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Allogenic forcing of the late Quaternary Rhine–Meuse fluvial record: the interplay of sea-level change, climate change and crustal movements

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ABSTRACT

The Rhine–Meuse system in the west-central Netherlands is a continental-scale fluvial system bordered by an extremely wide continental shelf. Consequently, late Quaternary eustatic sea-level changes have resulted in dramatic shoreline displacements, by as much as 800 km. In addition, changes in climate have been severe, given the latitudinal and palaeogeographic setting of the Rhine–Meuse system. We investigated the relative importance of these allogenic controls on fluvial aggradation and incision during the last two glacial–interglacial cycles. We used optical dating of quartz from $\sim$30 samples in a cross-section perpendicular to the palaeoflow direction, allowing us to correlate periods of aggradation and incision with independent records of sea-level change, climate change and glacio-isostatic crustal movements. We found the long-term aggradation rate to be $\sim$8 cm kyr^{-1}, a value similar to previous estimates of tectonic subsidence rates in the study area. Several excursions from this long-term aggradation trend could be identified for the last glacial–interglacial cycle. Dry climatic conditions with relatively high sediment supply induced aggradation during oxygen-isotope stages (OIS) 4 and 3. Build-up of a glacio-isostatic forebulge during OIS 2 is a likely cause of incision around the Last Glacial Maximum, followed by an aggradation phase during forebulge collapse. Sea-level highstands during OIS 5 have likely resulted in the aggradation of coastal prisms, but only minor, basal estuarine deposits have been preserved because these coastal prisms were prone to erosion during ensuing sea-level falls. Overall, the sedimentary record is dominated by strata formed during time intervals when the study area was completely unaffected by sea-level control, and our evidence shows that the falling-stage systems tract has the highest preservation potential. Our study highlights the importance of considering the complex interplay of both upstream and downstream controls to obtain a comprehensive understanding of the evolution of basin-margin successions.

INTRODUCTION

An insight into the response of rivers to allogenic forcing is important for understanding past and present-day fluvial morphodynamics. This is of relevance for predicting the behaviour of rivers in the context of global environmental change, for interpreting the ancient rock record and for a better understanding of the stratigraphic architecture of resource reservoirs (e.g. Shanley & McCabe, 1994; Blum & Törnqvist, 2000). Late Quaternary fluvial sediments constitute the most promising natural archive for improving the understanding of the response of fluvial systems to external forcing, because (1) the Quaternary has seen high-frequency, high-amplitude changes in climate and eustatic sea level; (2) there are independent, high-resolution records of climate and relative sea-level (RSL) change and (3) detailed chronological frameworks of fluvial deposits are feasible.

Investigations of the response of fluvial systems are often troubled by the lack of chronological control, making the identification of phases of incision and aggradation problematic. This precludes correlation of the fluvial stratigraphic record with independent records of climate and

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RSL change. Studies of Quaternary fluvial successions have traditionally focused on the role of climate in inland contexts, typically in settings with limited long-term preservation potential (e.g. Brakenridge, 1980; Starkel, 1991; Knox, 1995; Vandenberghe, 2003; and many others). In order to assist the interpretation of the ancient rock record, there is a particular need for studies that consider the complex interplay of upstream and downstream controls in basin-margin settings that produce sedimentary successions that can serve as near-modern analogues (cf. Blum & Törnqvist, 2000). The complex interplay and temporal and/or spatial dominance of forcing factors has received surprisingly little attention; notable exceptions include the research on the Colorado River in Texas (Blum, 1993; Blum & Price, 1998) and the Po River in Italy (Amorosi *et al.*, 2004; Amorosi & Colalongo, in press).

In this study, we investigate the response of a continental-scale fluvial system to sea level, climatic and isostatic forcing during the past two glacial–interglacial cycles. The aims of our investigation are (1) to determine the long-term fluvial aggradation rate and link it to tectonic subsidence rate; (2) to identify periods of incision or aggradation and to decipher what forcing factors were responsible and (3) to obtain an insight into the preservation potential of the deposits formed under the influence of the different forcing factors.

The late Quaternary fluvial record of the Rhine–Meuse system in the west-central Netherlands (Fig. 1) is an excellent place to investigate the above-mentioned issues because (1) tectonic subsidence in this area facilitates preservation of deposits; (2) the location of the study area just landward of the high-stand shoreline ensures that the fluvial system was affected by eustatic sea-level changes; (3) the catchment of the Rhine–Meuse system (northwest Europe) experienced dramatic climatic changes during the last two glacial–interglacial cycles; and (4) the study area is bordered by an extremely wide continental shelf, leading to dramatic shifts in shoreline position during eustatic sea-level changes. Thus, we expect that the relative role of forcing factors in this area changed considerably during glacial–interglacial cycles.

This study is part of our continuing investigation to understand the late Quaternary record of the Rhine–Meuse system in the west-central Netherlands. Preliminary results of our studies, mainly based on a single core that is also used for this study, were discussed by Törnqvist *et al.* (2000). Since then we recovered two more cores from the late Quaternary deposits and investigated these in detail. Based on these new data combined with information stored in the DINO database of the Netherlands Institute of Applied Geoscience TNO, Busschers *et al.* (in press) discuss the sedimentology of the deposits and the palaeogeographical evolution of the area. In the present publication we build on that paper to unravel the response of the Rhine–Meuse system to allogenic forcing. The same data set was previously used by Törnqvist *et al.* (2003) to determine the timing of the last sequence-boundary formation in our study area.

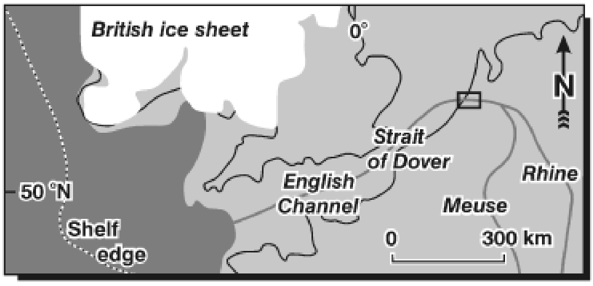

Fig. 1. Location of the research area (box) and extension of the Rhine–Meuse system (dark grey lines) across the continental shelf during relative sea-level lowstand. The position of the shoreline (land-sea boundary indicated by light/dark grey transition) relative to the shelf edge is shown as well as the maximum extent of the British ice sheet at 22 ka BP (both after Lambeck, 1995).

THE RHINE–MEUSE SYSTEM

Our study area is located in the west-central Netherlands near the present shoreline (Fig. 1). It is situated in the southern part of the subsiding North Sea Basin. According to earlier work, the hinge line of the Rhine system is located about 125-km upstream of our study area (near the Dutch–German border; Zonneveld, 1963; Törnqvist, 1995). Palaeogeographic reconstructions for the lower Rhine–Meuse system during the Quaternary have been provided elsewhere (e.g. Zagwijn, 1974, 1989; Gibbard, 1988, 1995; Bridgland & D'Olier, 1995; Bridgland & Gibbard, 1997; Busschers *et al.*, in press). Here we provide a brief summary, derived from that by Törnqvist *et al.* (2000).

During oxygen isotope stage (OIS) 6 the Fennoscandian Ice Sheet covered the northern half of the Netherlands and forced the Rivers Rhine and Meuse to take a route roughly similar to their present course. After melting of this ice sheet, the River Rhine reoccupied a more northerly route, draining into glacially scoured basins. Its southerly course (through our study area) was later reoccupied, probably during the first part of the last glacial (Van de Meene & Zagwijn, 1978; De Gans & Van Gijssel, 1996), but the exact timing of changes in drainage direction has yet to be provided. The River Meuse approximately followed its present course throughout the late Quaternary (e.g. Zagwijn, 1974; Van de Meene & Zagwijn, 1978).

It is generally assumed that drainage during RSL lowstands has been through the Strait of Dover since the Elsterian/Anglian glaciation (e.g. Bridgland & Gibbard, 1997). During RSL lowstands the system crossed the slightly uplifting Weald–Artois Anticline at the Strait of Dover (e.g. Lagarde *et al.*, 2003) before reaching the shoreline. Because of an extremely wide continental shelf, the low-stand shoreline was situated $\sim$800-km downstream from the present shoreline (Lambeck, 1995; Fig. 1).

During the Holocene (OIS 1), a coastal prism, associated with RSL rise and highstand, was built in the western Netherlands (Pons, 1957; Zonneveld, 1957). A similar feature most likely developed during the RSL highstands of OIS 5 (Törnqvist *et al.*, 2000). However, since probably only the River Meuse delivered sediment to the west-central Netherlands at that time, coastal prisms during OIS 5 were presumably smaller than their Holocene counterpart.

METHODS

This paper is based on a detailed study of five cores (up to 60 m in depth; only the upper 40 m is used for this paper) in an S–N transect, roughly parallel to the present Dutch shoreline and perpendicular to the drainage direction of the Rhine–Meuse system (Fig. 2). The transect forms a cross-section through most of the Rhine–Meuse palaeovalley fill in the west-central Netherlands, and penetrates the full late Quaternary sedimentary record in this area.

A mechanized bailer-drilling unit (Oele *et al.*, 1983) of the Netherlands Institute of Applied Geoscience TNO was used for coring; oriented, undisturbed cores with a 10-cm diameter were obtained. The cores were split under subdued red-light conditions; one half was brought into the light, photographed and used to produce lacquer peels. A macroscopic description was made of grain size, gravel content, sedimentary structures, biogenic content and the nature of facies transitions.

From the other half of the split cores, samples were taken for optical dating. This dating method makes use of the optically stimulated luminescence (OSL) signal of quartz or feldspar minerals to determine the burial age of sand or silt-sized grains (see reviews by Aitken, 1998; Wallinga, 2002a). Quartz optical dating results have been shown to be in good agreement with independent ages (e.g. Murray & Olley, 2002). Limited light exposure of the grains during fluvial transport might lead to improper resetting of the OSL signal and consequently to an overestimation of the burial age. However, such offsets have been shown to be minor for large river systems (Wallinga, 2002a). The validity of quartz optical dating results for Rhine–Meuse deposits has been demonstrated by comparison with independent age control up to 13 ka (Wallinga *et al.*, 2001).

The chronology of the deposits was investigated using optical dating of samples from three of the cores (referred to in the text as Wassenaar, Leidschendam and Delft). We applied optical dating to the quartz fraction of 28 samples. All samples were water washed and treated with 10% HCl and 30% H_2O_2 to remove carbonates and organic material. After drying, the samples were sieved and subsequently density separated using an aqueous solution of sodium polytungstate to extract the potassium-rich feldspar fraction lighter than 2.58 g cm^{-3}. The denser fraction was treated with concentrated (40%) hydrofluoric acid for 40 min to obtain clean quartz separates and to etch away the outer 10 μm of the quartz grains.

Measurements were made on an automated Risø TL/OSL reader, using an internal ^{90}Sr/^{90}Y β-source (Bøtter-Jensen *et al.*, 2000). The sample grains were mounted on stainless-steel discs using silicone spray. Blue light-emitting diodes (LEDs) were used for stimulation of the quartz (20 s at 110 °C (Delft and Wassenaar samples) or 40 s at 125 °C (Leidschendam samples)) and the resulting luminescence signal was detected through 9 mm of Schott U-340 filters (detection window 250–390 nm). The improved Single-Aliquot Regenerative-dose (SAR) protocol (Murray & Wintle, 2000) was used for estimation of the equivalent dose. Five different preheat temperatures in the range 120–280 °C (all for 10 s) were used for the Delft and Wassenaar samples, with three aliquots measured for each temperature. The flat part of the preheat plateau (see Aitken, 1998) was used for the equivalent-dose determination for these samples (Table 1). The test–dose response was measured after heating to 120 °C. We performed a dose recovery test (cf. Wallinga *et al.*, 2000) on the Delft and Wassenaar samples to confirm that the SAR procedure accurately determines a laboratory dose given to the samples prior to any heating (dose recovery ratio 1.02 ± 0.01, for a 10 s 200 °C preheat). For the Leidschendam samples, a 10-s preheat at 260 °C was routinely used, in combination with heating of the test dose to 160 °C. Preheat plateaux obtained on a number of samples from the Leidschendam core confirmed that equivalent doses were independent of preheat temperature around 260 °C (Wallinga *et al.*, 2001). The purity of the quartz samples was tested by stimulation with an infrared laser diode (emitting at 830 nm) at 50 °C (Aitken, 1998).

The natural dose rate was estimated using high-resolution gamma spectrometry (Murray *et al.*, 1987) on combined bulk samples taken just above and just below the sample used for equivalent-dose determination (results in Table 1). All our samples have been saturated with water throughout their lifetimes, which diminishes the dose rate (Aitken, 1985). The low-dose rates for our samples allow us to use quartz optical dating for a relatively long time range (~150–250 kyr).

RESULTS

Simplified sedimentary logs of the cores are presented in Fig. 2; a full account of the sedimentology and lithostratigraphy is presented elsewhere (Busschers *et al.*, in press) and will not be repeated here. Broadly, the succession can be divided into three informal lithostratigraphical units. The lowermost, fine-grained unit (denoted as pre-OIS 6 mud in Fig. 2) is of lesser importance to this study and is regarded as substratum. Infrared stimulated luminescence dating of feldspar indicates a minimum age of 350 ka for this unit (Törnqvist *et al.*, 2000). A Matuyama age (> 778 ka; Tauxe *et al.*, 1996) is inferred for this unit based on correlation with deposits of reversed magnetic polarity (Zagwijn *et al.*, 1971; Kasse, 1996).

The middle unit is of particular interest to the present study because it has long been considered to represent fluvial deposition during the last glacial plus one or more previous glacial periods (e.g. Doppert *et al.*, 1975;

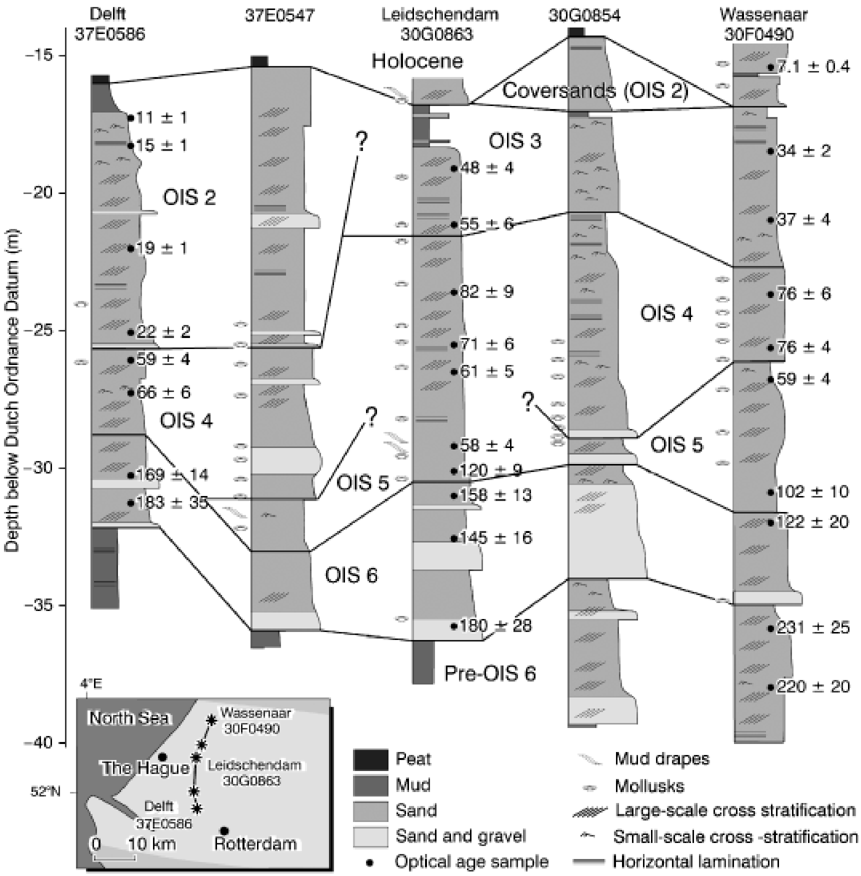

Fig. 2. Cross-section perpendicular to the late Quaternary Rhine–Meuse palaeovalley, located near the present Netherlands' coast (for location see inset; study area indicated in Fig. 1). Optical ages are labelled with 1σ confidence intervals; isochrons are based on a combination of optical ages, lithologic and sedimentologic features. Aeolian coversand deposits were encountered in core 30G0854 (Busschers *et al.*, in press); these are ascribed to OIS 2 based on earlier studies (e.g. Van der Hammen *et al.*, 1967).

Verbraeck, 1984; Bosch & Kok, 1994). This unit (OIS 6–OIS 2 and pre-OIS 6 sand and gravel in the two northernmost cores in Fig. 2) extends to a depth of 32 m-OD (Dutch Ordnance Datum) in the southern part of the cross-section and > 55-m OD in the Wassenaar core (Busschers *et al.*, in press). The sedimentology and geochronology of the middle unit are discussed below.

The uppermost unit (above 13–16 m-OD, denoted as Holocene in Fig. 2) consists predominantly of fine-grained coastal-plain deposits. These strata were formed during more or less continuous aggradation since about 8 ka in response to Holocene RSL rise. The depositional history of this unit has been discussed in detail elsewhere (e.g. Beets & Van der Spek, 2000; and references therein). The basal part of this unit is formed by peat, although estuarine (Leidschendam) or fluvial (Wassenaar) channels have locally eroded the basal peat. An optical age of ∼7 ka was obtained on sample Wassenaar I, taken from the basal part of a fluvial channel belt, in agreement with the ^{14}C age obtained on reworked organic material (Table 2).

The middle lithostratigraphical unit (OIS 6–OIS 2 and pre-OIS 6 sand and gravel in Fig. 2) consists primarily of large-scale cross-stratified, medium to coarse, sometimes gravelly sands, interpreted as fluvial Rhine and/or Meuse deposits. It is incised deeply into underlying strata in the Wassenaar core (> 55 m-OD; not shown here, see Busschers *et al.*, in press). In the other cores, the base of this unit is encountered between 34 and 39 m-OD, where it can easily be identified as a distinct, erosional contact with the underlying substratum. A coarse-grained lag deposit normally overlies the contact; we correlate the lag deposit with that encountered in the Wassenaar core at 35 m-OD. The geochronological data for the middle unit are presented in Table 1 and Fig. 2. For our interpretation, we use rounded age ranges derived from optical dating based on 1σ confidence intervals; we relate formation of the deposits to oxygen-isotope stages (using OIS age boundaries as indicated in Fig. 3).

Optical ages of samples Wassenaar IX and X (200–260 ka) suggest that the lower part of the Wassenaar core was deposited prior to the maximum ice extent of OIS 6. Extreme caution is, however, needed in interpreting these ages. The validity of quartz optical ages in this range has so far not been demonstrated and small errors in the sensitivity-change correction can result in large errors in the age estimate in this high-dose range (Murray *et al.*, 2002). Further research is therefore needed to allow inferences about the timing of these deposits. The basal part of the middle unit in the other cores is relatively coarse grained and rarely contains shell remains. Optical ages (Delft VII and VIII: 150–220 ka; Leidschendam VIII–X: 130–210 ka; Wassenaar VIII: 100–140 ka) suggest deposition during OIS 6. The resolution of optical ages for this time interval does not allow inferences on deposition in relation to RSL changes or the timing of maximum ice extent.

The transition to the overlying package (OIS 5 and 4; Fig. 2) is characterized in all but the Delft core by the occurrence of reworked marine shell rubble, mud pebbles

Table 1. Quartz optical dating results.

Sample	Depth (m)	Depth (m-OD)	Grain size (μm)	Radionuclide concentration* (Bq kg^{-1}) ^{238}U	^{226}Ra	^{232}Th	^{40}K	Dose rate*,† (Gy ka^{-1})	Equivalent dose (Gy)	Preheat‡ (°C)	Optical age (ka)
Delft I	14.75	17.29	90–180	33 ± 5	35.8 ± 0.5	39.1 ± 0.5	483 ± 10	2.19 ± 0.07	24.6 ± 1.0	160–280	11.2 ± 0.6
Delft II	15.75	18.29	180–212	13 ± 2	8.8 ± 0.2	9.1 ± 0.2	316 ± 5	1.05 ± 0.03	15.5 ± 0.9	160–280	14.8 ± 0.9
Delft III	19.5	22.04	212–250	8 ± 2	8.8 ± 0.2	10.1 ± 0.2	318 ± 6	1.17 ± 0.03	22 ± 2	160–240	18.5 ± 1.4
Delft IV	22.55	25.09	212–250	11 ± 4	9.1 ± 0.3	9.9 ± 0.3	357 ± 7	1.12 ± 0.03	25 ± 2	160–280	22 ± 2
Delft V	23.55	26.09	212–250	9 ± 4	6.9 ± 0.3	7.7 ± 0.2	212 ± 5	0.74 ± 0.02	43 ± 3	160–280	59 ± 4
Delft VI	24.75	27.29	212–250	5 ± 3	5.8 ± 0.2	6.3 ± 0.2	177 ± 5	0.62 ± 0.02	41 ± 3	160–280	66 ± 6
Delft VII	27.75	30.29	250–300	9 ± 2	5.6 ± 0.2	6.6 ± 0.2	212 ± 4	0.69 ± 0.02	117 ± 9	160–280	169 ± 14
Delft VIII	28.75	31.29	212–250	14 ± 4	7.5 ± 0.3	8.3 ± 0.2	249 ± 6	0.84 ± 0.03	152 ± 29	160–280	183 ± 35
Leidschendam I	14.85	19.12	180–212	10 ± 3	9.5 ± 0.5	9.6 ± 0.4	297 ± 13	1.11 ± 0.06	54 ± 3	260	48 ± 4
Leidschendam II	16.90	21.17	180–212	6 ± 3	9.5 ± 0.3	10.8 ± 0.3	249 ± 7	0.99 ± 0.05	54 ± 5	260	55 ± 6
Leidschendam III	19.35	23.62	180–212	6 ± 2	9.5 ± 0.4	7.5 ± 0.3	200 ± 10	0.78 ± 0.05	64 ± 6	260	82 ± 9
Leidschendam IV	21.25	25.52	180–212	12 ± 3	12.8 ± 0.3	14.1 ± 0.3	341 ± 7	1.29 ± 0.06	92 ± 7	260	71 ± 6
Leidschendam V	22.25	26.52	180–212	6 ± 3	6.9 ± 0.5	8.2 ± 0.4	253 ± 13	0.92 ± 0.05	56 ± 3	260	61 ± 5
Leidschendam VI	24.95	29.22	180–212	8 ± 3	9.8 ± 0.2	10.7 ± 0.2	314 ± 7	1.13 ± 0.05	66 ± 3	260	58 ± 4
Leidschendam VII	25.85	30.12	180–250	12 ± 3	6.8 ± 0.3	6.7 ± 0.2	237 ± 6	0.86 ± 0.04	104 ± 6	260	120 ± 9
Leidschendam VIII	26.75	31.02	180–250	10 ± 2	6.5 ± 0.4	8.2 ± 0.4	201 ± 10	0.79 ± 0.05	124 ± 7	260	158 ± 13
Leidschendam IX	28.30	32.57	180–212	11 ± 2	5.5 ± 0.3	7.4 ± 0.2	190 ± 7	0.74 ± 0.04	107 ± 10	260	145 ± 16
Leidschendam X	31.50	35.77	180–212	7 ± 3	6.9 ± 0.3	8.2 ± 0.2	240 ± 6	0.87 ± 0.04	156 ± 23	260	180 ± 28
Wassenaar I	15.25	15.45	180–212	11 ± 4	7.3 ± 0.3	8.3 ± 0.3	329 ± 7	1.05 ± 0.03	7.4 ± 0.4	160–280	7.1 ± 0.4
Wassenaar II	18.30	18.50	180–212	9 ± 4	9.8 ± 0.3	12.5 ± 0.3	257 ± 6	1.04 ± 0.03	35 ± 2	160–280	34 ± 2
Wassenaar III	20.80	21.00	180–212	13 ± 4	11.1 ± 0.3	11.5 ± 0.3	323 ± 8	1.10 ± 0.03	41 ± 4	160–280	37 ± 4
Wassenaar IV	23.50	23.70	250–300	6 ± 3	5.1 ± 0.2	5.4 ± 0.2	148 ± 4	0.53 ± 0.02	40 ± 3	160–280	76 ± 6
Wassenaar V	25.45	25.65	212–250	5 ± 3	5.0 ± 0.2	5.9 ± 0.2	180 ± 5	0.61 ± 0.02	46 ± 2	160–280	76 ± 4
Wassenaar VI	26.60	26.80	180–212	11 ± 3	8.6 ± 0.2	9.7 ± 0.2	247 ± 5	0.87 ± 0.02	51 ± 4	160–280	59 ± 4
Wassenaar VII	30.70	30.90	180–212	13 ± 4	7.7 ± 0.3	8.9 ± 0.3	294 ± 7	0.95 ± 0.03	96 ± 9	160–240	102 ± 10
Wassenaar VIII	31.80	32.00	250–300	9 ± 3	5.9 ± 0.2	7.0 ± 0.2	176 ± 5	0.61 ± 0.02	75 ± 12	160–240	122 ± 20
Wassenaar IX	35.65	35.85	250–300	9 ± 4	5.1 ± 0.2	6.5 ± 0.2	184 ± 5	0.61 ± 0.02	141 ± 14	160–280	231 ± 25
Wassenaar X	37.80	38.00	212–250	13 ± 4	5.2 ± 0.2	6.6 ± 0.2	236 ± 6	0.74 ± 0.02	164 ± 14	160–280	220 ± 20

*Spectral data from high-resolution gamma spectroscopy converted to activity concentrations and infinite matrix dose rates using the conversion data given by Olley *et al.* (1996).
†The natural dose rate was calculated from the infinite matrix dose rate using attenuation factors given by Mejdahl (1979), and includes a contribution from cosmic rays (Prescott & Hutton, 1994). All dose rates calculated for a water content of 20 ± 2% (based on a porosity of 34 ± 3%; Weerts, 1996) using attenuation factors given by Zimmerman (1971).
‡Range of preheat temperatures used for equivalent-dose determination. See main text for explanation.

Table 2. Radiocarbon ages.

Sample	Lab number	Depth (m)	Depth (m-OD)	Analysed fraction	^{14}C age	Calculated age range* (BP, 2σ)
Delft	UtC-10757	13.8	16.3	Terrestrial macrofossils (*Scirpus* nuts)	8220 ± 60	9422–9013
Wassenaar	UtC-10758	16.0	16.2	Terrestrial macrofossils and charcoal	6889 ± 49	7816–7612

*Calibrated using the calibration data set of Stuiver *et al.* (1998) and the Groningen radiocarbon calibration program (Van der Plicht, 1993).

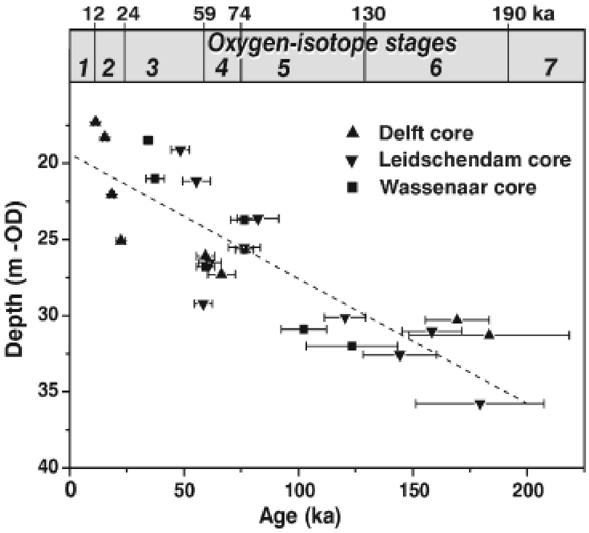

Fig. 3. Age–depth plot of samples in this study optically dated to the period of 200–10 ka. The slope of the linear trend line (0.082 ± 0.011 m kyr^{-1}; n = 25; r^2 = 0.72) indicates the average aggradation rate over the period 200–10 kyr. Also shown are the OIS age boundaries (cf. Pisias *et al.*, 1984; Martinson *et al.*, 1987).

and reworked organic matter. Faunal analysis of shell remains in the Leidschendam core (Törnqvist *et al.*, 2000) revealed several taxa that are characteristic of OIS 5e. Although most shells are not intact, they are relatively well preserved. In the Leidschendam core, a number of inclined mud drapes (around 29-m OD) were encountered; palaeoecological analysis (Törnqvist *et al.*, 2000) showed that these contain diatoms typical of an estuarine environment, and a pollen spectrum suggesting an Early Weichselian (OIS 5a–5d) age. Törnqvist *et al.* (2000) inferred an OIS 5a (~80 ka) age for these deposits because estuarine deposits are very unlikely to have formed during OIS 4-2 when RSL was much lower and the shoreline had retreated to a position hundreds of kilometres from our study area. The optical age obtained on sample Leidschendam VI (58 ± 4 ka) from this level is likely an underestimate (cf. Törnqvist *et al.*, 2000, 2003; Wallinga *et al.*, 2001). Moreover, older optical ages were obtained for overlying sediments, suggesting that the age obtained on sample Leidschendam VI is erroneously young.

The OIS 5/4 transition is characterized by a channel lag in combination with a sudden increase in the concentration of marine shell rubble. The transition is distinct in the two northerly cores, but cryptic in the Leidschendam core; it may be represented by a minor gravel lag at 25.2 m-OD but this correlation remains speculative. The OIS 4 deposits overlying the channel lag are interpreted as fluvial channel deposits containing reworked marine shells from underlying OIS 5 estuarine strata (Bennema & Pons, 1952; Bosch & Kok, 1994; Törnqvist *et al.*, 2000; Busschers *et al.*, in press). We interpret the great amount of shell rubble in basal OIS 4 deposits as an indication that a large body of shell-bearing marine or estuarine deposits was eroded at the time of formation of these strata. This is in agreement with the previously inferred removal of the OIS 5a coastal prism following RSL fall at the OIS 5/4 transition (cf. Törnqvist *et al.*, 2000, 2003). In the Delft core only a few shells are present, possibly because transgression proceeded less far inland in the southern part of our transect during OIS 5.

We regard it highly unlikely that the section we interpret as OIS 4 deposits was formed during the OIS 5a highstand because the deposits lack any direct evidence of formation during a period of high RSL. Moreover, optical ages (samples Delft V and VI: 55–70 ka; Leidschendam III: 75–90 ka; and Wassenaar IV and V: 70–80 ka) are in agreement with deposition during OIS 4 or the OIS 5/4 transitions.

In the Delft core, the transition from OIS 6 deposits (samples Delft VII and VIII: 150–220 ka) to OIS 4 deposits (samples Delft V and VI: 55–70 ka) is readily identified by optical dating, but the unconformity is sedimentologically indistinct (cf. Miall & Arush, 2001). In contrast, a clear erosional boundary is found between sediments deposited during OIS 4 and those deposited during OIS 2 (samples Delft I–IV: 10–25 ka). In sample Delft I (11.2 ± 0.6 ka), small amounts of pumice were encountered, likely from the Laacher See volcanic eruption that took place around 13 ka (Friedrich *et al.*, 1999).

Transitions are less clear in the Leidschendam core (cf. Törnqvist *et al.*, 2000). In spite of the optical ages and the detailed sedimentological analysis, we are not able to pinpoint unambiguously the transition from estuarine to fluvial deposits. Sample Leidschendam VII (120 ± 9 ka) indicates an OIS 5 age and estuarine mud drapes at 29 m-OD point to an OIS 5a age for sample Leidschendam VI (Törnqvist *et al.*, 2000). Mud drapes are lacking above this level but the position of the OIS 5/4 transition remains uncertain; we tentatively suggest that the transition should be placed at 25.2 m-OD where an indistinct lag deposit overlies deposits with planar bedding. Optical ages above this lag (samples Leidschendam I–III: 45–90 ka) are mostly consistent with deposition after OIS 5. We therefore infer that these deposits are fluvial and were formed during OIS 4 and 3.

For the Wassenaar core, sample VII (102 ± 10 ka) indicates an OIS 5 age. Sample VI was assigned to OIS 5 based on its stratigraphic position below the channel lag interpreted as the OIS 5/4 transition. Positioning in OIS 5 implies that the optical age obtained on this sample (59 ± 4 ka) underestimates the depositional age. Age underestimation is corroborated by the older ages obtained on overlying samples. These (Wassenaar IV and V: 70–80 ka) likely represent OIS 4 or the OIS 5/4 transition. The possibility of slight optical age overestimation because of poor bleaching cannot be ruled out for these samples. Using small aliquots, slightly skewed dose distributions were obtained (Wallinga, 2001), which may be an indication that not all grains had their OSL signal completely reset prior to deposition (Olley *et al.*, 1999; Wallinga, 2002b). However, offsets of more than a few thousand years are unlikely, considering the small offsets encountered in modern and known-age fluvial samples in the Rhine–Meuse system and elsewhere (reviewed in Wallinga, 2002a). The upper part of the middle unit in the Wassenaar core (samples II and III) is clearly younger (30–40 ka) and was deposited during OIS 3.

It is striking that similar age reversals occur around the OIS 5/4 transition in the Leidschendam and Wassenaar cores. These age reversals were also found for measurements that used the infrared stimulated luminescence signal from feldspar (Wallinga *et al.*, 2001), suggesting that the problems are caused by errors in the estimation of the external dose rate (shared by the two methods) rather than the equivalent dose assessment. However, the causes for erroneous dose-rate determination remain unclear (see also Wallinga *et al.*, 2001; Törnqvist *et al.*, 2003); no indications for disequilibrium were found in the decay chains.

In spite of uncertainties in the exact positioning of the isochrons, especially with regard to the OIS 5/4 transition in the Leidschendam core, the key feature (OIS 4 strata overlying OIS 5 or OIS 6 strata and underlying OIS 2 or OIS 3 strata) remains valid for any age model that honours the optical ages plus the sedimentology, and it is this key feature that we build on in our discussion below.

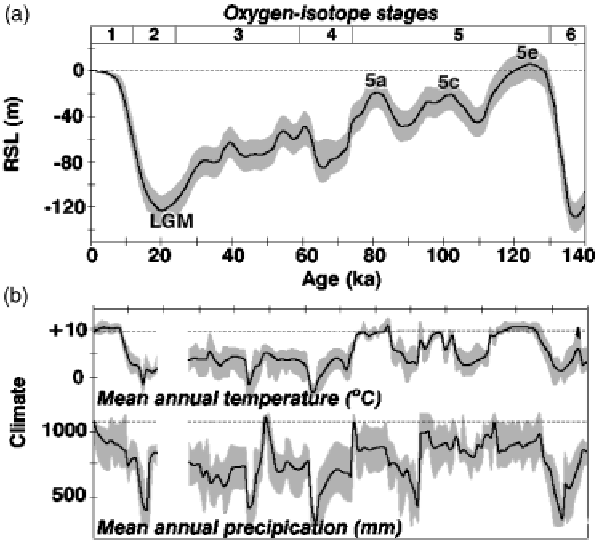

Fig. 4. Global sea-level change (a) and northwest European climate change (b) during the past 140 kyr. Sea-level record derived from North Atlantic and Equatorial Pacific benthic oxygen-isotope data (Waelbroeck *et al.*, 2002). Mean annual temperature and precipitation data derived from a pollen record from La Grande Pile, France (Guiot *et al.*, 1989), just outside the Meuse drainage basin. Horizontal, stippled lines indicate present-day conditions. Error envelopes in all records taken from the original sources. LGM, last glacial maximum.

DISCUSSION

Tectonic subsidence and aggradation rate

We use the term 'tectonics' for all long-term movements that can be considered to be more or less constant for the time interval of interest (~250 kyr). Processes involved include compaction of sediments in the deeper subsurface, long-term sediment isostasy, as well as plate tectonic forces and mantle processes (e.g. Kooi *et al.*, 1998). We do not include glacio- and hydro-isostasy as these act on shorter time scales and will be discussed in a later section.

Our study area subsides relative to areas just upstream and farther downstream (Kooi *et al.*, 1998). Quantitative estimates of the subsidence in the study area are sparse. Combining compaction, isostatic and tectonic movements as determined for the Quaternary by Kooi *et al.* (1998, Fig. 4), we obtain an average subsidence rate for our study area relative to the hinge line of approximately 0.10 mm yr^{-1}.

Relative subsidence in our study area provides accommodation for fluvial sediments; aggradation is required to keep the longitudinal profile unchanged through time. In the absence of other forcing factors, and provided that sediment supply is sufficient to fill the accommodation created, the aggradation rate is equal to the subsidence rate. In Fig. 3 we have plotted the depth vs. the optical ages obtained on the samples for the period OIS 6–2. The plot indicates that the points are scattered around a linear trend ($n = 25$; $r^2 = 0.72$). The slope of the line suggests an average aggradation rate of 8.2 ± 1.1 cm kyr^{-1} (0.08 mm yr^{-1}) over the period 200–10 ka. We propose that aggradation occurred in response to accommodation creation because of relative subsidence; the aggradation rate is very similar to the 0.10 mm yr^{-1} relative subsidence rate for the area that we infer based on information presented by Kooi *et al.* (1998).

We acknowledge that the time period over which we estimate the aggradation rate is possibly too short to allow a direct connection to accommodation creation. It is likely that at least part of OIS 3 and 2 deposits will be removed in the event of future RSL fall. If we ignore these deposits in our calculation we obtain a slightly lower aggradation rate; the value of 8.2 cm kyr^{-1} calculated above should be interpreted as a maximum value. We do not attempt to assess differences in aggradation rate for the three cores, as there are very few data points per core for such an exercise.

Scatter around the trend line is to be expected, considering that our data points derive from samples with variable positions relative to the coeval floodplain surface (i.e. ranging from channel-belt tops to bases). Moreover, excursions from the long-term aggradation trend have occurred, as is discussed in the following sections.

Excursions from the aggradation trend

Optical dating permits us to identify periods of incision (sensu Salter, 1993) or aggradation during the last glacial–interglacial cycle, and to correlate these with independent records of allogenic forcing. Based on our cross-section (Fig. 2) we identify two phases of incision: around the OIS 5/4 transition (discussed at length by Törnqvist *et al.*, 2003) and around the OIS 3/2 transition. The incision phases were followed by periods of aggradation during OIS 4 and 3 (74–24 ka), and during OIS 2 (24–12 ka). Besides these two periods, the Holocene is also characterized by high aggradation rates with ~15 m of deposition during the past 8 kyr.

The identified periods of incision or aggradation are the result of changes in sea level, climate or isostasy, or a combination of these. Below, we discuss what forcing factor was responsible for the observed excursions from the long-term aggradation trend. We do not attempt to quantify aggradation or incision rates for relatively short periods; there are very few data points and the precision is too limited for such an exercise.

For our discussion we will make use of the concepts of accumulation space and preservation space. Following Blum & Törnqvist (2000), we define preservation space as the area located below a conceptual boundary formed by a channel-bed profile graded to low-stand conditions. The preservation space boundary is fixed in time and space, at least for the time period we are concerned with (last ~200 kyr). Accumulation space is the area below the longitudinal profile of the floodplain surface. As the longitudinal profile shape may change in response to changes in external forcing, accumulation space is variable in time.

Even in a subsiding setting, the majority of fluvial deposits are cannibalized before reaching preservation space. Deposits formed in deeper parts of the channel will be over-represented in the stratigraphic record because the deepest deposits are most likely to subside below the preservation space boundary.

Sea-level forcing

Relative sea-level forcing of fluvial systems has received immense attention in the literature and forms the backbone of sequence-stratigraphic principles. Although early studies (e.g. Posamentier & Vail, 1988) suggested a nearly linear relationship between RSL change and fluvial longitudinal profile evolution, more recent work has demonstrated a much higher level of complexity of fluvial responses (reviewed by Blum & Törnqvist, 2000).

Global sea level (Fig. 4a) showed large-amplitude changes during the Quaternary because of build-up and decay of continental ice sheets (e.g. Waelbroeck *et al.*, 2002). For the Rhine–Meuse system, RSL fall resulted in

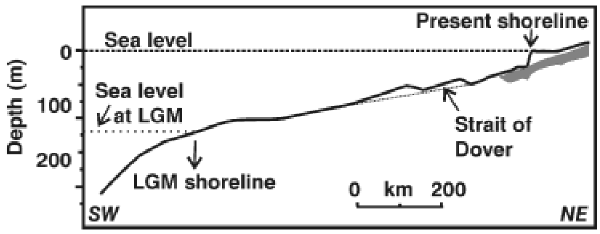

Fig. 5. Approximation of the longitudinal profile of the Rhine–Meuse system during low-stand conditions. The profile follows the drainage route of the Rhine–Meuse system (indicated in Fig. 1) through the Strait of Dover and English Channel. Late Pleistocene fluvial channel deposits from the Rhine–Meuse system are indicated in grey shading. The figure is compiled from our results in combination with published data (Hydrografisch Bureau, 1982, 1983; Berendsen and Stouthamer, 2000; Laban & Rijsdijk, 2002; Lericolais *et al.*, 2003). The bulges in the profile near the Strait of Dover are probably Holocene deposits; thus, there are no indications of inflection points in the longitudinal profile (indicated by the dotted line).

a dramatic extension of the river across the exceptionally wide, low-gradient continental shelf (e.g. Gibbard, 1995; Lambeck, 1995; Mulder & Syvitski, 1996). The long profile of the system during low-stand conditions is shown in Fig. 5. Large-scale fluvial incision in the English Channel during the last glacial cycle is unlikely, as sea level did not fall below the shelf edge during the Last Glacial Maximum (Figs 1 and 5) (Lambeck, 1995). Consequently, the area near the high-stand shoreline is the reach that is most likely to be strongly affected by RSL change.

Our study area is located just landward of the present shoreline and was also located near the coast during the highstands of OIS 5 (see Törnqvist *et al.*, 2000; Busschers *et al.*, in press). During the Holocene RSL highstand a coastal prism developed with a maximum thickness of 15 m to more than 30 m in the western Netherlands (e.g. Beets & Van der Spek, 2000). We expect that coastal prisms also formed during the OIS 5a, 5c and 5e highstands (Törnqvist *et al.*, 2000), although these were likely smaller than their Holocene counterpart because only the Meuse River contributed to the sediment input to the area during OIS 5 (Van de Meene & Zagwijn, 1978). In our study area, we encountered some erosional remnants of estuarine deposits related to OIS 5 RSL highstands. However, predominantly fine-grained high-stand strata, comparable with the Holocene coastal prism, seem to be absent. We deduce that RSL fall following highstands caused incision into the coastal prism (cf. Posamentier *et al.*, 1992; Blum & Price, 1998; Talling, 1998; Törnqvist *et al.*, 2000, 2003), after which lateral migration of the channels resulted in complete removal of coastal-prism deposits (see also next section).

The Rhine–Meuse system in the west-central Netherlands aggraded during the last glacial period of RSL fall (OIS 4–2). Although, in principle, it is possible that this net aggradation occurred in response to regression over a low-gradient shelf, we believe that the combination of tectonic accommodation creation and climatic circumstances is a more likely cause. Sea-level forcing is an unlikely cause of incision at the OIS 3/2 transition, because (1) sea level did not fall below the shelf edge during the LGM; (2) if a knickpoint was created, migration of this knickpoint would not reach our study area because it is located more than 800-km upstream; and (3) knickpoint migration would be hampered by shallow bedrock in the Strait of Dover area.

We conclude that RSL changes caused lengthening and shortening of the channel, which, because of geometric principles, induced aggradation and incision in our research area in periods just before, during and shortly after RSL highstands. In our transect, only a few deposits related to RSL forcing have been preserved as they are formed well above the preservation boundary and the duration of OIS 5 highstands was too short for subsidence to allow these strata to enter preservation space. This illustrates that preservation of deposits formed during dominant RSL forcing (i.e. during RSL highstand) is determined by the duration of the highstand and the rate of tectonic subsidence.

Climatic forcing

Although few would dispute that climate affects fluvial systems through discharge and sediment yield, the response of rivers to climate change is highly complex because of feedbacks that involve vegetation, permafrost and other variables (e.g. Vandenberghe, 2003). Moreover, response times of the fluvial system may be too long to record rapid climate changes (e.g. Castelltort & Van Den Driessche, 2003).

Palaeoclimate reconstructions for the last glacial in north-west Europe (e.g. Guiot *et al.*, 1989; Huijzer & Vandenberghe, 1998) not only point to mean annual temperatures considerably lower than present but also to much lower precipitation rates (Fig. 4b). A recent model study by Bogaart *et al.* (2003) has demonstrated that the combination of cold and dry conditions in settings like our study area leads to a reduced mean annual discharge. In addition, increased sediment supply is predicted because of a less dense vegetation cover, the development of permafrost and reduced soil water storage. The combination of these conditions causes channel aggradation. Thus, the dominantly cold and dry climatic conditions during much of OIS 4–2 are a likely explanation for the net aggradation in our study area at that time. We emphasize that this aggradation took place in spite of RSL fall during this period.

We hypothesize that rapid climate changes (millennial-scale Dansgaard–Oeschger events) as documented for the North Atlantic (Johnsen *et al.*, 1992; Bond *et al.*, 1993) and also distinctly featuring in high-resolution continental paleoclimate records in Europe (e.g. Genty *et al.*, 2003) resulted in substantial variations in discharge and sediment supply of the Rhine–Meuse system. Such variations may well have resulted in cut-and-fill cycles with alternating floodplain aggradation and degradation, operating within the overall aggradational context of OIS 4–2. Hence, we anticipate a stratigraphy characterized by valley fills or nested valley fills (cf. Holbrook, 2001) that occur within the sequence representing the last glacial–interglacial cycle. Unfortunately, our chronological data and the density of our drillings do not allow us to test this hypothesis, as detailed subdivision of fluvial strata beyond the level of oxygen–isotope stages is presently not possible. Finally, we cannot exclude the possibility that fluvial incision phases have been triggered by the catastrophic drainage of proglacial lakes associated with the Alpine Ice Sheet in the hinterland, comparable with what has been proposed for the Mississippi Valley (e.g. Knox, 1996; Blum *et al.*, 2000).

Recent work by Costard *et al.* (2003) has demonstrated that thermal erosion of frozen river banks can be extremely effective, with values for the present-day Siberian Lena River as high as $40\,\mathrm{m\,yr^{-1}}$. Given the fact that our study area was located within the range of continuous permafrost during substantial portions of the last glacial (Huijzer and Vandenberghe, 1998), it is likely that lateral fluvial erosion in the Rhine–Meuse palaeovalley was considerable, thus contributing to the rapid degradation of the OIS 5a high-stand coastal prism.

We conclude that cold and dry climatic conditions during OIS 4–2 caused net aggradation in our study area during that period. Climate oscillations, such as D–O events, probably induced short periods of incision and aggradation during OIS 4–2, giving rise to a highly punctuated sedimentary record.

Isostatic forcing

We use the term isostasy for crustal movements that are caused by ice and water loading and unloading because of the growth and decay of large ice masses during the Quaternary glacial–interglacial cycles. These isostatic movements occur on a time scale of 10^3–10^4 years. The impact of isostatic movements on fluvial systems has so far received little attention. On the US–Canadian border, isostatic tilting was identified as the cause for drainage reversal of Lake Agassiz (see Fisher, 2003; and references therein). Maddy & Bridgland (2000) proposed that post-glacial isostatic rebound might be responsible for a late Anglian (OIS 12, according to these authors) incision phase of the Thames in its middle reach.

During the Weichselian glaciation (OIS 5–2), the Fennoscandian Ice Sheet reached its south-western maximum extent around 22 ka; it halted in northern Germany some 500 km from our study area (see e.g. Houmark-Nielsen & Kjær, 2003; Ehlers & Gibbard, 2004). The melting of the south-western part of the ice sheet (Denmark, Germany) occurred between 20 and 15 ka (e.g. Boulton *et al.*, 2001; Houmark-Nielsen & Kjær, 2003). According to isostatic models, our study area was located on the far shoulder of the uplifting glacial forebulge during OIS 2 (Fig. 6); uplift in the area may have been more than 10 m (Lambeck *et al.*, 1998, Fig. 11). Kiden *et al.* (2002) have recently demonstrated that forebulge collapse is recorded in Holocene RSL curves for the western Netherlands and Belgium.

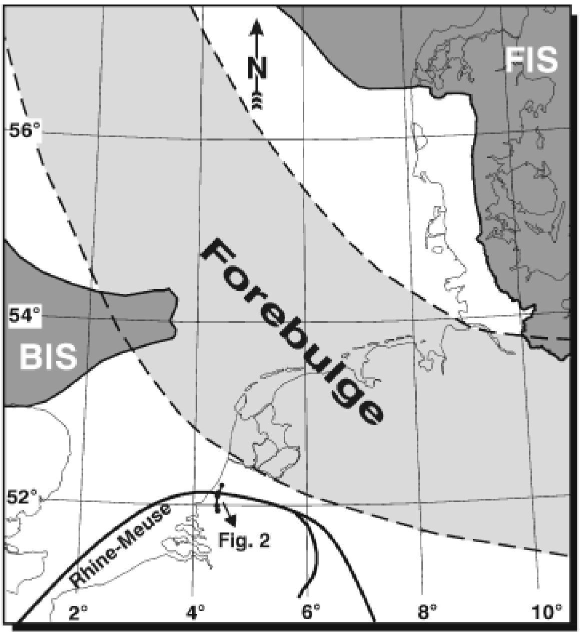

Fig. 6. Approximate position of the last glacial maximum glacial forebulge (Kiden *et al*., 2002) and ice sheets (Ehlers & Gibbard, 2004; BIS, British Ice Sheet; FIS, Fennoscandian Ice Sheet). Our study area is located on the far shoulder of the glacial forebulge which implies that the area was uplifted and tilted southward during ice build-up.

We propose that updoming of the research area during maximum ice extent resulted in a southward shift, as well as incision of the fluvial system. We note that the basal OIS 2 deposits are found 4 m lower than the base of OIS 3 deposits, indicating at least 4 m of fluvial incision, assuming constant channel depth. The trend in optical ages of the OIS 2 deposits in the Delft core indicates that the OIS 2 system was aggrading; lateral migration of a deep single-storey system would give rise to deposits of equal age for this unit. The aggradation of the OIS 2 system may be explained by accommodation creation because of collapse of the glacial forebulge in combination with climatic forcing. We conclude that isostatic forcing is likely (partly) responsible for the incision–aggradation event observed for the southern part of our transect during OIS 2.

SYNTHESIS AND CONCLUSIONS

Tectonic creation of preservation space has allowed the build-up of a 20-m thick sedimentary succession of predominantly fluvial strata in the west-central Netherlands during the past two glacial cycles. This succession is overlain by ~15 m of coastal-prism deposits formed in response to Holocene RSL rise. Two brief phases of incision were identified for the last glacial cycle. Incision around the OIS 5/4 transition was caused by RSL fall, whereas incision around the OIS 3/2 transition may be the result of updoming of a glacial forebulge. The period of OIS 4–3 was characterized by enhanced aggradation because of the dominantly cold and dry climatic conditions. After the LGM (OIS 2), a combination of climatic conditions and collapse of the glacial forebulge induced high aggradation rates.

Few deposits resulting from RSL forcing were preserved in our study area in spite of its location close to the present shoreline. The reasons for this are that (1) during the majority of the time, the shoreline was too far away from our study area to allow RSL changes to affect the fluvial system, and (2) although the thickness of high-stand deposits may have been considerable, their preservation potential is very low as they are prone to erosion during subsequent RSL fall. The dominant role of strata unaffected by RSL change in our study area highlights the importance of considering upstream controls of stratigraphic sequences. In addition, the chronostratigraphic relationships in our study area demonstrate that in case of a future RSL fall, the deposits from OIS 4 have the highest potential to survive erosion and to enter the stratigraphic record. We consider this an important finding, because it demonstrates the important role of the falling-stage systems tract and runs counter to widely used models that infer sediment bypass during RSL fall (e.g. Emery & Myers, 1996; Posamentier & Allen, 1999; Coe & Church, 2003).

It has been well established, following the synthesis by Shanley & McCabe (1994), that there exists a considerable downstream change in the relative importance of forcing factors in shaping stratigraphic successions, with climate change dominating the updip settings, giving way downdip to RSL control. Here, we show that associated with this spatial variability, the relative role of climate change vs. RSL change may vary dramatically through time, depending primarily on location relative to the coeval shoreline.

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