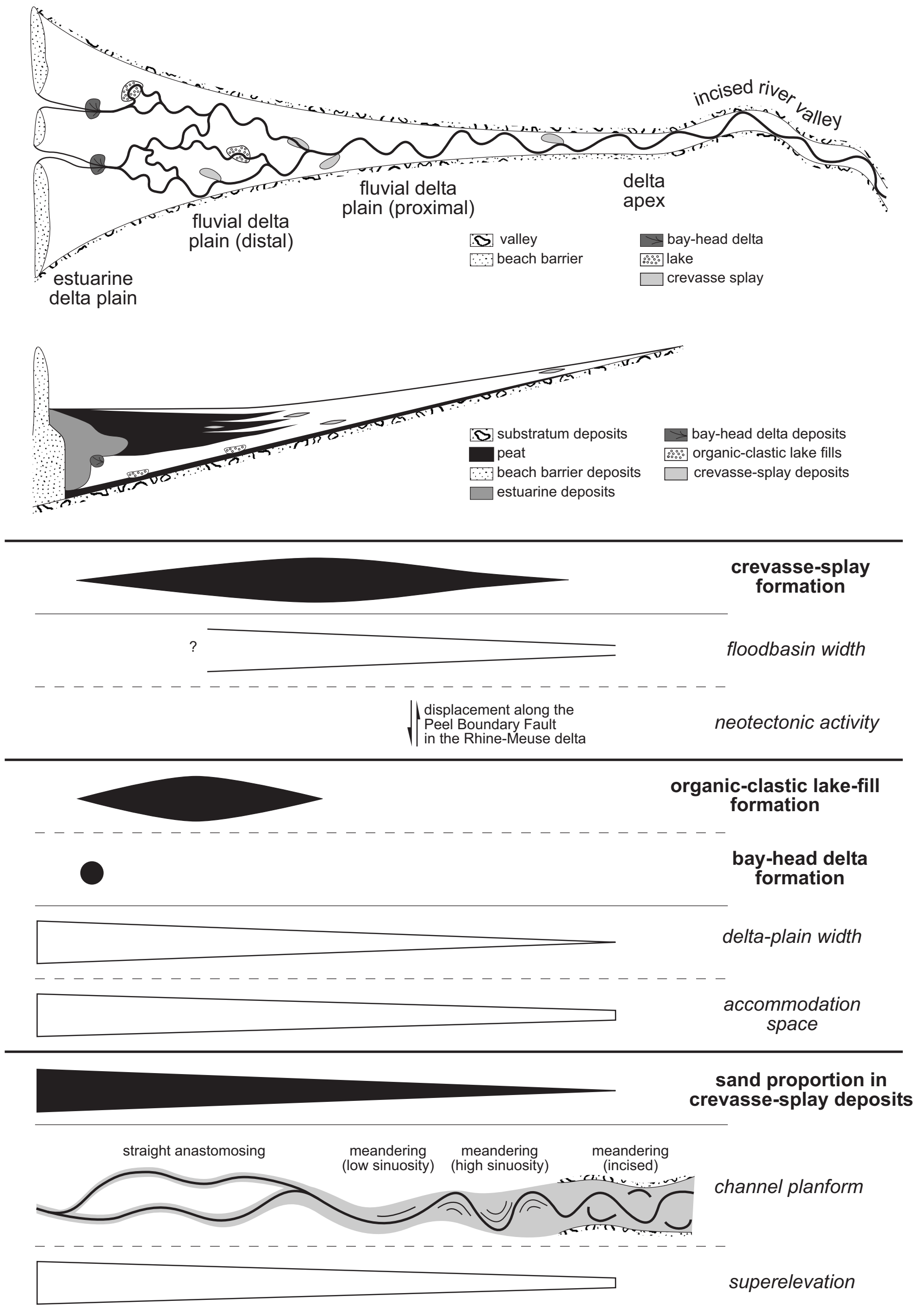
Spatial and temporal variability of crevasse-splay deposit proportion

Crevasse-splay deposits are present throughout the delta plain, with highest proportions within individual time slices occurring 50-100 km downstream of the associated delta apex (Fig. 5.3A). We propose that preservation of deposits, width of floodbasins, and avulsion-related factors, affect the distribution of coarse-grained overbank deposits.

Preservation of delta-plain deposits changed during the Holocene, which partly explains CSDP trends. Erkens (2009) published preservation values for floodbasin deposits, which are assumed to be comparable to those of coarse-grained overbank deposits because they are usually encased in floodbasin deposits. He showed that preservation increases from 57% in cross sections A-C to 74% and 76% in cross sections D-E and F-H respectively. Correction for erosion thus means that the CSDP in cross sections A-C would increase relatively to those in cross sections D-E, but the effect on the trends in our results, however, would be minimal. For example, the difference between crevasse-splay deposits that formed between 7000 and 5000 cal yr BP in cross sections A-C and D-E would become smaller (Fig. 5.3A), but an increasing CSDP in downstream direction would still exist. Hence, also other factors than preservation control CSDP variability.

The width of floodbasins, which is approximated by the spacing between (palaeo) channels, is likely to control the CSDP too (Fig. 5.7, 5.8). Narrow floodbasins limit splay development and hence are characterized by relatively low CSDP (Fig. 5.8A). The CSDP will increase when floodbasin width increases (Fig. 5.8B), but when floodbasins become very wide (Fig. 5.8C), crevasse-splay deposits will not be able to occupy the entire width of the floodbasin ultimately leading to reduction of the CSDP. Measurements of floodbasin widths on palaeogeographic maps (Erkens, 2009, his Appendix D) indeed suggest that lateral distances between palaeochannels of the same age overall increases in downstream direction (Fig. 5.9). Further, optimum CSDP values for consecutive time slices were found in cross section D (7000-5000 cal yr BP), cross section C (5000-3000 cal yr BP) and cross sections A-B (3000-800 cal yr BP) (Fig. 5.3A). The floodbasin widths that correspond with these optima are remarkably similar and range between 3.1 and 3.6 km (Fig. 5.9). We, therefore, tentatively propose that average floodbasin widths of 3.1-3.6 km yield optimum proportions of crevasse-splay deposits (cf. Fig. 5.8B) in the Rhine-Meuse delta.

Other factors that control the spatial and temporal distribution of crevasse-splay deposits are probably related to avulsion. Avulsion often is initiated by levee breaching and the subsequent formation of a crevasse splay in the adjacent floodbasin. The factors that control the spatial distribution of avulsions, therefore, likely have also greatly influenced the spatial distribution of crevasse-splay deposits. Avulsion locations in the Rhine-Meuse delta are controlled by sea-level rise (upstream shift delta apex) and tectonic activity (Stouthamer and Berendsen, 2000; Stouthamer and Berendsen, 2007). Cohen et al. (2005)


incised river valley
delta apex
fluvial delta plain (proximal)
fluvial delta plain (distal)
estuarine delta plain
valley
beach barrier
bay-head delta
lake
crevasse splay
substratum deposits
peat
beach barrier deposits
estuarine deposits
bay-head delta deposits
organic-clastic lake fills
crevasse-splay deposits
crevasse-splay formation
?
floodbasin width
displacement along the Peel Boundary Fault in the Rhine-Meuse delta
neotectonic activity
organic-clastic lake-fill formation
bay-head delta formation
delta-plain width
accommodation space
sand proportion in crevasse-splay deposits
straight anastomosing
meandering (low sinuosity)
meandering (high sinuosity)
meandering (incised)
channel planform
superelevation

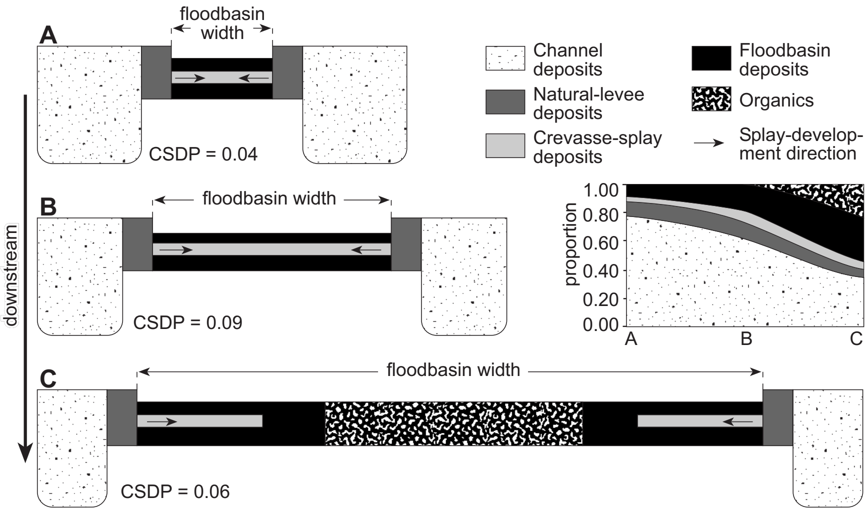


Figure 5.8. Proposed schematic relation between floodbasin width and the proportion of crevasse-splay deposits (CSDP). Floodbasin width is relatively small in cross section A, which limits the areal extent of crevasse-splay deposits. The proportion of crevasse-splay deposits increases farther downstream (cross section B) as floodbasin width increases. When floodbasin width increases, CSDP decreases in favour of floodbasin deposits. Note that the proportions are comparable to those found in the Rhine-Meuse delta.

showed that neotectonic activity also influenced the position of crevasse-splay deposits. They found an abrupt increase of the abundance of crevasse-splay deposits across the Peel Boundary Fault (Fig. 5.1). Owing to differential subsidence, aggradation, and hence the formation of accommodation space, started earlier in time and at higher pace in the Roer Valley Graben downstream of the fault compared to the Peel Block, upstream of the fault (Berendsen and Stouthamer, 2000). In the vicinity of this fault, CSDP apparently is chiefly controlled by, and positively related to the quantity of accommodation space.

Spatial and temporal variability of OCLFP

Organic-clastic lake fills are limited to the distal delta plain. An earlier published example of organic-clastic lake fills near Schoonhoven (Bos, 2010; Chap. 3), which is

← Figure 5.7. Schematic relation between observed spatial trends (in plan view and longitudinal cross section) of crevasse-splay deposits, organic-clastic lake fills, bay-head delta deposits and the proportion of sandy lithofacies in crevasse-splay deposits (bold text in the diagram) along with their contingent controlling factors (italic text in the diagram). Note that the downstream-changing proportion of sandy lithofacies in crevasse-splay deposits accounts only for the period 7000-5000 cal yr BP. Thereafter, the proportion of sandy lithofacies in crevasse-splay deposits is fairly constant in downstream direction. Channel planform has been modified from Wolfert (2001).

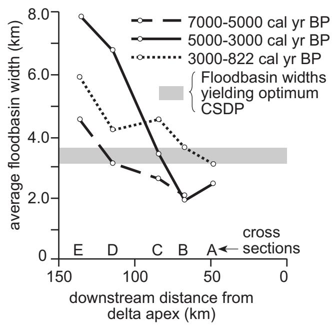


Figure 5.9. Average floodbasin widths in the Rhine-Meuse delta for three time slices and differentiated for cross sections (Fig. 5.1). Indicated is range for average floodbasin widths that coincides with optimum CSDP values. Measurements were carried out on existing palaeogeographic maps (Erkens, 2009).

located between cross sections E and F and was formed between 7000 and 5000 cal yr BP, fits well in the observed distribution pattern of the studied cross sections (Fig. 5.3B), both in time and space.

High rates of base-level rise and wide floodplains facilitate the formation of organic-clastic lake fills. Until the palaeovalley is filled, also low alongstream gradients of the palaeovalley has a positive effect thereon. These factors provide the necessary conditions for the formation of accommodation space. When the volume of clastic-sediment input cannot fill the created accommodation space, extensive peat-forming wetlands will form, which in turn are suitable areas for the development of organic-clastic lake fills (Bos, 2010; Chap. 3). In the central delta plain, OCLFP decreases with time (Fig. 5.3E), corresponding to decreasing rates of base-level rise.

Spatial and temporal variability of bay-head delta deposits

Bay-head delta deposits were found in cross section H (Hijma et al., 2009). Other studies, nonetheless, suggest that this unit is not a unique phenomenon in the Holocene Rhine-Meuse delta. Hijma et al. (2009), for example, expect bay-head delta deposits to be present also at other locations, either in between cross sections or not recognized due to low data quality and low data resolution. Further, a recently formed fluvial-tidal sediment body, containing $151*10^6$ m^3 sand, that is present slightly eastward of cross section F (Kleinhans et al., 2010), also can be regarded as bay-head delta deposits. The spatial distribution of bay-head delta deposits, however, is strongly limited to the relatively narrow transition zone between the fluvial and marine realm (Fig. 5.7). Their relatively large volume, nonetheless, especially of the sand bodies, points to the significance of these deposits, e.g., in the light of reservoir properties.

5.5.2 Explanation of temporal and spatial lithofacies variability

Spatial trends of sand-body proportions in coarse-grained overbank deposits

Stouthamer (2001b), tentatively proposed that the composition of crevasse-splay deposits would become finer-grained in downstream direction. Contrastingly, the composition of crevasse-splay deposits and of coarse-grained overbank deposits in general, although variable within individual time slices, does not show clear spatial trends (e.g., Fig. 5.4F, I). An exception are crevasse-splay deposits that formed between 7000 and 5000 cal yr BP, whereby the proportion of sandy lithofacies remarkably increases from 0.04 (cross section B) to 0.41 (cross section G) in downstream direction (Fig. 5.4E).

Channel planform is believed to partially determine the composition of crevasse-splay deposits in the Rhine-Meuse delta (Stouthamer, 2001b). Helical flow in meandering-river channels transports sandy bed-load to inner bends, leaving little sand for diversion into crevasse channels, which preferentially are positioned in outer bends (Stouthamer, 2001b). Hence, crevasse-splay deposits forming in these settings are relatively fine-grained. Because this mechanism is not effective or absent in straight-anastomosing river channels, crevasse-splay deposits along these systems are generally coarser-grained. In the Rhine-Meuse delta, straight-anastomosing river systems along with large-scale crevassing were dominant in the distal central delta plain between 8000 and 4000 cal yr BP, whereas meandering rivers were dominant in the proximal delta plain and along the edges of the distal delta plain. The relatively large proportion of sand in crevasse-splay deposits that formed between 7000 and 5000 cal yr BP in the distal delta-plain (Fig. 5.4E), therefore, can be associated with the dominance of straight-anastomosing rivers (Fig. 5.7).

Superelevation differences between straight-anastomosing and meandering river systems might also contribute to differences in lithofacies composition of coarse-grained overbank deposits (Fig. 5.7). Superelevation is the elevation difference between the water level in the river channel at bankfull discharge (i.e., levee crest) and the surface level in the adjacent floodbasin (Heller and Paola, 1996). Because lateral migration in meandering river systems obstruct natural levees in the outer bend to reach maximum potential heights we assume that superelevation in these systems is low compared to straight-anastomosing river systems. Consequently, floodbasins of straight-anastomosing rivers tend to be deeper and have the largest water depths during high-discharge events. When water depth in floodbasins is comparatively large, flow velocity will drop strongly to almost zero owing to flow divergence. As a result, transported sediment will be deposited. This mechanism will not be effective when water depth is small, which we hypothesize to be the case for meandering-river floodbasins. Then, flow divergence is hampered, resulting in partially channelized flow. Sand, therefore, will be predominantly deposited in the channels or transported further downstream. As a consequence, crevasse-splay deposits in meandering river systems are finer-grained than their counterparts in straight-anastomosing river systems. Furthermore, large superelevation (straight-anastomosing rivers) facilitates deep incision of crevasse channels, which enlarges the chance that they tap coarse-grained bed material of the trunk channel and yielding relatively coarser-grained splay deposits (Slingerland and Smith, 1998). Splays along meandering rivers, oppositely, will be relatively fine-grained. However, superelevation incorporates elevation of floodbasin surface, which does not optimally approximate potential gradients of crevasse-splay channels. Instead, the groundwater level in floodbasins relative to bankfull water level in the trunk channels probably better relates to potential gradients of crevasse-splay channels. We, therefore, contend that fluvial style and superelevation (using floodbasin groundwater level) influence the lithofacies composition of coarse-grained overbank deposits.

The properties of the substratum probably also affect the lithofacies composition of coarse-grained overbank deposits. Crevasse-splay deposits that exist at or near the base of Holocene successions are relatively sandy in various locations in the Rhine-Meuse delta, which is attributed to reworking of sandy substratum material (Stouthamer, 2001b; Cohen et al., 2009). Especially when a channel has the ability to erode topographically elevated substratum deposits (e.g., aeolian river dunes) a potentially significant volume of sand is added to the upstream supplied sediment (Cohen et al., 2009).

Temporal trends of sand-body proportions in coarse-grained overbank deposits

Coarse-grained overbank deposits within the Holocene Rhine-Meuse delta-plain succession tend to become finer grained with decreasing age (Fig. 5.5A, 5.6), as has been proposed for crevasse-splay deposits by Stouthamer (2001b).

What possibly controls the temporal fining of crevasse-splay deposits is decreasing water depth in floodbasins with younger age. Water depth is expected to be larger when rates of base-level rise are high, thereby enhancing trap efficiency, which would explain the decreasing proportion of sand in crevasse-splay deposits with time. This explanation is supported by observations by Van der Woude (1981), chiefly based on lithogical and palynological analyses who found, predominantly water logged floodbasins before 4700 cal yr BP and much drier conditions thereafter. The decreasing rate of base-level rise probably is reflected in the temporal fining trend of crevasse-splay deposits.

Alternatively, downstream fining might have had a larger imprint on the sediment composition of younger deposits because the delta expanded with time in longitudinal direction. This means that the pathway for sediment to reach downstream portions of the delta plain increases with time, which probably means that fining of sediment is more effective during younger periods. As a consequence, sediments on average become finer-grained with time. The longer average distance to the sediment source during later periods, also means that incorporation of sandy substratum material is more limited than during earlier periods because in the distal delta plain, younger channels are not incised into the substratum. The effect of this mechanism, however, is not supported by the spatial distribution of the lithofacies composition. The anticipated downstream fining of crevasse-splay deposits is not reflected in our results (Fig. 5.4C, E, G, I).

We, therefore, propose that temporal fining of coarse-grained overbank deposits in the Holocene Rhine-Meuse delta is controlled by temporally decreasing floodbasin water depth.

5.5.3 Influence of coarse-grained overbank deposits on reservoir properties

Predictions of reservoir exploitability, which partly depends on reservoir volume (= volume of connected sand bodies) and connectedness, are commonly based on numerical simulation models. These so-called reservoir models, especially recent ones (e.g., Pyrcz et al., 2009) recognize the potential of crevasse-splay deposits as reservoir rock, but in general lack field data for model conditioning. Hence, the notion that coarse-grained overbank deposits may form >10% of the deposits in distal parts of delta plains (cross section H in Fig. 5.3B, C) underlines that they should not be ignored in reservoir models. In this section, we discuss the significance of overbank sand bodies for reservoir volume and connectedness.

Reservoir volume

In the Holocene Rhine-Meuse delta plain channel- and overbank-sand bodies volumetrically form 40% (Erkens, 2009) and 3% (this study) respectively of the succession. The proportion of channel-sand bodies decreases in downstream direction from 70% (cross section A) to 30% (cross section E), whereas the proportion of overbank sand increases in downstream direction from 1% (cross section A) to 11% (cross section H), implying that the contribution of overbank sand bodies to the reservoir volume increases in downstream direction. The preserved sand body of the bay-head delta deposits in cross section H, for example, has a volume of ~120 million m^3 (Hijma et al., 2010), which accounts for approximately 11% of the total fluvial sediment in that part of the delta plain. In cross section H, channel deposits form 21% of the fluvial deposits (Erkens, 2009), which means that 34% (= (11/(11+21))*100%) of the reservoir volume in that part of the delta is formed by the bay-head delta deposits. Similarly, preserved sand bodies of organic-clastic lake fills in the relatively isolated Angstel-Vecht area (Fig. 5.1), have a combined volume of 44 million m^3 (Bos, 2010; Chap. 3), whereas the volume of channel sand bodies is 99 million m^3. Here, overbank sand bodies account for 31% of the reservoir volume. Another example is the Gelderse IJssel, where 60% of the Holocene deposits comprises sand bodies, which are included in channel deposits (43%) and in coarse-grained overbank deposits (17%) (based on Cohen, unpublished data). In the Gelderse IJssel area (Fig. 5.1), sand in coarse-grained overbank deposits comprises 39% of the total reservoir volume. Upstream of these deposits, the IJssel has eroded the locally elevated substratum, which explains the large proportion of sand in the coarse-grained overbank deposits.

These examples show that sand bodies contained in overbank deposits may form up to 39% of total volume of reservoir rocks in distal delta plains.

Reservoir connectedness

The connectedness ratio (CR) is a principle parameter for alluvial architecture, especially concerning reservoir characterization as it partly determines the exploitability of reservoirs, and is a measure for the degree to which individual channel-sand bodies are amalgamated (e.g., Leeder, 1978). The CR is defined as 'the length of horizontal contact between connected channel belts divided by the total horizontal width (or length) of channel-belt deposits in the transect' within a single cross-valley transect (Mackey and Bridge, 1995), and can be expressed as follows

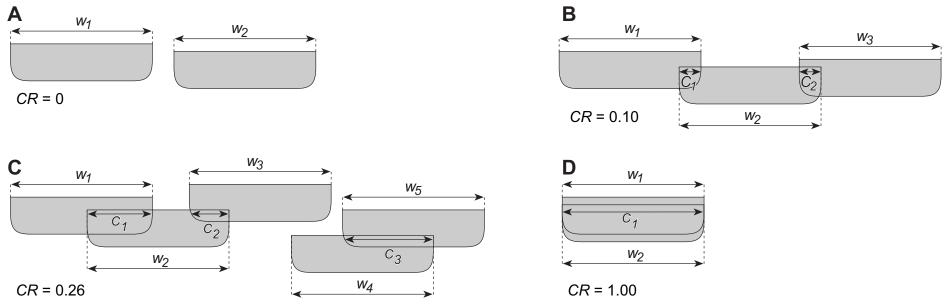


Figure 5.10. Schematic visualisation of possible channel-sand body configurations. The connectedness ratio (CR) partly reflects this configuration. CR ranges from '0' (individual channel-sand bodies are completely isolated) to '1' (channel-sand bodies totally overlap). In the Rhine-Meuse delta plain, CR values were found to range between 0.08 (cross section E in Fig. 5.1) and 0.21 (cross section A in Fig. 5.1) (Gouw, 2008) although in model simulations, higher CR values (up to 0.69) were found (Mackey and Bridge, 1995).

$$CR = \frac{\sum_{i=1}^{n-1} c_i}{\sum_{j=1}^{n} w_j} \qquad \text{[Eq. 5.1]}$$

Wherein c is the shared horizontal cross sectional width of two individual channel-sand bodies, w is the width of an individual channel-sand body and n the number of channel-sand bodies in the cross section. The CR ranges from 0, whereby individual sand bodies are completely isolated, to theoretically 0.5, which means that individual sand bodies in a cross section are all connected over the full width, i.e., vertically stacked (Fig. 5.10). In the latter case, the CR is by definition set to 1 (cf. Mackey and Bridge, 1995). Low CR values, thereby, indicate the presence of multiple smaller reservoirs, whereas high CR values indicate that the sediment body is a single but often bigger reservoir. The latter obviously being associated which much lower costs for exploitation and therefore being more interesting from an economic point of view.

Strictly, CR involves only sand that is deposited in fluvial channels. Practically, though, reservoir connectedness potentially also benefits from sand bodies contained in overbank deposits. Especially when the areal extent of the individual sand bodies is large, they may be able to connect isolated channel belts (Fig. 5.11). Despite the relatively large total volume of sand contained in crevasse-splay deposits, we consider them not very suitable as connector because individual units are relatively small and with sand concentrated in channels that are often isolated in cross sections from other reservoir elements. Laterally extensive sand bodies in organic-clastic lake fills and bay-head delta deposits, in contrast, may form effective connectors for otherwise unconnected channel-sand bodies. In this way, they potentially yield higher CR values (Fig. 5.9). In addition, the presence

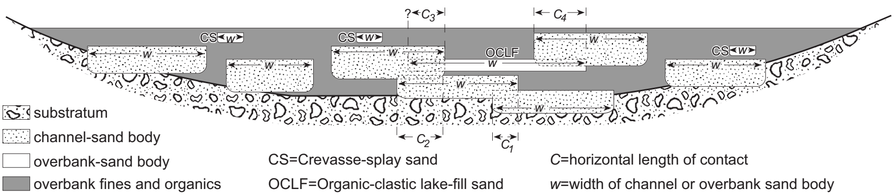


Figure 5.11. Schematic cross section illustrating the effect of overbank sand bodies on the connectedness ratio (CR) in distal delta-plain successions. The CR of channel-sand bodies, using only the widths of the channel-sand bodies and C1 and C2, is 0.09, which is realistic for the distal Rhine-Meuse delta plain (Gouw, 2008). The CR increases to 0.17 when overbank-sand bodies are considered as reservoir elements, thereby using the width of all sand bodies in the cross section and C1 to C4. Note that the left-hand edge of the organic-clastic lake fill cannot be determined exactly. The CR, therefore, may vary slightly from what has been calculated. Nevertheless, including overbank-sand bodies in CR calculations, will lead to higher and more realistic CR values.

of overbank sand bodies promotes the local development of wider channel-sediment bodies. Bos et al. (2009; Chap. 2) showed that lateral channel migration is enhanced by the presence of sandy subsoils contained in organic-clastic lake fills, ultimately resulting in wider channel-belt deposits. Apart from larger reservoir volumes, it potentially also enlarges CR values because younger channel belts more likely are connected to these wider channel-sand bodies. Hence, the CR sensu stricto, i.e., excluding sand bodies contained in overbank deposits, likely underestimates the 'true' reservoir connectedness in areas where organic-clastic lake fills or bay-head delta deposits occur in the subsoil, i.e., in distal delta plains.

5.6 CONCLUSIONS

Coarse-grained overbank deposits in the Holocene Rhine-Meuse delta form 7.1% of fluvial delta-plain volume (including organics) and are composed of crevasse-splay deposits (4.6%), organic-clastic lake fills (1.0%) and bay-head delta deposits (1.5%).

Proportionally, crevasse-splay deposits are most abundant mid-way the delta apex and the coast but they occur throughout the fluvial delta-plain succession. The CSDP was found to be related with floodbasin width and affected by neotectonic activity. Narrow floodbasins restrict crevasse-splay deposit formation, which yields relatively low CSDP. Wider floodbasins facilitate the full development of crevasse-splay deposits, yielding higher CSDP. Our results suggest that when average floodbasin width exceeds 3.6 km CSDP decreases because the increased space is not occupied by crevasse-splay deposits, but rather by floodbasin deposits and organics.

Organic-clastic lake fills are restricted to the distal delta plain. The distribution of organic-clastic lake fills is associated with extensive peat accumulations, which form when accommodation-space creation outpaces sedimentation. This requires high rates of base-level rise (occurring before ~5000 cal yr BP in the Rhine-Meuse delta) and/or wide delta plains (present in the distal Rhine-Meuse delta plain). In the central delta, organic-clastic lake fills formed before 5000 cal yr BP, when base-level rise controlled delta-plain formation. Extended delta-plain width facilitated the formation of organic-clastic lake fills after 3000 cal yr BP in areas adjacent to the central delta plain.

Bay-head delta deposits, by definition, are confined to the upper estuary.

In general, it seems justified to conclude that distribution patterns of coarse-grained overbank deposits reflect spatial variability, rather than temporal variability

Volumetrically, sand bodies averagely form 26-30% of coarse-grained overbank deposits. The proportion of sand bodies in overbank deposits generally decreases with younger age. Because the upstream limit of aggradation shifted upstream over time, this temporal trend is also reflected spatially: on average, coarse-grained overbank deposits in the distal delta plain contain more sand than those in the proximal delta plain. We attribute the lithofacies variability of coarse-grained overbank deposits to variations in channel planform, superelevation and substratum properties. In meandering river systems crevasse splays deposits are relatively fine-grained owing to helical flow, which transports relatively coarse-grained bed material to the crevasse-splay channel. Owing to the absence of helical flow in straight-anastomosing river systems crevasses along these systems are relatively coarse-grained. Furthermore, large superelevation values stimulate crevasse channels to incise deeper, thereby enlarging the proportion of bed-load material that is contained in these crevasse-splay deposits. Substratum properties, in addition, may also influence the composition of crevasse-splay deposits, especially when they form just downstream of an area where the channel incised into sandy substratum.

Because of the relatively large volume of sand bodies in overbank deposits, especially in distal delta plains, ignoring coarse-grained overbank deposits in reservoir models potentially leads to an underestimation of reservoir volumes with 39%. Furthermore, organic-clastic lake fills and bay-head delta deposits potentially form connectors in distal delta plains, thereby increasing the connectedness ratio.

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6 Synthesis

The objective of this thesis is to analyze and explain the architecture, facies distribution and formation of coarse-grained overbank deposits, with special attention to organic-clastic lake fills, and organics in the distal delta plain of the Holocene Rhine-Meuse system. To meet this objective, three specific research goals were formulated, which were explored in the preceding research chapters (Chap. 2-5).

The first research goal was to describe the geometrical properties and depositional facies of organic-clastic lake fills, to understand how they form and to determine their influence on delta-plain architecture and facies composition of fluvial deposits. Organic-clastic lake fills typically are composed of clastic deposits that coarsen upward and overlie organic-lacustrine sediment (gyttja). They contain up to 30% of sandy deposits. Their thickness ranges between 2 and 5 m and surface areas are up to 25 km^2. It is proposed that organic-clastic lake fills preferentially occur at the base of aggradational sequences, i.e., in the lower part of Transgressive Systems Tracts (TST) and High-stand Systems Tracts (HST) according to sequence stratigraphic nomenclature. Owing to trapping of the coarsest sediment grains in lakes (predominantly silt and sand), channel deposits of fluvial systems downstream of lakes are relatively fine-grained and are accompanied by relatively thin natural-levee deposits. The large proportion of sand that is contained in organic-clastic lake fills, in addition, provides weak channel banks compared to clay and peat, which are usually found in distal delta plains. Channels that transverse organic-clastic lake fills, therefore, have a tendency to meander instead of being straight.

The second research goal was to investigate spatial sedimentary and botanical variability of the basal peat layer, including all organics at the base of the Holocene succession, of the distal delta plain and to assess the controlling factors. Organics were divided into organic lake sediment (gyttja) and in situ peat and further subdivided according to sedimentary characteristics and botanical content respectively. It was shown that this key enabled identification of five sedimentary- and botanical-organic facies in archived borehole descriptions. Database querying along with interpolation techniques resulted in a distribution map of basal-peat organic facies for the distal delta plain. The followed approach and applied methods appeared to be successful in obtaining consistent and geologically meaningful distribution patterns of organic facies. The dominant organic facies in the basal peat layer was reed peat (63% of the surface area), wood peat (8%) and gyttja (29%). Organic facies variability of the basal peat layer indicates the presence of three hydrological domains that characterized the onset of Holocene aggradation in the Rhine-Meuse delta: marine-dominated, fluvial-dominated and seepage-dominated. Because organic facies are an indicator for a number of environmental conditions (e.g., hydrological setting, nutrient availability and salinity), identification of these facies can be used, for example, to better interpret sedimentary facies of overlying sediment. Identification of organic-clastic lake fills, for example, will improve when the distribution of gyttja is known.

The third research goal was to analyze and explain spatial and temporal distribution patterns of coarse-grained overbank deposits and their lithofacies composition at delta scale. Coarse-grained overbank deposits occupy 7.1% of the fluvial delta-plain succession. Crevasse-splay deposits were found to be present throughout the fluvial part of the delta plain, although their proportion is partly controlled by lateral floodbasin width. Both narrow and wide floodbasins tend to have relatively low proportions of crevasse-splay deposits whereas intermediate floodbasin widths, 3.1-3.6 km in the central Rhine-Meuse delta, yield optimum crevasse-splay deposit proportions (CSDP). It was proposed that crevasse splays cannot optimally develop in narrower floodbasins and they cannot fully accommodate wider floodbasins. Organic-clastic lake fills form during periods with high rates of base-level rise, or when the floodplain is wide. The lithofacies composition of crevasse-splay deposits and organic-clastic lake fills is found to be controlled by channel planform, superelevation and, less dominantly, by the local composition of the substratum. Finally, we found that coarse-grained overbank deposits increase reservoir volume and connectedness ratio. At present, the quantitative significance of these deposits, however, is insufficiently incorporated in reservoir simulation models, which leads to underestimation of reservoir volume.

In this chapter, the results and conclusions that are presented in the preceding chapters are integrated. Thereby, the focus is on the architecture and lithofacies of distal delta-plain successions, because the ensemble of the individual research chapters provides additional value especially on these issues. Further, I discuss opportunities for practical application of the work presented in this thesis. Finally, I give suggestions for future research.

6.1 ARCHITECTURE AND LITHOFACIES COMPOSITION OF DISTAL DELTA-PLAIN SUCCESSIONS

6.1.1 Proportions of architectural elements: channel-belt deposits, overbank deposits and organics

In delta plains, there is a difference in the relative proportion of channel-belt deposits (CDP), overbank deposits (ODP) and organics (OP) in downstream direction. In the Rhine-Meuse delta, for example,,the channel-belt deposit proportion is high (> 0.6) in the proximal delta plain and low (< 0.3) in the distal delta plain (Gouw, 2008). The proportion of overbank deposits (ODP) is relatively constant throughout the delta plain. Organics, in contrast are nearly absent in the proximal delta plain, but have a proportion of at least 0.3 in the distal delta plain of the Rhine-Meuse delta (cross section E, Fig. 1.2) (Gouw, 2008). Organics in regions adjacent to the central delta plain, especially where delta-plain width is relatively large, may account for an even larger volumetric proportion. In the Angstel-Vecht area (Fig. 3.1), for example, the proportion of organics (OP) in the fluvial domain is 0.45 (Bos, 2010; Chap. 3; Appendix 2). In areas fringing the (distal) delta plain, organics may extensively accumulate during relatively long periods. When these areas eventually are annexed by the laterally expanding delta, organics are the dominant architectural element. Fluvial sedimentation along with compaction of organics, however, is expected to strongly reduce the OP soon thereafter in favour of the CDP and the ODP.

6.1.2 Channel planform

The ubiquity of organics *(1, italic* numbers in this chapter refer to corresponding numbers in Figure 6.1), has far-reaching consequences for the architecture of distal delta-plain successions. The high resistance to erosion of peat accumulations, especially when compacted, provides relatively erosion-resistant banks, which restrict lateral migration of rivers. Organics, thereby, enhance the development of rivers with straight channels *(2)* (Bos

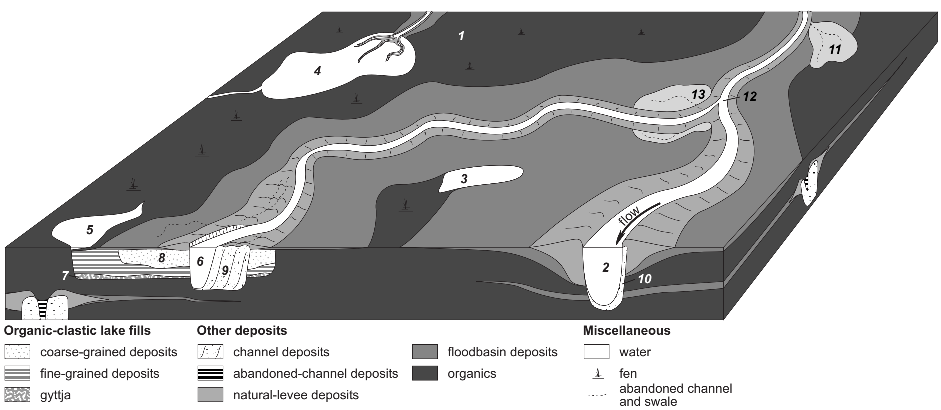

Figure 6.1. Block diagram of distal delta-plain architecture. Adapted from Miall (1985) and Makaske (2001). Italic numbers are referred to in the text.

et al., 2009; Chap. 2). Also stream power, which is predominantly controlled by discharge and channel gradient, affects channel planform. In delta plains, stream power usually decreases downstream in association with decreasing gradient and diversion of flow into multiple channels, which progressively limits lateral migration of river channels. Channel planform, thus, tends to change from meandering in upstream high-gradient reaches to straight in downstream low-gradient reaches. This is confirmed by field examples from the Rhine-Meuse delta (Törnqvist, 1993b; Berendsen and Stouthamer, 2001; Makaske et al., 2007; Gouw, 2008), the Lower Mississippi Valley, USA (Gouw and Autin, 2008), the Saskatchewan River in the Cumberland Marshes, Canada (Smith et al., 1989) and the Columbia River, Canada (Smith, 1983; Makaske et al., 2002). Field examples from megafans, supported by model experiments, also suggest that channel sand body width (w) and associated width/thickness ratio (w/t) tend to decrease in downstream direction in delta plains (Nichols, 1987; Hirst, 1991; Hickson et al., 2005; Gibling, 2006).

The presence of extensive peat accumulations *(1)* not only limits lateral migration of fluvial channels, but it also is accompanied by the presence of lakes *(3-5)*, which subsequently facilitate the formation of organic-clastic lake fills (Bos et al., 2009; Bos, 2010; Chap. 2, 3). The relatively large proportion of sand therein means that bank stability of river channels that are encased in these deposits is low compared to those of channels that are encased in organics. As a result, channels traversing organic-clastic lake fills have a tendency to migrate laterally. The presence of organic-clastic lake fills, therefore, promotes the development of meandering channels *(6)*, which is in contrast to the straight channel planform *(2)* that is dominant in organic-rich distal delta plains.

Changes in channel planform are likely accompanied by changes in overbank architecture. Weerts and Bierkens (1993), for example, showed that overbank-deposit thickness of straight rivers in the Rhine-Meuse delta, varied over much shorter distances perpendicular to the associated channel belt than overbank-deposit thickness of meandering rivers. Furthermore, Van Asselen (Van Asselen, in press) showed that natural-levee deposits in organic-rich delta plains, i.e., areas where straight-anastomosing river systems prevail, are thick compared to those in areas with small amounts of organics. The susceptibility of peat for compaction leads to the enhanced creation of accommodation space when clastic sediment covers the peat. This accommodation-space surplus generally is filled by more clastic deposits, which leads to relatively thick natural-levee deposits. However, natural-levee deposits of river channels in peat areas, such as those in the Angstel-Vecht area (Fig. 2.1), show a tendency to thin in downstream direction owing to increasing shortage of clastic sediment farter from the source (Bos et al., 2009; Chap. 2). In that area, peat lakes trapped large volumes of sediment, which means that little sediment was available for levee building along channel sections downstream of those lakes.

6.1.3 Lakes and lake fills

Distal delta plains accommodate lakes, bounded by emerged fluvial levees *(3)*, by peat banks in fen areas *(4, 5)*, or both. The size of these lakes varies considerably.

Accurate measurements of present lakes and recent lake fills in both the Rhine-Meuse delta (Bos, 2010; Chap. 3) and the Cumberland Marshes (Smith and Pérez-Arlucea, 1994; Pérez-Arlucea and Smith, 1999), indicate that most lakes do not exceed areas of 15 km^2. Furthermore, early Holocene lakes of up to 40 km^2, inferred from interpolated subsurface data, have been found in the basal peat layer (Chap. 4). The presence of lakes has consequences for delta-plain architecture and lithofacies. Under sediment-starved conditions gyttja will form in these lakes given a minimum water depth of ~0.5 m *(7)*, whereas organic-clastic lakes form when fluvial deposits are diverted to the lake *(4, 5)* (Bos et al., 2009; Bos, 2010; Chap. 2, 3). Both gyttja and organic-clastic lake fills have distinct geomechanical and hydrological properties that differ from non-lacustrine organics (i.e., peat) and non-lacustrine fluvial deposits. Erosion resistivity of gyttja, for example, which is composed of small particles, is expected to be less than the erosion resistivity of peat in situ. Plant fibres in the latter increase the tensile strength of the material and thereby increase the resistivity to erosion. Organic-clastic lake fills contain significant portions of sand *(8)* in contrast to other overbank deposits. The sandy organic-clastic lake fills enables fluvial channels to migrate laterally and thus to adopt a meandering channel planform *(6)*. The resulting channel-sediment body *(9)* is wide compared to those of straight-anastomosing channel-sediment bodies *(10)*.

The configuration of organic-clastic lake fills in the delta-plain succession may differ from those of crevasse-splay deposits. Crevasse-splay deposits always form at one side of the trunk channel *(11)*. When the crevasse-channel evolves into a new fluvial channel, thereby forming a bifurcation *(12)* and potentially an avulsion, the crevasse-splay deposits may be dissected by a fluvial channel *(13)*. As a consequence, dissected crevasse-splay deposits always occur in association with two fluvial channels: a trunk channel and a matured crevasse-splay channel. Organic-clastic lake fills, in contrast, may be dissected by the feeding channel that initially debouches into the lake and, after filling the lake, traverses and dissects the organic-clastic lake fill *(6)*. Organic-clastic lake fills that are dissected by a channel, therefore, may occur in association with a single fluvial channel.

6.2 PRACTICAL APPLICATIONS

6.2.1 Subsurface modelling

Ancient delta-plain successions are reservoirs for hydrocarbons and water. The architecture and facies composition of these deposits, to a large extent, determine the economic feasibility to exploit these reservoirs. Due to the generally difficult accessibility of reservoir rocks, predictions of reservoir characteristics, such as volume and connectedness ratios, usually are based on model simulations (Mackey and Bridge, 1995). These models often are tested with data and characteristics of extensively studied outcrop or modern analogues. Outcrops, though providing great sedimentary and architectural detail on parts of the depositional system (i.e., delta plain), often do not cover entire delta plains. Maybe even more important, accurate time control on the deposits is not available, obstructing investigation of relationships between architecture and processes. Modern deposits provide the opportunity to obtain these relationships, but knowledge of architectural characteristics in general is limited. The Rhine-Meuse delta, however, is a unique exception as it combines the presence of an architectural framework, extensive palaeogeographic reconstructions, and excellent time control, owing to the existence of large geological databases and a long scientific research history (Berendsen and Stouthamer, 2001; Weerts et al., 2005; Berendsen, 2007). Hence, knowledge and understanding of the composition and three-dimensional geometry of the Rhine-Meuse delta plain is of practical use as a calibration tool for reservoir models. Especially models of delta-plain successions that have abundant organic accumulations may be improved by comparison with distal delta-plain architecture and lithofacies of the Holocene Rhine-Meuse delta (Chap. 3, 4). Furthermore, these characteristics potentially are of value for the interpretation of any fluvial rock record that includes coarse-grained overbank deposits and/or organics. Early Pleistocene delta-plain deposits of the Rivers Rhine and Meuse (Waalre Formation), for example, contain organics and coarse-grained overbank deposits (Westerhoff et al., in press). This, along with strong similarities in tectonic setting, suggests that interpretation of ancient Rhine-delta deposits could benefit from knowledge of modern Rhine-delta deposits. In general, interpretation of organics and overbank deposits may be helpful in obtaining insight in the delta-plain configuration and the controlling factors on delta formation such as the rate of base-level rise. For example, organic facies within ancient deposits (i.e., Waalre Formation, may provide insight in the vicinity of fluvial systems or the coast. Identification of bay-head delta deposits or organic-clastic lake fills also gives rough indications of the distance to the coastline as they occur in the estuary and the distal delta plain respectively.

The results presented in this thesis are ready for incorporation in the 3D subsurface model of the Netherlands, which is being built at TNO Built Environment and Geosciences, Geological Survey of The Netherlands. The development of this model, which serves both commercial and scientific purposes, replaced the construction of printed geological maps of the Dutch subsurface, which was abandoned in the 90's. The model applies architectural principles to constrain interpolations of subsurface data. The general concepts presented in this thesis, for example the lithofacies composition of organic-clastic lake fills (chapter 3) and the classification of organic facies in the basal peat layer (chapter 4) are of value for the interpretation of deposits in other regions in the Netherlands. Further, the results present here (Chap. 2-4) could be used as direct input for the model. The spatial distribution of fluvial deposits in the Angstel-Vecht area (Fig. 2.5) and the Schoonhoven area (Fig. 3.3), for example, could be used as boundary conditions for the model. Furthermore, the extent of organic facies in the basal peat layer could be incorporated. The 3D subsurface model is of great practical significance as it serves geotechnical (e.g., (rail-)road construction) and hydrological (e.g., groundwater-flow prediction) purposes.

6.2.2 Public information

Part of the results, notably the lacustrine origin of part of the fluvial deposits along with insights from the palaeogeographic reconstruction of the Angstel-Vecht area, has proved valuable for geoheritage conservation and tourism-related purposes. The Aetsveldse organic-clastic lake fill (Fig. 3.3), for example, currently has the status of geoheritage site in the province of Noord-Holland. Although this initiative has not been based on this thesis, the understanding of this site has benefited from insights presented in this thesis. Further, the general understanding of the Angstel-Vecht area is currently used to inform the public in a web portal serving inhabitants and visitors of the area (www.aardkundigewaarden.nl/aardkundigemonumenten/home.php).

6.3 DIRECTIONS FOR FUTURE RESEARCH

6.3.1 Origin of delta-plain lakes

This study mainly focussed on lake fills whereby the existence of lakes in delta plains primarily has been taken for granted. Bos et al. (2009; Chap. 2) briefly discuss the formation of lakes in the Angstel-Vecht area, but the mechanism behind the formation of delta-plain lakes remains not very well known. Understanding this mechanism will be useful to better determine the position of lakes and lake fills in delta plains. A fundamental difference exists between lakes that form as a consequence of base-level rise and those that are the result of seasonally fluctuating ground-water levels. During periods of base-level rise, lakes form as the result of relative groundwater rise. Lakes could come into existence in substratum depressions or in basins that form as a result of differential aggradation rates. Examples of the latter are lakes that form in floodbasins between natural levees. In the distal Rhine-Meuse delta plain, these environments probably existed abundantly between ~7500 and 6000 cal yr BP (Van der Woude, 1984; Chap. 4). Deficit of clastic sediment relative to the created accommodation space resulted in the existence of both peat-forming wetlands and—large—lakes, dissected by a number of fluvial distributary channels. When base-level rise is small or absent, small lakes could temporarily exist when groundwater levels fluctuate (seasonally). During winter, groundwater levels are high, which promote the existence of small lakes. Then, wave erosion could become effective leading to enlargement of these lakes. There is a positive relation between lake size, affecting the maximum downwind length (fetch) and the maximum energy of individual waves and thus the potential erosive power exerted by these waves. This means that, given the presence of erodible lake banks, large lakes will become larger until water depth becomes the factor limiting wave size. Abovementioned processes, nevertheless, still do not explain the arrangement of lakes along the ice-pushed ridge in the Angstel-Vecht area (Fig. 2.5). Bos et al. (2009; Chap. 2) tentatively proposed that the seepage-associated peat type partly controlled lake formation. In the Angstel-Vecht area, the zone where seepage water exfiltrated shifted to higher positions over time in response to base-level rise. This caused initially oligotrophic conditions in the zone where the lakes formed, to become mesotrophic. These changing environmental conditions possibly have limited peat growth for a while, thereby facilitating the initiation of lakes, which afterwards became larger as a result of wave erosion. Another explanation could be the temperature difference in winter between cold surface water and relatively warm seepage water (personal communication Gerard Kruse). Lakes in the seepage zone, consequently, were less prone to become ice-covered in winter, the storm season, which means that those lakes were much more vulnerable to erosion than lakes outside the seepage zone. I regard the latter explanation promising and suggest that future research on this subject focuses on comparable settings, both in similar and in different climatic regions, and on model studies aiming to reconstruct seepage-water temperature during the formation of the lakes.

6.3.2 Spatial-temporal distribution patterns of sedimentary- and botanical-organic facies

This thesis includes the first steps towards understanding the spatial and temporal distribution of organic lithofacies and botanical facies within a delta plain (Chap. 4). The setup of a classification key in combination with satisfactory results obtained from archived borehole data, provides a strong starting point to further extend our knowledge on delta-plain organics. Mapping the spatial and temporal distribution of organic facies within the entire Rhine-Meuse delta plain becomes feasible. There is, however, an essential difference between the basal peat layer that has been analysed in this thesis and other organics in the delta-plain succession. The first is not prone to subsidence because it directly covers uncompactable substratum, in contrast to overlying organic beds that subside due to compaction of underlying (peat) accumulations. This strongly limits age determination beyond the (radiocarbon) sample scale, which subsequently obstructs temporal analyses of organic-facies variability at delta-scale. Recently, however, it was quantitatively shown that peat compaction in delta plains depends on organic-matter content, stress imposed on the material and peat type (Van Asselen et al., 2010). When these factors are known for a single borehole, it would become possible to estimate the time-depth relation for that location. Application thereof on a large set of boreholes would enable construction of time surfaces—analogues to time lines in cross sections—throughout the delta. These time surfaces enable the differentiation of palaeoenvironments in areas prone to subsidence, such as floodbasins, which potentially are of great value for understanding delta evolution. Coupling fluvial processes or events (e.g., avulsion) with floodbasin characteristics, such as vegetation type and water depth, on a Holocene time scale would then become feasible. It also enables reconstruction of surface gradients perpendicular to fluvial channels, which is of importance for understanding avulsion and the formation of crevasse splays.

6.3.3 Sediment transport in palaeochannels

Sediment transport is a key issue in fluvial sedimentology. Many studies have measured and calculated sediment transport in active rivers on relatively small time scales, i.e.

10^{-2} to 10^{1} years, which is why there is good knowledge of relations between sediment transport, discharge and channel morphology (see Kleinhans, 2005 and references therein). Sediment transport in palaeochannels, however, received much less attention because the necessary input parameters for sediment transport models, such as channel dimensions and channel gradient, often cannot directly be measured. Bos (2010; Chap. 3), however, showed that channel dimensions of straight channels (assuming that channel-belt width approximates palaeochannel width) and channel gradient could be estimated on the basis of cross sections. When combined with the volume and ages of an associated delta-sediment body, it was shown that the formative duration of that particular sediment body could be calculated. Given the wealth of data, it seems feasible to apply such an approach also to older deposits in the Rhine-Meuse delta, thereby enabling to identify formative duration of crevasse-splay deposits under different conditions, e.g., during periods of higher rates of relative sea-level rise. These investigations will yield better insight in, for example, the formative duration of crevasse-splay deposits and the effect of the size of feeding channel of the crevasse splay on the size of the resulting crevasse-splay deposits. Subsequently, these characteristics could be incorporated in architectural simulation models, which will result in more accurate predictions of the arrangement and size of crevasse-splay deposits in delta-plain successions.

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Appendix 1

Duration and characteristics of fluvial regime – discharge and sediment transport – related to mouth-bar deposition in organic-clastic lake fills

There is a strong relation between the characteristics of fluvial channels – sediment supply and period of activity – and the sediment bodies they produce upon debouching into, for example, a lake. In order to calculate minimum timescales for delta sedimentation from alluvial architecture – e.g., width/thickness ratio, deposited sediment volume – a number of computations have to be made, primarily based on fluid dynamics. The relevant equations (numbers refer to those as used by Kleinhans 2005) are discussed below and are embedded in a more fundamental context in a publication by Kleinhans (2005). A brief overview of the parameters used in the model (App. 3) is given in Table A1.1.

Flow computations

Water flow through a channel can be characterized by a set of parameters such as discharge, flow velocity, hydraulic roughness, Reynolds and Froude numbers and bed shear stress.

Discharge (Q) is the product of the channel width (W), channel depth (h) and the flow velocity (u) averaged over the width and depth of the channel:

$$Q = hWu \qquad (1)$$

The hydraulic radius instead of water depth is used to account for the effect of banks on the water depth and is expressed as

$$R_h = \frac{Wh}{W + 2h} \qquad (2)$$

The flow velocity, averaged over the depth and width, is given by the Darcy-Weisbach equation:

$$u = \sqrt{\frac{8ghS}{f}} \qquad (3)$$

where g = the acceleration due to gravity, h is depth (hydraulic radius) and S = water surface slope or channel bed surface, given that the flow is steady and uniform. f = the friction factor, which is a measure for the hydraulic roughness.

Hydraulic roughness is very relevant for flow calculations as it determines the water depth. However, it is hard to determine as it depends on complexly interacting variables such as flow turbulence and roughness elements on the channel bed (e.g., grains and bed forms). A widely used predictor for the friction factor is the White-Colebrook friction law:

$$\sqrt{\frac{8}{f}} = 5.74\ log_{10}(12.2\frac{h}{k_s}) = 5.74\ log_{10}(\frac{h}{k_s}) + 6.24 \quad (8)$$

where k_s = Nikuradse roughness length, which depends on characteristics of the grain-size distribution (e.g., D_{50}) or on bed-form height.

The Reynolds and Froude numbers describe two elementary characteristics of flow. The Reynolds number differentiates between laminar and turbulent flow and relates to the flow velocity (u), water depth (h) and the fluid viscosity (v) as

$$Re_f = \frac{uh}{v} \quad (14)$$

For $Re_f > 500$, flow is turbulent, which is valid for all rivers.

The Froude number indicates whether flow is subcritical ($Fr < 1$) or supercritical ($Fr > 1$).

$$Fr = \frac{u_c}{\sqrt{gh}} \quad (15)$$

where u_c = critical flow velocity.

The bed shear stress affects sediment transport and is defined as

$$\tau_c = \rho u^{*2} = \rho gh \sin S \quad (16)$$

in which ρ = fluid density (1000 kg/m^3 for fresh water), u^* = flow velocity and S = channel gradient. If u is written as the Darcy-Weisbach equation (eq. 3), the bed shear stress relates to the hydraulic roughness as follows

$$\tau_c = \frac{1}{8}\ \rho f_c u_c^2 \quad (17)$$

The bed shear stress may be divided in two components: bed-form related and grain-related. Of these two, only grain-related bed shear stress is relevant for sediment transport, which can be calculated from equation (17) where the friction factor (f) is defined by the White-Colebrook friction law (eq. 8). The hydraulic roughness length is commonly chosen as $k_s = 2.5D_{50}$ (Kleinhans 2005).

Table A1.1. Overview of the input and calculated parameters in the discharge and sediment transport model (results are presented in Appendix 3).

symbol	*units*	*description*
Parameters and variables		
g	m/s^2	Gravity
W	m	Width of channel
h	m	Depth of channel
S	m/m	Slope upstream channel
k_s	m	Nikuradse roughness length
	kg/m^3	Fluid density
	Centigrade	Fluid temperature
	-	Dynamic viscosity
	-	Kinematic viscosity
D_{50}	m	Median grain size
D_{90}	m	90th percentile grain size
	kg/m^3	Sediment density
λ		Porosity
Discharge calculations		
R_h	m	Hydraulic radius (eq. 2 Kleinhans)
	-	Width/depth ratio
f	-	Friction factor (eq. 8 Kleinhans)
u	m/s	Velocity (eq. 3 Kleinhans)
Fr	-	Froude number (eq. 15 Kleinhans)
Re	-	Reynolds number (eq. 14 Kleinhans)
Q_w	m^3/s	Discharge (eq. 1 Kleinhans)
Bed-load transport		
f'	-	Grain friction factor (eq. 8 Kleinhans)
τ'	Pa	Grain shear stress (eq. 17 Kleinhans)
θ'	-	Shields parameter (grain related) (eq. 22 Kleinhans)
	-	Shields criterion for incipient motion
Φ_b	-	Non-dimensional transport rate (eq. 30 Kleinhans)
R	-	Relative submerged density (in eq. 20 Kleinhans)
	m^2/s	Specific volumetric transport rate
q_b	m^3/yr	Volumetric transport rate
	-	Water/sediment ratio
	-	Total/bedload ratio
Total load transport (suspension dominated)		
τ	Pa	Total shear stress (eq. 16 Kleinhans)
θ	-	Shields parameter (total) (eq. 22 Kleinhans)
Φ_s	-	Non-dimensional total transport rate (eq. 37 Kleinhans)
	m^2/s	Specific volumetric transport rate (eq. 29 Kleinhans)
q_s	m^3/yr	Volumetric transport rate
	-	Water/sediment ratio
Erosion/sedimentation		
	m^2	Deposited sediment delta surface
	m^3	Deposited sediment delta volume (sand)
	m^3	Deposited channel-belt volume (sand)
Time-scale of formation assuming bed load		
	yr	Delta volume/bed-load flux
	yr	CB volume/bed-load flux
Time-scale of formation assuming suspended load		
	yr	Delta volume/total sediment flux (eq. 54 Kleinhans)
	yr	CB volume/sediment flux

Sediment mobility

Preferentially, non-dimensional variables and parameters are used to describe bed-load sediment transport. These variables either describe the flow or the sediment size. The grain size is described by

$$D^{*}=D_{50}\sqrt[3]{\frac{Rg}{\upsilon^{2}}} \qquad (20)$$

where $R = (\rho_s - \rho)/\rho$ is relative submerged density, ρ = the liquid density and ρs is the sediment density (2650 kg/m^3 for quartz sand).

Commonly applied flow parameters are based on current velocity, shear stress (τ) and grain shear stress (τ', only skin friction). The shear stress is represented by the non-dimensional "Shields parameter":

$$\theta= \frac{\tau}{(\rho_s-\rho)gD_{50}} \qquad (22)$$

which also is valid for grain shear stress (θ').

Sediment transport

Non-dimensional sediment transport is defined as

$$\Phi= \frac{q_{b\ or\ s}}{(Rg)^{0.5}D_{50}^{\ 1.5}} \qquad (29)$$

wherein qb or s = bed load or suspended-load sediment transport rate (m^2/s, cubic m per m width per second) excluding pore space.

A predictor for bed-load sediment transport near the beginning of motion was derived by Meyer-Peter and Mueller (1948)

$$\Phi_b= 8(\theta'-\theta_{cr})^{1.5} \qquad (30)$$

A well-known predictor for suspended-load sediment transport has been published by Engelund and Hansen (1967) and is given below

$$\Phi_s= \frac{0.4}{f}\ \theta^{2.5} \qquad (37)$$

wherein Φs = non-dimensional suspended-load sediment transport and f = Darcy-Weisbach coefficient related to the total roughness.

Timescale of channel and delta formation

The minimum timescale of delta/channel formation is defined by

$$T_s= \frac{V_s}{(1-\lambda)Q_s} \qquad (54)$$

wherein V_s = volume of sediment (including pores), which is either eroded from a channel or deposited in a delta and $Q_s = (q_b + q_s)W_s$ with W_s = width that part of the channel floor that transports sediment (Kleinhans 2005).

Input parameters and variables

W = 180 m. This is the minimum width of the Angstel channel belt and consequently the maximum width of the palaeochannel, which is measured on a recent geological-geomorphological map of the area (Fig. 3.5B, Chap. 3).

h = 8 m. Three corings penetrated the channel belt of the Angstel, recovering channel-belt deposits that were 6.0, 9.1 and 10.3 m thick.

S = 3.6 cm/km. The slope of the upstream channel was derived from the Gradient of the Top of Sand (GTS) of the channel belt deposits of the Aa and Angstel (Fig. A1.1). Berendsen (1982) showed that GTS lines can be used to approximate the gradient of palaeochannels. We selected all corings that recorded the top of Aa or Angstel channel-belt deposits (n=103). Within sections of 2 km along the palaeochannel – as measured along abandoned channel deposits – the point with highest detected channel-belt sediment was selected. These points (n=10) were plotted as a function of the distance from Utrecht along the paleo channel. The gradient of the trendline was used as gradient of the palaeochannel.

Although R2 of the trend line is relatively low (0.27), a gradient of 3.6 cm/km is not unusual for channels in distal parts of delta plains. Values between 3.0 and 10 cm/km (Makaske et al. 2007; Törnqvist et al. 1993) were reported for palaeochannels in the Rhine-Meuse delta.

λ ~ 34% for Holocene channel-belt sediment in the Rhine-Meuse delta (Weerts 1996).

Water temperature = 10.4 °C. This is the 1909-1910 average (based on daily measurements), which are the earliest documented measurements (Rijkswaterstaat 2009) of water temperature in the Rhine.

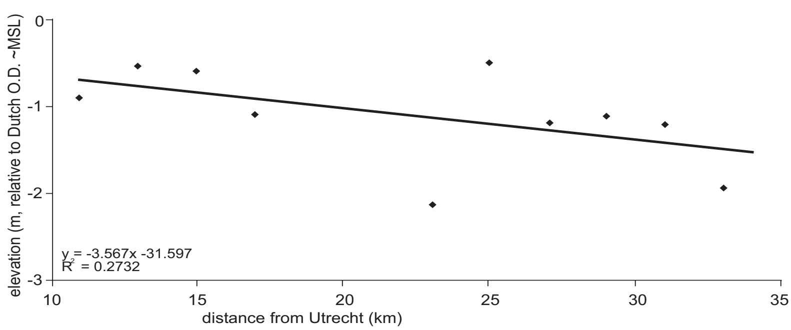


Figure A1.1. GTS-line of the Aa-Angstel.

Appendix 2

Volume calculations of facies in the Angstel-Vecht area

Table A2.1. Volume calculations of organic-clastic lake fill facies in the Angstel-Vecht area.

	av. thick-ness (m)	*thickness error (m)*	*area (m^2)*	*volume (m^3)*	*vol (km^3)*	*% of cl. dep*
Fluvial deposits						
channels	8,3	0,40	12032954	99873517	0,100	31
levees	2,7	0,15	3322842	8971675	0,009	3
floodbasin	2,4	0,15	21793083	52303400	0,052	16
thin floodbasin	0,2	0,05	31196375	6239275	0,006	2
crevasse	3,5	0,50	2100746	7352611	0,007	2
lakes (sandy)	4,0	0,20	21229667	84918668	0,085	26
lakes (clayey)	3,8	0,20	16695380	63442445	0,063	20
			108371048	**323101591**	**0,32**	
Sand in organic-clastic lake fills						
av. thickness of sand (sandy oclf)	1,7			36090434		
av. thickness of sand (clayey oclf)	0,5			8347690		
				44438124		
Organics (in fluvial domain)						
underlying levees dep	1,6		3322842	5316548	0,005	
underlying floodb dep	3,1		21793083	67558558	0,068	
underlying thin floodb dep	5,2		37148880	193174173	0,193	
				266049279	**0,266**	
Proportions						
CDP* (channel deposits proportion)	0,17					
ODP* (overbank deposits proportion)	0,38					
OP* (organic proportion)	0,45					
oclf/fl dep.	0,46					
oclf/overbank dep	0,66					
oclf-sand/oclf	0,30					
oclf-sand/overbank	0,20					
oclf-sand/total sand	0,31					

Appendix 3

Sand transport and minimum deposit-formation time scale

Table A3.1. Overview of input parameters, constants and variables for discharge and sediment-transport calculations in the Aetsveldse organic-clastic lake fill and its feeding channel, the Angstel channel belt (based on Kleinhans, 2005).

				preferred	sc 1	sc 2	sc 3	sc 4
		ch width relative to width channel belt dep.		0,75	0,5	0,5	0,5	0,5
		ch depth relative to thickness cb dep.		0,75	0,5	0,63	0,75	0,88
		ch gradient relative to cb gradient						
Coordinates	1	Longitude	deg	05°01'E	05°01'E	05°01'E	05°01'E	05°01'E
	2	Latitude	deg	52°17'N	52°17'N	52°17'N	52°17'N	52°17'N
Discharge & Sediment Transport Input Values	3	Gravity, g	m/s2	9,81	9,81	9,81	9,81	9,81
	4	Channel width, W	m	135	90	90	90	90
	5	Channel depth, h	m	6	4	5	6	7
	6	Slope of upstream channel	m/m	3,60E-05	3,60E-05	3,60E-05	3,60E-05	3,60E-05
	7	Roughness length, k_s	m	1,50E-01	1,50E-01	1,50E-01	1,50E-01	1,50E-01
	8	Fluid density	kg/m^3	1000	1000	1000	1000	1000
	9	Fluid temperature	Centigrade	10	10	10	10	10
	10	Dynamic Viscosity		1,3E-03	1,3E-03	1,3E-03	1,3E-03	1,3E-03
	11	Kinematic Viscosity		1,3E-06	1,3E-06	1,3E-06	1,3E-06	1,3E-06
	12	Median Grain Size (D_{50})	m	3,00E-04	3,00E-04	3,00E-04	3,00E-04	3,00E-04
	13	90th percentile grain size D_{90} (+ skin friction)	m	1,00E-03	1,00E-03	1,00E-03	1,00E-03	1,00E-03
	14	Sediment density	kg/m^3	2650	2650	2650	2650	2650
	15	Porosity	-	0,34	0,34	0,34	0,34	0,34
Discharge calculations	16	Hydraulic radius, Rh	m	6	4	5	5	6
	17	width/depth ratio (line 5/4)	-	23	23	18	15	13
	18	Friction factor, f	-	0,03	0,02	0,02	0,03	0,03
	19	Friction factor, C	m^0.5/s	55	63	58	55	52
	20	Velocity, u	m/s	0,8	0,7	0,7	0,8	0,8
	21	Froude Number, Fr	-	0,10	0,12	0,11	0,10	0,09
	22	Reynolds Number, Re	-	3,3E+06	2,0E+06	2,5E+06	3,0E+06	3,5E+06
	23	Discharge, Q_w	m^3/s	6,3E+02	2,6E+02	3,3E+02	4,1E+02	4,8E+02
	24	Discharge, Q_w	m^3/day	5,4E+07	2,3E+07	2,9E+07	3,5E+07	4,2E+07
Bedload transport	25	Grain friction factor, f	-	0,01	0,01	0,01	0,01	0,01
	26	Grain Shear Stress	Pa	1	1	1	1	1
	27	Shields Parameter (grain-related)	-	0,16	0,15	0,15	0,15	0,15
	28	Shields Criterion for incipient motion	-	0,03	0,03	0,03	0,03	0,03
	29	Nondimensional transport rate	-	0,38	0,34	0,35	0,35	0,35
	30	Relative submerged density, R	-	1,65	1,65	1,65	1,65	1,65
	31	Specific volumetric transport rate	m^2/s	1,2E-05	1,1E-05	1,1E-05	1,1E-05	1,1E-05
	32	Bedload transport	m^3/day	140	83	85	86	86
	33	Bedload transport	m^3/yr	5,1E+04	3,0E+04	3,1E+04	3,1E+04	3,1E+04
	34	water/sediment ratio	-	3,9E+05	2,7E+05	3,4E+05	4,1E+05	4,8E+05
	35	total/bedload ratio	-	5,1	2,7	4,0	5,5	7,2
Total load transport (suspension dominated)	36	Total Shear stress	Pa	2	1	2	2	2
	37	Shields Parameter (total)	-	0,4	0,3	0,4	0,4	0,5
	38	Nondimensional total transport rate	-	1,9	0,9	1,4	1,9	2,5
	39	Specific volumetric transport rate	m^2/s	6,2E-05	2,9E-05	4,4E-05	6,1E-05	8,0E-05
	40	Total load transport	m^3/day	718	227	340	472	622
	41	Total load transport	m^3/yr	2,6E+05	8,3E+04	1,2E+05	1,7E+05	2,3E+05
	42	water/sediment ratio	-	7,6E+04	9,9E+04	8,5E+04	7,5E+04	6,7E+04
Erosion/sedimentation	43	Deposited sediment delta surface	m^2	2,4E+07	2,4E+07	2,4E+07	2,4E+07	2,4E+07
	44	Deposited sediment delta volume (sand)	m^3	2,6E+07	2,6E+07	2,6E+07	2,6E+07	2,6E+07
	45	Deposited channel-belt volume (sand)	m^3	4,8E+06	3,2E+06	4,0E+06	4,8E+06	5,6E+06
Formation assuming bedload	46	Delta volume / bedload flux	years	5,0E+02	8,5E+02	8,3E+02	8,2E+02	8,2E+02
	47	CB volume / bedload flux	years	9,4E+01	1,1E+02	1,3E+02	1,5E+02	1,8E+02
Formation assuming suspended load	48	Formative duration	years	9,9E+01	3,1E+02	2,1E+02	1,5E+02	1,1E+02
	49	CB volume / total sediment flux	years	1,8E+01	3,9E+01	3,2E+01	2,8E+01	2,5E+01

sc 5	sc 6	sc 7	sc 8	sc 9	sc 10	sc 11	sc 12	sc 13	sc 14	sc 15	sc 16
0,5	0,75	0,75	0,75	0,75	0,75	0,75					
	0,5	0,63	0,75	0,75	0,88		0,5	0,63	0,75	0,88	
			0,76	1,26							
05°01'E	05°01'E	05°01'E	05°01'E	05°01'E	05°01'E	05°01'E	05°01'E	05°01'E	05°01'E	05°01'E	05°01'E
52°17'N	52°17'N	52°17'N	52°17'N	52°17'N	52°17'N	52°17'N	52°17'N	52°17'N	52°17'N	52°17'N	52°17'N
9,81	9,81	9,81	9,81	9,81	9,81	9,81	9,81	9,81	9,81	9,81	9,81
90	135	135	135	135	135	135	180	180	180	180	180
8	4	5	6	6	7	8	4	5	6	7	8
3,60E-05	3,60E-05	3,60E-05	2,70E-05	4,50E-05	3,60E-05	3,60E-05	3,60E-05	3,60E-05	3,60E-05	3,60E-05	3,60E-05
1,50E-01	1,50E-01	1,50E-01	1,50E-01	1,50E-01	1,50E-01	1,50E-01	1,50E-01	1,50E-01	1,50E-01	1,50E-01	1,50E-01
1000	1000	1000	1000	1000	1000	1000	1000	1000	1000	1000	1000
10	10	10	11	12	10	10	10	10	10	10	11
1,3E-03	1,3E-03	1,3E-03	1,3E-03	1,2E-03	1,3E-03	1,3E-03	1,3E-03	1,3E-03	1,3E-03	1,3E-03	1,3E-03
1,3E-06	1,3E-06	1,3E-06	1,3E-06	1,2E-06	1,3E-06	1,3E-06	1,3E-06	1,3E-06	1,3E-06	1,3E-06	1,3E-06
3,00E-04	3,00E-04	3,00E-04	3,00E-04	3,00E-04	3,00E-04	3,00E-04	3,00E-04	3,00E-04	3,00E-04	3,00E-04	3,00E-04
1,00E-03	1,00E-03	1,00E-03	1,00E-03	1,00E-03	1,00E-03	1,00E-03	1,00E-03	1,00E-03	1,00E-03	1,00E-03	1,00E-03
2650	2650	2650	2650	2650	2650	2650	2650	2650	2650	2650	2650
0,34	0,34	0,34	0,34	0,34	0,34	0,34	0,34	0,34	0,34	0,34	0,34
7	4	5	6	6	6	7	4	5	6	6	7
11	34	27	23	23	19	17	45	36	30	26	23
0,03	0,02	0,02	0,03	0,03	0,03	0,03	0,02	0,02	0,03	0,03	0,03
49	63	59	55	55	52	50	63	59	55	52	50
0,8	0,7	0,8	0,7	0,9	0,8	0,8	0,7	0,8	0,8	0,8	0,8
0,09	0,12	0,11	0,09	0,11	0,10	0,09	0,12	0,11	0,10	0,10	0,09
4,0E+06	2,1E+06	2,7E+06	2,9E+06	3,9E+06	3,8E+06	4,3E+06	2,2E+06	2,8E+06	3,4E+06	4,0E+06	4,7E+06
5,6E+02	4,0E+02	5,1E+02	5,4E+02	7,0E+02	7,4E+02	8,6E+02	5,4E+02	6,9E+02	8,5E+02	1,0E+03	1,2E+03
4,8E+07	3,4E+07	4,4E+07	4,7E+07	6,1E+07	6,4E+07	7,4E+07	4,6E+07	6,0E+07	7,3E+07	8,7E+07	1,0E+08
0,01	0,01	0,01	0,01	0,01	0,01	0,01	0,01	0,01	0,01	0,01	0,01
1	1	1	1	1	1	1	1	1	1	1	1
0,15	0,16	0,16	0,12	0,20	0,16	0,16	0,16	0,16	0,17	0,17	0,17
0,03	0,03	0,03	0,03	0,03	0,03	0,03	0,03	0,03	0,03	0,03	0,03
0,35	0,36	0,37	0,22	0,57	0,38	0,39	0,37	0,39	0,40	0,40	0,41
1,65	1,65	1,65	1,65	1,65	1,65	1,65	1,65	1,65	1,65	1,65	1,65
1,1E-05	1,1E-05	1,2E-05	6,9E-06	1,8E-05	1,2E-05	1,2E-05	1,2E-05	1,2E-05	1,3E-05	1,3E-05	1,3E-05
86	133	137	81	210	142	143	182	190	196	199	202
3,1E+04	4,8E+04	5,0E+04	3,0E+04	7,7E+04	5,2E+04	5,2E+04	6,7E+04	6,9E+04	7,1E+04	7,3E+04	7,4E+04
5,6E+05	2,6E+05	3,2E+05	5,8E+05	2,9E+05	4,5E+05	5,2E+05	2,5E+05	3,1E+05	3,7E+05	4,4E+05	5,0E+05
9,2	2,6	3,8	4,3	6,0	6,7	8,4	2,5	3,6	4,9	6,4	8,0
3	1	2	2	3	2	3	1	2	2	2	3
0,6	0,3	0,4	0,3	0,5	0,5	0,6	0,3	0,4	0,4	0,5	0,6
3,2	0,9	1,4	0,9	3,4	2,6	3,3	0,9	1,4	2,0	2,6	3,3
1,0E-04	2,9E-05	4,4E-05	3,0E-05	1,1E-04	8,1E-05	1,0E-04	3,0E-05	4,5E-05	6,2E-05	8,2E-05	1,0E-04
788	344	516	350	1254	946	1201	461	692	963	1272	1615
2,9E+05	1,3E+05	1,9E+05	1,3E+05	4,6E+05	3,5E+05	4,4E+05	1,7E+05	2,5E+05	3,5E+05	4,6E+05	5,9E+05
6,1E+04	1,0E+05	8,6E+04	1,3E+05	4,8E+04	6,8E+04	6,2E+04	1,0E+05	8,6E+04	7,6E+04	6,9E+04	6,3E+04
2,4E+07	2,4E+07	2,4E+07	2,4E+07	2,4E+07	2,4E+07	2,4E+07	2,4E+07	2,4E+07	2,4E+07	2,4E+07	2,4E+07
2,6E+07	2,6E+07	2,6E+07	2,6E+07	2,6E+07	2,6E+07	2,6E+07	2,6E+07	2,6E+07	2,6E+07	2,6E+07	2,6E+07
6,4E+06	3,2E+06	4,0E+06	4,8E+06	4,8E+06	5,6E+06	6,4E+06	3,2E+06	4,0E+06	4,8E+06	5,6E+06	6,4E+06
8,3E+02	5,3E+02	5,2E+02	8,8E+02	3,4E+02	5,0E+02	5,0E+02	3,9E+02	3,7E+02	3,6E+02	3,6E+02	3,5E+02
2,1E+02	6,7E+01	8,0E+01	1,6E+02	6,3E+01	1,1E+02	1,2E+02	4,8E+01	5,8E+01	6,8E+01	7,7E+01	8,7E+01
9,0E+01	2,1E+02	1,4E+02	2,0E+02	5,7E+01	7,5E+01	5,9E+01	1,5E+02	1,0E+02	7,4E+01	5,6E+01	4,4E+01
2,2E+01	2,6E+01	2,1E+01	3,8E+01	1,1E+01	1,6E+01	1,5E+01	1,9E+01	1,6E+01	1,4E+01	1,2E+01	1,1E+01

Appendix 4

LOI values

Table A4.1. LOI-values for individual samples

Core	*depth (cm)*	*LOI (%)*	*facies*
B25G1055	235	40	1c
B25G1055	79	42	1c
B25G1055	100	59	1c
B25G1055	105	64	1c
B25G1055	110	83	1c
B25G1057	535	38	1c
B25H0739	48	47	1c
Hedel.2	3,245	40	1c
HoonkomLeebrug.1	1,625	60	1c
HoonkomLeebrug.2	2,575	60	1c
HoonkomLeebrug.4	5,175	60	1c
B38D0432	480	30	1p
B38D0432	455	31	1p
B38D0432	450	32	1p
B38D0432	300	34	1p
B38D0432	145	37	1p
B38D0432	470	37	1p
B38D0432	385	38	1p
B38D0432	445	38	1p
B38D0432	370	38	1p
B38D0432	435	39	1p
B38D0432	390	39	1p
B38D0432	425	39	1p
B38D0432	395	40	1p
B38D0432	475	40	1p
B38D0432	430	42	1p
B38D0432	330	44	1p
B38D0432	320	44	1p
B38D0432	380	45	1p
B38D0432	375	46	1p
B38D0432	325	47	1p
B38D0432	440	49	1p
B38D0432	355	49	1p
B38D0432	365	49	1p
B38D0432	305	51	1p
B38D0432	140	52	1p
B38D0432	315	53	1p
B38D0432	135	56	1p
B38D0432	335	58	1p
B38D0432	350	60	1p
B38D0432	310	62	1p
B38D0432	420	69	1p
B38D0432	105	70	1p
B38D0432	360	71	1p
B38D0432	100	71	1p
B38D0432	130	75	1p
B38D0432	125	76	1p
B38D0432	110	76	1p
B38D0432	115	77	1p

Core	*depth (cm)*	*LOI (%)*	*facies*
B38D0432	415	79	1p
B38D0432	120	79	1p
B38E0306	540	34	1p
B38E0306	220	34	1p
B38E0306	225	35	1p
B38E0306	440	36	1p
B38E0306	435	37	1p
B38E0306	225	40	1p
B38E0306	445	42	1p
B38E0306	439	42	1p
B38E0306	180	42	1p
B38E0306	535	43	1p
B38E0306	505	43	1p
B38E0306	215	45	1p
B38E0306	525	47	1p
B38E0306	230	47	1p
B38E0306	480	48	1p
B38E0306	475	49	1p
B38E0306	210	52	1p
B38E0306	465	52	1p
B38E0306	235	53	1p
B38E0306	485	54	1p
B38E0306	530	55	1p
B38E0306	185	56	1p
B38E0306	200	57	1p
B38E0306	205	57	1p
B38E0306	460	57	1p
B38E0306	455	58	1p
B38E0306	534	59	1p
B38E0306	450	59	1p
B38E0306	510	59	1p
B38E0306	470	60	1p
B38E0306	245	61	1p
B38E0306	245	62	1p
B38E0306	240	62	1p
B38E0306	195	65	1p
B38E0306	490	67	1p
B38E0306	190	68	1p
B38E0306	493	70	1p
B38E0306	515	70	1p
B38E0306	495	71	1p
B38E0306	192	72	1p
B38E0306	520	76	1p
B38E0306	67	78	1p
Bazeldijk.2	0,955	60	1p
DeTreeft.2	2,195	40	1p
DeTreeft.3	2,615	40	1p
DeTreeft.4	2,925	40	1p
LangeAvontuur.3	3,055	70	1p

Table A4.1. continued

Core	depth (cm)	LOI (%)	facies
LangeAvontuur.4	5,15	70	1p
Leerdam.1	2,965	60	1p
Leerdam.2	2,725	60	1p
Leerdam.3	2,585	60	1p
Noordeloos.1	4,695	60	1p
Noordeloos.2	3,735	60	1p
Noordeloos.3	2,815	60	1p
Noordeloos.4	2,685	60	1p
OosterwijkI.1	2,155	70	1p
OosterwijkI.2	1,26	70	1p
OosterwijkI.3	0,8	70	1p
OosterwijkII.2	1,55	60	1p
OosterwijkIII.1**	3,825	70	1p
OosterwijkIII.2	2,235	70	1p
OosterwijkIV.2	2,385	70	1p
Polsbroek.1	2,09	80	1p
Polsbroek.2	4,04	80	1p
Schelluinen.1	3,125	60	1p
Voetakkers.1	1,21	40	1p
Waardenburg.3	5,325	35	1p
Wijngaarden.1	0,57	70	1p
Wijngaarden.2	1,12	70	1p
Wijngaarden.3	1,605	70	1p
Wijngaarden.4	2,105	70	1p
Wijngaarden.5	2,605	70	1p
Wijngaarden.6	3,105	70	1p
Wijngaarden.7	3,63	70	1p
Wijngaarden.8	4,115	60	1p
Wijngaarden.9	4,605	60	1p
Wijngaarden.10	5,105	60	1p
Wijngaarden.11	5,605	60	1p
Wijngaarden.12	6,115	60	1p
Wijngaarden.13	6,605	60	1p
Wijngaarden.14	7,005	60	1p
Wijngaarden.15	7,535	60	1p
Wijngaarden.16	8,175	40	1p
Wijngaarden.17	8,62	40	1p
Wijngaarden.18	9,105	40	1p
Wijngaarden.19	9,105	40	1p
Zeedijk.5	4,515	60	1p
Zeedijk.6	5,625	60	1p
Zoowijk1.0	2,045	40	1p
Zoowijk1.1	2,92	40	1p
Zoowijk1.2	4,39	40	1p
Hedel.1	1,735	40	2p
Hedel.3	4,755	40	2p
Heldam.1	3,08	40	2p
Heldam.2	2,73	40	2p
Heldam.3	2,27	40	2p
Heldam.4	1,94	40	2p
Heldam.5	1,52	40	2p
HoonkomLeebrug.3	4,275	60	2p
B38E0306	110	53	2p
B38E0306	125	58	2p
B38E0306	70	58	2p
B38E0306	105	59	2p
B38E0306	130	60	2p
B38E0306	120	61	2p
B38E0306	126	62	2p
B38E0306	135	63	2p
B38E0306	85	68	2p
B38E0306	95	69	2p
B38E0306	115	69	2p
B38E0306	80	70	2p
B38E0306	75	71	2p
B38E0306	90	73	2p
B38E0306	90	75	2p
Bazeldijk.1	3,44	60	2p
DeWoerd.6	3,125	40	2p
DeWoerd.7	3,425	40	2p
DeWoerd.8	3,725	40	2p
LangeAvontuur.1	1,005	40	2p
LangeAvontuur.2	1,95	40	2p
Leerdam.4	2,495	60	2p
Leerdam.5	2,415	60	2p
Leerdam.6	1,125	60	2p
Leerdam.7	0,88	60	2p
Ochten.1*	3	40	2p
Ochten.2	3,8	40	2p
Ochten.3	4,1	40	2p
Ochten.4	4,3	40	2p
Ochten.5	4,5	40	2p
Ochten.6	4,8	40	2p
Ochten.7	5,1	40	2p
OosterwijkII.1	3,425	60	2p
OosterwijkIV.1	3,315	70	2p
Polsbroek.3	5,27	70	2p
Polsbroek.4***	5,99	70	2p
Polsbroek.5***	6,865	70	2p
Polsbroek.6***	9,01	70	2p
Schelluinen.2	4,815	60	2p
Tienhoven.2	3,245	70	2p
Tienhoven.3	2,27	70	2p
Vlaardingen.1	6,51	40	2p
Vlaardingen.2	6,56	40	2p
Vlaardingen.3	6,7	40	2p
Vlaardingen.4	8,1	40	2p
Vlaardingen.5	8,17	40	2p
Vlaardingen.6	8,365	40	2p
Vlaardingen.7	8,255	40	2p
Voetakkers.2	1,84	40	2p
Voetakkers.3	3,415	55	2p
Waardenburg.1	2,105	35	2p
Waardenburg.2	2,625	35	2p
Zeedijk.1	0,78	40	2p
Zeedijk.2	1,95	40	2p
Zeedijk.3	2,61	40	2p
Zeedijk.4	3,075	60	2p
Zoowijk1.3	6,41	40	2p
Zoowijk1.4	6,59	40	2p
Zoowijk1.5	7,245	40	2p
Zoowijk1.6	7,48	40	2p
Zoowijk2.1	1,45	40	2p
Zoowijk2.2	2,41	40	2p
B25G1055	125	76	3c
B25G1055	130	76	3c

Table A4.1. continued.

Core	*depth (cm)*	*LOI (%)*	*facies*
B25G1055	120	77	3c
B25G1055	115	79	3c
B25G1055	170	83	3c
B25G1055	135	84	3c
B25G1055	180	85	3c
B25G1055	145	86	3c
B25G1055	175	86	3c
B25G1055	140	87	3c
B25G1055	185	87	3c
B25G1055	150	87	3c
B25G1055	155	88	3c
B25G1055	160	88	3c
B25G1055	195	89	3c
B25G1055	210	89	3c
B25G1055	190	89	3c
B25G1055	215	90	3c
B25G1055	220	90	3c
B25G1055	225	91	3c
B25G1055	230	92	3c
B25G1055	165	92	3c
B25G1057	532	88	4c
B25H0739	410	78	4c
B25H0739	391	80	4c
B25H0739	395	86	4c
B25H0739	385	86	4c
B25H0739	434	69	5c
B25H0739	431	80	5c
B25H0739	415	86	5c
B25H0739	420	89	5c
B25H0739	427	91	5c
B25G1057	495	32	6c
B25G1057	475	39	6c
B25G1057	470	39	6c
B25G1057	460	39	6c
B25G1057	465	49	6c
B25G1057	480	49	6c
B25G1057	490	49	6c
B25G1057	485	51	6c
B38D0432	560	30	6p
B38D0432	783	30	6p
B38D0432	885	31	6p
B38D0432	814	32	6p
B38D0432	575	33	6p
B38D0432	810	35	6p
B38D0432	895	36	6p
B38D0432	895	37	6p
B38D0432	885	38	6p
B38D0432	905	39	6p
B38D0432	890	39	6p
B38D0432	910	42	6p
B38D0432	910	43	6p
B38D0432	920	43	6p
B38D0432	915	44	6p
B38D0432	916	44	6p
B38D0432	785	45	6p
B38D0432	790	45	6p
B38D0432	610	46	6p
B38D0432	805	53	6p
B38E0306	250	43	6p
B38E0306	870	46	6p
B38E0306	868	49	6p
B38E0306	869	51	6p

Table A4.2. Statistics of LOI results.

	1c	*1p*	*2p*	*3c*	*4c*	*5c*	*6c*	*6p*
Average	54	55	51	86	84	83	43	41
min	38	30	35	76	78	69	32	30
max	83	80	75	92	88	91	51	53
count	11	131	70	22	5	5	8	24

Table A4.3. Key for organic facies codes.

facies code	*organic facies*
1	wood peat
2	reed peat
3	reed-sedge peat
4	sedge peat
5	oligotrophic peat
6	gyttja

suffix 'c' means that it originates from a core in the coversand area
suffix 'p' means that it originates from a core in the palaeovalley

Appendix 5

Distribution of cores that penetrate the basal peat layer in the Zuid-Holland study area

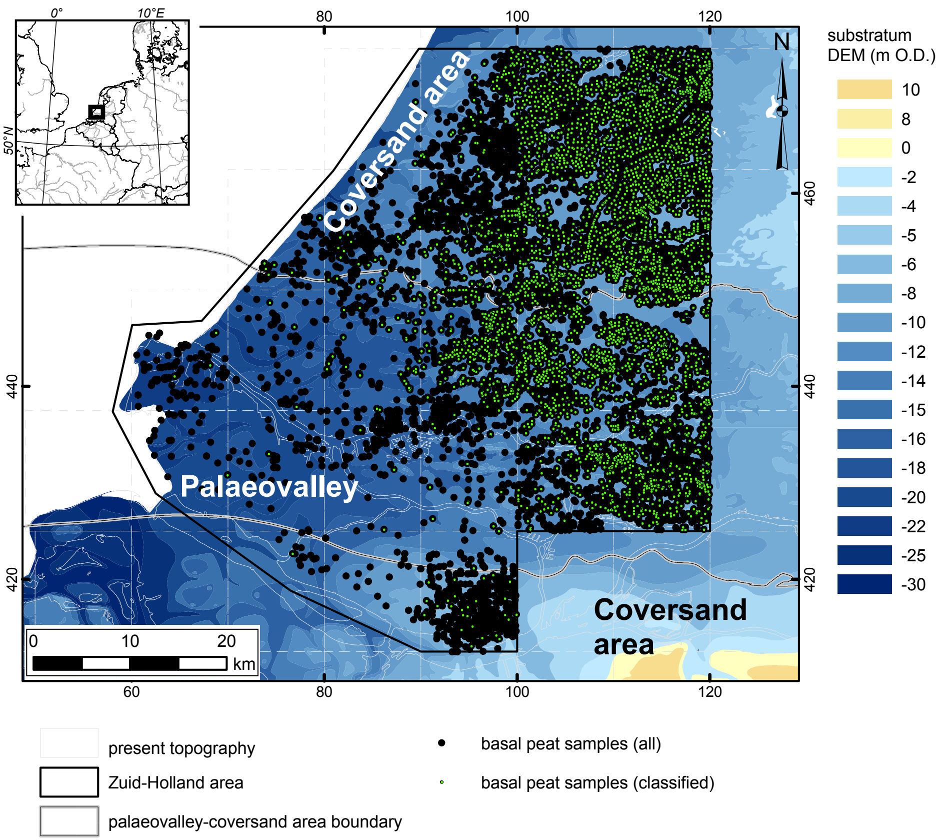


Fig. A5.1. Distribution of cores in the DINO-database that penetrate the basal peat layer in the Zuid-Holland area. Indicated are the positions of cores used to map the organic facies distribution of the basal peat layer.

Appendix 6

Interpolation of organic facies by means of Indicator Kriging or Inverse Distance Weighting

The organic-facies distribution map was based on interpolation of each facies separately. The statistics of these interpolation procedures are summarized in Table A6.1. The following steps were carried out on these raster files.

Step 1. Normalize the probability values for each cell to 1.
- Sum probability of the organic facies for each cell → e.g., <dekz_sum_prob>,
- Divide the probability value for each organic facies with <XX_sum_prob> → e.g., <gy_dekz_norm>
- This map could be used for displaying organic facies with probabilities of over 50%. With lower values, another facies may have a higher probability in that pixel. This requires the following steps.

Step 2. Obtain for each cell the organic facies with the highest probability
- Calculate for each organic facies where this facies has higher probabilities than the other organic facies → <dekz_gy>

Step 3. Indicate where the highest probability is at least 0.60
- Use 'con' on <gy_dekz_norm> where <dekz_gy> is > 0 → <gy_dekz_Mask>

Step 4

Remove parts were interpolation results are likely not accurate due to low data density.
- Only display those areas that are within a 500 m radius from the nearest data point.

Table A6.1. Interpolation properties for organic facies variability in the basal peat layer.

Area	*Organic facies*	*Range (m) [direction 1]*	*Range (m) [direction 2]*	*Sill*
	Wood peat	350	-	0.12
Palaeovalley	Reed peat	475 [120°]	420 [210°]	0.19
	Gyttja	250 [90°]	1 [180°]	0.16
	Wood peat	190	-	0.095
Coversand	Reed peat	IDW		
	Gyttja	IDW		

The neighborhood used was 2500 m, and the number of data points (n) required was set at 2<n<10.

Appendix 7

Results of macro-remains analyses on two cores in the Angstel-Vecht area

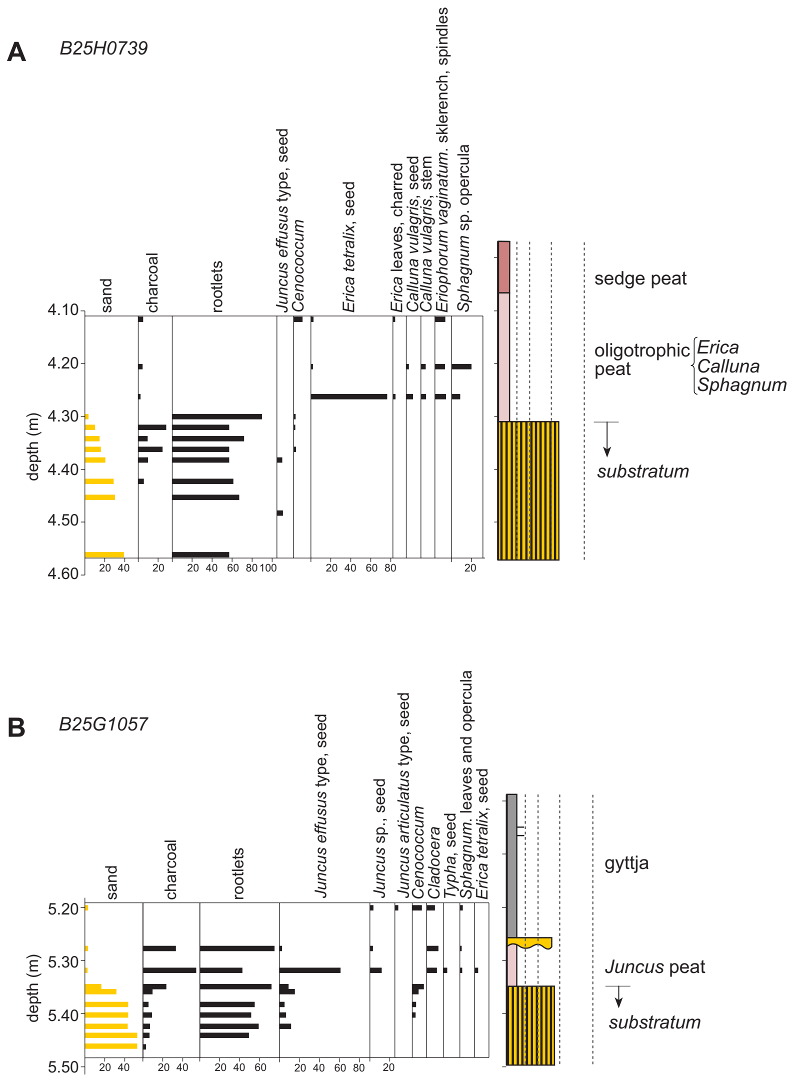


Figure A7.1. Macro-remains analyses of basal peat layer successions in the coversand area. The location is indicated in Figure 4.1 (Chap. 4). Analyses by Bas van Geel

Appendix 8

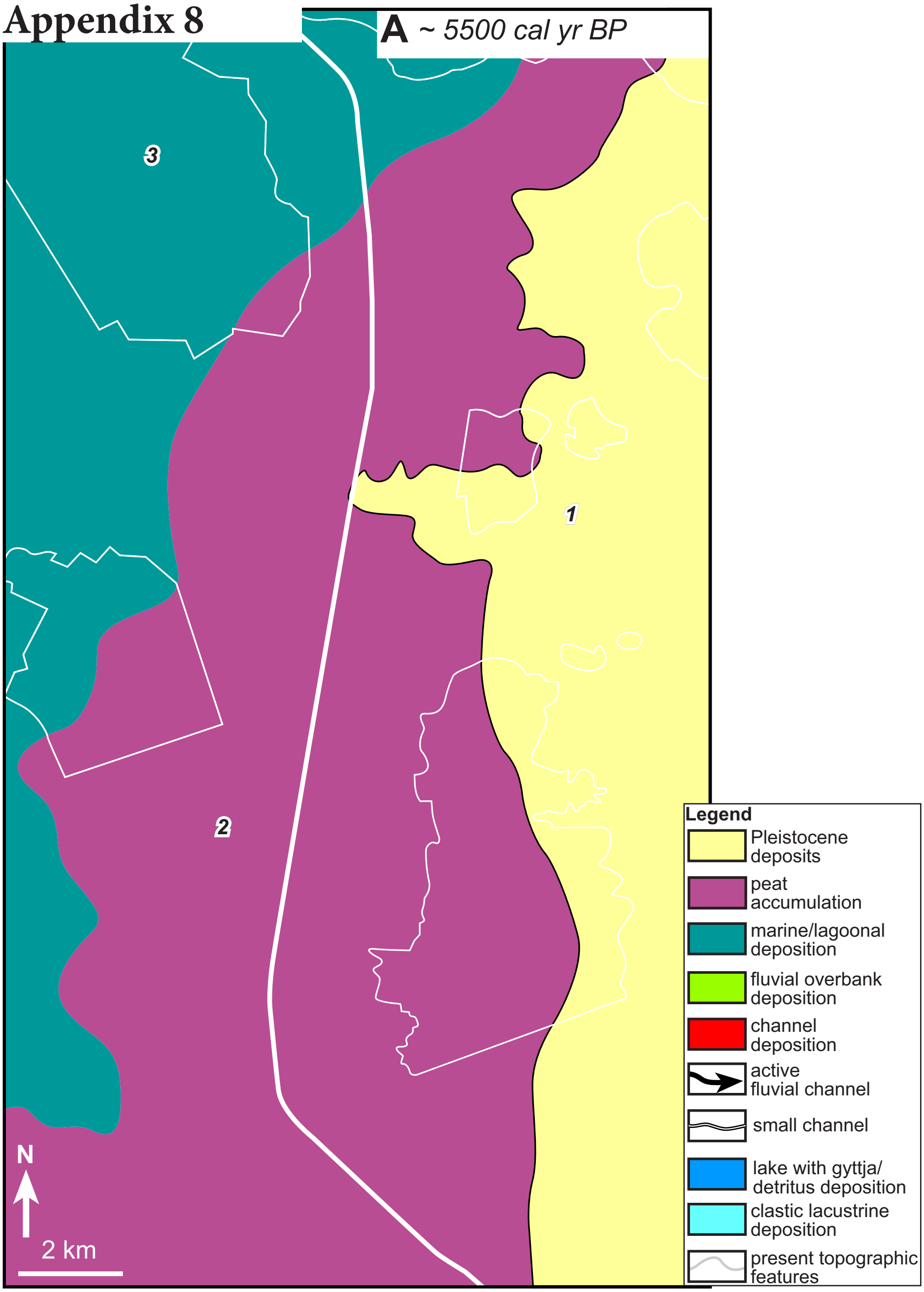


Figure A8.1. Palaeogeography of the Angstel-Vecht area ~ 5500 cal yr BP (Fig. 2.9A, Chap. 2). Pleistocene deposits surface in the east whereas wetlands and lagoonal environments prevrail in the west.

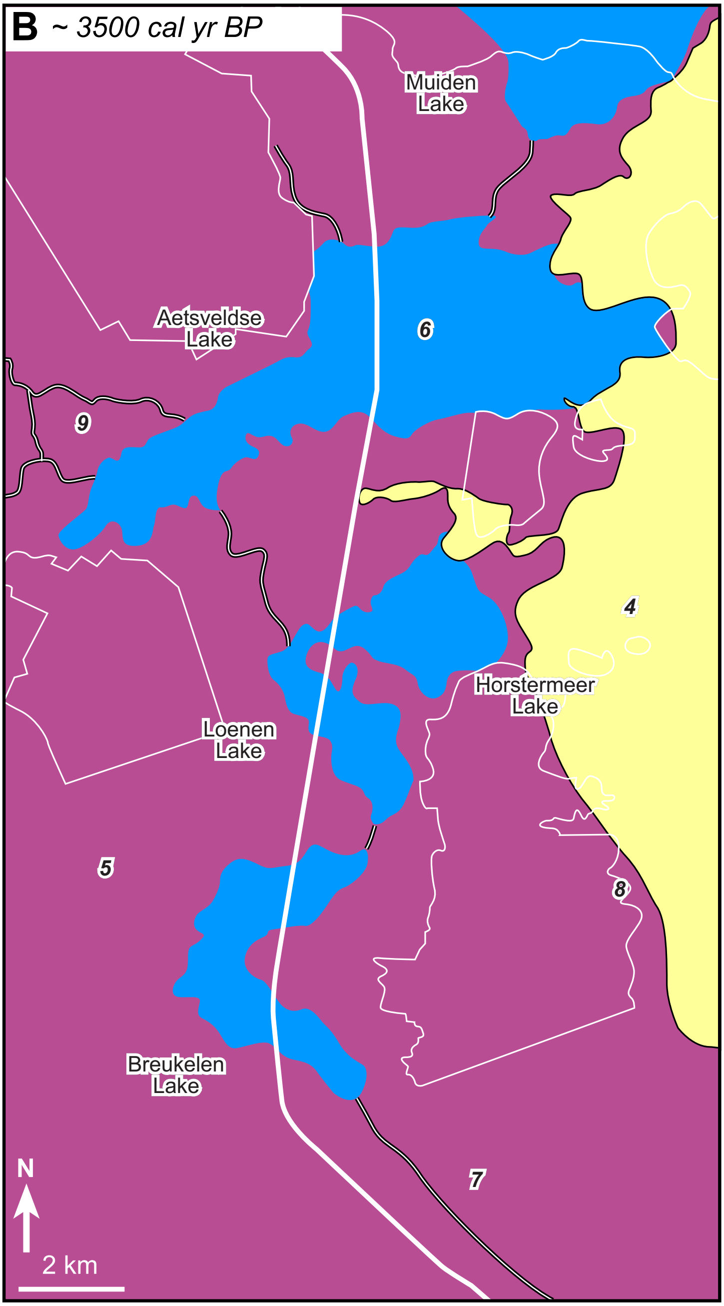


Figure A8.2. Palaeogeography of the Angstel-Vecht area ~ 3500 cal yr BP (Fig. 2.9B, Chap. 2). Pleistocene deposits surface in the east. The western part is characterized by extensive peat-forming wetlands and lakes. Further a small and/or intermittend connection with the Rhine is established.

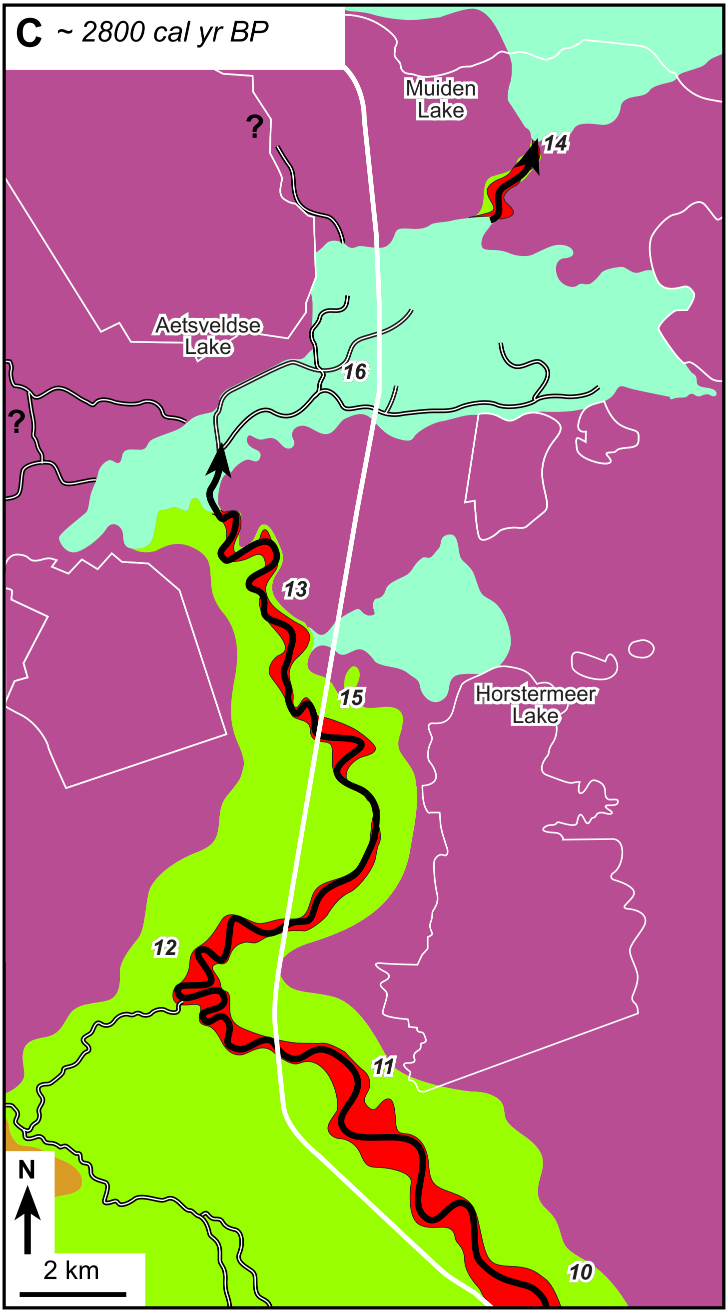


Figure A8.3. Palaeogeography of the Angstel-Vecht area ~ 2800 cal yr BP (Fig. 2.9C, Chap. 2). The Angstel-Vecht system functioned as a Rhine branch resulting in the formation of organic-clastic lake fills.

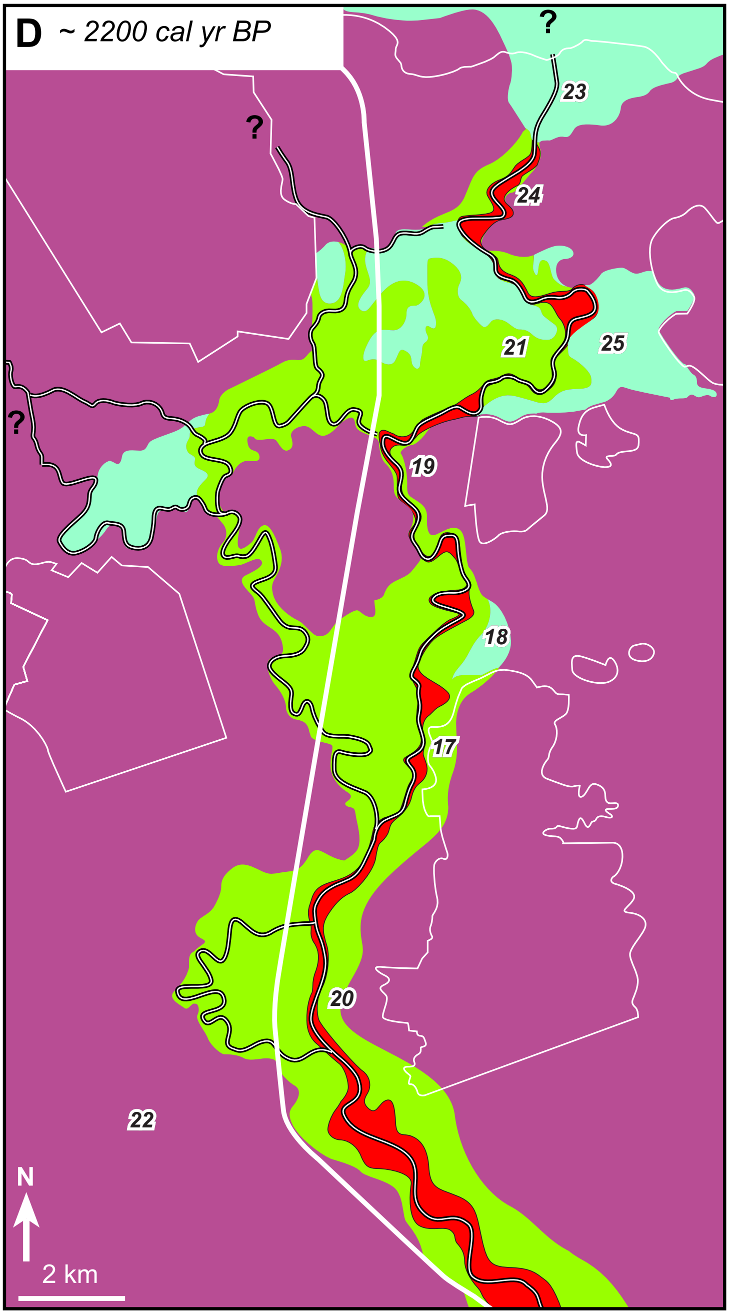


Figure A8.4. Palaeogeography of the Angstel-Vecht area ~ 2200 cal yr BP (Fig. 2.9D, Chap. 2). By this time, two avulsion had occurred, fluvial sedimentation had ceased and the lakes were largely filled.

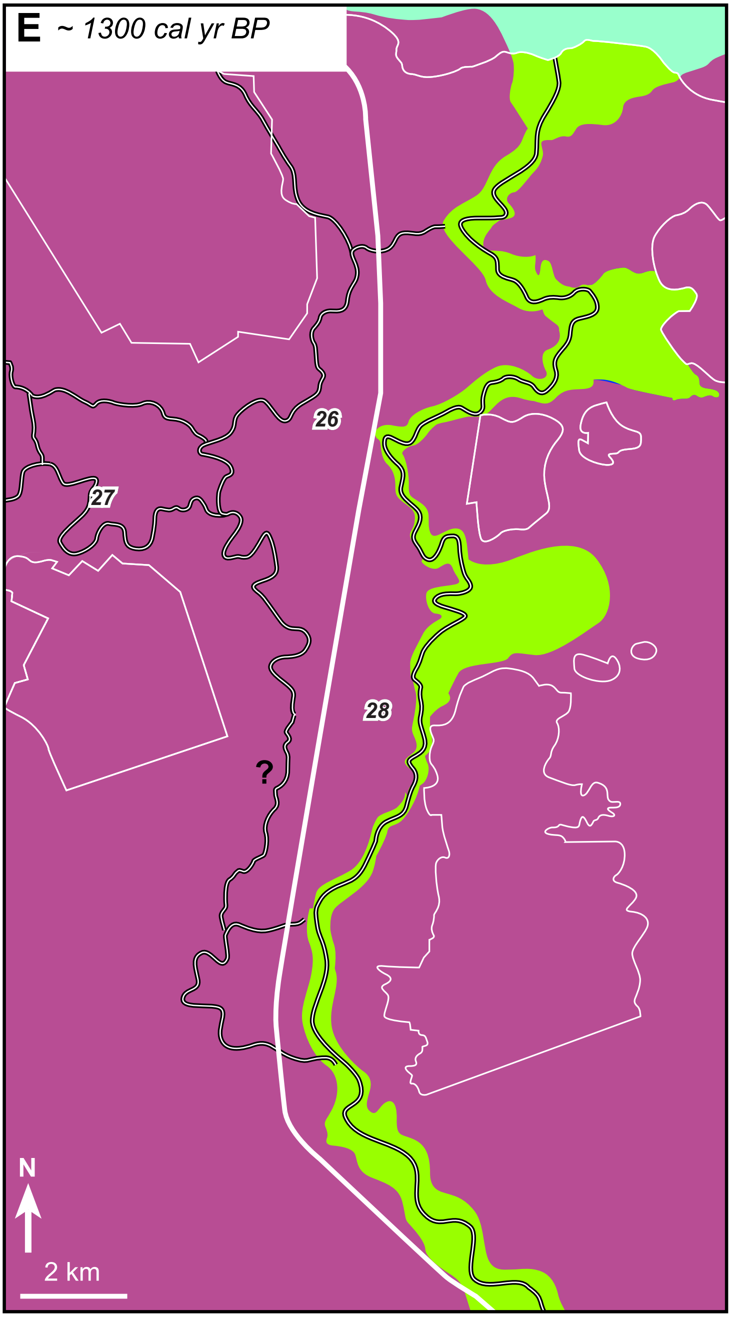


Figure A8.5. Palaeogeography of the Angstel-Vecht area ~ 1300 cal yr BP (Fig. 2.9E, Chap. 2). By this time, Rhine influence had substantially diminished; peat-forming wetlands now largely cover the fluvial deposits.

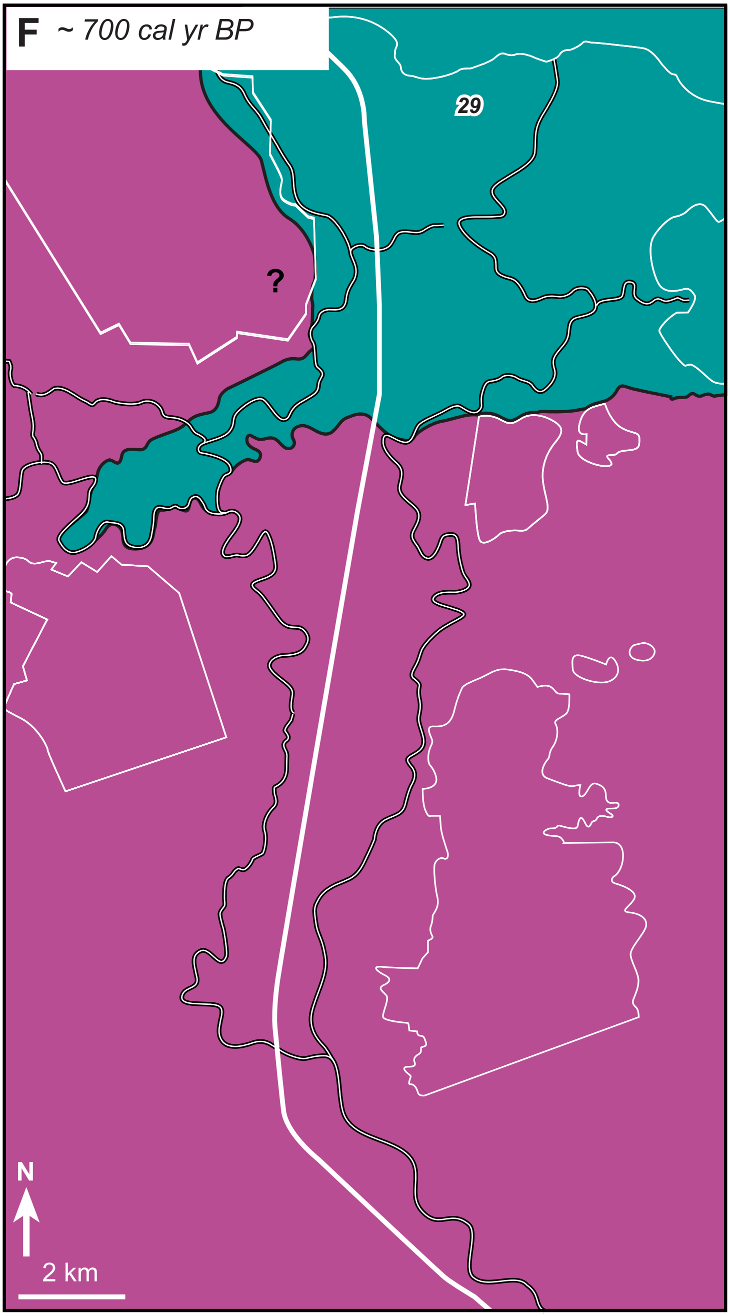


Figure A8.6. Palaeogeography of the Angstel-Vecht area ~ 700 yr BP (Fig. 2.9F, Chap. 2). Rhine influence definitely ceased AD 1122. The fluvial channel now functioned as predominantly as drainage channels for seepage water from the ice-pushed ridge.

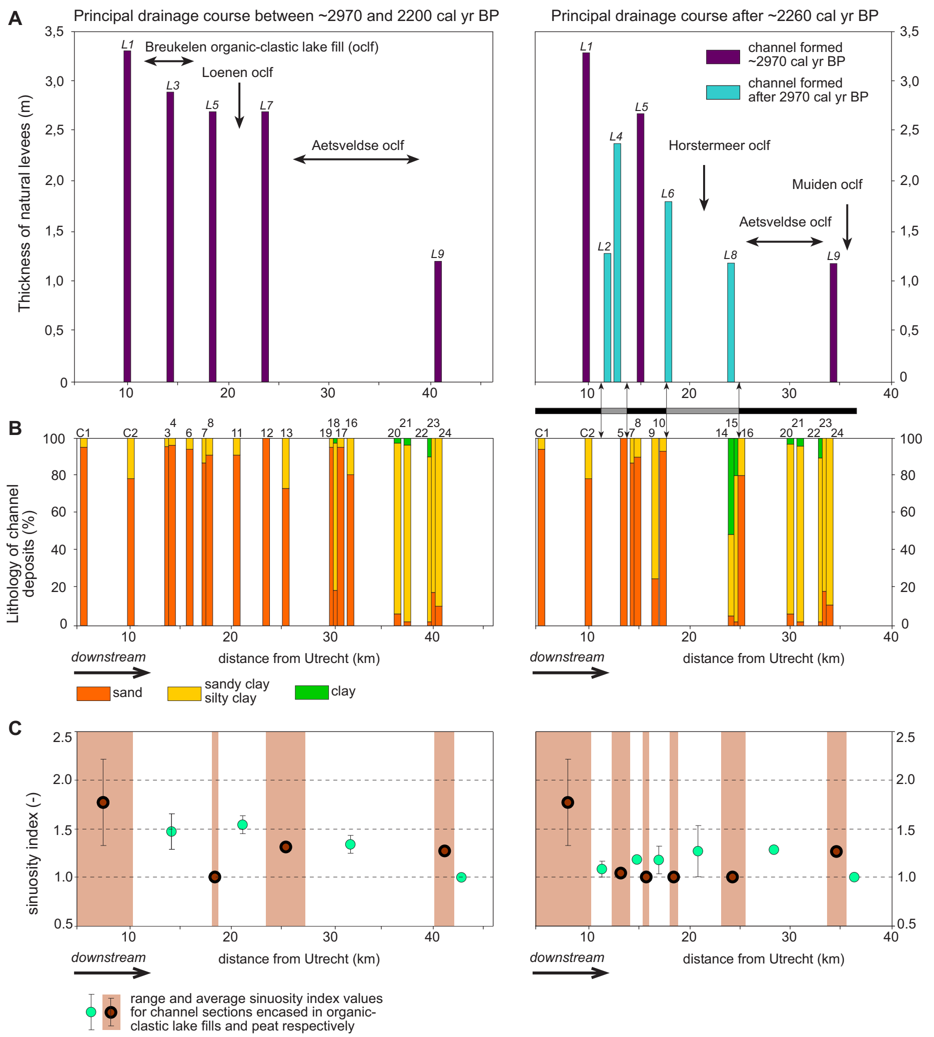


Figure A8.7 (Fig. 2.10, Chap. 2).

A *Graph showing the thickness of levees in downstream direction following the initial river course (outlined in Fig. A8.3) and the river course that includes the channels that formed at a later stage and follows the present River Vecht. The position of the organic-clastic lake fills relative to the measurement locations is indicated. The locations of the measurements are shown in Fig. 2.5 (Chap. 2).*

B *The lithological composition of channel deposits for the same river courses as outlined in graph A.*

C *The sinuosity index values of channel sections in the Angstel-Vecht area. The channel sections are either encased in organics or in organic-clastic lake fills. Sinuosity values indicate whether fluvial channels are straight (<1.3) or meandering. The channel sections encased in organic-clastic lake fills have higher sinuosity index values compared with those that are encased in organics (predominantly peat).*

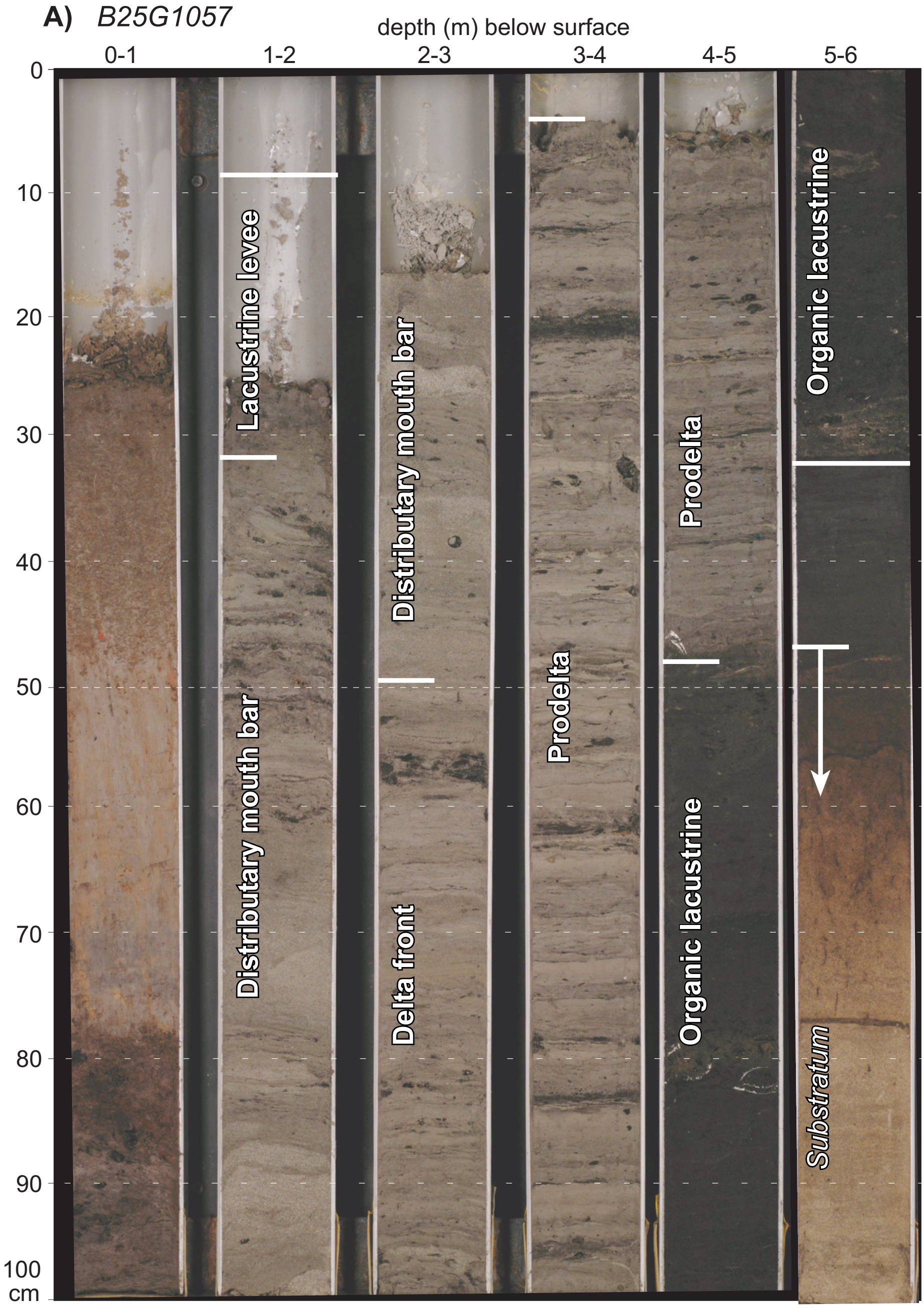
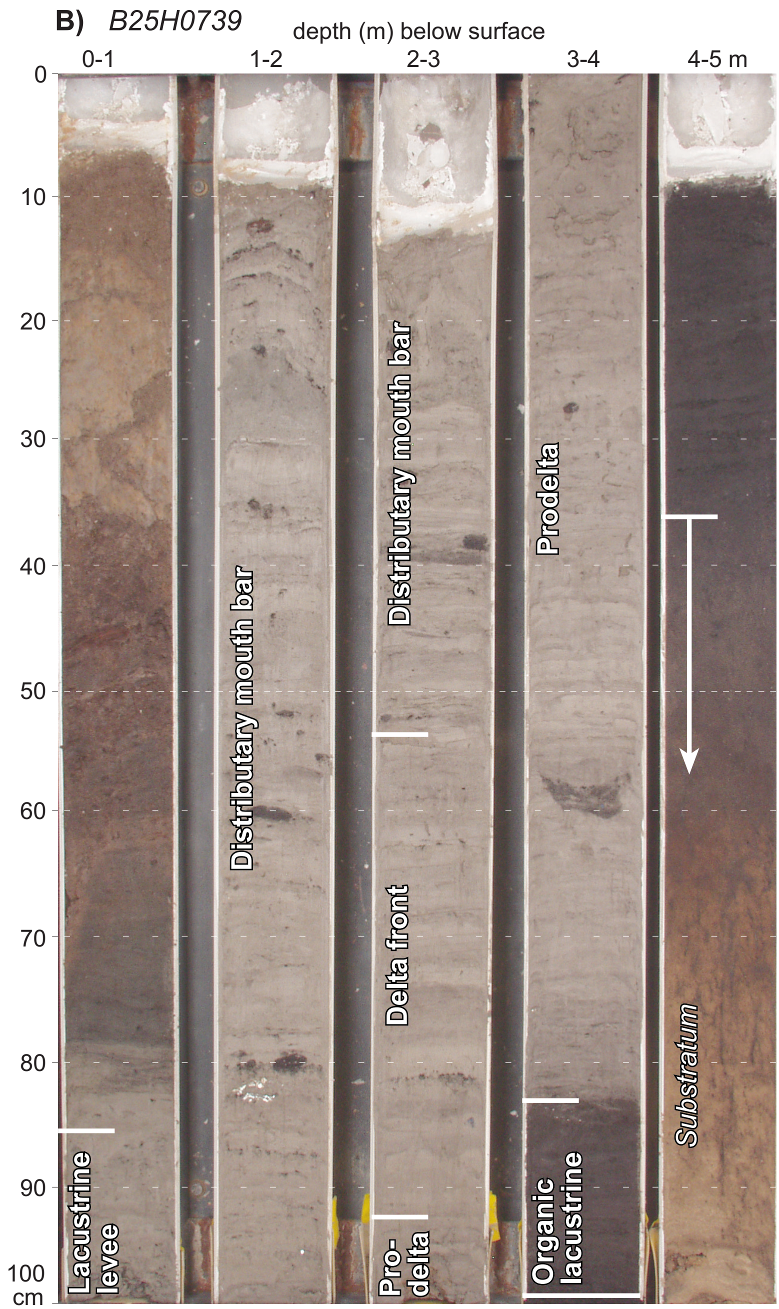


Figure A8.8. Pictures of mechanical cores (Fig. 3.6, Chap. 3) in A) sandy organic-clastic lake fills and B) clayey organic-clastic lake fills. See Fig. 3.3A-B (Chap. 3) for the location and sedimentary logs. Sur-



face elevation of the cores is -1.09 O.D. (A) and -1.78 O.D. (B). The interpretation of the sections that were not recovered – empty parts of the tubes – is based on manual cores from the same location.

A

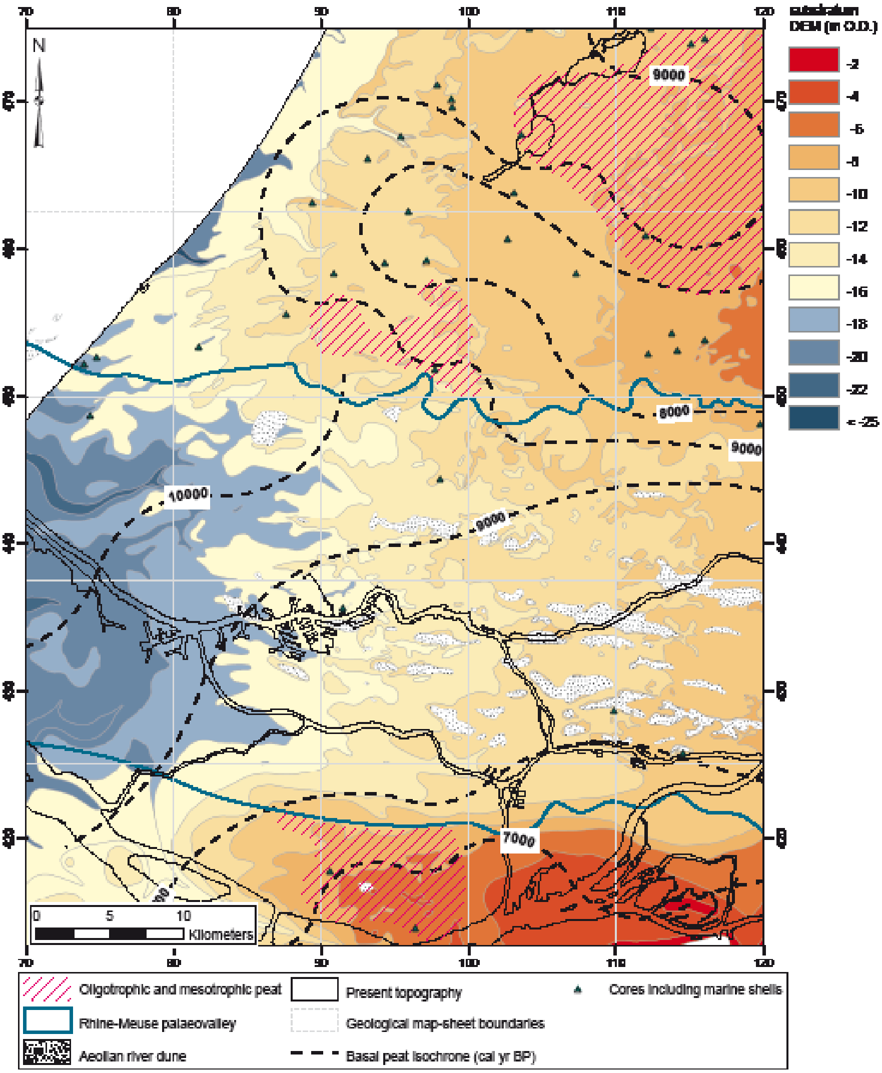
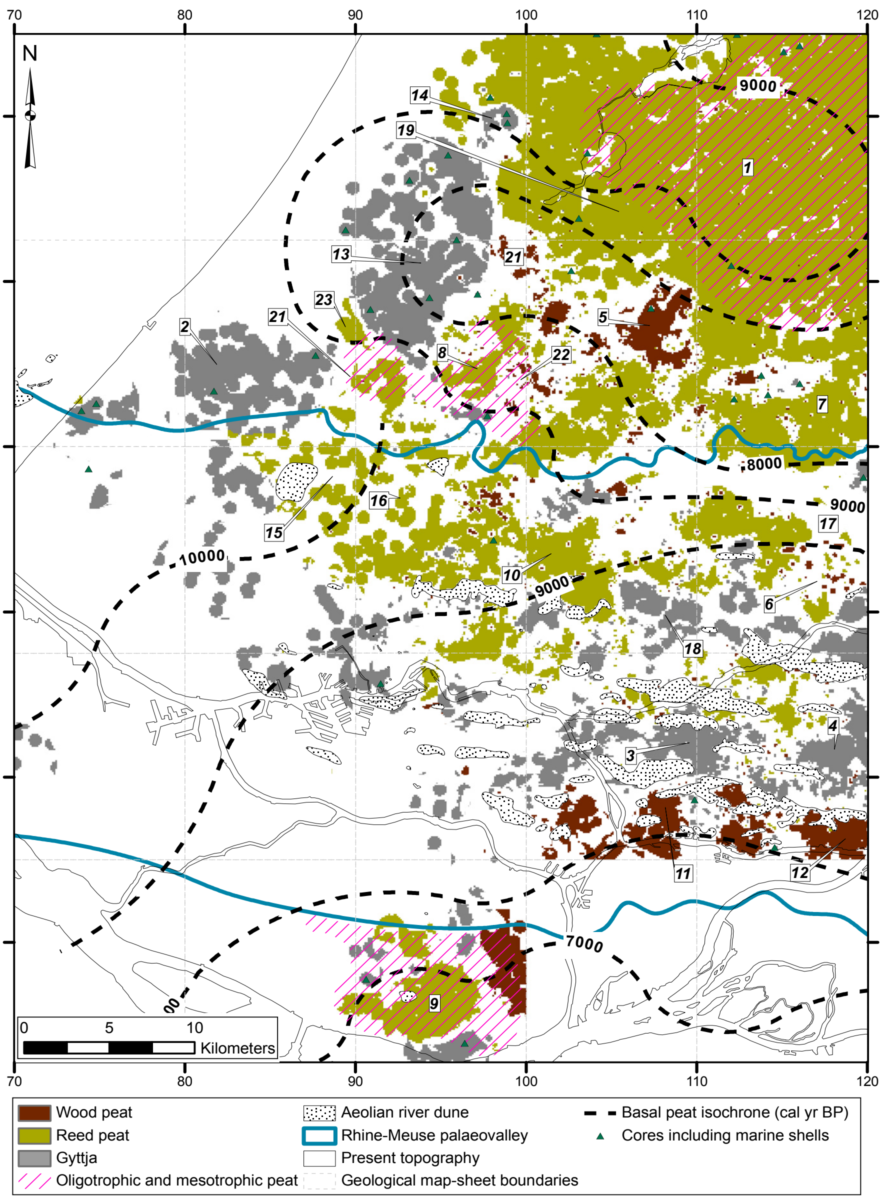


Figure A8.9. (Fig. 4.7, Chap. 4). A. DEM of the substratum. B. (next page) Spatial distribution of organic facies in the (diachronous) basal peat layer in the distal part of the Holocene Rhine-Meuse delta. Isochrones (corresponding to palaeogroundwater levels of Cohen, 2005) indicate the onset of Holocene aggradation and hence the maximum age of the basal peat. Facies are displayed for locations where the distance to the nearest core is less than 500 m, which corresponds to the maximum range of spatial correlation. The area shown corresponds to Fig. A8.9A, which enables comparison of the spatial distribution of organic facies in the basal peat layer and the topography of the substratum. Italic numbers are referred to in the text (Chap. 4).

B


N
9000
8000
9000
10000
9000
7000
0 5 10
Kilometers
Wood peat
Reed peat
Gyttja
Oligotrophic and mesotrophic peat
Aeolian river dune
Rhine-Meuse palaeovalley
Present topography
Geological map-sheet boundaries
Basal peat isochrone (cal yr BP)
Cores including marine shells

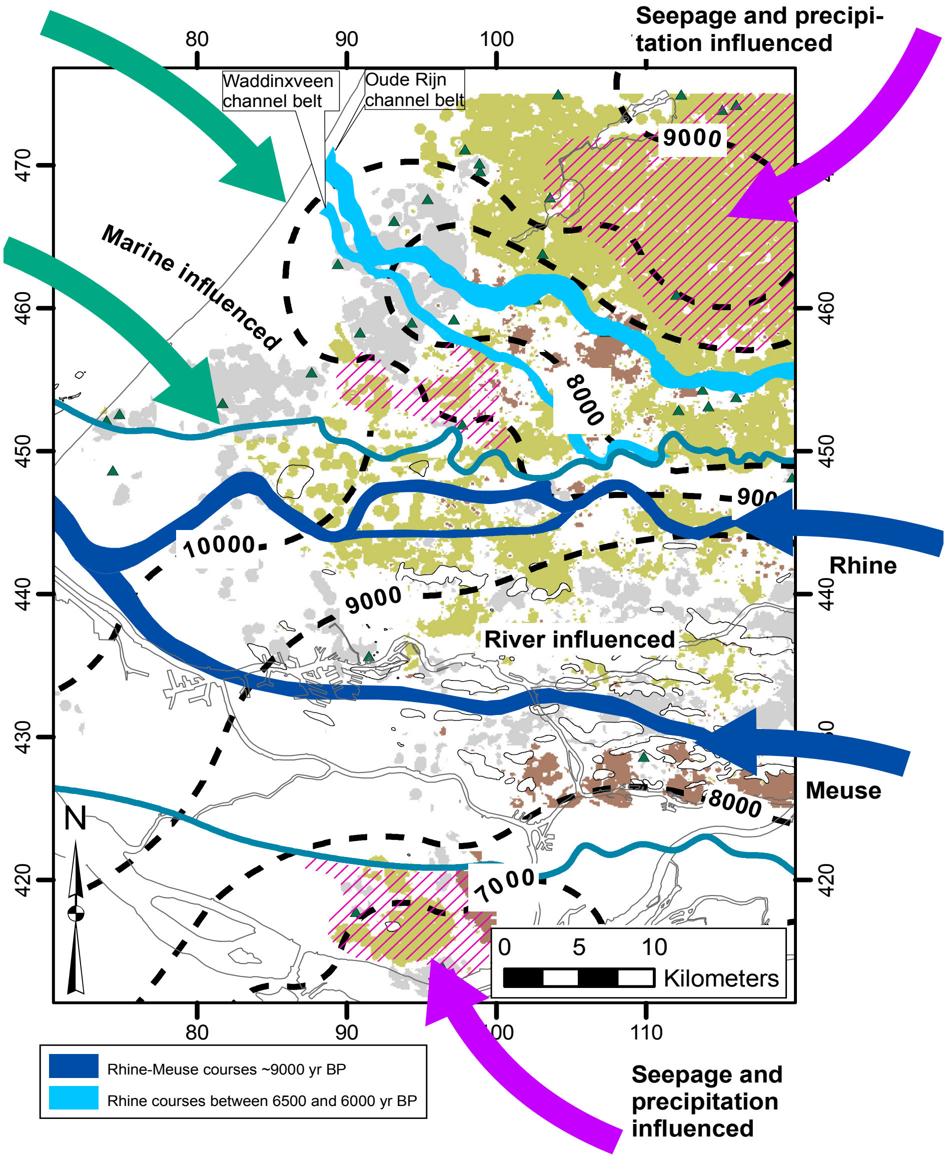


Figure A8.10. (Fig. 4.8, Chap. 4). Dominant hydrological regimes controlling the onset of Holocene organic aggradation as inferred from the facies composition of the basal peat. Only legend elements that differ from Fig. A8.9B are explained. The position of fluvial channels is after Hijma (2009). Between 9000 and 6500 cal yr BP the Rhine is positioned in the palaeovalley, corresponding to the situation at 9000 cal yr BP.

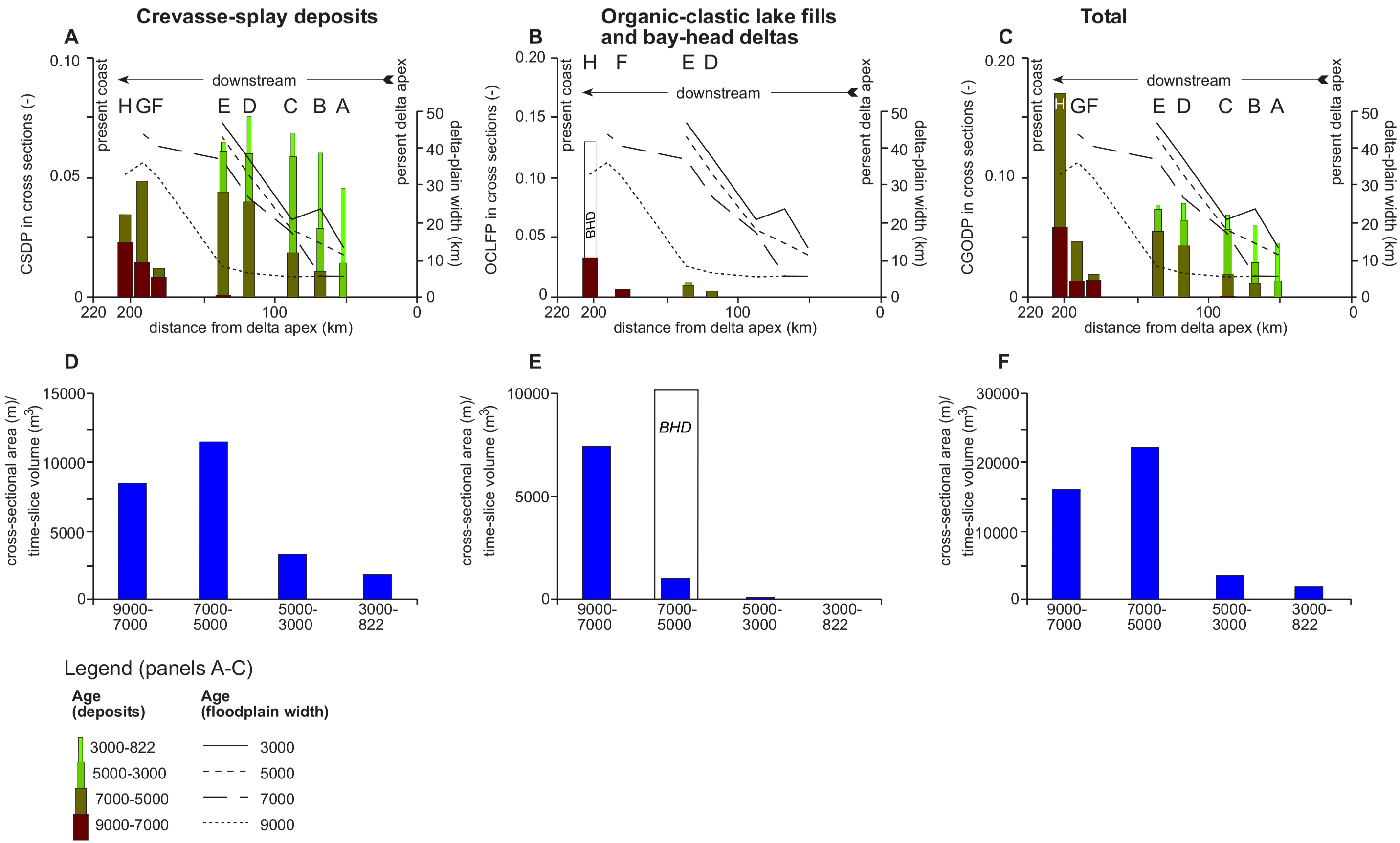
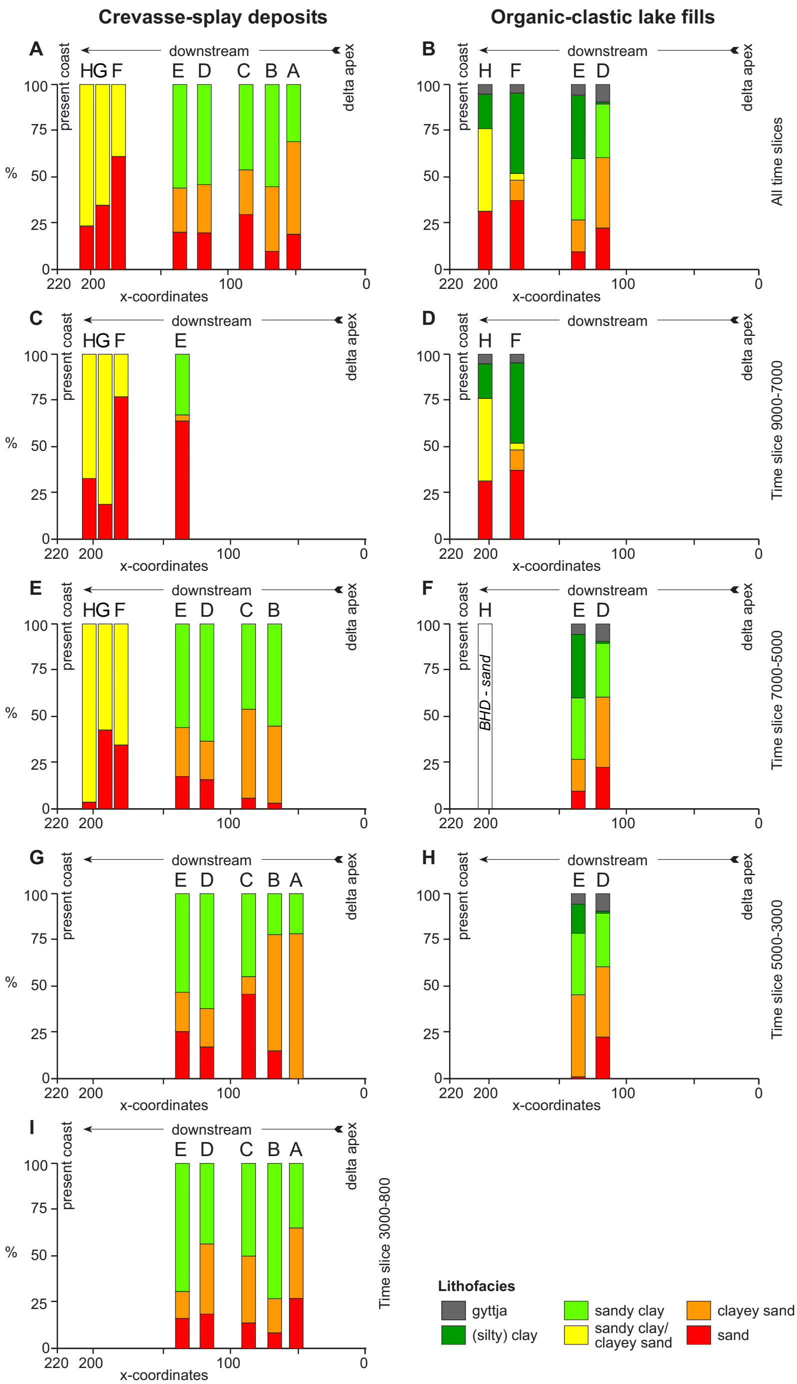


Figure A8.11. (Fig. 5.3, Chap. 5). Spatial (A-C) and temporal (D-F) distribution of the proportions of crevasse-splay deposits (left-hand panels), and organic-clastic lake fills and bay-head delta deposits (middle panel) and their summed values. Floodplain width is given in diagrams A-C.



Crevasse-splay deposits
Organic-clastic lake fills
All time slices
Time slice 9000-7000
Time slice 7000-5000
Time slice 5000-3000
Time slice 3000-800
downstream
present coast
delta apex
x-coordinates
BHD - sand
Lithofacies
gyttja
(silty) clay
sandy clay
sandy clay/
clayey sand
clayey sand
sand

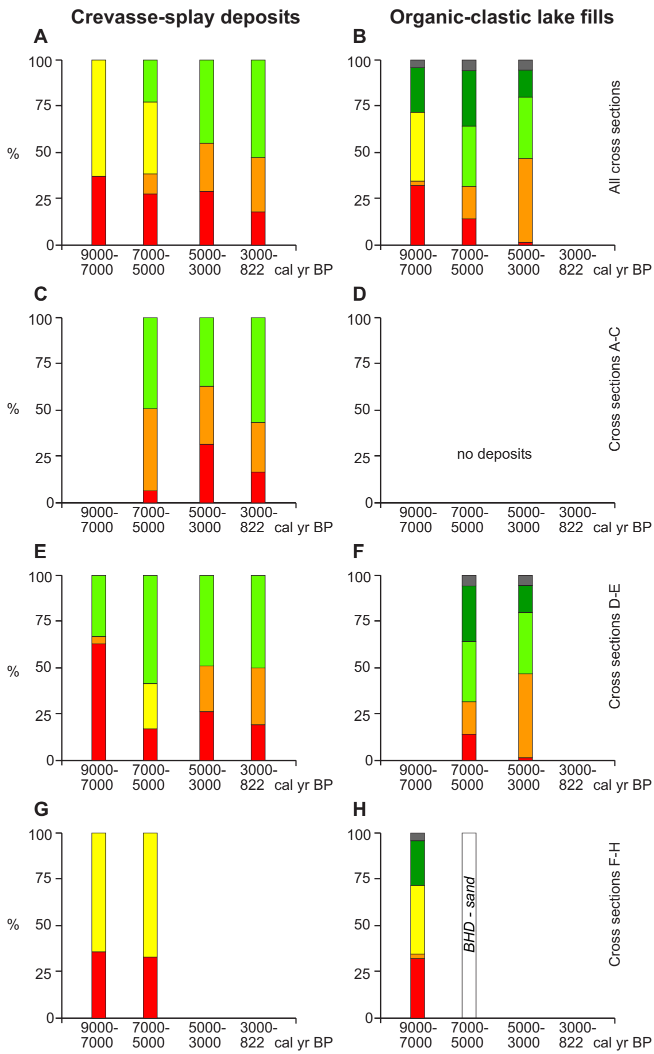


Figure A8.13. (Fig. 5.5, Chap. 5). Lithofacies composition of crevasse-splay deposits and organic-clastic lake fills related to time slices. Diagrams A and B indicate composition for the whole delta. The other diagrams give the composition for three parts of the delta, being the proximal delta plain (summed values of cross sections A-C), the intermediate delta plain (summed values of cross sections D and E) and the distal delta plain (summed values of cross sections F-H). For legend, see figure A8.12.

← Figure A8.12. (Fig. 5.4, Chap. 5). Lithofacies composition of crevasse-splay deposits and organic-clastic lake fills related to longitudinal position in the delta plain. Diagrams A and B indicate composition for the whole delta. The other diagrams give the composition for various periods, being 9000-7000 cal yr BP (C and D), 7000-5000 cal yr BP (E-F), 5000-3000 cal yr BP (G and H) and 3000-800 cal yr BP (I). Organic-clastic lake fills did not form after 3000 cal yr BP in the central delta plain.

Summary

Introduction (Chap. 1)

Delta plains are of enormous significance for present-day society, because they provide ample space for habitation and agriculture. Delta-plain deposits are a source rock for water and fossil fuels such as oil and gas. Therefore, the composition of delta-plain deposits has been studied intensively over the past decades. The focus of these studies was on proximal parts of delta plains and on channel-belt deposits. This means that architectural elements that are confined to distal delta plains, especially non-channel deposits, received relatively little attention, despite the fact that non-channel sediments and organics may form up to 70% of distal delta plains successions. So far neglected issues relate to organic accumulations, lake fills (both organic and clastic) and coarse-grained overbank deposits (crevasse-splay deposits, organic-clastic lake fills and bay-head delta deposits). The Rhine-Meuse delta is an excellent area to study these issues, because of the availability of well-defined architectural and palaeogeographic frameworks. Further, the deposits of interest are at or near the surface, which enables optimal use of the available geological data for this area.

The general objective of this thesis is to analyze and explain the architecture, facies distribution, age and origin of coarse-grained overbank deposits, with special attention for organic-clastic lake fills, and organics in the distal Holocene Rhine-Meuse delta plain. This objective has been specified by the formulation of research goals, which are:

1. To describe the geometrical properties and depositional facies of organic-clastic lake fills, to understand how they form and to determine their influence on delta-plain architecture and facies composition.
2. To investigate the sedimentary and botanical variability of organics in the basal peat layer in the distal delta plain and to assess the factors that controlled their spatial distribution patterns.
3. To analyze and explain spatial and temporal distribution patterns of coarse-grained overbank deposits and their lithofacies composition at delta scale.

In order to depict the influence of lakes and their fills on the development and architecture of fluvial systems, I reconstructed the palaeogeographic development in the Angstel-Vecht area (Chap. 2). To obtain insight in the facies composition, architecture and formation of organic-clastic lake fills, I investigated these deposits in the central Rhine-Meuse delta (Schoonhoven area) and an area fringing the central delta (Angstel-Vecht area) (Chap. 3). The sedimentary and botanical variability of the basal peat layer was described, first, by designing a key for classifying organics and, second, by applying this key on archived borehole descriptions and interpolating the resulting subset of organic facies. This enables the production of a distribution map of organic facies in the basal peat layer (Chap. 4). I used existing valley-wide cross sections to determine the delta-scale spatial and temporal distribution and lithofacies composition of coarse-grained overbank deposits (Chap. 5).

Organic-clastic lake fills (Chap. 2 and 3)

The influence of the presence of lakes and organic-clastic lake fills on the evolution of a fluvial system has been investigated by means of a palaeogeographic reconstruction of the Angstel-Vecht area (Chap. 2). On the basis of manual cores, I constructed a geological map and cross sections. Age control on the period of formation of the identified deposits was obtained by using ^{14}C, OSL, pollen, archaeological artefacts and historical sources. It was illustrated that delta-plain lakes affect downstream architecture and facies composition of fluvial deposits. Owing to trapping of the coarsest sediment grains in lakes (predominantly silt and sand), channel deposits downstream of lakes are relatively fine-grained and are accompanied by relatively thin natural-levee deposits. The large proportion of sand that is contained in organic-clastic lake fills, in addition, provides weak channel banks compared to the clay and peat usually found in distal delta plains. Channels that traverse organic-clastic lake fills, therefore, have a tendency to meander instead of being straight.

Sedimentary analyses were carried out to describe the depositional facies of organic-clastic lake fills (chapter 3). Mechanical cores were collected for log description and a geological map and cross sections were used to determine the architecture. Further, the formative duration of the Aetsveldse organic-clastic lake fill in the Angstel-Vecht area was determined by applying sediment-transport calculations on the feeding channel (=Angstel), and using lake-fill sediment volume and time constraints for its formation. The lacustrine setting is strongly reflected in the facies composition and the architecture of organic-clastic lake fills. The base of organic-clastic lake fills is composed of organic-lacustrine sediment (gyttja), which is overlain by laminated clastic deposits, including significant volumes of distributary mouth bar deposits (mainly sandy lithofacies). It is proposed that organic-clastic lake fills preferentially occur at the base of aggradational sequences, i.e., in the lower part of Transgressive Systems Tracts (TST) and High-stand Systems Tracts (HST) according to sequence stratigraphic nomenclature.

The presence of a palaeogeographic and architectural framework provided the possibility to calculate the formative duration of the Aetsveldse organic-clastic lake fill (chapter 3). It was shown that the formative duration of a deltaic sand body could be calculated by comparing their volume with sediment transport rates, which were deduced from channel dimensions of the feeding channel. The formative duration was found to be ~99 yr within 770 year (Chap. 3). The intermittency I_f, which is defined as the fraction of time that a river has or exceeds bankfull discharge, thus is 0.13 (=99/770)—which is similar to the I_f modern Rhine distributaries.

Finally, a synthesizing conceptual model is presented that describes the development of these deposits (Chap. 3).

Spatial distribution and sedimentary and botanical variability of organics in the basal peat layer in the distal delta plain (Chap. 4)

The field characteristics of sedimentary and botanical organic facies in the peat layer at the base of the Holocene succession were investigated by designing a classification key for organic facies, primarily on the basis of concise macroscopic descriptions of cores. Organics were divided into organic lake sediment (gyttja) and *in situ* peat and further subdivided according to sedimentary characteristics and botanical content respectively. It was shown that this key enabled identification of four types of sedimentary organic facies (algal, detrital, calcareous and siderite gyttja) and four botanical-organic facies (wood, reed, sedge and oligotrophic peat).

Application of this key on archived borehole descriptions resulted in determination of the distribution of organic facies in the basal peat layer in the distal part of the Rhine-Meuse delta in ~4000 cores. These data were interpolated (indicator kriging and IDW), which resulted in a distribution map of basal-peat organic facies. It was feasible to obtain the spatial distribution and spatial proportion of reed peat (63%), wood peat (8%) and gyttja (29%). Identification of sparsely recognized oligotrophic and mesotrophic peat types (sedge) as well as shell species was very helpful for interpreting the organic-facies distribution. Organic facies variability in the basal peat layer indicates the presence of three main hydrological domains that characterized the onset of Holocene aggradation in the Holocene Rhine-Meuse delta: marine-dominated, fluvial-dominated and seepage-dominated.

Spatial and temporal distribution patterns of coarse-grained overbank deposits (Chap. 5)

The spatial and temporal distribution patterns of coarse-grained overbank deposits and their lithofacies composition were evaluated in eight existing valley-wide cross sections, which resulted in identification of crevasse-splay deposits, organic-clastic lake fills and bay-head delta deposits. The presence of time lines in these cross sections allowed for temporal differentiation, thereby enabling identification of temporal distribution patterns. Coarse-grained overbank deposits occupy 7.1 % of the Holocene Rhine-Meuse delta-plain. Crevasse-splay deposits were found to be present throughout the fluvial part of the delta plain, although their proportion is partly controlled by lateral floodbasin width. Both narrow and wide floodbasins tend to have relatively low proportions of crevasse-splay deposits whereas intermediate floodbasin widths, 3.1-3.6 km in the Rhine-Meuse delta, yield optimum crevasse-splay deposit proportions (CSDP). Organic-clastic lake fills are associated with extensive accumulation of organics (Chap. 2, 3 and 5), which only occurs when sedimentation of clastic deposits is outpaced by the creation of accommodation space. In aggrading delta plains, this situation is restricted to periods with relatively high rates of base-level rise, or to wide floodplains. The lithofacies composition of crevasse-splay deposits and organic-clastic lake fills is found to be controlled by channel planform, superelevation and, less dominantly, by the local composition of the substratum. Outer-bend crevasse channels in meandering rivers receive relatively small amounts of sand because helical flow transports river bed load towards the inner side of the bend. Furthermore, superelevation in meandering river systems tends to be relatively low, which limits incision depths of crevasse-splay channels, thereby again restricting significant amounts of bed load to be diverted to the crevasse splay. In straight-anastomosing river systems, helical flow is less prominent and crevasse channels are generally deeper incised owing to larger superelevation values. Coarse-grained overbank deposits along anastomosing river channels contain more sand relative to those associated with meandering river systems. Finally, we found that coarse-grained overbank deposits have a positive effect on reservoir volumes and connectedness ratios.

Synthesis (Chap. 6)

This chapter integrates results and conclusions of the preceding research chapters. This thesis illustrates the significance of organics in build-up of the distal Rhine-Meuse delta; in the Angstel-Vecht area, for example, the proportion of organics is 0.45. The ubiquity of organics and the lakes that are associated with peat-forming wetlands have had a great influence on fluvial processes and thereby affected facies composition and architecture of fluvial deposits in distal delta plains. The potentially large volume of sand in organic-clastic lake fills (up to 44%) means that bank stability is low compared to that in surrounding peat areas and that fluvial channels therefore more easily adopt a meandering channel planform. Further, downstream of lakes natural levee deposits are thin and channel deposits are fine-grained relative to their upstream counterparts. The architecture of organic-clastic lake fills differs from the architecture of crevasse-splay deposits. Generally, the former is associated with a single channel by which they are dissected and may occur anywhere in the distal delta plain. In contrast, crevasse-splay deposits are present at one side of a fluvial channel, and may only be dissected by their own channel, but then, they exist at a bifurcation of two channels.

Application of the results of this thesis can improve 3D architectural models and numerical models that simulate subsurface geometry. Simulation models that are used, for example, to analyse hydrocarbon reservoirs can be calibrated with quantitative results regarding architecture and facies composition. Further, interpretation of deposits in ancient settings can be improved by application of the insights presented in this thesis, especially regarding organic facies and organic-clastic lake fills. Recognition of these deposits will also be helpful in reconstructing the configuration of the distal parts of ancient delta plains.

Finally, directions for future research are suggested. First, the issue of lake formation in peat-forming wetlands has not been solved completely. Studies in other regions and model exercises could shed light on the processes involved. Second, identification of spatial and temporal distribution patterns of organic facies at delta scale potentially could be extended to more proximal regions and to organics that are not contained in the basal peat layer at the base of the Holocene succession, i.e., included in sediments overlying this basal peat layer. Third, the use of palaeochannel dimensions to calculate palaeoflow conditions is promising thereby enabling the calculation of the formative duration of coarse-grained overbank deposits. It is feasible to apply this method in the central Rhine-Meuse delta, which will improve the understanding of the formation of these deposits.

Samenvatting

Introductie (Hfst. 1)

Deltavlakten zijn van enorme betekenis voor onze samenleving. Ze liggen stroomopwaarts van riviermondingen. Het zijn gebieden met vruchtbare landbouwgrond en mede daardoor vaak dichtbevolkt. Daarnaast worden veel van onze fossiele brandstoffen gewonnen uit (zandige) geulafzettingen in oude, tegenwoordig op grote diepte liggende deltaïsche-afzettingen. Vanwege het enorme economische belang is er veel wetenschappelijk onderzoek verricht naar de architectuur (de 3D opbouw in de grond) van deze afzettingen. De meeste aandacht ging hierbij uit naar zandige geulafzettingen. Omdat afzettingen in benedenstroomse gedeelten van deltavlakten (in Nederland: ruwweg ten westen van Woerden) tot wel 70% uit andere grondsoorten kan bestaan bleven afzettingen in dergelijke gebieden relatief onderbelicht. Deze andere grondsoorten zijn afzettingen buiten rivierbeddingen (komafzettingen) en grond bestaande uit organisch materiaal (organische accumulaties). Deze laatste bestaat uit resten van planten die ter plekke zijn gegroeid (*in situ* veen) en in water bezonken plantenresten en algen (gyttja). Uitgebreidere kennis van benedenstroomse deltavlakten kan onder andere van pas komen bij bouwprojecten (o.a. gebouwen, infrastructuur), en heeft ook als gevolg dat we de potentie van gas- en oliereservoirs beter kunnen benutten.

Er bestaan nog veel vragen over de ontwikkeling en samenstelling van de organische accumulaties en komafzettingen, en dan vooral grofkorrelige (zandige) afzettingen. Deze laatste bestaan uit grofkorrelige, veelal zandige injecties vanuit de rivier de kom in (crevasse-afzettingen), en kleinere, lokale delta's die zich vormen in meren (organische-klastische meerafzettingen) en op plekken waar riviertakken zich verbreden (zogenaamde estuariumbekkendelta's, zoals de Biesbosch in Nederland). Organische-klastische meerafzettingen zijn een mix van organische deeltjes (veenbrokjes, blaadjes) en klastische deeltjes (zand en klei). Dit proefschrift beschrijft en verklaart de architectuur, de samenstelling en de vorming van grofkorrelige komafzettingen en organische accumulaties. De Rijn-Maasdelta is een uitermate geschikt gebied om bovenstaande onderwerpen te onderzoeken, omdat de positie en ontwikkeling van riviergeulen sinds 9.500 v. Chr. in hoofdlijnen erg goed bekend is. Bovendien zijn er veel organische accumulaties en organische-klastische meerafzetingen en liggen deze vaak aan of nabij het oppervlak. Hierdoor kan, naast de uitgebreide geologische databases, optimaal gebruik worden gemaakt geomorfologische (≈reliëf) kenmerken. Dit geldt vooral voor het Angstel-Vechtgebied ten noorden van Utrecht waar veel van dit onderzoek plaatsvond. Voor de eveneens onderzochte basisveenlaag is vooral gebruik gemaakt van de databases. De basisveenlaag ligt juist diep (tot ~25 m -NAP) aan de basis van de delta en ontstond na de laatste ijstijd toen door zeespiegelstijging het onderzoeksgebied vernatte. Er zijn drie onderzoeksdoelen geformuleerd:

1. Het beschrijven van de geometrie (vorm) en het afzettingsmilieu van organische-

klastische meerafzettingen, het vormingsproces begrijpen en de invloed ervan bepalen op de 3D-opbouw van deltavlakte-afzettingen en de sedimentologische eigenschappen van delta-afzettingen,

2. Het onderzoeken van de variabiliteit van de basisveenlaag in het benedenstroomse gedeelte van de delta (uit welke plantenresten en klastische deeltjes bestaat deze) en het verklaren van de ruimtelijke verbreiding van typen veen/gyttja in de basisveenlaag,
3. Het analyseren en verklaren van de ruimtelijke en temporele verbreiding van grofkorrelige komafzettingen en de variatie in sedimentologische opbouw voor de gehele Rijn-Maasdelta.

De volgende stappen zijn ondernomen om deze onderzoeksdoelen te bereiken:

- De paleogeografie van het Angstel-Vechtgebied is beschreven op basis van een geologische kartering en dateringen (Hfst. 2).
- In het centrale gedeelte van de Rijn-Maasdelta (het gebied rond Schoonhoven) en in een gebied grenzend aan de centrale delta (Angstel-Vechtgebied) zijn organische-klastische meerafzettingen onderzocht (Hfst. 3).
- Op basis van uitvoerig macroscopisch beschreven sedimentkernen is een classificatiesleutel ontwikkeld om typen organische accumulaties te bepalen. Deze sleutel is vervolgens toegepast op gearchiveerde en digitaal beschikbare beschrijvingen van boringen (Hfst. 4). De analyse heeft een verbreidingskaart opgeleverd van de verschillende typen organische accumulaties.
- De ruimtelijke en temporele verbreiding van grofkorrelige komafzettingen en de variabiliteit van de samenstelling daarvan zijn geanalyseerd op basis van geologische dwarsprofielen over de gehele delta (Hfst. 5).

Organische-klastische meerafzettingen: paleogeografie van het Angstel-Vechtgebied (Hfst. 2)

De invloed van meren op de ontwikkeling van een riviersysteem is onderzocht aan de hand van een paleogeografische reconstructie van het Angstel-Vechtgebied. De reconstructie beslaat het Holoceen, dat wil zeggen sinds 9.500 v. Chr. Op basis van handboringen zijn een geologische kaart en dwarsprofielen gemaakt. De ouderdom van de afzettingen is bepaald met onder andere ^{14}C-dateringen, archeologische vondsten en historische bronnen. De Angstel en Vecht ontstonden ongeveer 1000 v. Chr. en drongen een moerasgebied binnen. De daar al aanwezige meren werden in hoog tempo opgevuld met organische-klastische meerafzettingen. Ongeveer 200 v. Chr. nam de activiteit van deze rivieren sterk af. Veel later, in 1122 AD, werd de verbinding met de Rijn zelfs helemaal verbroken. De geulen voerden daardoor nog slechts lokale neerslag en kwelwater uit de Utrechtse Heuvelrug af. Dit onderzoek toont aan dat meren in deltavlakten de architectuur en de sedimentologische samenstelling van riviersedimenten beïnvloeden. Doordat de meest grove sedimentdeeltjes (silt en zand) in meren bezinken, zijn riviergeulafzettingen benedenstrooms van meren relatief fijnkorrelig en zijn de daaraan gekoppelde oeverwalafzettingen relatief dun. Bovendien zorgt het relatief grote aandeel van zand in organische-klastische meerafzettingen ervoor dat de oeverstabiliteit daar relatief laag is. Rivierbochten kunnen zich daardoor relatief gemakkelijk naar buiten toe verplaatsen, zodat een meanderend rivierpatroon kan ontstaan. Dit in tegenstelling tot rivieren die zich vormen in een veen- of kleigebied. De stevige oevers in dergelijke gebieden zorgen ervoor dat deze rivieren zich over het algemeen niet lateraal verplaatsen en een zogenaamd recht rivierpatroon aannemen.

Organische-klastische meerafzettingen: architectuur, samenstelling en vormingsproces (Hfst. 3)

Om het precieze afzettingsmilieu te bepalen van organische-klastische meerafzettingen zijn sedimentanalyses uitgevoerd op machinaal gestoken boorkernen uit het Angstel-Vechtgebied en uit de omgeving van Schoonhoven. De architectuur en samenstelling zijn afgeleid uit een geologische kaart en dwarsprofielen, gemaakt op basis van handboringen. De sedimentaire structuren, waaronder de horizontale gelaagdheid, en de populatie van diatomeeën (kiezelwieren) wijzen op een lacustrien (=meer) afzettingsmilieu. De verticale opbouw van het sediment bevestigt dit: aan de basis organische meerafzettingen (gyttja), gevormd toen de Angstel en Vecht nog niet bestonden, met daar bovenop gelaagde klastische afzettingen, afkomstig van de Angstel en Vecht. Deze laatste bestaan voor 23-44% uit zandbankafzettingen die zijn gevormd in voormalige mondingen van kleine riviergeultjes. Organische-klastische meerafzettingen in het Angstel-Vechtgebied zijn doorgaans niet dikker zijn dan 5 m en niet groter dan 25 km^2. Aan de hand van onder meer de dikte en breedte van de voormalige aanvoergeulsedimenten (Angstel) is een schatting gemaakt van de transportcapaciteit voor water en sediment. Vervolgens is berekend hoeveel tijd er nodig is geweest om de Aetsveldse organische-klastische meerafzetting te vormen, uitgaande van de maximale geulafvoer. Deze zogenaamde vormingsperiode is 99 jaar. Op basis van dateringen (Hfst. 2) is bekend in hoeveel kalenderjaren de vorming heeft plaatsgevonden: 770 jaar. Daaruit volgt dat de Angstel 13% (=(99/770)*100) van de tijd maximale geulafvoer heeft gehad. Dat is vergelijkbaar met de 16% van de huidige Rijn. Uitgaande van bovenstaande bevindingen en eerder verricht onderzoek is een conceptueel model ontwikkeld voor de vorming van organische-klastische meerafzettingen.

Ruimtelijke verbreiding en sedimentologische en botanische variabiliteit van organische accumulaties in de basisveenlaag in de benedenstroomse deltavlakte (Hfst. 4)

In de eerste fase van dit onderzoek zijn beschrijvingen van sedimentaire en botanische eigenschappen van organische accumulaties gemaakt. Vervolgens zijn deze verwerkt in een classificatiesleutel. Organische accumulaties worden op hoofdlijnen onderverdeeld in gyttja en *in situ* veen. Verdere onderverdeling kan aan de hand van sedimentologische eigenschappen (bij gyttja) en botanische samenstelling (bij veen). Op basis van de classificatiesleutel kunnen in gearchiveerde boorbeschrijvingen vier gyttjatypen (algen-, detritus-, kalk- en siderietgyttja) en vier veentypen (bos-, riet-, zeggeveen en oligotroof veen) worden onderscheiden.

In het benedenstroomse gedeelte van de Rijn-Maasdelta kon met behulp van de classificatiesleutel in ~4000 boringen, met daarin de basisveenlaag, het type veen/gyttja bepaald worden. Met behulp van interpolatietechnieken (indicator kriging en IDW) werd van deze puntdata een vlakdekkende kaart gemaakt. Uit de resultaten blijkt dat de basisveenlaag voornamelijk uit rietveen (63%), bosveen (8%) en gyttja (29%) bestaat. Verder komen sporadisch oligotroof veen, iets vaker zeggeveen en soms ook schelpen voor. De resultaten geven aanleiding in de basisveenlaag drie verschillende hydrologische regio's te onderscheiden: regio's gedomineerd door marine condities, regio's gedomineerd door condities gerelateerd aan rivierwater, en regio's gedomineerd door kwelwater.

Ruimtelijke en temporele verspreidingspatronen van grofkorrelige komafzettingen (Hfst.5)

De ruimtelijke en temporele verbreidingspatronen van grofkorrelige komafzettingen en hun lithologische samenstelling zijn geanalyseerd in acht noord-zuid georiënteerde dwarsprofielen van de Rijn-Maasdelta. Dit heeft geresulteerd in identificatie van crevasse-afzettingen, organische-klastische meerafzettingen en estuariumbekkendelta's. De tijdslijnen in de profielen maakten het vervolgens mogelijk om de afzettingen onder te verdelen naar vormingsperiode zodat ook temporele trends zichtbaar konden worden gemaakt. Het blijkt dat grofkorrelige afzettingen 7.1% van de totale Rijn-Maasdelta-afzettingen vormen. Crevasse-afzettingen komen vrijwel overal voor in de delta. Het aandeel ervan varieert wel enigszins en is deels afhankelijk van de breedte van de kommen (≈de laterale afstand tussen twee riviergeulen). Bij zowel brede als smalle kommen is het aandeel van crevasse-afzettingen relatief laag, terwijl kombreedten van 3.1 tot 3.6 km in de Rijn-Maasdelta optimum-waarden opleveren voor het aandeel van crevasse-afzettingen. Organische-klastische meerafzettingen komen hoofdzakelijk voor in gebieden met grootschalige veenvorming. Dergelijke gebieden komen alleen dan voor wanneer de relatieve zeespiegel snel stijgt en/of wanneer de deltavlakte erg breed is. In de Rijn-Maasdelta zijn deze alleen te vinden in het benedenstroomse gebied. Estuariumbekkendelta's komen uitsluitend voor op de overgang van benedenstroomse deltavlakte naar estuarium, in Nederland grofweg ten westen van Gorkum.

De lithologische samenstelling van grofkorrelige komafzettingen wordt hoofdzakelijk bepaald door geulpatroon, superelevatie (het hoogteverschil tussen rivier en kom bij maximale geulafvoer) en in bepaalde gevallen ook door de locale samenstelling van het onderliggende sediment. Crevasse-afzettingen langs meanderende rivieren worden doorgaans gevoed vanuit de buitenbocht. Ze bevatten meestal maar weinig zand, omdat de zogenaamde helicoïdale stroming sediment vanuit de buitenbocht naar de binnenbocht transporteert waardoor er relatief weinig zand vanuit de buitenbocht de crevassegeul in kan gaan. Superelevatie is relatief klein in meanderende riviersystemen waardoor de insnijdingsdiepte van crevassegeulen wordt beperkt. Ook dit heeft als gevolg dat minder zand dergelijke crevasse-afzettingen bereikt. In anastomoserende riviersystemen is de helicoïdale stroming minder sterk en zijn crevassegeulen gemiddeld genomen dieper ingesneden vanwege grotere superelevatie ten opzichte van meanderende riviersystemen.

Dit verklaart het relatief zandige karakter van grofkorrelige komafzettingen in anastomoserende riviersystemen. Eén van de conclusies van dit onderzoek is dat organische-klastische meerafzettingen, en zeker ook estuariumbekkendelta's, relatief erg veel zand bevatten, in ieder geval tot wel 44%. Een andere belangrijke conclusie is dat grofkorrelige komafzettingen van belang zijn voor de olie- en gaswinning: de reservoirs zijn groter wanneer dergelijke afzettingen aanwezig zijn, maar de reservoirs zijn ook beter met elkaar verbonden, wat de winbaarheid aanzienlijk doet toenemen.

Synthese (Hfst. 6)

In dit hoofdstuk worden resultaten en conclusies van de onderzoekshoofdstukken (Hfst. 2-5) in hun samenhang gepresenteerd. Dit proefschrift toont aan dat organische accumulaties aan de randen van de Rijn-Maasdelta van grote betekenis zijn geweest voor de ontwikkeling ervan. In het Angstel-Vechtgebied bijvoorbeeld, vormen ze 45% van de afzettingen van de afgelopen 11.500 jaar. De grote veengebieden met de daarbij behorende meren en meeropvullingen hebben fluviatiele processen beïnvloed en daarmee ook de vorm en de samenstelling van het riviersediment. Het grote volume zand in organische-klastische meerafzettingen (tot 44%) zorgt lokaal voor een verlaagde oeverstabiliteit ten opzichte van het omringende veengebied. Daardoor kunnen rivieren die organische-klastische meerafzettingen doorsnijden makkelijker hun bochten uitbreiden en zullen daardoor eerder een meanderend karakter aannemen. Een ander effect van meren op fluviatiele systemen is dat ze sediment afvangen waardoor benedenstrooms van deze meren oeverwallen dunner zijn en geulafzettingen minder zandig. Daarnaast is het zo dat de architectuur van organische-klastische meerafzettingen verschilt van die van crevasse-afzettingen. Organische-klastische meerafzettingen zijn gerelateerd aan één riviergeul die veelal de meerafzettingen doorsnijdt. Crevasse-afzettingen komen juist voor aan de zijkant van een riviergeul en worden dus niet doorsneden, tenzij door de eigen geul, maar dan ligt de crevasse-afzetting op de plek van een voormalige bifurcatie, een splitsing in de riviergeul.

De resultaten van dit onderzoek kunnen worden toegepast in zowel conceptuele modellen van deltavlakte-afzettingen als in computersimulatiemodellen die de ondergrond van delta's beschrijven. Simulatiemodellen die gebruikt worden bij olie- en gasonderzoek, kunnen bijvoorbeeld verbeterd worden doordat de percentages zand, veen, enz., in olie- en/of gashoudende lagen beter kunnen worden bepaald. Ook bij onderzoek naar de vorming van oude delta's leveren de inzichten uit dit proefschrift een belangrijke bijdrage. Het identificeren van organische accumulaties en organische-klastische meerafzettingen, bijvoorbeeld, geeft informatie over de vorm en de oriëntatie van de delta. Dat volgt uit de observatie dat deze afzettingen in het benedenstroomse deltavlakten voorkomen, en dus een maat zijn voor bijvoorbeeld de afstand tot de kust.

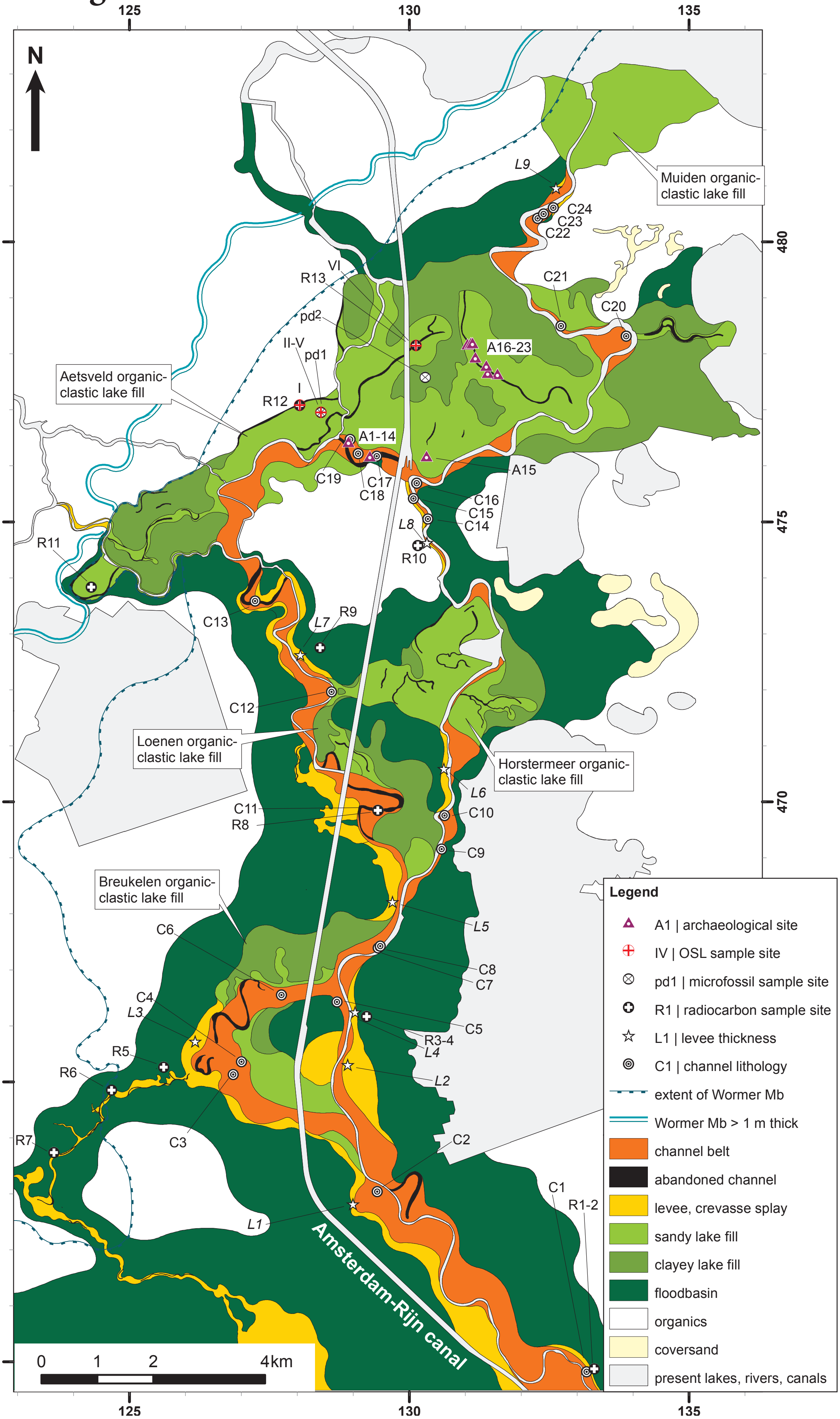
Ten slotte worden suggesties gedaan voor toekomstig onderzoek. De vorming van meren in laagveengebieden is deels verklaard in dit proefschrift; een vervolgonderzoek zou een vergelijkingsstudie moeten bevatten van verschillende gebieden, evenals een modelstudie naar het effect van kwel op de vorming van meren in veengebieden. Een ander onderwerp waar aandacht naar uit zou kunnen gaan, is het identificeren van verschillende typen organische accumulaties. In dit onderzoek is dat gedaan voor een gedeelte van de basisveenlaag; het zou waardevol zijn om dat uit te breiden naar de hele delta en naar bovenliggende organische accumulaties, om zo een compleet ruimtelijk en temporeel beeld te krijgen van de verspreiding van typen veen/gyttja. Vervolgonderzoek zou ook gedaan kunnen worden naar de berekening van stromingskenmerken van voormalige riviergeulen, waardoor de vormingsperiode van bijvoorbeeld organische-klastische meerafzettingen en crevasse-afzettingen kan worden berekend. Als dit op deltaschaal is toe te passen, zal dit bijdragen aan het inzicht in de vorming van deltavlakten. Gezien de huidige problematiek in veel delta's (bodemdaling, zeespiegelstijging, erosie) is een beter begrip van organische-klastisch afzettingen noodzakelijk, omdat dit onderzoek voor het eerst aantoont dat dergelijke afzettingen een groot deel van het totale pakket deltavlakte-afzettingen vormen.

About the author

Ingwer Bos was born and raised in Driebergen-Rijsenburg, positioned on the transition between the Pleistocene landscape of the Utrechtse Heuvelrug and the Holocene fluvio-deltaic area. He graduated from high school in 1997. Before he moved to Amsterdam to pick up studies, he lived one year in New Zealand, milking cows. Between 1998 and 2004, Ingwer studied physical geography at the University of Amsterdam. In this period, he conducted fieldwork in a wide variety of environments, ranging from Dutch inland aeolian dunes to fluivo-glacial regions in the Alps and semi-arid terrains in Spain. After his graduation he carried out a 3-months geomorphological-mapping fieldwork in Austria for geoheritage-conservation purposes. In 2005, Ingwer started the PhD research that resulted in this thesis. Ingwer is currently employed at Stichting Landschap Noord-Holland.

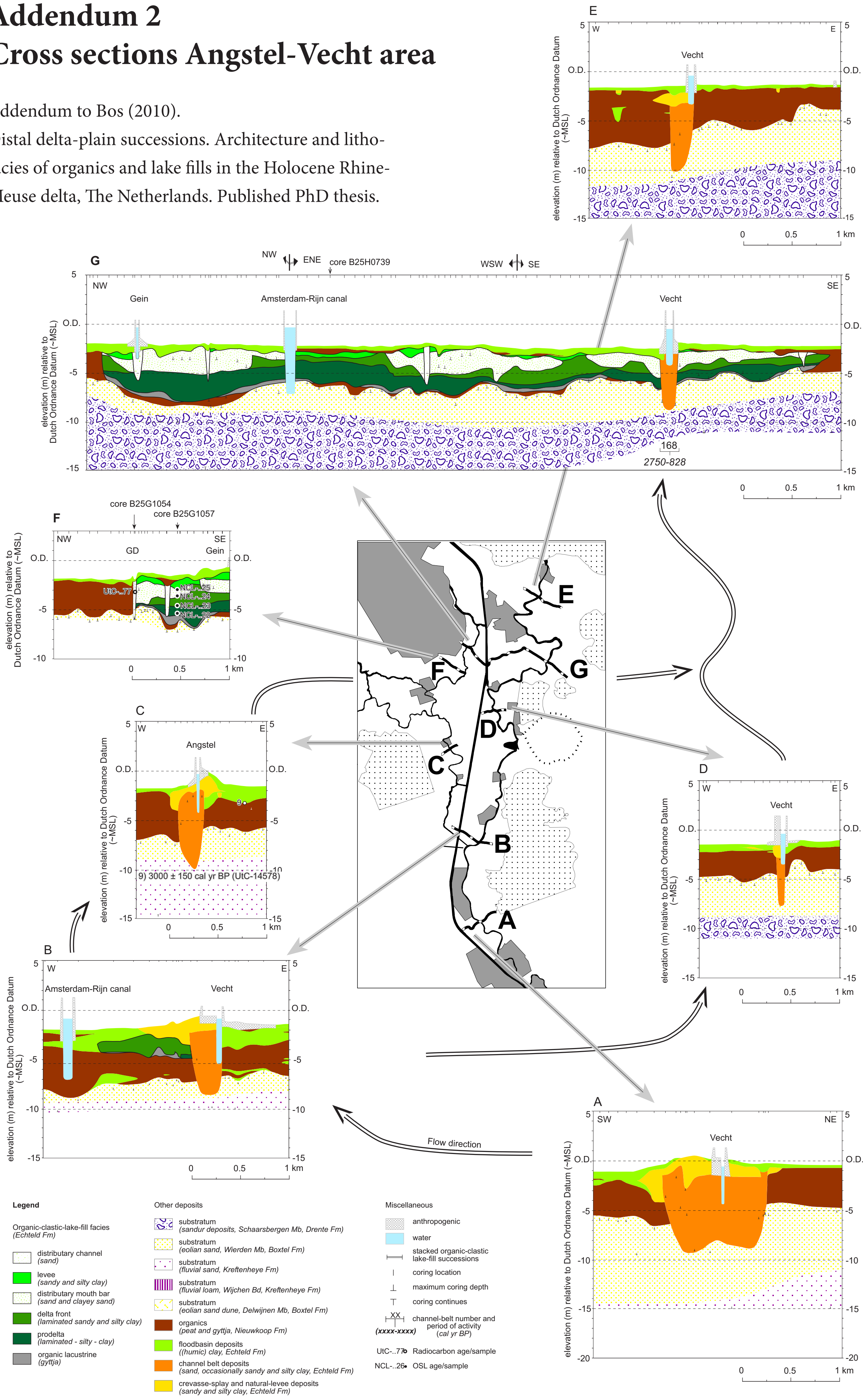
Addendum 1
Geomorphogenetic map of the Angstel-Vecht area

Addendum to Bos (2010).
Distal delta-plain successions. Architecture and lithofacies of organics and lake fills in the Holocene Rhine-Meuse delta, The Netherlands. Published PhD thesis.



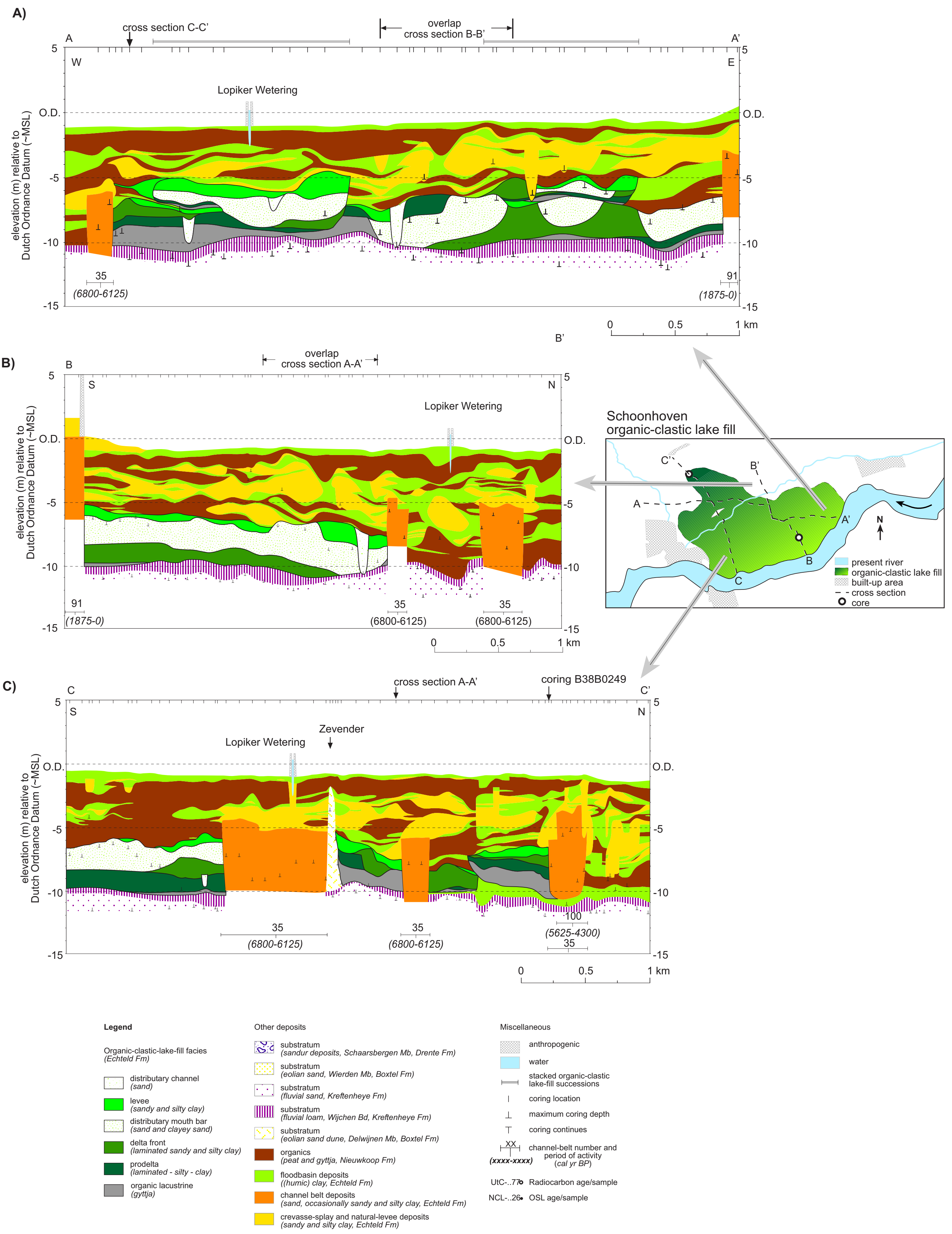
Addendum 2
Cross sections Angstel-Vecht area

Addendum to Bos (2010).
Distal delta-plain successions. Architecture and litho-facies of organics and lake fills in the Holocene Rhine-Meuse delta, The Netherlands. Published PhD thesis.



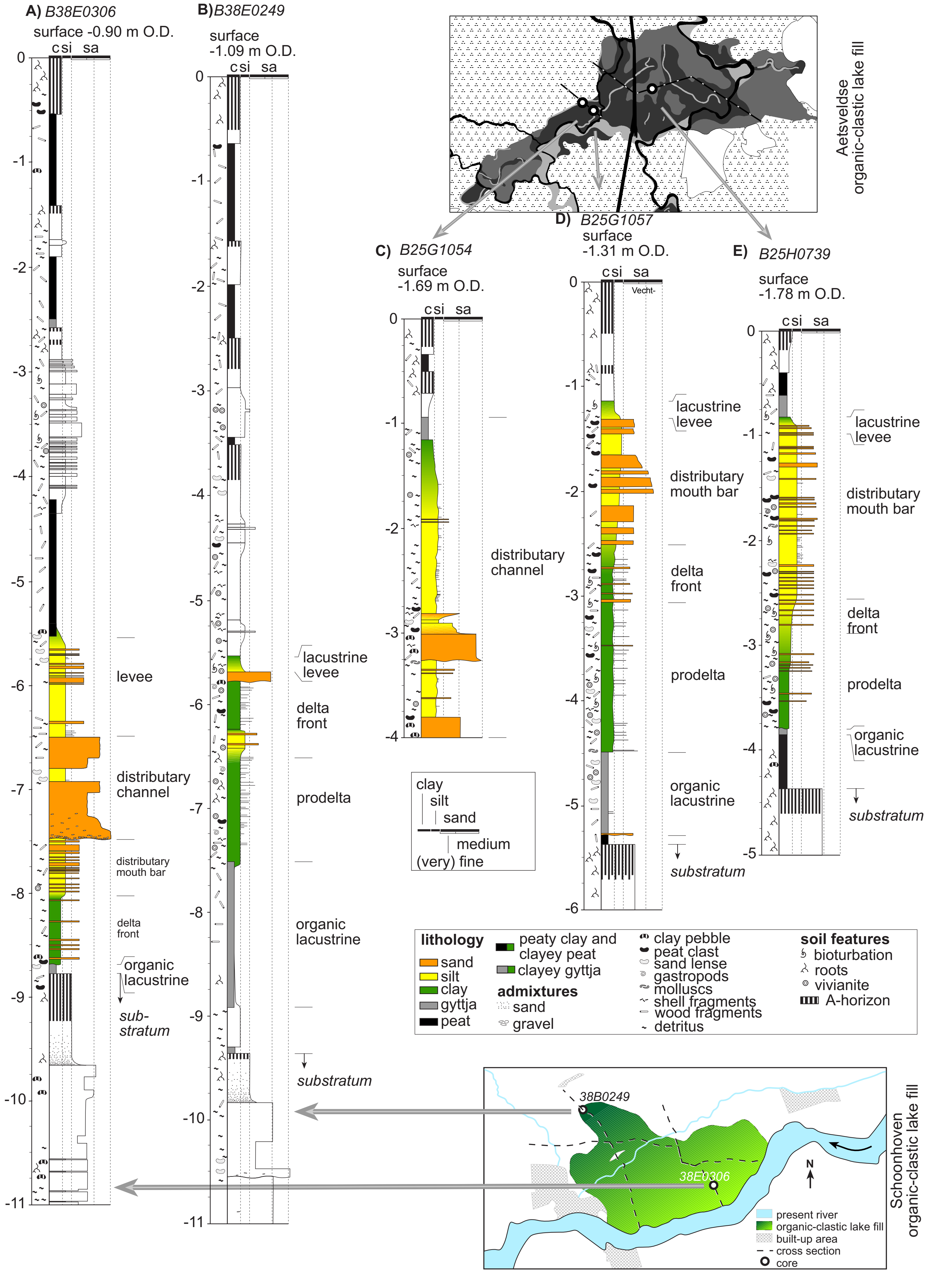
Addendum 3
Cross sections Schoonhoven area

Addendum to Bos (2010).
Distal delta-plain successions. Architecture and lithofacies of organics and lake fills in the Holocene Rhine-Meuse delta, The Netherlands. Published PhD thesis.



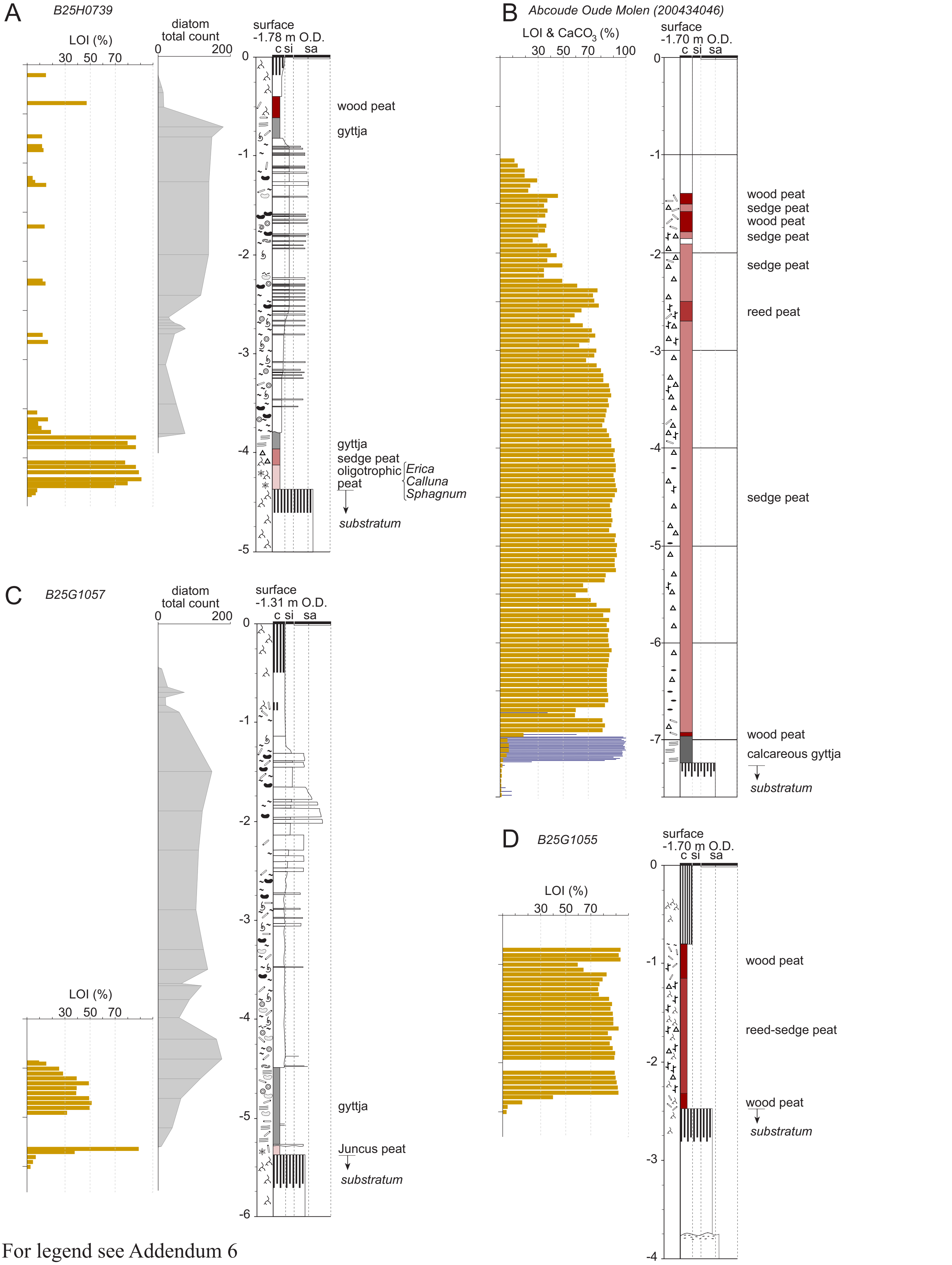
Addendum 4
Sedimentary logs: lake fills

Addendum to Bos (2010).
Distal delta-plain successions. Architecture and lithofacies of organics and lake fills in the Holocene Rhine-Meuse delta, The Netherlands. Published PhD thesis.



Addendum 5
Sedimentary logs: organics
Schoonhoven area

Addendum to Bos (2010).
Distal delta-plain successions. Architecture and litho-facies of organics and lake fills in the Holocene Rhine-Meuse delta, The Netherlands. Published PhD thesis.



For legend see Addendum 6

Addendum 6
Sedimentary logs: organics
Angstel-Vecht area

Addendum to Bos (2010).
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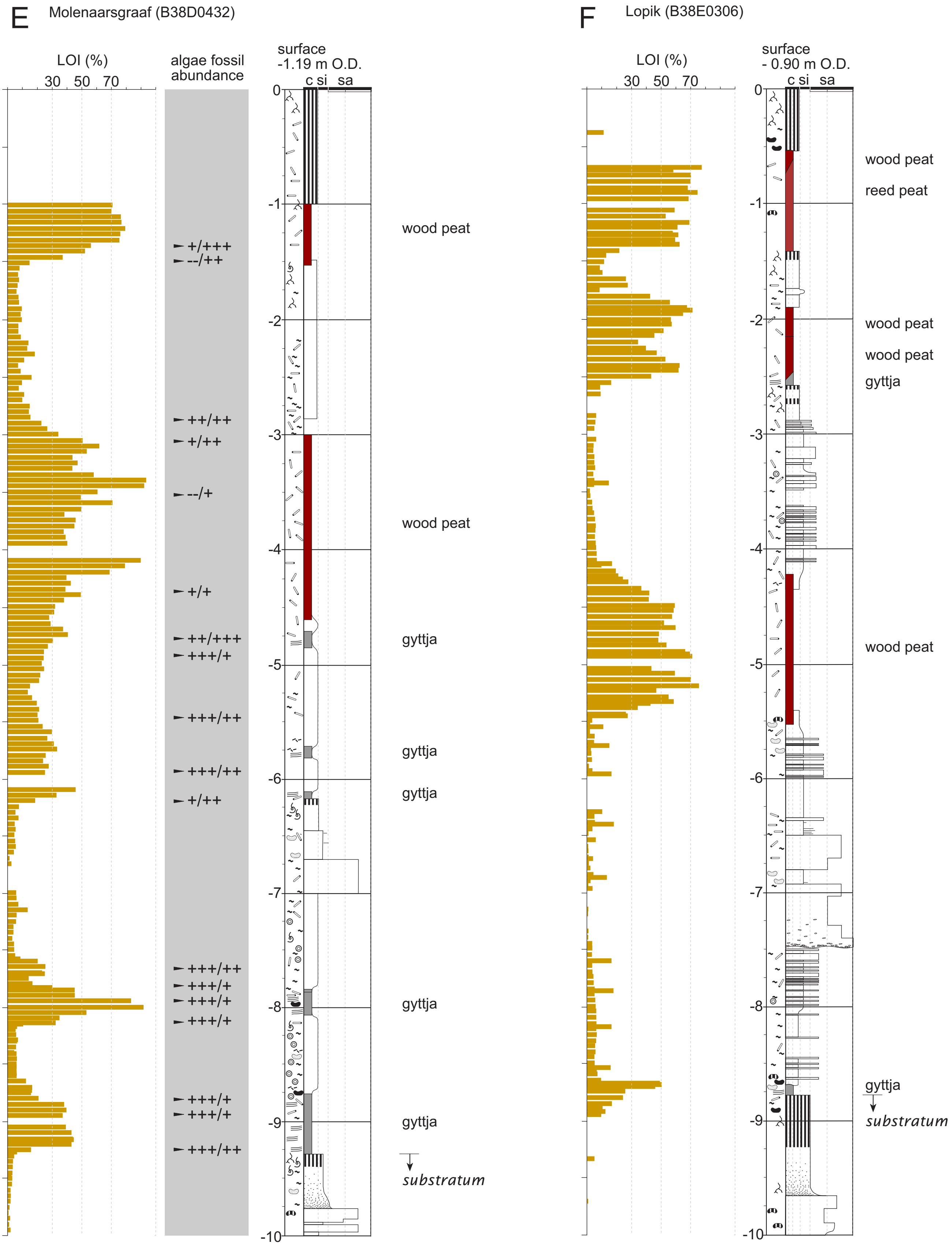

algae fossils abundance
diatoms / chrysophyceae statospores
+++ = very abundant
++ = moderately abundant
+ = scarce
-- = absent
► = sample

admixtures
sand
gravel
clay pebble
peat clast
sand lense
gastropods
molluscs
shell fragments
detritus

botanical content
reed fragments
sedge fragments
Menyanthes seed
wood fragments
oligotrophic pl. fragm.

other features
bioturbation
roots
vivianite
lamination
A-horizon

organic facies
wood peat
reed peat
sedge peat
oligotrophic peat
gyttja

clay (c)
silt (si)
sand (sa)
medium
fine

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Geomorphogenetic map of the Angstel-Vecht area

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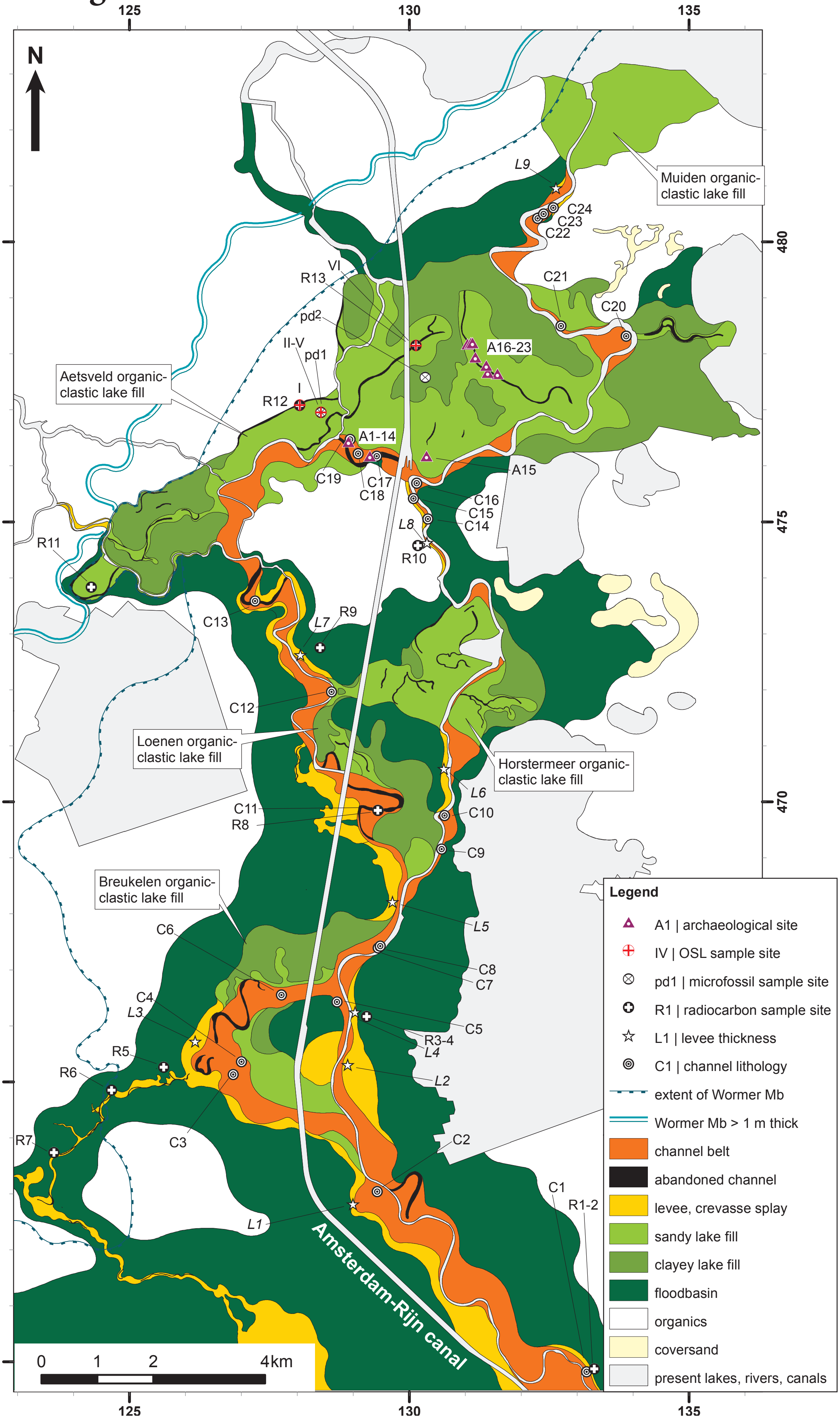

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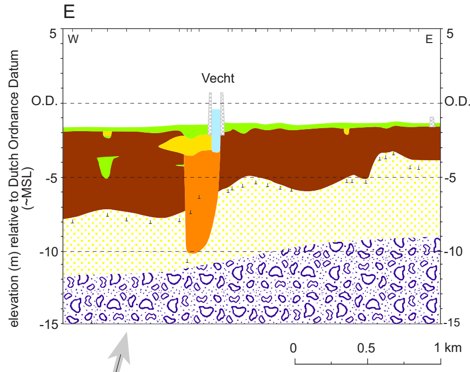
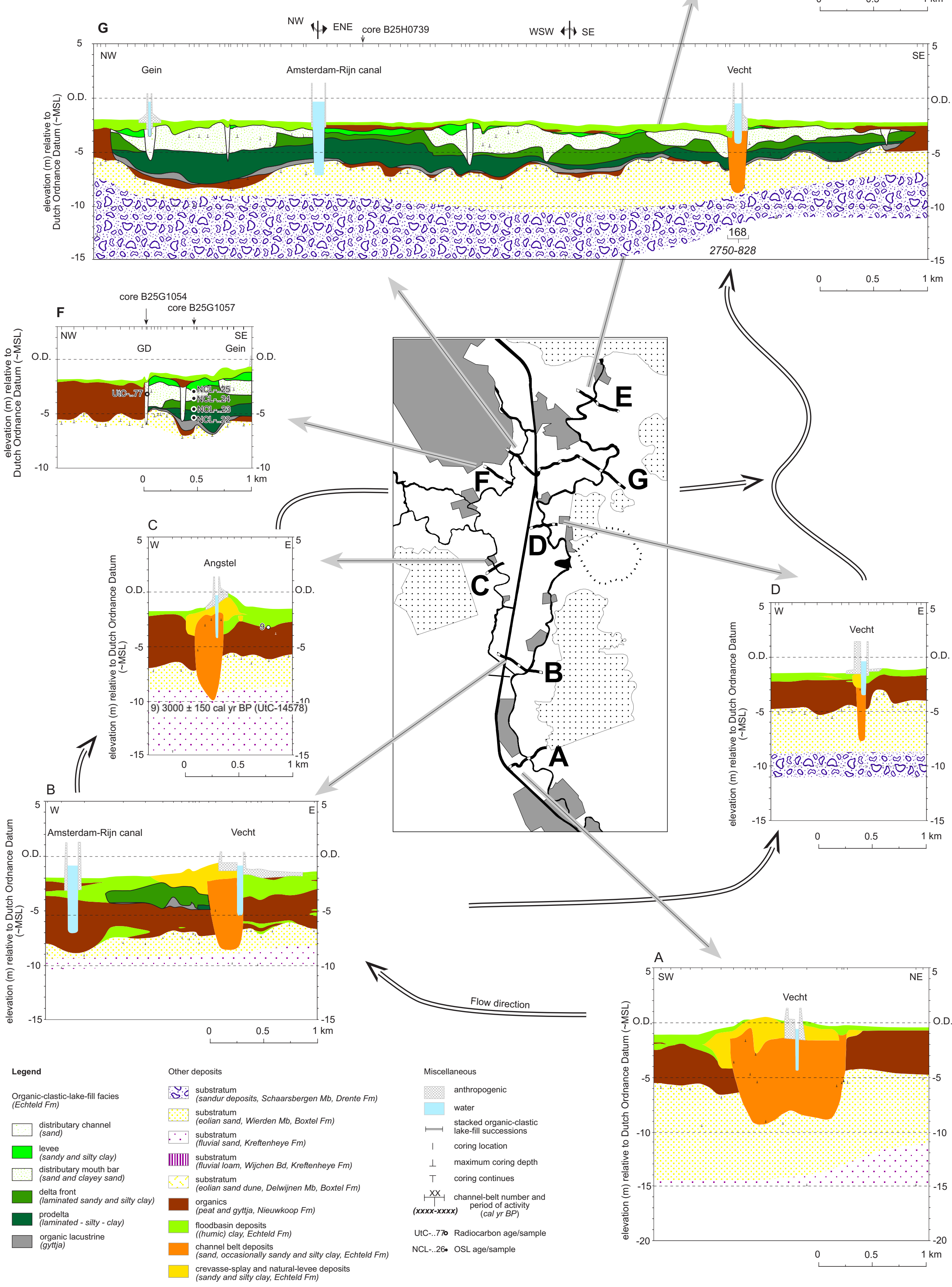
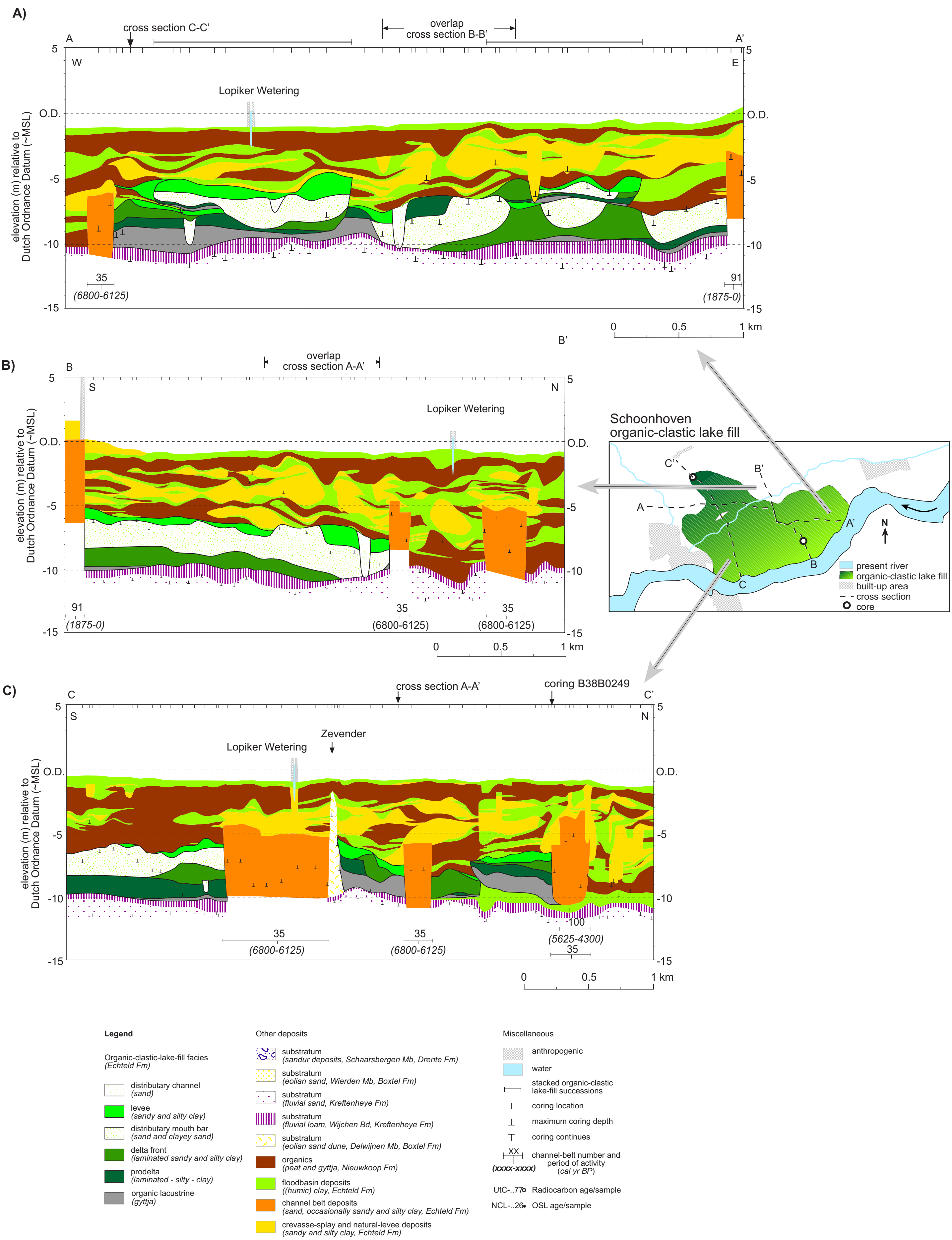




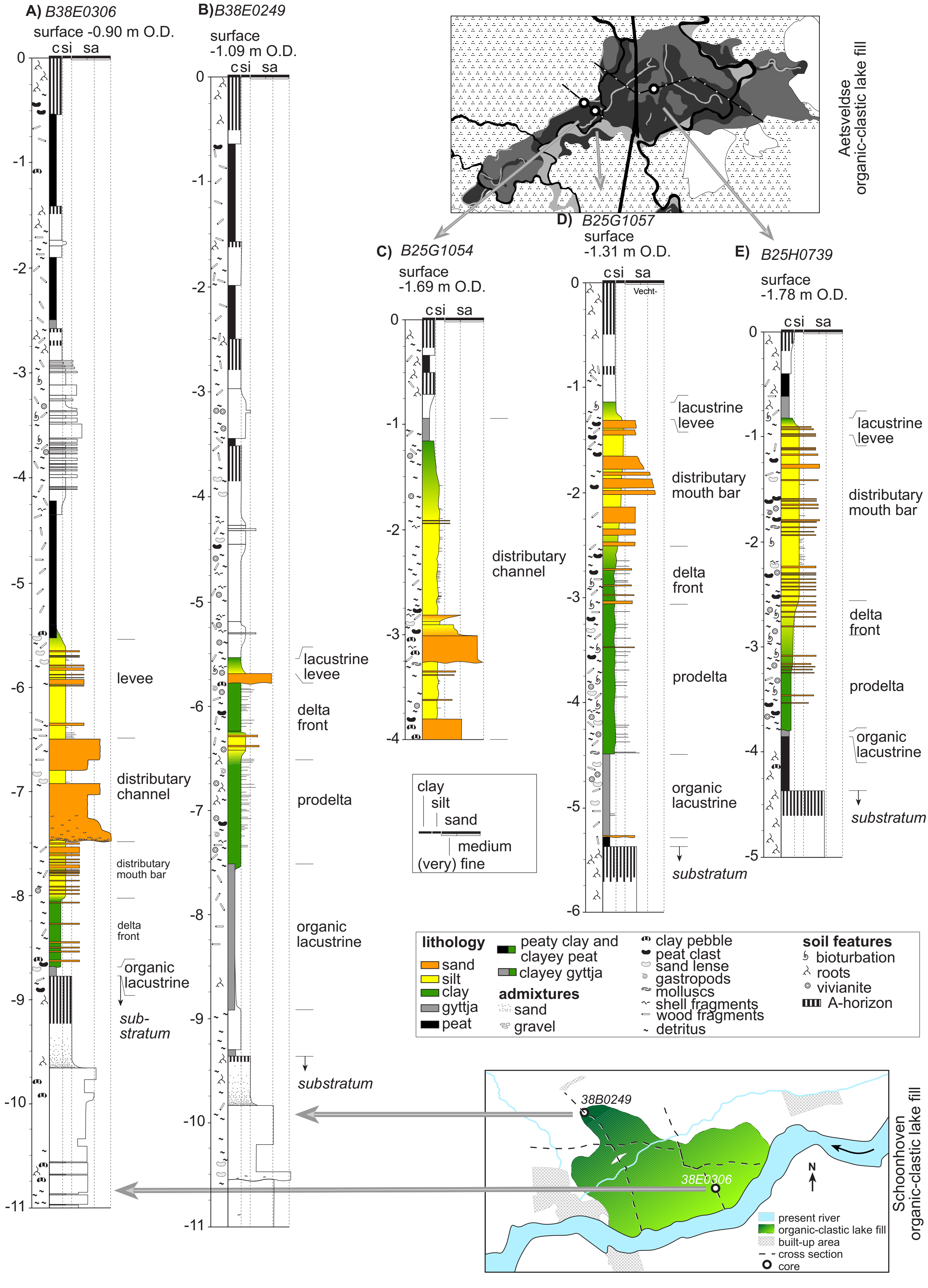
Addendum 3
Cross sections Schoonhoven area

Addendum to Bos (2010).
Distal delta-plain successions. Architecture and litho-facies of organics and lake fills in the Holocene Rhine-Meuse delta, The Netherlands. Published PhD thesis.



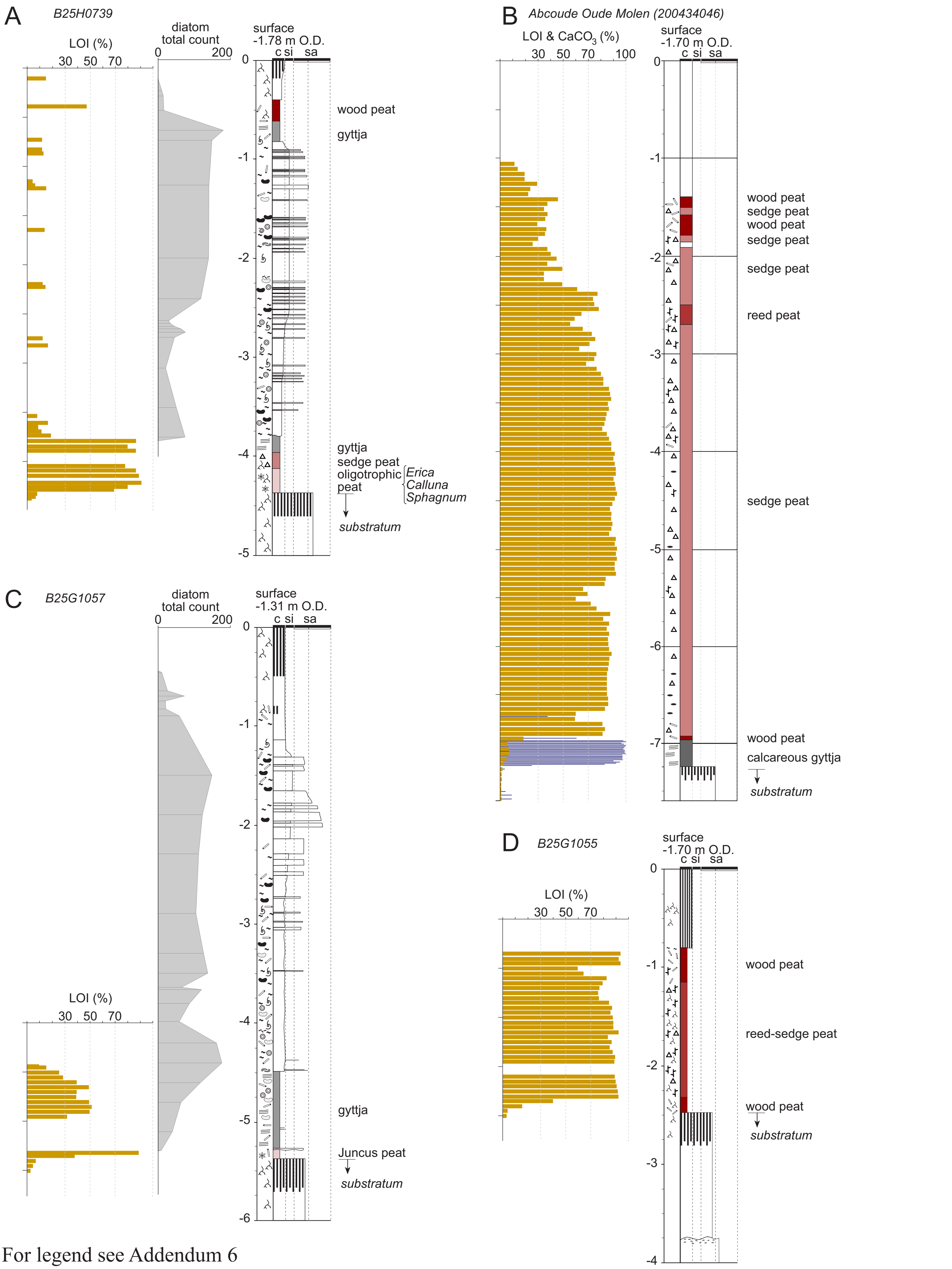
Addendum 4
Sedimentary logs: lake fills

Addendum to Bos (2010).
Distal delta-plain successions. Architecture and litho-facies of organics and lake fills in the Holocene Rhine-Meuse delta, The Netherlands. Published PhD thesis.



Addendum 5
Sedimentary logs: organics Schoonhoven area

Addendum to Bos (2010).
Distal delta-plain successions. Architecture and litho-facies of organics and lake fills in the Holocene Rhine-Meuse delta, The Netherlands. Published PhD thesis.



For legend see Addendum 6

Addendum 6
Sedimentary logs: organics
Angstel-Vecht area

Addendum to Bos (2010).
Distal delta-plain successions. Architecture and lithofacies of organics and lake fills in the Holocene Rhine-Meuse delta, The Netherlands. Published PhD thesis.