LATE QUATERNARY EVOLUTION OF THE MEUSE FLUVIAL SYSTEM AND ITS SEDIMENT COMPOSITION

A RECONSTRUCTION BASED ON BULK SAMPLE GEOCHEMISTRY AND FORWARD MODELLING

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Leonardus Antonius Tebbens

PROEFSCHRIFT

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STELLINGEN

Behorende bij het proefschrift:

"Late Quaternary evolution of the Meuse fluvial system and its sediment composition. A reconstruction based on bulk sample geochemistry and forward modelling"

van L.A. Tebbens

1. De abrupte klimaatverbetering tijdens het Laat-glaciaal komt in de rivierdynamiek van de Maas in Noord-Limburg tot uiting als snelle verticale geulinsnijding, gevolgd door laterale stroomvlakteverlaging. Graduele klimaatverslechtering is te herkennen als een stilstandfase in de geulinsnijding en leidt uiteindelijk tot aggradatie. (Dit proefschrift).

2. De vorming van het Laatglaciale erosieterras ("Vierlingsbeek terrace; Transitional system") tussen Meerlo en Boxmeer wordt door Huisink (1997) ten onrechte toegeschreven aan het Vroeg-Bølling. (Dit proefschrift).

Huisink, M. 1997. Late-glacial sedimentological and morphological changes in a lowland river in response to climatic change: the Maas, southern Netherlands. Journal of Quaternary Science 12: 209-223.

3. Op een tijdschaal van 1000 tot 10.000 jaren is de bulk mineralogische en geochemische samenstelling van gereduceerde, klastische en kleiïge tot siltige fluviatiele sedimenten een gevoelige graadmeter voor klimatologische veranderingen in de verweringsprocessen en rivierdynamiek in het bovenstroomse drainage-gebied. (Dit proefschrift).

4. De aanrijking aan Al_2O_3 en de depletie van K_2O, MgO en TiO_2 in Holocene fijnkorrelige sedimenten ten opzichte van glaciale equivalenten afgezet in restgeulen van de Maas in Noord-Limburg is het gevolg van postglaciale bodemvorming en mineraalverwering. (Dit proefschrift).

5. Een goede evaluatie van de individuele invloed van tektoniek, klimaat- en zeespiegelveranderingen op de lange-termijn (10^3-10^6 jaren) rivierdynamiek vereist dat men het riviersysteem in zijn gehele vierdimensionale context bestudeert. (Dit proefschrift).

6. Hypothesen met betrekking tot lange-termijn rivierdynamiek kunnen uitstekend worden getoetst met de forward-modellering techniek. (Dit proefschrift).

7. De grote invloed van eolische influxen op de glaciale rivierdynamiek van Noord-amerikaanse en Noordwest-europese riviersystemen staat in schril contrast tot de geringe hoeveelheid onderzoek die naar deze beïnvloeding gedaan is.

Ashley, G.J & Hamilton, T.D. 1993. Fluvial response to Late Quaternary climatic fluctuations, Central Kobuk Valley, Northwestern Alaska. Journal of Sedimentary Petrology 63: 814-827.

8. Het klimaat zou wel eens sneller kunnen veranderen (<20 jaar) dan de termijn waarop een door de politiek beoogde of geïnitieerde vermindering van de uitstoot van broeikasgassen effect sorteert.

Dansgaard, W., White, J.W.C. & Johnsen, S.J. 1989. The abrupt termination of the Younger Dryas climate event. Nature 339: 532-533.

Severinghaus, J.P., Sowers, T., Brook, E., Alley, R.B. & Bender, M.L. 1998. Timing of abrupt climate change at the end of the Younger Dryas interval from thermally fractionated gases in polar ice. Nature 391: 141-146.

9. De ondergang van de steden Sodom en Gomorrah moet niet worden toegeschreven aan een goddelijke bestraffing van zondiging tegen de seksuele moraal door hun bevolking, maar aan een aardse klimaatverslechtering die mogelijk gevolgd werd door een zware aardbeving van tektonische oorsprong.

Nissenbaum, A. 1994. Sodom, Gomorrah and the other lost cities of the plain – A climatic perspective. Climatic change 26: 435-446.

10. Een dogmatische afkeuring van homoseksuele/lesbische liefdesrelaties en leefstijlen door orthodox-gelovigen is om twee redenen principieel onjuist. Ten eerste omdat zíj er over oordelen en niet God zelf, en ten tweede omdat zij bij dit waarde-oordeel hun eigen visie of geloofsinterpretatie als uitgangspunt nemen, en niet de voor iedereen onbekende bedoelingen van God met het leven en de menselijke seksualiteit.

11. Er is in Nederland voldoende belastinggeld en ruimte te vinden om asielzoekers en witte illegalen op te vangen, en wel bij die rijke mensen die uit angst voor een waardedaling van hun landhuizen hun vermogen aanwenden voor de aankoop van een leegstaand potentieel asielzoekerscentrum, óf bij hen die massaal de vermogensbelasting ontvluchten door zélf of hun geld naar het buitenland te verhuizen.

12. Integratie van minderheden in de samenleving vereist geen aanpassing van die minderheden, maar respect van zowel de meerderheid als de minderheden voor andere leefvormen en een andere cultuur dan die van henzelf.

13. "De objectieve wetenschappelijke waarheid" bestaat niet omdat iedere wetenschapper een eigen waarheid publiceert die is gebaseerd op zijn of haar subjectieve aannames en inzichten.

14. Al is de toepassing nog zo snel, de fundamentele wetenschap achterhaalt haar wel.

15. De agressieve benaderingswijzen en versluierde verkoopmethoden van telemarketing-bedrijven zijn "zinloos gebeld".

Aan mijn ouders

Denkend aan Holland
zie ik breede rivieren
traag door oneindig
laagland gaan...

(Uit: Verzameld werk, H. Marsman)

VOORWOORD

Met het gereedkomen van dit proefschrift sluit ik een leerzame en soms turbulente periode af. Deze periode kende vanaf november 1993 diverse toppen en dalen. De eerstgenoemden waren het onderzoeksthema zelf, de wetenschappelijke uitdagingen daarin en de vele nieuwe mensen die ik heb leren kennen tijdens het onderzoek aan de vakgroep en tijdens congresbezoek. De laatstgenoemden waren de door het College van Bestuur opgelegde amputatie van de leerstoel Geologie en Mineralogie en de diverse reorganisaties aan het Laboratorium voor Bodemkunde en Geologie. Hierdoor is de analytische ondersteuning van mijn onderzoeksproject meer dan eens onder zware druk komen te staan. Met name het derde en vierde jaar heb ik ervaren als een soort survivaltocht door de universitaire jungle van de jaren negentig, die veel geëist heeft van mijn motivatie, doorzettingskracht en incasseringsvermogen. Want wat doe je als - door overmacht - je analyses na drie jaar voor de derde keer opnieuw gemeten moeten worden? Doorgaan dus! In dit voorwoord wil ik al die mensen bedanken die mij tijdens deze survivaltocht op de een of andere wijze ondersteund hebben. Allereerst bedank ik natuurlijk mijn promotor Salle Kroonenberg en co-promotor Tom Veldkamp.

Beste Salle, de geologie en het landschap hebben mij altijd mateloos geïnteresseerd, zoals je welbekend is. Deze interesse werd in mijn jeugd in goede banen geleid via de boekenserie "Hoe en Waarom" (over Vulkanen) en daarna door de Time-Life serie over de Planeet Aarde en "Nederland in het Holoceen". Na het vanuit thuis gestimuleerde eerste bezoek aan de open dag, deden de naam van de vakgroep (Bodemkunde en *Geologie*) en enthousiaste studie-introducties in het eerste jaar mij definitief besluiten Bodemkunde te gaan studeren. Dat er zoiets als een studie Fysische Geografie bestond, bleek pas tijdens de gecombineerde LUW/UvA-excursie Geomorfologie naar Oostenrijk. Met veel plezier denk ik terug aan o.a. jouw colleges, excursies en practica waarin de samenhang tussen lange-termijn veranderingen aan het aardoppervlak en de invloed daarvan op de landschaps- en bodemontwikkeling werd uitgelegd. Uiteindelijk heb ik mijn interesses kunnen inzetten op dit dankzij jou geïnitieerde NWO-project. Je aanvankelijk enthousiaste begeleiding is er in de afgelopen jaren noodgedwongen een van grotere afstand geworden door de maatregelen van het CvB. Gelukkig kwamen de artikelen tijdens de schrijffase altijd snel en voorzien van nuttig commentaar terug. Bovendien wist je samen met Tom mijn koudwatervrees voor het modelleren te relativeren en me te stimuleren om toch vooral uit te zoomen in plaats van in te zoomen: het juiste duwtje in de juiste richting op het juiste moment! Hartelijk dank!

Beste Tom, als co-promotor heb je pas vanaf mijn tweede jaar je belangrijke en grote inzet voor mijn onderzoek kunnen laten gelden. Als mijn dagelijks begeleider kon ik ten allen tijde op je terugvallen en dat heeft er mede toe bijgedragen dat het onderzoek ondanks diverse tegenslagen toch is voltooid. Jouw constructieve bijdrage aan de modelleringstudie en de *previews* van mijn soms uitgebreide en "overcomplete" (!) concept-artikelen hebben de resultaten in dit proefschrift nog beter boven het regionale niveau helpen uittillen. Als geen ander versta jij de kunst om het juiste evenwicht te bewaren tussen veldwerk en modellering en om de voor menigeen abstracte PEM-plaatjes te vertalen naar concrete veldsituaties. Zo blijkt maar weer dat een op basis van grote lijnen gestoelde samenwerking tussen de detaillisten uit het 'veld-kamp' en de modelleurs veel kan opleveren. Heel erg bedankt voor al je hulp!

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is. Hans Huisman, ook jou bedank ik voor de fijne collegialiteit en alle goede discussies over geochemie. De woorden pyriet en sideriet zal ik voor altijd met jou associëren!

Mijn collega's van de faculteit Aardwetenschappen aan de VU in Amsterdam ben ik zeer erkentelijk voor de eerste introducties in "hun" veldwerkgebied van de Laatglaciale Maas en de inzage in diverse boorprofielen en palynologische gegevens. Dit werk bracht me op het idee mij te concentreren op de goed bewaard gebleven restgeulopvullingen in Noord-Limburg. Margriet Huisink, Kees Kasse, Wim Hoek, Sjoerd Bohncke, Jef Vandenberghe, Kaj Beets en Nicolaj Walraven: dank voor alle informatie en (veld-)discussies! Over de Laatglaciale rivierdynamiek van de Maas zijn we vast nog niet uitgepraat en dat is maar goed ook: het houdt de geest scherp voor nieuwe inzichten. Net zo essentieel voor dit onderzoek waren de gesprekken die ik voerde met medewerkers van het NITG-TNO. "Maasexpert" Meindert van den Berg, Patrick Kiden en Wim Westerhoff: dank voor jullie informatie en tevens voor de ^{14}C-dateringen van Dubbroek, Linksstraat en De Berckt. Torbjörn Törnqvist (Universiteit Utrecht) en dr. Van de Plicht van het isotopenlab in Groningen dank ik voor de goede adviezen met betrekking tot de interpretatie van ^{14}C-dateringen aan kleiïge gyttja-monsters.

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Tijdens de veldwerkperiode hebben meerdere studenten en vakgroepmedewerkers in het veld geassisteerd met de vele boringen. Geuch de Boer, Arie van Dijk, Peter Droogers, Corine van Griethuysen, Karin Groenestein, Monica Krol, Lammert Kooistra, Erik Meyles, Irina Overeem, Thom Pape en Piet Peters, allen hartelijk bedankt! Rienk Miedema, Arie van Dijk en Toine Jongmans worden tevens bedankt voor respectievelijk het verstrekken van informatie over het Broekhuizen veldpracticumgebied en het nemen en prepareren van de wellicht in de toekomst nog te bekijken micromorfologische monsters in de groeve Bosscherheide.

Jan van Doesburg zorgde voor de XRD-metingen, terwijl Bram Kuyper en in een later stadium Jan Huting de XRFS-metingen en hernieuwde calibraties verzorgden. Peter Buurman adviseerde bij kleimineralogische interpretaties en granulometrische bepalingen met de Laser Grainsizer. Dank voor jullie inzet! De studenten Marjan van Leerdam en Farida Berrad namen een deel van het malen en smelten van de monsters op zich in het kader van een afstudeervak Geologie en Mineralogie. Dank daarvoor. Farida, ici je voudrais te remercier beaucoup encore pour tes efforts plus grands. Ce n'etait pas facilement pour toi pour preparer des plusieures analyses pendant le Ramadan et tandis qu'on souffre d'attaques migraines. Merci!

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Wageningen, april 1999.

CONTENTS

GENERAL INTRODUCTION

1 GENERAL INTRODUCTION

1.1 GLOBAL AND SOCIAL CONTEXT

Satellite observations of the temperature of the Earth's surface recently confirmed ground data that our Planet is truly warming (Hansen et al., 1998). These observations are now also in line with model studies that have been predicting already for some years that anthropogenic enhanced concentrations of greenhouse gases in the Earth's atmosphere will result in global warming. Global warming, to which most people will refer to as the Greenhouse Effect, is suggested to cause global feedback mechanisms in the atmosphere, hydrosphere, cryosphere and the lithosphere, to which scientists refer to as climate change (Goodess et al., 1992). The increasing awareness of human impact on global climate has ironically raised concern with respect to the rebound effect of the short- and long-term climatic impact on the human environment. Higher temperatures related to global warming will certainly affect the hydrological cycle. Late Quaternary Heinrich events (Bond et al., 1993) and the present-day recurrent El Niño events demonstrate that global circulation patterns of oceanic water and water vapour periodically change to affect the redistribution and availability of water on regional to global scales. The resulting changing weather patterns are known to influence the amount and distribution of precipitation and run off on the continents (cf. Knox, 1996). Temperature and water availability directly control the weathering intensity and erosion rates of rocks in the landscape.

Rivers carry the most important part of continental run off and weathering products to their depositional alluvial basins. In view of most silted-up river deltas being fertile and therefore densely populated agricultural areas, it is clear that any influence of climate change on both the quantity and quality of the discharge and sediment flux of river systems has important social relevance. This applies especially to regions and countries that derive their existence almost exclusively from alluvial deposits, such as the river deltas of the Mississippi (southern Louisiana), the Brahmaputra (Bangladesh) or the Rhine/Meuse (The Netherlands). The recent 1993 and 1995 flooding events in the Netherlands caused considerable damage and social unrest (Bleichrodt and Ensinck, 1994; Goudriaan, 1995). Fluvial deposition of allegedly polluted sediment (Leenaers, 1989; Rang and Schouten, 1989; Winkels, 1997) stresses the need for thorough environmental research and touches the fundamental issue of finding natural background values for potentially dangerous heavy metals (Wolterbeek et al., 1996; Darnley, 1997).

1.2 FLUVIAL DYNAMICS

Knowledge on the influence of climatic variability on fluvial dynamics is of crucial value to assess and predict future changes in the quantity and quality of river discharges and sediment fluxes. Only by quantifying and understanding the fundamental geological processes controlling the fluvial system in its natural undisturbed habitat, one can begin to reconstruct any human or future climatic impact on the river system. The fact that present-day fluvial processes in Western industrialised areas are far from natural and undisturbed requires that natural fluvial behaviour be reconstructed from pre-industrial fluvial sedimentary records or archives (e.g. Benito et al., 1996). Several workers have used this concept ("the past as a key to the future") to describe fluvial dynamics on different time-scales. A reconstruction of the temporal evolution of sediment fluxes and sediment composition directly involves a study of fluvial system evolution. Schumm (1977), Bull (1991) and Vandenberghe (1995) stress the importance of time-scale and geomorphological thresholds in characterising fluvial response to climate change. Thus, a true 4-D approach is required: the three-dimensional landscape and the fourth dimension of time. Until now, the niche of quantifying and modelling river response to climate change on a drainage basin-wide scale has remained largely uninhabited. Even less attention has been paid to systematic temporal changes in the quantity and quality (or composition) of river sediment fluxes.

Sediment flux on different time-scales

Earlier research dealing with shorter time-scales (10^0-10^2 years) involved evaluations of flooding hazards for the present-day rivers, e.g. flow forecasting for the River Meuse (Berger, 1992) and quantifying the impact of climatic change on the discharges of the River Rhine (Kwadijk, 1993). Van der Weijden and Middelburg (1989), Winkels (1997) and Asselman (1997) studied the present-day suspended load of the River Rhine. Middelkoop (1997) reconstructed the evolution of embanked floodplains in the Netherlands. Authors in Petts et al. (1989) treated historical changes in large alluvial rivers of Western Europe to focus on environmental, ecological and river regulation issues. Finally, authors in Benito et al. (1998) give the latest state of knowledge concerning palaeohydrology and flooding events.

Most long-term studies qualitatively reconstruct the development of river morphodynamics and sedimentology in specific river reaches. Van den Berg (1996) focussed on the climo-tectonic evolution of the Meuse sedimentary record in the upper and middle reaches on a temporal scale of 10^3-10^6 years. He found that the tectonic regime controls preservation of the sedimentary record, by determining whether terracing or stacking of sedimentary unit takes place. The major and minor unconformities separating individual sedimentary units are believed to result from the external climatic forcing of sediment supply and river discharge. Törnqvist (1993) reconstructed channel pattern development and channel belt avulsion rates in the lower part of the Holocene Rhine-Meuse delta to conclude a dominant role for sea-level rise controlling fluvial aggradation. Berendsen et al. (1995), Kasse et al. (1995), Weerts and Berendsen (1995), Makaske and Nap (1995), Makaske (1998) and Huisink (1999) reconstructed the palaeogeography and morphodynamics of different sections of the Dutch Rhine, Meuse and Vecht river systems. Channel patterns at the Weichselian-Holocene transition were found to switch from braiding to middle reach meandering and lower reach anastomosing styles. Starkel et al. (1991) and Frenzel et al. (1995) found comparable results in their reconstruction of geomorphological, palaeohydrological and sedimentological developments for Temperate Zone river systems for the last 15,000 years.

Recent insights from sequence stratigraphy (Leeder and Stewart, 1996) and the study of long fluvial longitudinal profiles or complete catchments (Merritts et al., 1994; Rose, 1995) have shown that the results of specific river reaches can not be extrapolated unconditionally to the scale of the whole drainage basin. Complex-response behaviour of the fluvial system (Schumm, 1977; Bull, 1991) requires an integrated basin-wide analysis. Forward modelling techniques have proven their great value to capture all dimensions of fluvial dynamics (Veldkamp and Van Dijke, 1998, 1999). Weltje (1994) applied inverse end-member modelling techniques to reconstruct sediment dispersal paths and source area (provenance) characteristics from a sedimentary record for which the exact nature of sediment mixing is unknown.

Sediment composition

The composition of clastic sedimentary rocks is defined best as their bulk mineralogy and because each mineral has its own chemical composition, the bulk mineralogy directly relates to the bulk geochemistry. Johnsson (1993) thoroughly described a system of intimately interrelated parameters, controlling the bulk mineralogy of fluvial sediments. The following introductory review is largely based on this work. Johnsson distinguished four main controls on sediment composition: 1. Provenance and source rock composition, 2. Physical and chemical weathering in the source rock area, 3. Erosion and transport to and within the river and 4. Diagenetic controls after deposition or burial. These controls either alter the mineralogical composition or change the relative abundance of the individual rocks or minerals in time and in space. The tectonic and climatic setting of a river drainage basin and its tributary catchments determine how the weathering, erosion and transport processes mutually influence each other and which of them will be most important. They form the links between the main controls.

Provenance and source rock composition always form the starting point. The tectonic setting directly determines the composition by the type of magmatism from which the rocks originate. Likewise, metamorphic rocks demonstrate mineralogical and compositional differentiation according to the pressure and temperature gradients these rocks have (re)equilibrated to. Plate tectonic movements may initiate phases of orogenesis (crustal thickening) leading to regional uplift or rift valley formation (crustal extension) leading to subsidence (Fuchs et al., 1983; Ziegler, 1994). Via the generation of relief, tectonic movements indirectly decide which rocks and minerals become available for weathering and how long they are exposed to the reactive Earth's surface. During physical or chemical weathering, two aspects are of major importance, namely the intensity and the duration of weathering. Weathering intensity firstly depends on the initial chemical and physical characteristics of the minerals involved and secondly on the climatic conditions to which they are exposed. The physical and chemical stability of source rock minerals in saprolite or soil environments specifies which of them will be most subject to weathering. E.g. in a given climatic setting, micas composed of individual sheets or internally lamellar feldspars will be more susceptible to breakdown than monomineralic quartz or zircon. Alternating freeze-thaw cycles in a (peri-)glacial climate actively enforces physical breakdown of rock fragments or minerals to enlarge the amount of reactive surface areas for chemical reactions. Higher temperatures and higher availability of water or organic acids under temperate or tropical warm and humid climatic conditions greatly accelerate chemical weathering processes.

Weathering may be intense, but physical and chemical weathering will barely have effect if the time to affect source rocks, soils or sediments is too short. Therefore, weathering duration is a very important parameter during sediment production and storage. The main factor determining the residence time of weathering material in a river drainage basin is the stability of weathering mantles (saprolites and their topping soils) in the landscape. Climate and tectonics provide several feedback mechanisms to this landscape stability. Rapid tectonic uplift produces steep slopes and accentuated relief. Mass wasting processes like slope failure, sheet erosion and soil creep in a high-relief landscape contribute to constant erosion and regeneration of weathering mantles, leaving little time for chemical weathering processes to act on exposed rocks and minerals. In this situation, sediment supply is said to be weathering-limited and the detrital sediment composition will reflect the source rocks to great extent. Interglacial warm and humid climatic conditions favour soil formation and enables vegetation growth. A fully developed forest vegetation cover greatly reduces superficial splash erosion and trees tend to stabilise loose weathered material with their rooting systems. A stable landscape with deep and mature soils will be the result. Now, sediment supply is transport-limited and the composition of the weathering material will chemically adjust to the prevailing climatic conditions. Apart from the provenance characteristics, the ultimate detrital sediment composition will bear a climatic signature as well. However, excessive rainfall will induce water logging and slumping of weathering material. Similarly, the waterlogged active thaw-layer during summer under glacial cold-stage permafrost conditions will give rise to enhanced gelifluction rates in low- and high-relief areas. Extreme tropical weathering is likely to wipe out all provenance characteristics.

Once erosion has removed weathered material from the source areas, transport within streams, tributaries and main river controls the fate of sediment composition. Climatic fluctuations in river discharge and sediment supply in time and in space determine whether sediment is to be eroded, to stay in transport or to be deposited. Uplift or subsidence along the longitudinal river gradient causes river transport to be speeded up or slowed down to cause grain-size gradients in downstream direction. The process of abrasion during transport might progressively break down large monomineralic grains and polymineralic rock fragments to smaller sizes. As this is largely a physical process, the composition is not changed considerably, but it can effect the process of hydrodynamic sorting (see below). Grain-size further varies laterally and vertically at one site or river section owing to facies differences in the local depositional environment. Size, shape and density are the basic properties of mineral grains to exert influence on the process of hydrodynamic sorting. Examples are the predominance of quartz in the coarse sand fractions, the concentration of clay minerals and phyllosilicates in the clay and fine silt size fractions and the concentration of heavy minerals in sandy to gravelly placers (Moura and Kroonenberg, 1990; Veldkamp and Kroonenberg, 1993). Hydrodynamic sorting during transport and sedimentation thus tends to concentrate typical suites of minerals in specific grain-size classes (Irion, 1991) to make grain-size in itself a powerful geochemical differentiating factor. Apart from sorting, transport processes intimately mix the mineral supply of a variety of weathering materials and soils from the tributary catchments to generate an average composition for the sediment flux in the main river (e.g. Ottesen, 1989; Veldkamp, 1991). Should the relative contributions of one or more tributary catchment change significantly as a consequence of climatic change, then the sediment flux of the main river is likely to change too.

After deposition of the sediments, sediments may be temporarily stored on the alluvial plain to await new transport, rather than be completely buried. Chemical weathering and erosion can again intervene to further modify the sediment composition (e.g. Johnsson, 1990). If sediments become out of reach of the eroding river or are completely buried, they are not exposed anymore to atmospheric influence or the open aqueous environment. Curtis (1990) argues that "preservation of original soil or sediment compositions and their mineralogy will be favoured by events that effectively isolate the mineral assemblage from aqueous solution through-flow (closed systems preserving bulk composition) and subsequent burial at shallow depths and in low geothermal gradients". Climatic signals are therefore likely to survive best in bulk compositional data and in the relative distribution of climatically sensitive minerals. In all other situations, the mineral assemblage will reequilibrate to a new thermodynamic stability state that fits the new chemical and physical environment. Low temperature and pressure reequilibration mainly involves early diagenetic microbially mediated redox reactions (Curtis, 1995).

1.3 OBJECTIVES

Degens et al. (1991) pointed out that rivers are not steady-state systems, but will reflect global climatic change. They envision that climatic changes will be expressed in the physical and chemical properties of the river sediments. Chamley (1989) advocated the study of Quaternary fine-grained river deposits to quantify compositional changes. He stated that "mineralogical data combined with lithology may provide useful information on the contribution of rocks versus soils, the climate, the importance of alluviation from different tributaries and geological formations of a given drainage basin, and the morphological and tectonic changes in the course of time". Ibbeken and Schleyer (1991) studied present-day river sediments at the active plate margin of Calabria (southern Italy) to find mainly provenance-related differences. Pioneering work by Kroonenberg (1990), Moura and Kroonenberg (1990), Veldkamp (1991), Hakstege et al. (1993) and Veldkamp and Kroonenberg (1993) has shown that the bulk geochemical composition of Quaternary river sediments is also likely to vary due to climatic change.

The main objective of this thesis is to establish whether climatic change has measurable effects on the bulk geochemical composition of fluvial sediments. For this purpose, we will narrow the definition of climate change to changes in temperature and the availability of water (precipitation and discharge) in time. Temperature and water availability are the key parameters determining the intensity of weathering and erosion processes. Since the completion of ice-core drilling for the Greenland Ice Core Project (GRIP) and the Greenland Ice Sheet Project (GISP2), a wealth of information on North Atlantic climatic change over the last 250,000 years (250-ka) has become available. The ice cores demonstrate that climate change is the rule in the North Atlantic region, and that our stable present-day Holocene climate is rather an exception. Johnsen et al. (1992), Dansgaard et al. (1993), GRIP Members (1993) and Grootes et al. (1993) demonstrate frequent, large and abrupt climatic changes at a time-scale of 10^3 to 10^5 years. Their regional to global extent is confirmed by North Atlantic marine sedimentary records (Long et al., 1988; Bond et al., 1993; McManus et al., 1994) and Southern Hemisphere ice-cores (cf. Bender et al., 1994). Gehrke et al. (1996) demonstrated that stadial to interstadial (10^3-10^4 years) and glacial to interglacial (10^5 years) climatic variability was reflected in clay mineral assemblages of oceanic cores SW of Iceland, due to differential input from continental erosion and ice-rafted detritus. With respect to the effect of climate on the composition of fluvial clastic sediments, Curtis and Douglas (1993) and Curtis (1995) expected major changes to occur only on a time-scale of $>10^3$ years.

A subgoal of the research is to determine which fluvial sediments preserve the palaeoclimatic signal best and in what way. The local fluvial energetic conditions, influenced by relief, sediment supply and discharge themselves, determine whether gravel, coarse or fine sands, silts or clay is deposited. Chamley (1989) and Curtis (1990) recommended analysing the composition of fine-grained clastic sediments for evidence of the climatic record. Most fine-grained sediment supply to rivers is ultimately derived from weathering and soil formation processes that are climate-controlled (e.g. Singer, 1984; Righi and Meunier, 1995). Simultaneously however, Singer (1984) and Hillier (1995) noted that the palaeoclimatic interpretation of clay mineral assemblages in sediments is anything but straightforward and should consider multiple feedback mechanisms and diagenetic overprinting processes. Nevertheless, large sedimentary basins integrate the clay mineral response of their source areas to climate change, so that in favourable cases climatic signals appear to be preserved (Ottesen, 1989; Hillier, 1995).

Parrish et al. (1993) call for a more rigorous calibration of the climatic controls on sedimentation and on the formation of sedimentary palaeoclimatic indicators. The fluvial system has a notorious reputation for the occurrence of alternating phases of aggradation and degradation in space and in time, creating series of hiatuses in the sedimentary records at the profile level (Miall, 1996). The analysis of clastic fine-grained sediments implies by definition that one concentrates on the preserved sedimentary record of the fluvial system only. Furthermore, the processes of fluvial erosion and deposition and the preservation potential of sediments are not solely dependent on climatic changes, but on tectonics and sea-level changes as well (cf. Veldkamp and Van den Berg, 1993; Törnqvist, 1993; Merritts et al., 1994). Therefore, it is important to make a link to the external forcing of fluvial dynamics (see outline above) to address complex-response fluvial behaviour. The second subgoal of this project is to find out why the sediments one is interested in have been laid down there in the first place and to find an answer to the question how they relate to internal fluvial dynamics.

1.4 THE RIVER MEUSE CASE STUDY

Tectonic and climatic setting

The availability of an extensive climatic and palaeohydrological database within the northwest European context has drawn our focus to the Temperate-Zone River Meuse. The River Meuse has its origin in Northern France and drains rocks in Belgium and the Netherlands to end up in the southern part of the North Sea Basin (Fig. 1.1). The Meuse is a medium-sized river with a length of 874-km and a total drainage basin of 33,000-km^2. The present-day drainage area is entirely situated in the humid temperate climatic zone, with mean annual temperatures and precipitation varying from 6-°C and ~1400-mm/yr in the sloping Ardennes low mountain range in Belgium to 9-°C and ~800-mm/yr in the low-lying countries. Mean monthly discharge near the Belgian-Dutch border amounts 339-m^3/s. The river is rain-fed making it very sensitive to climatically induced variations in discharge. Lowest discharges are normally recorded in July and highest in December. The absence of upstream glaciers causes low summer base-flow to a minimum of 0-m^3/s, while prolonged or high precipitation combined with snow melting can produce severe downstream flooding at a maximum recorded discharge of 3120-m^3/s in December 1993 (Berger and Mugie, 1994; Bleichrodt and Ensinck, 1994).

Extensive study of the River Meuse has provided ample literature on different geological, geomorphological, sedimentological and morphodynamic aspects (cf. Pons, 1957; Van den Broek and Maarleveld, 1963; Paulissen, 1973; Teunissen, 1983; Juvigné and Renard, 1992; Bohncke et al. 1993; Berendsen et al., 1995; Kasse et al., 1995; Van den Berg, 1996; Huisink, 1997, 1999). The tectonic context is a slowly uplifting hinterland with maximum altitudes around 600-m and minor neotectonic movements (Van den Berg, 1994). The slow uplift preserved glacial aggradational terraces deposited in the Dutch part of the Meuse delta about every 100-ka (Veldkamp and Van den Berg, 1993). Van den Berg (1994), Van den Berg et al. (1994) and Van den Berg (1996) provide thorough information on the tectonic framework and climo-tectonic fluvial response of the River Meuse. Therefore, excellent opportunities exist to evaluate sediment compositional changes in their long-term fluvio-systematic context, i.e. within the context of fluvial response to long-term (10^3-10^5 years) external and internal forcing at the spatial scale of the entire drainage basin.

Outline of the thesis

This thesis will report on a systematic inventory on fine-grained infillings of residual channels of the River Meuse in the southern Netherlands. Late Weichselian to Holocene residual channels and fine-grained clastic to organic infillings are well preserved in the depositional Venlo Graben in the northern part of the province of Limburg. The residual channels comprise the low-energy environments necessary to capture snap shots of suspended load deposition during flooding events. Their low-lying, wet positions favour organic matter accumulation to provide ample dating possibilities and the reduced conditions minimise post-depositional pedogenetic overprinting processes. Part I contains the empirical results from the geomorphological fieldwork and the laboratory analyses. The origin of the residual channels and fine-grained sediments in the North Limburg study area will be reconstructed in Chapter 2. Systematic ^{14}C-dating, granulometric and bulk and clay geochemical laboratory analyses followed an extensive sampling campaign to describe river dynamics and sediment composi-

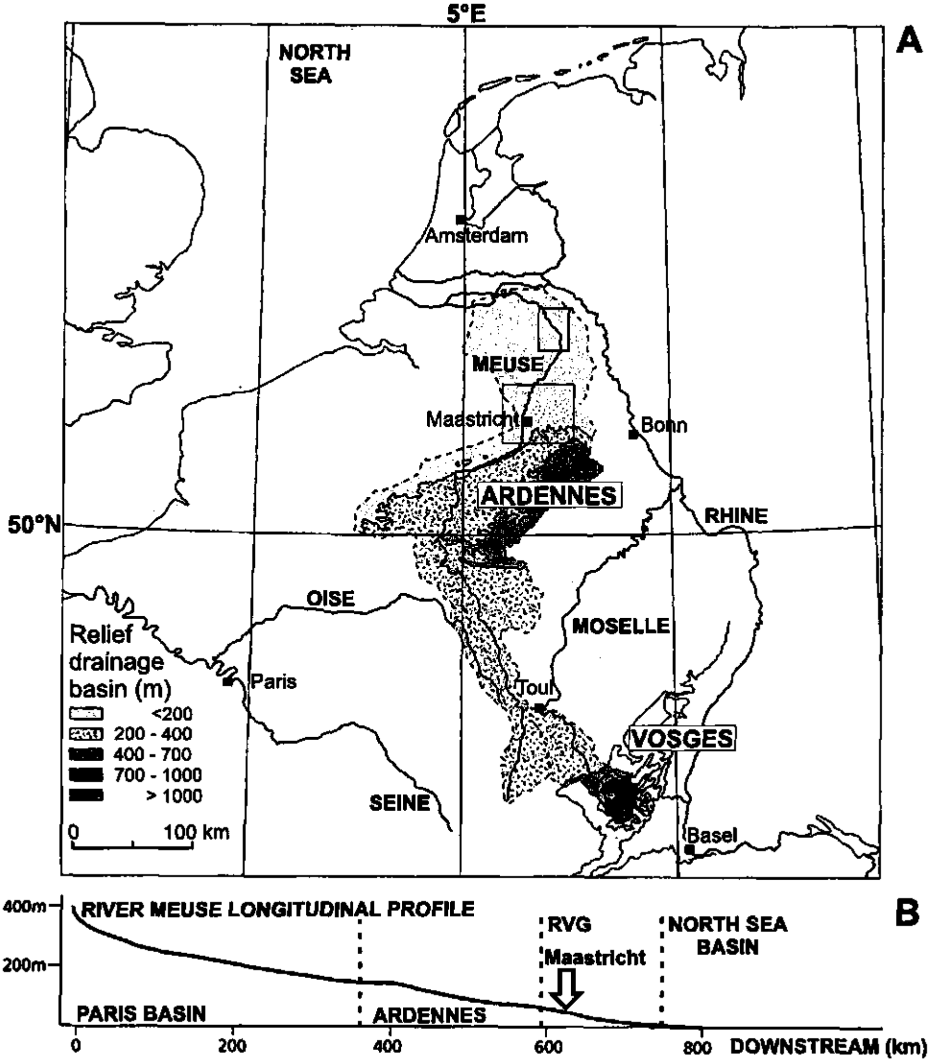

Figure 1.1 *(A)* The Northwest-European setting of the River Meuse drainage basin. At 250-ka BP, the Moselle captured the Upper Meuse Vosges catchment near Toul. The upper rectangle indicates the North Limburg study area of the Meuse lower reach (this thesis). The lower rectangle indicates the Maastricht terrace flight area of Van den Berg (1996). ***(B)*** The present-day longitudinal profile of the Meuse. The major tectonic domains are indicated. RVG = Roer Valley Graben (Adapted from Van den Berg, 1996).

tion on a time-scale of 10^3-10^4 years (Ch. 3-4). Part II presents the results from a forward-modelling case study to understand the results from the North Limburg study area within the broader fluvio-systematic context (Ch. 5-7).

Chapter 2 discusses the origin of the residual channels and their infillings. Fieldwork data are presented to give insight into the regional geological setting and fluvial dynamics of the Meuse lower reach study area in North Limburg. The ^{14}C dating of the organic residual channel infillings provided the chronostratigraphical framework for both the timing of incisional phases and the (approximate) ages of residual channel infillings or clayey units intercalated between organic layers. Next, the incisional history of the River Meuse is reconstructed to put the residual channels and their infillings within the context of Late Glacial and Early Holocene landscape evolution and river morphodynamic developments.

Chapter 3 and 4 describe the main and trace element geochemistry of the studied residual infillings. Laser granulometrical data, X-Ray Fluorescense Spectroscopy (XRFS)- and X-Ray Diffraction (XRD)-measurements are used to describe the natural bulk and clay geochemical variation within a set of ~640 samples of fine-grained residual channel infillings. More specifically, Chapter 3 describes the natural compositional variation to enable comparison of trace element contents from allegedly polluted sediments with the values of pre-industrial sediments. Chapter 4 zooms in on the search for a suitable sedimentary palaeoclimatic indicator. The focus lies on the main constituents and on the possibilities and limitations (e.g. overprinting effects) of the use of clay-related constituents as a palaeoclimatic indicator. Differences between Late Glacial and Holocene samples are interpreted in terms of changing detrital clay mineral supply.

Chapter 5 forms the onset to quantify river sediment fluxes and sediment composition on a temporal scale of 10^3-10^5 years and on the spatial scale of the whole drainage basin. The evolution of the Meuse longitudinal profile is modelled in response to the external forcing factors of tectonics, sea-level and climate change. The changes in sediment flux, leading to erosional and depositional phases in the fluvial system, are predicted on a basin-wide scale for the period 250-0 ka BP (Before Present). Results from Chapter 2 are used to tune the model.

Finally, the modelled sediment fluxes from Chapter 5 are integrated with bulk geochemical data in **Chapter 6** to predict the downstream and long-term (10^3-10^5 years) evolution of sediment composition. The main goal of this chapter is to evaluate whether observed compositional differences (Chapter 3, 4) are solely the consequence of chemical weathering (transport-limited sediment supply) or can alternatively be attributed to changes in the relative contributions of tributary catchments (weathering-limited sediment supply) or both. The synthesis (**Chapter 7**) attempts to integrate all information and provides some suggestions for applications in other river systems and recommendations to further research.

PART I

RESIDUAL CHANNELS AND THEIR INFILLINGS IN THE RIVER MEUSE LOWER REACH

THE 15–0 ka BP WINDOW FROM NORTHERN LIMBURG, THE NETHERLANDS

CHAPTER 2

FLUVIAL INCISION AND CHANNEL DOWNCUTTING AS A RESPONSE TO LATEGLACIAL AND EARLY HOLOCENE CLIMATE CHANGE: THE RIVER MEUSE LOWER REACH

Based on: Tebbens, L.A., Veldkamp, A., Westerhoff, W. and Kroonenberg, S.B. 1999.

Fluvial incision and channel downcutting as a response to Lateglacial and Early Holocene climate change: the lower reach of the River Meuse, The Netherlands.

J. Quaternary Science **14**: 59-75.

2 FLUVIAL INCISION AND CHANNEL DOWNCUTTING AS A RESPONSE TO LATE-GLACIAL AND EARLY HOLOCENE CLIMATE CHANGE: THE RIVER MEUSE LOWER REACH

ABSTRACT

Detailed fieldwork and new extensive ^{14}C-dating of residual channel infillings provide data for the reconstruction of the Late-glacial channel downcutting and incision history of the Venlo-Boxmeer lower reach of the River Meuse (Maas) in the southern Netherlands. Within a period of 500-1300 years after Late-glacial climatic amelioration, the Meuse responds to increased discharges and decreased sediment supply by adjusting the width/depth ratio of its channels. Two main phases of channel downcutting are followed by two main phases of floodplain lowering and narrowing, indicating net floodplain degradation by the fluvial system as a non-linear response to Late-glacial and Early Holocene climate change.

Some 1300 years after initial Late-glacial warming, channels downcut rapidly during the Early Bølling (13.3-12.5 ka BP) and adopt a high-sinuosity meandering style. Channel downcutting pauses around 11.9-ka BP, possibly in response to rising groundwater levels and/or the Older Dryas cooling event. Between 11.9 and 11.3-ka BP a new floodplain is formed. Then, lateral erosion takes over and initiates a first phase of 2.6-m floodplain lowering during the Late Allerød. Gradual climate deterioration during the Allerød effects that soils and vegetation cover progressively break up from 11.3 to 10.9 ka BP. The Meuse gradually adjusts to an increased ratio of sediment supply over transport capacity through higher width/depth ratios. Main channels become shallower and adopt a low-sinuosity pattern, finally culminating in a braiding river system during the Younger Dryas. The final Holocene warming results within 500 years in renewed rapid channel downcutting by a single low-sinuosity channel during the Early Preboreal, followed by a second phase of 1.8-2.8 m floodplain lowering.

2.1 INTRODUCTION

Perturbation of fluvial systems takes place by tectonics, climate and sea level changes at different time-scales. Fluvial response to climate change is fundamentally dependent on time-scale (Vandenberghe, 1995). At a time-scale of 100-ka, tectonic uplift and glacial sea-level lowstands trigger river incision by lowering the base level of erosion of the river. Tectonic uplift is essential in preserving glacially aggraded river terraces (Veldkamp and Van den Berg, 1993). A high-frequency cyclicity (parasequences) within these fluvial terraces may be related to climatic variability on a time-scale of 1 to 10-ka (Van den Berg, 1996). At this time-scale, long stable periods alternate with short instability phases occurring at climatic transitions, with vegetation, soil cohesion and run off as important deterministic variables. The self-organised dynamics of the fluvial system respectively local climate (flood frequency) and sedimentary thresholds and last but not least human impact become increasingly important on a time-scale of 0.1 to 1-ka (Macklin et al., 1992a,b).

Climatic forcing of fluvial behaviour has been demonstrated for the River Meuse in the south-eastern Netherlands (Kasse et al., 1995; Van den Berg, 1996), the downstream part of the Rhine-Meuse delta (Törnqvist, 1993; Törnqvist et al., 1994; Berendsen et al., 1995) and other north-west European river systems like the Vistula in Poland (Starkel et al., 1996) and the Somme in France (Antoine, 1994). Many north-western European rivers incised as a response to Late Weichselian/Holocene climatic change (Starkel, 1991; Walker, 1995), but exact timing and causal relationships of incisional phases are still insecure. Climatically induced river incision occurs when the local balance between river discharge and sediment load is disturbed and the erosional-depositional threshold is exceeded (Schumm, 1977; Bull, 1991). This threshold separates modes of aggradation (net deposition) and degradation (net erosion). It is defined as the ratio of power available (stream power) to power needed (resisting power) for entrainment and transport of sediment load. Whenever this ratio is >1, net erosion takes place. Both stream power and resisting power contain variables as river discharge, gradient, river width, depth and sediment load characteristics, that are largely influenced by climate and local conditions on a time-scale of 1 to10-ka.

Following the definition of Bull (1991), the Late-glacial Meuse terraces can be regarded as fill-cut terraces, formed by Pleniglacial valley aggradation and subsequent channel downcutting into the alluvium. The palaeofloodplain surface becomes a (paired) terrace level when the remnants of the former valley floor are left (see Fig. 2.1). Incision by the river that abandons its old floodplain to form a new, lower elevated floodplain is then by definition the terrace forming process. Thus, when we speak of incision or degradation, we mean a lowering of the floodplain or valley floor. In this paper, we will attempt to unravel the timing between climate change, channel downcutting and floodplain degradation phases in the fluvial system on a 1-10 ka time-scale. We will focus on the Late Weichselian and Early Holocene incision history of the lower reach of the Meuse by studying Late-glacial floodplains, terrace gradients and residual channels for a 40-km stretch just upstream the present-day terrace intersection.

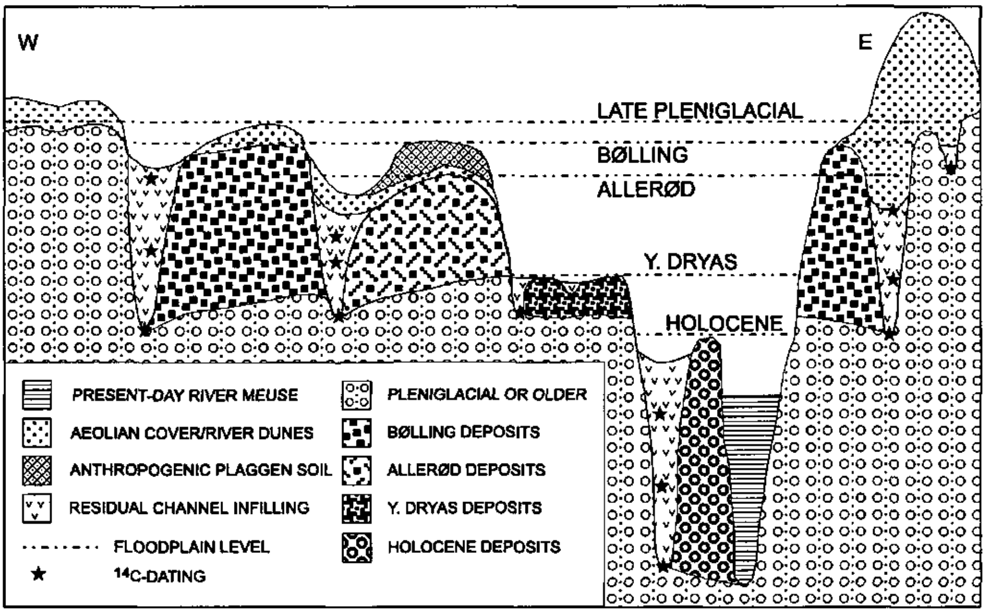

Figure 2.1 Schematic representation of the Late-glacial and Holocene palaeofloodplains, residual channels (not to scale), dated residual infillings and aeolian covers.

2.2 LATE-GLACIAL TERRACE STRATIGRAPHY

The Venlo-Boxmeer transect is situated in the lower reach of the River Meuse in the southeastern Netherlands (Fig. 2.2). The transect is oriented mainly parallel to the main faults of the Venlo Graben, which forms a subsiding part of the uplifting Peel-Venlo block, located upstream of the hinge line of the North Sea Basin (Van den Berg et al., 1994). Late-glacial fill-cut terraces as well as Late-glacial and Early Holocene fine-grained sediments have been well preserved here. The area has been the subject of many studies and therefore a wealth of information on the Late-glacial terrace stratigraphy and morphodynamics of the Meuse is already available.

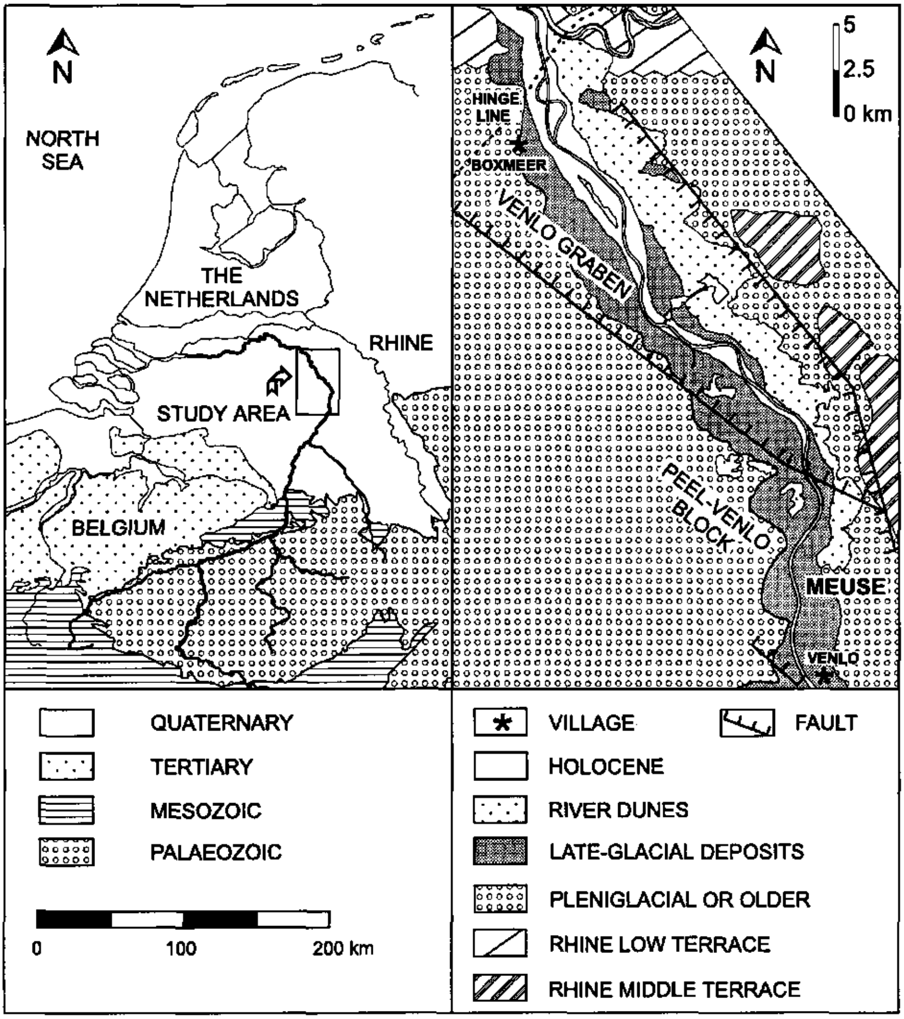

Figure 2.2 The drainage area of the River Meuse (left) and the Venlo-Boxmeer lower reach (right), showing the extent of both Weichselian and Holocene Meuse sediments and of aeolian and Rhine deposits. Main tectonic features after Van den Berg (1994).

The reconstruction of the Late Weichselian and Early Holocene incision and channel downcutting phases requires a precise chrono- and lithostratigraphic framework. Pons and Schelling (1951), Schelling (1951) and Pons (1957) described a fluvial low terrace with a loam cover ("Hochflutlehm") dating from before the Allerød Interstadial. Van den Broek and Maarleveld (1963) described a Late Pleniglacial level, three Allerød levels and a Younger Dryas level based on geomorphological and pedological investigations. Miedema et al. (1983) emphasised that terrace levels can be separated unambiguously only on the basis of elevation, leading Jongmans and Miedema (1986) to distinguish four levels of pre-Bølling, Allerød, Younger Dryas and Holocene ages. Recently, Kasse et al. (1995) and Huisink (1997) described 2 Pleniglacial terraces, a transitional Bølling level, an Allerød level and a Younger Dryas level in a review on the morphological changes of the Late Weichselian Meuse. Finally, Van den Berg (1996) distinguished a Pleniglacial level, two Allerød levels, a Younger Dryas level and a Holocene level for the same area.

From the foregoing, it appears that there has been considerable debate on the occurrence and distribution of the Late-glacial floodplain levels. Emphasis has largely been put on the floodplain morphology, rather than the channel dynamics. Recent ^{14}C-dating and fieldwork by the present authors yielded data that were not always consistent with present insights, indicating that previous studies still do not yield a complete picture. In this chapter, we propose a (partial re)-interpretation of new and existing data to enable a reconstruction of Late-glacial river activity and in this way the incisional and channel downcutting history of the Meuse lower reach.

2.3 METHODS

Gradient lines

Late-glacial palaeofloodplains and residual channels have been identified on the 1:50,000 geomorphological and soil maps of the Venlo-Boxmeer area (Stiboka, 1975, 1976; Buitenhuis and Wolfert, 1988; Wolfert and De Lange, 1990). Combining palaeofloodplain level remnants with similar residual channel surface patterns and with the ^{14}C-ages of the oldest residual channel infillings yielded the terrace map of Figure 2.3.

We constructed gradient lines and determined net Late-glacial incision from the difference between the palaeofloodplain levels (Fig. 2.4). The top elevation of different floodplain levels was obtained from 1:10,000 topographical height maps from the Dutch Ordnance Survey. At some sites, aeolian coversands (up to 2-m) or anthropogenic plaggen soils (up to 1.2-m) were observed to cover older palaeofloodplains (see Fig. 2.1). These covers are also visualised as separate units on the geomorphological and soil maps. Sandy brown top soils on the lowest floodplain level are not anthropogenic, but originate from periodic Holocene flooding and sedimentation followed by pedogenic and tilling homogenisation (Jongmans and Miedema, 1986). Because the height difference between successive floodplain levels in this area is only 1-2 m, the presence of aeolian covers, anthropogenic soils or both can lead to erroneous interpretations of palaeofloodplain top elevations. Where necessary, we corrected the topographic heights for these covers using field borings and information from the soil map legend or from literature.

Owing to river sinuosity, it can be difficult to determine the downstream distances from individual palaeofloodplain sampling points. For calculating the palaeofloodplain gradients, we projected floodplain top level sampling points (corrected for covers) perpendicular to the Late-glacial palaeovalley axis (Fig. 2.3) and performed linear regression. North of the village of Beugen, the Meuse flows into a former Late-glacial branch of the River Rhine that is now occupied by the Meuse tributary Niers. The branch is situated in a subsiding area downstream of the North Sea Basin hinge line (Van den Berg et al., 1994) and downstream of the present terrace

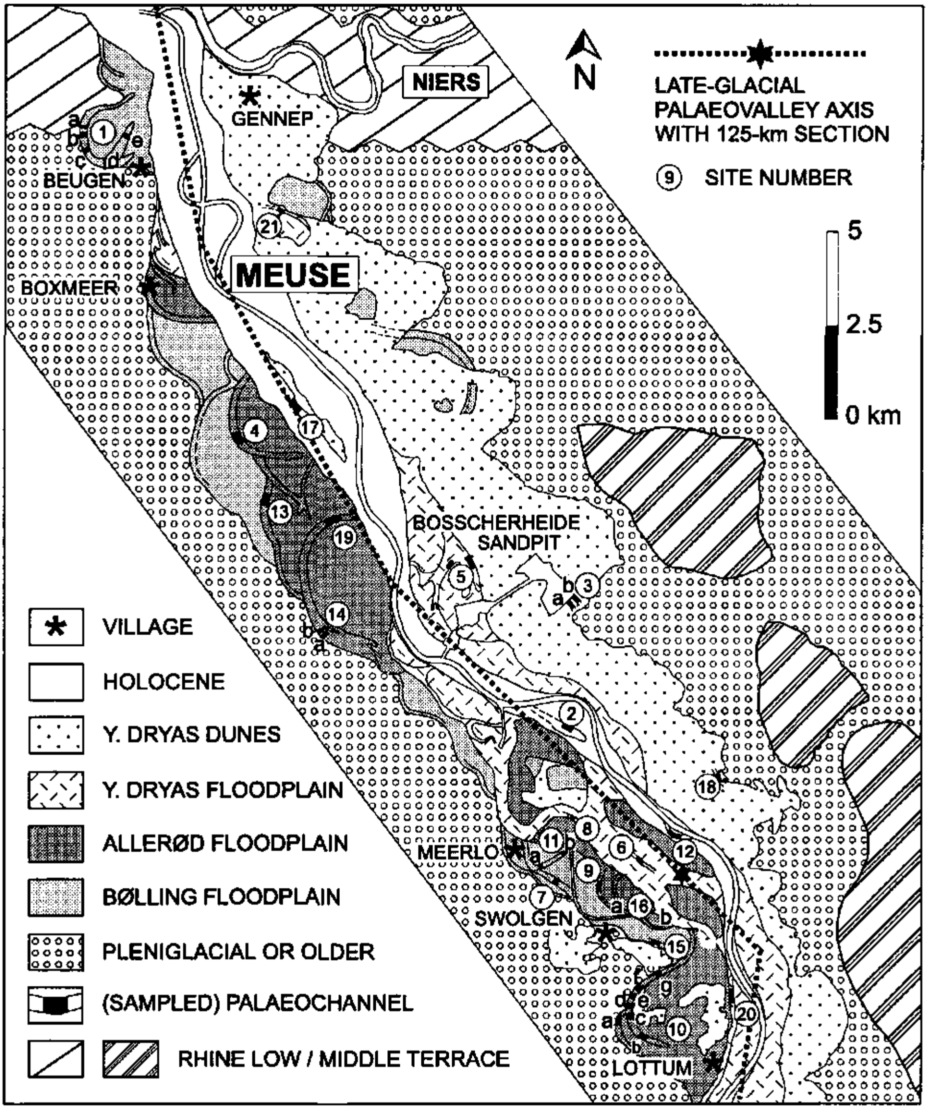

Figure 2.3 The Late-glacial terrace stratigraphy for the Venlo-Boxmeer stretch. The dashed line indicates the Late-glacial palaeovalley axis, whereas the star depicts the location of the ideal cross-section (Fig. 2.5, page 22-23) at 125-km downstream of the Dutch-Belgian border. Information on the sampling sites and ^{14}C-datings can be found in Table 2.1 on page 21.

intersection. A prototype meander has developed here (Kasse et al., 1995), but we excluded this area north of Boxmeer from the calculation of our gradient lines to rule out any Late-glacial Rhine influence or subsidence effects on floodplain elevation. Allowing for their floodplain gradients, the surface elevations and depths of the sampled residual channels can be plotted to form an ideal "cross-section" for any point along the palaeovalley axis. Figure 2.5 shows this

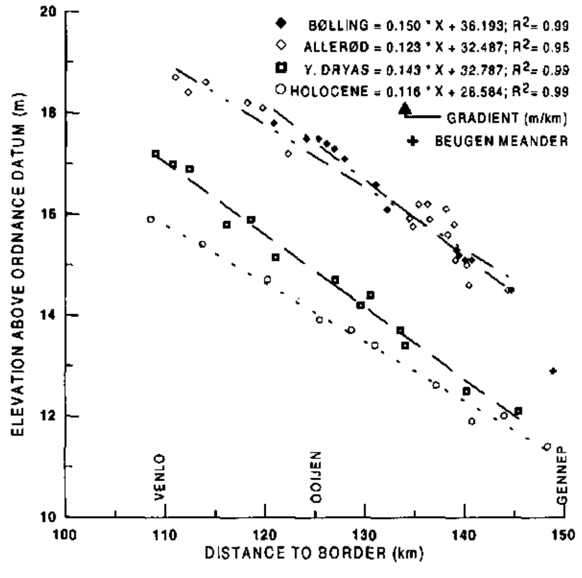

Figure 2.4 Late-glacial & Holocene floodplain gradients for the 40-km Venlo-Boxmeer stretch.

compilation and projection of the floodplain levels and the sampled main residual channels for a point midway between Venlo and Boxmeer (125-km downstream of the Dutch-Belgian border, 3-km east of the village of Meerlo).

^{14}C-Chronostratigraphy

Samples from the base of organic beds in residual channel fills are very suitable for dating the end of channel activity and yield a minimum age for palaeofloodplain activity. The start of organic infilling of a residual channel is synchronous with the end of overbank sedimentation (Törnqvist and Van Dijk, 1993). AMS-dated terrestrial macrofossils are to be preferred to obtain a well-framed chronostratigraphy (Lowe and NASP Members, 1995; Wohlfahrt, 1996). In this research, however, we had to rely on ^{14}C-dates of bulk-sediment samples for both economic and practical reasons. Törnqvist et al. (1992) concluded that special care should be taken in interpreting bulk ^{14}C-ages of gyttjas and strongly clayey samples. They found bulk ^{14}C-ages being 0.2-0.6 ka too old (Table 2.1) with respect to terrestrial macrofossils of the same stratigraphical level owing to uptake of bicarbonate by aquatic plants (hardwater effect) and reworking of older organic material (reservoir effect). Dissolved inorganic carbon (DIC) in bicarbonate-rich water originates partly from equilibrium with air-CO_2 with 100% ^{14}C-activity and partly from "old" dissolved carbonate minerals with zero ^{14}C-activity. The ^{14}C-activity of aquatic plants or submerged parts taking up this DIC was found 60-85% of modern activity (Marčenko et al., 1989). Consequently, any radiocarbon age based on a bulk sample with a high contribution of aquatic plants (with low ^{14}C-activity) will be too old. On the other hand, Berendsen et al. (1995) found an age of 10720 ± 60 (GrN-15023) for a calcareous clayey gyttja at the base of a Meuse Allerød channel. The lime fraction of the same sample gave a younger age of 10690 ± 230 BP (GrN-16038). This suggests that the calcium carbonate was precipitated synsedimentary and that ageing effects due to uptake of "old" bicarbonate ions might be of minor importance not only for this sample, but also for other calcareous gyttja samples.

Table 2.1 Sample information and conventional ^{14}C-ages of bulk material. TQ-samples from Törnqvist et al. (1992).

Site-no. & Name (See Figure 2.3)	Sample	Lab.-number	^{14}C-age (yrs. BP) (bulk)	Coord. (1:25,000 top.-map)	Elevation above ordn.-datum (m)	Depth (m)	Material	δ^{13}C (‰) (Bulk sample)	TQ: ^{14}C(bulk)-^{14}C (AMS) = # ageing yrs.
1c: Beugen	BEU 4-28	GrN-23416	12330 ± 170	191.815/410.240	10.8	3.10-3.19	calcareous green yellow gyttja	-31.93	
1c: "	BGN-97-1	UtC-6893	8390 ± 70	191.815/410.240	10.8	3.22-3.25	AMS on *Betula* and *Carex* seeds	-27.00	Too young?!
2: Blitterswijck	BLI 1-26	GrN-23417	5340 ± 80	204.865/394.825	12.7	3.63-3.67	clayey peat	-29.91	
2: "	BLI 1-42	GrN-23418	7320 ± 80	204.865/394.825	12.7	5.01-5.07	peaty dark gray gyttja	-29.56	
3b: Bosscherheide	BOS 3-8	GrN-21882	13760 ± 70	204.825/398.300	19.6	4.50	twigs, fine leaves and moss	-28.27	
4: Groeningen	GRO 1-1	GrN-22003	10220 ± 60	196.300/402.350	12.8	2.30-2.40	slightly calcareous green gyttja	-29.91	
4: "	GRO 1-14	GrN-22004	11280 ± 80	196.300/402.350	12.8	3.50-3.70	slightly calcareous green gyttja	-30.71	
5: Heuloërbroek	NITG-TNO	GrN-18615	10600 ± 60	201.600/392.500	13.1	2.26-2.30	gyttja	-	
7: Keuter	KEU 1-6	GrN-22005	10630 ± 120	204.475/390.800	15.6	1.10-1.16	peat	-31.95	
7: "	KEU 1-14	GrN-22006	11060 ± 100	204.475/390.800	15.6	1.86-1.96	calcareous black gyttja	-28.70	
7: "	KEU 1-23	GrN-22007	11270 ± 240	204.475/390.800	15.6	2.71-2.82	slightly calcareous d.gray clayey gyttja	-29.54	
7: "	KEU 1-32	GrN-22008	11780 ± 270	204.475/390.800	15.6	3.83-3.95	calcareous black clayey gyttja	-29.49	
8: Linksstraat	NITG-TNO	GrN-18612	10850 ± 70	205.050/392.500	13.9	2.72-2.76	gyttja	-	
9: Looiveld	LOO 1-25	GrN-22009	10950 ± 280	205.800/390.450	18.6	5.74-5.81	yellowish green calcareous gyttja	-29.10	
10c: Lottum	LOT 29-0D	GrN-23420	7550 ± 80	206.425/386.975	17.6	1.55-1.60	black peat	-29.93	
10c: "	LOT 29-0F	GrN-23421	10890 ± 150	206.425/386.975	17.6	2.19-2.30	peaty gyttja	-32.13	
10c: "	LOT 29-12	GrN-23422	11640 ± 220	206.425/386.975	17.6	2.30-2.44	gyttja	-34.32	
11a: Meerlo	MEE 2-4	GrN-22010	9590 ± 90	203.975/391.250	15.4	2.07-2.14	peat	-30.42	
11a: "	MEE 2-13	GrN-22011	11320 ± 210	203.975/391.250	15.4	3.10-3.20	calcareous d.gray-green clayey gyttja	-30.46	
11a: "	MEE 2-20	GrN-22012	12500 ± 120	203.975/391.250	15.4	3.80-3.90	calcareous d.gray-green clayey gyttja	-30.03	
12: Ooijen	OOI 4-5	GrN-22013	3480 ± 40	207.225/392.350	13.5	1.20-1.30	clayey peat	-28.26	
12: "	OOI 4-17	GrN-22014	6800 ± 50	207.225/392.350	13.5	2.40-2.47	non-calcareous brownish gray gyttja	-28.42	
12: "	OOI 4-25	GrN-22015	9170 ± 200	207.225/392.350	13.5	3.17-3.27	non-calcareous sandy gyttja	-28.46	
13: Schafferden	SCH	GrN-22016	11260 ± 110	196.600/401.075	14.2	2.50-2.53	non-calcareous sandy gyttja	-31.08	
14b: Smakt	SMA 2-1	GrN-22300	5660 ± 60	197.950/397.650	15.1	0.88-0.96	peat	-29.60	
14b: "	SMA 2-4	GrN-22018	9800 ± 100	197.950/397.650	15.1	1.20-1.26	non-calcareous gyttja	-29.42	
14b : "	SMA 2-11	GrN-22019	10350 ± 130	197.950/397.650	15.1	1.84-1.91	non-calcareous dark brown gyttja	-28.99	
14b : "	SMA 2-21	GrN-22020	11170 ± 230	197.950/397.650	15.1	2.72-2.85	non-calcareous gr. brown clayey gyttja	-28.47	
15: Swolgen 1	SWO 1-5	GrN-22021	9270 ± 150	206.325/389/850	17.5	2.35-2.45	peat	-29.66	
16b: Swolgen 3	SWO 3-15	GrN-22022	10080 ± 110	206.725/389.600	17.4	3.12-3.22	non-calcareous greenish grey gyttja	-29.31	
17: Vortum	VOR 1-1	GrN-22024	6160 ± 90	197.275/403.650	11.3	2.00-2.10	peat	-29.08	
18: Westmeerven	WES 1-2	GrN-23423	8510 ± 60	208.865/393.715	17.0	1.60-1.70	brownish sandy gyttja	-29.43	
18: "	WES 1-8	GrN 23424	10590 ± 70	208.865/393.715	17.0	2.38-2.48	light brownish gyttja	-28.20	
18: "	WES 1-16	GrN-23425	11520 ± 90	208.865/393.715	17.0	3.30-3.35	peaty gyttja	-31.53	
Törnqvist et al. (1992)	TQ 3A	UtC-1133	3310 ± 80	-	-	-	*Alnus* peat	-29.7	60; 240
"	TQ 9A	UtC-1128	4670 ± 80	-	-	-	*Phragmites* peat	-28.8	170
"	TQ 13A	UtC-1297	5440 ± 50	-	-	-	strongly clayey *Phragmites* peat	-28.9	200; 110
"	TQ 14A	UtC-1125	4460 ± 70	-	-	-	calcareous, brownish yellow gyttja	-26.8	-10; 40; 220
"	TQ 15A	UtC-1298	4570 ± 120	-	-	-	brownish yellow gyttja	-30.5	600; 540
"	TQ 16A	UtC-1143	5490 ± 100	-	-	-	calcareous, str. clayey gr. brown gyttja	-29.5	390; 510

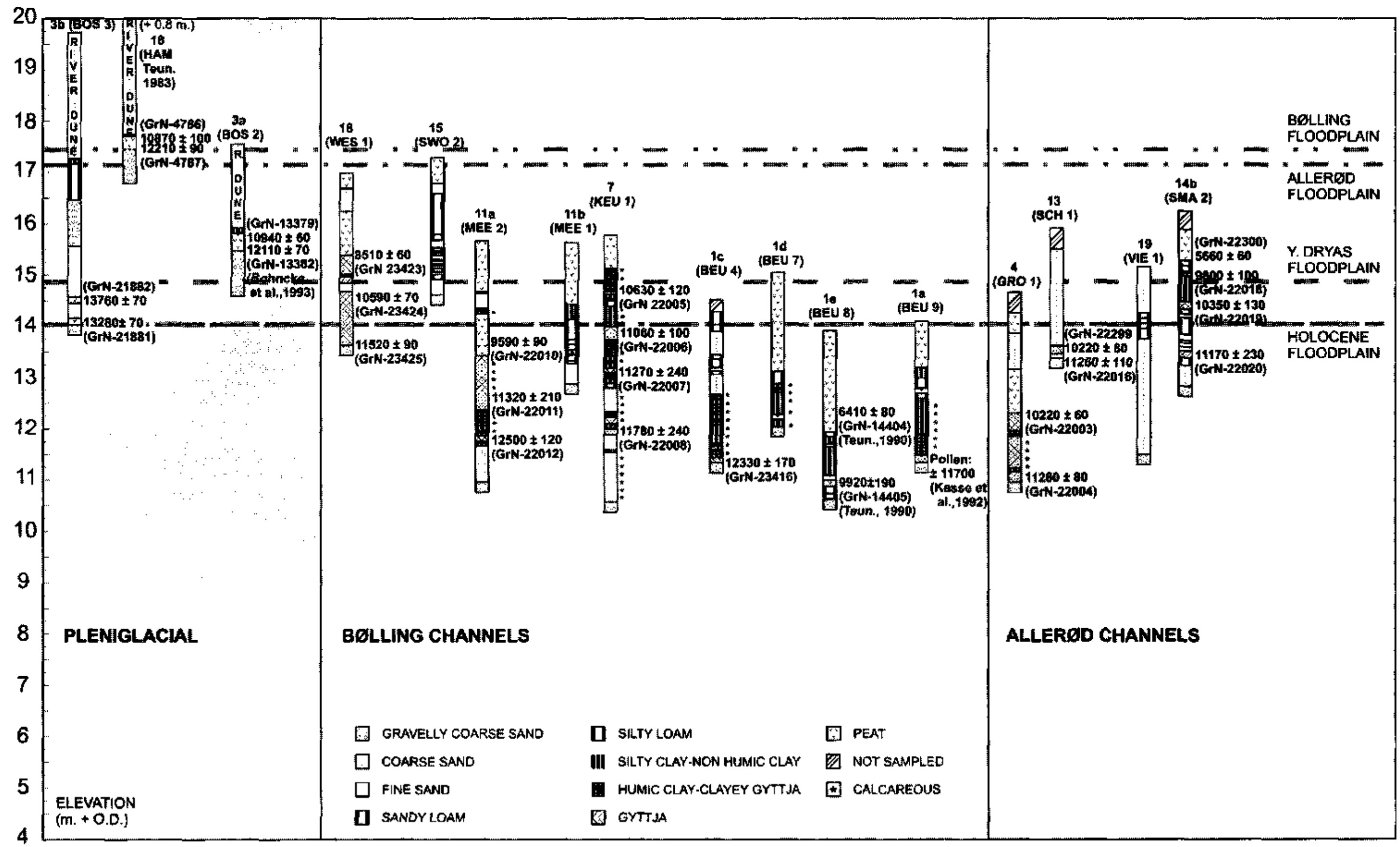

Figure 2.5 Projection or ideal cross-section of the floodplain elevations, main residual channels and infillings for a site 125-km downstream of the Dutch-Belgian border. (Elevations have been recalculated using the floodplain gradients from the linear fits in Figure 2.4).

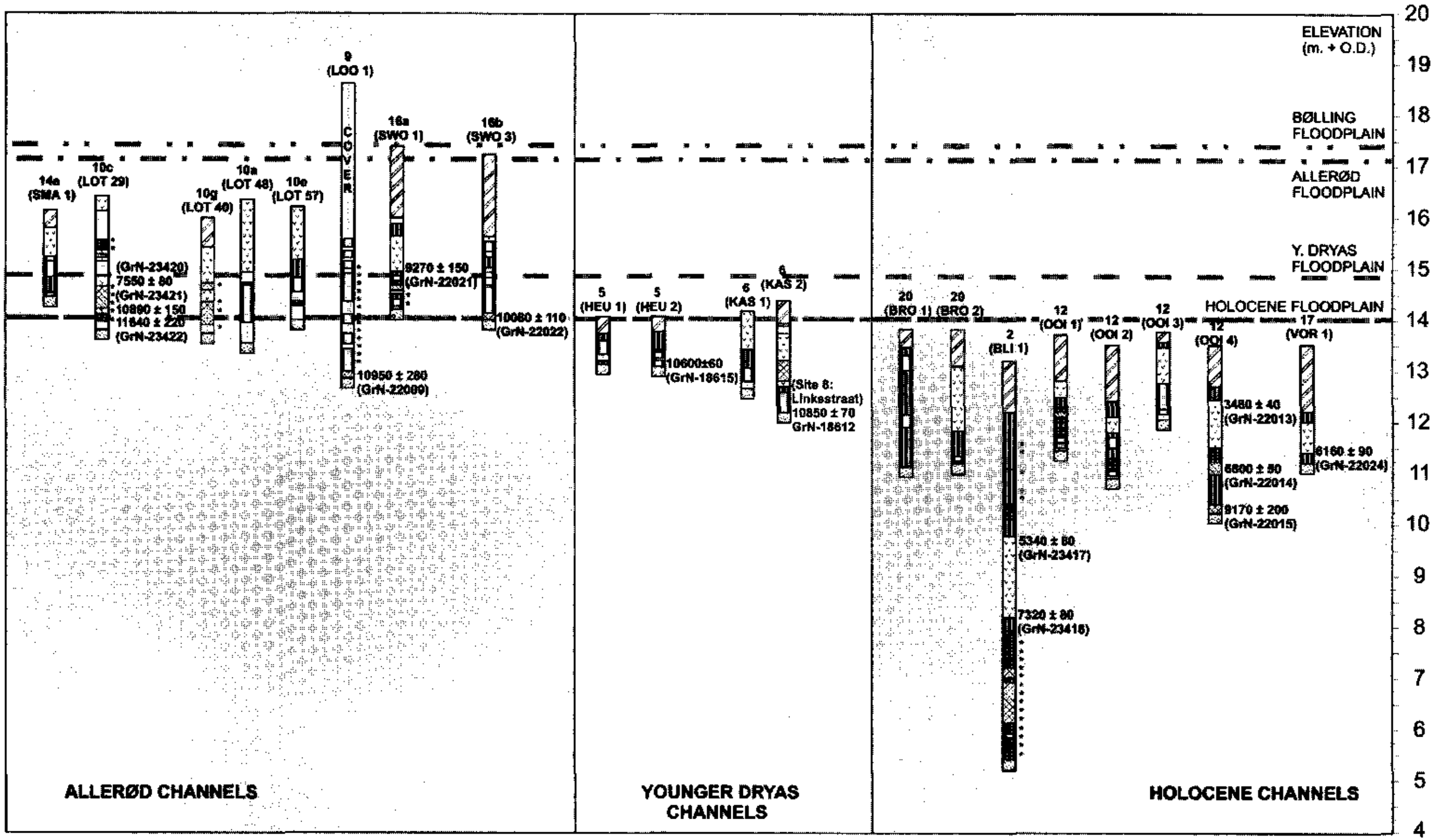

Figure 2.5 Continued

Some insight into hardwater and reservoir effects can be obtained by assessing the $\delta^{13}C$-values for the dated bulk material. The $\delta^{13}C$-values can serve to estimate the relative contributions of terrestrial and aquatic plants. During respiration, terrestrial plants take up CO_2 from the air, leading to bulk $\delta^{13}C$-values in the order of -25 to -30‰. True aquatic plants or submerged parts of helophytes taking up "old" dissolved inorganic carbon from water are characterised by very negative bulk $\delta^{13}C$-values of -32 to -47‰ (Marčenko et al., 1989). Most peat and gyttja samples in this research demonstrated $\delta^{13}C$-values between -28 and -30‰ (Table 2.1), suggesting a considerable contribution of terrestrial plants. Our $\delta^{13}C$-values are comparable to those measured by Törnqvist et al. (1992) and for this reason, we expect ageing effects will be of the same order of magnitude, i.e. 0.2 to 0.6-ka. Two samples at site 10c formed the exception to the rule with $\delta^{13}C$-values of -34.32 and -32.13. These samples possibly suffer from a hardwater effect over 0.6-ka. Nevertheless, it has to be stressed that the ^{14}C-dates both spatially and chronologically (successive dates within one coring) gave coherent results.

Carbon-14 plateaux exist around 12.7, 10.4 and 10-ka BP. The radiocarbon time scale can be calibrated against a calendar-year time scale using the dendro-calibration curve, but bulk-sediment based radiocarbon dates are likely to give erroneous ages when calibrated against a calendar-year chronology (Wohlfahrt, 1996). Therefore, we refer to uncalibrated ^{14}C-dates throughout this paper to facilitate direct comparison with literature and the chronostratigraphical subdivision of Mangerud et al. (1974). In this subdivision, the Bølling ranges from 13 to 12-ka BP, the Older Dryas from 12 to 11.8-ka BP, the Allerød from 11.8 to 10.8-ka BP and the Younger Dryas from 10.8 to 10-ka BP. It has to be stressed here that by using these terms, we refer to chronozones. These zones have been based on bulk ^{14}C-dates, like our age determinations, and should not be confused with biostratigraphical zones. Fore sake of clarity, we will refer to pollen or palynological ages when Bølling or Allerød biozones are involved.

2.4 RESULTS

Main residual channels and infillings

Most channels have not been filled in to the level of their corresponding meander pointbars and still form depressions on their floodplains. The top parts of clastic infillings of the Bølling/Allerød residual channels are 1-1.5 m below the Bølling/Allerød floodplain levels, while Younger Dryas and Holocene channels are almost filled up to their respective floodplain levels (Fig. 2.5). In general, the residual infillings show gradual texture fining, with younger channels containing the most clayey sediments. The deep Bølling/Allerød channels are filled in with calcareous gyttjas and strongly humic clays with sandy to sandy loam intercalations. A meander neck cut-off north of Beugen and the Younger Dryas channels predominantly contain clastic sandy and silty loams. Holocene channels mainly contain silty loams and silty to non- or slightly organic clays. Most residual channels have registered a period of peat formation, which often dates from the Atlantic period (e.g., 7320-5340 BP at site 2 and 6800-3480 BP at site 12). For the sake of brevity, only the infillings and ^{14}C-ages of main residual channels will be treated.

Pleniglacial/Bølling

A 1.2-m deep Pleniglacial residual channel in the Bosscherheide sandpit (site 3) was infilled with fine micaceous sand, like channels in the Panheel and Grubbenvorst exposures (Kasse et al., 1995; Van den Berg, 1996). Two distinct lime-free layers of organic matter at the base and the middle part of this channel have been dated (Fig. 2.6). The lower layer (small twigs) yielded 13280 ± 70 BP (GrN-21881), while the upper layer (small twigs, leaves and moss fragments) gave 13760 ± 70 BP (GrN-21882). A thin and horizontally deposited loam layer marks the top of

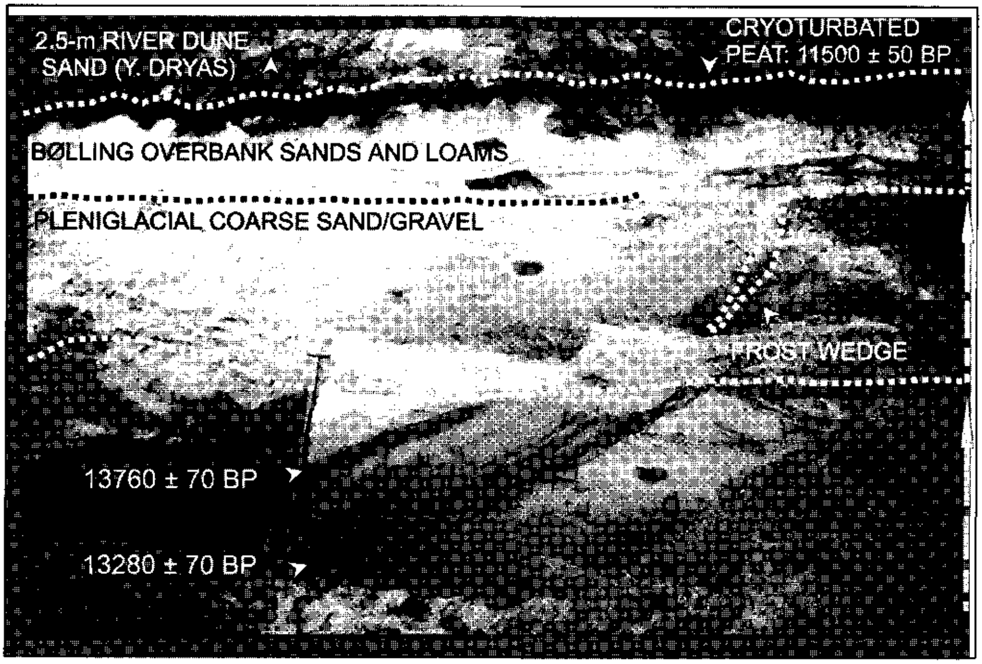

Figure 2.6 Late Pleniglacial residual channel in the Bosscherheide sandpit (Site 3b; Figures 2.3 and 2.5) filled in with organic material (dated 13760 ± 70 and 13280 ± 70 BP) and fine micaceous sand, erosively overlain by gravels and Bølling-age overbank sediments. Note horizontal loam at the right-hand edge of the channel, the frost wedge above it and the cryoturbated organic horizon of Allerød age. The spade is 1.12-m.

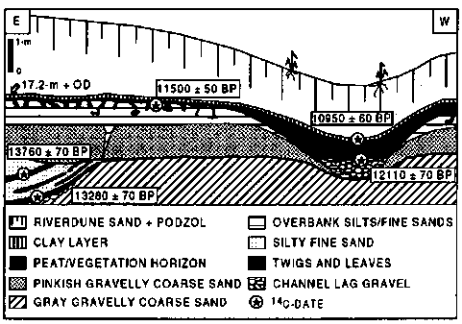

Figure 2.7 The lateral transition from a Late Bølling-Allerød peat in a 0.5-m residual channel (right) to a cryoturbated peaty horizon above the Late Pleniglacial channel (left) in the Bosscherheide sandpit (sites 3a & 3b in Fig. 2.6). The Bølling-Allerød ^{14}C-dates are taken from Bohncke et al. (1993).

the corresponding sedimentary unit on the right-hand side of the channel. This unit and the top of the channel are erosively overlain by a pinkish gravelly unit of 1.4-m, indicating that the floodplain still aggraded after 13280 BP. Grey overbank silts and fine sands in turn cover the pinkish gravelly unit. A frost wedge penetrating downwards from the top of the gravel unit and the textural difference with the overlying overbank sediments indicate a palaeofloodplain surface. A cryoturbated peaty horizon covers the overbank sediments. This horizon could be followed laterally to the peaty infilling of another residual channel with a depth of 0.5-m (Fig. 2.7). Bohncke et al. (1993) dated the basal peat in this channel at 12110 ± 70 BP (GrN-13382) and the base of the lateral extension of the peat over the overbank silts at 11500 ± 50 BP (GrN-12165). As no silts were present in the channel, the overbank silts and fine sands must have been deposited before 12110 BP, i.e. during the Bølling.

On top of the same floodplain, Teunissen (1983) dated a river dune sand blown over a peat layer between 12210 ± 90 BP (GrN-4787) and 10870 ± 100 BP (GrN-4786) near the Heerenven lake (site 18, the 1991 1:25,000 top. map erroneously indicates Westmeerven). A location nearby yielded an oldest peat age as early as 12760 ± 150 BP (GrN-4478). Thus, overbank silt sedimentation ended 12210 to 12110 BP and maybe as early as 12760 BP, confirming a Bølling age for floodplain formation and abandonment. The Heerenven lake itself is a remnant of a Weichselian Meuse branch belonging to this floodplain. Teunissen (1983) assumes the branch has been abandoned during the Late Allerød or early Younger Dryas, based on a Younger Dryas palynological age. Our oldest date of the Heerenven channel (11520 ± 90 BP; GrN-23425) suggests the channel has been abandoned earlier and that the channel was active until the Early Allerød. Palynological re-evaluation also points to an Early Allerød pollen age (Hoek, pers.com.). Most likely, the Heerenven channel has been abandoned later than the Bølling floodplain, due to rapid downcutting of channels (see below).

On the left bank (west) of the present Meuse, small remnants of the Bølling floodplain can be found near the villages of Beugen and Meerlo-Lottum. We dated the first gyttja infilling of the well-expressed neck cut-off palaeomeander near Beugen (site 1c) to 12330 ± 170 BP (GrN-23416). Kasse et al. (1995) found an Early Allerød pollen age. This corresponds to 11.7-11.5 ka BP according to the well-dated regional Late-glacial pollen zones for the Netherlands (Hoek, 1997). Both datings suggest that channel migration already ended during the Late Bølling. A later check by AMS dating on seeds from a new sample of the lowest part yielded a much younger (Late Boreal) age of 8390 ± 70 BP (UtC-6893). This date conflicts with palynological data and present interpretations of the presumed Late-glacial infillings, as it suggests Early Holocene activity and a Late Boreal to Atlantic infilling for the channel. A Late Boreal age is very unlikely, because pollen of Holocene thermophilous trees that are typical for the Late Boreal and Atlantic (e.g. *Corylus*, *Quercetum mixtum*, *Alnus*; Teunissen, 1990) are absent in both the gyttja and the overlying silty clay (Kasse et al., 1992). Moreover, the silty clay has a typical Younger Dryas signature with low pollen percentages of *Pinus* and *Betula* and high percentages of herbs, grasses and sedges (Hoek, pers. com). Thus, we have to assume the AMS sample has been contaminated with younger material during sample treatment (cf. Wohlfahrt et al., 1998) or measurement and reject its date.

Southwest of Meerlo, we dated a very humic calcareous clay at the base of a 4.7-m deep palaeomeander (site 11a) to 12500 ± 120 BP (GrN-22012). Although the δ^{13}C-value of -30.03 (Table 2.1) gives little reason to doubt, ageing effects cannot completely be ruled out. Allowing a maximum correction of 0.6-ka, an approximate age of 12.5-0.6 = 11.9-ka is found. Both dates suggest channel activity during the Bølling. However, pollen analysis (diagram Meerlo) of a basic infilling from the same channel indicated a Younger Dryas age, leading Kasse et al. (1995) to Allerød channel activity. This difference is probably due to different sampling sites, for the deepest Allerød age gyttja may have been missed during palynological sampling. Calcareous humic clay was deposited in the Meerlo channel till 11320 ± 210 BP (GrN-22011), followed by calcareous gyttja sedimentation, until at 9590 ± 90 BP (GrN-22010) peat started to grow.

South of Meerlo, we dated the oldest calcareous gyttja of a 5.3-m deep main channel near Keuter (site 7) at 11780 ± 270 BP (GrN-22008). The Geological Survey (NITG-TNO) dated the oldest calcareous gyttja in a well-expressed broad meander 5-km west of Venlo at Dubbroek (not on map) to 11830 ± 180 BP (GrN-18605). Because channel downcutting and migration must have occurred before channel infilling, we conclude that the Meuse was a fully developed and meandering river having deep channels already during the Late Bølling. However, this is in contrast to the interpretations of Huisink (1997).

Allerød

A small meander loop near Dubbroek (De Berckt, ~3-km southwest of Venlo; not in Figure 2.3) gave an age of 11460 ± 80 BP (GrN-18607) indicating that channels still meandered during the Early Allerød. The most extensive part of the Allerød floodplain can be found between Boxmeer and Swolgen. Here, well-curved channels have laterally eroded the slightly higher Pleniglacial and Bølling floodplain levels to the west. Just south of Boxmeer, the Allerød floodplain branched off to the east, where it has been eroded by the Holocene Meuse. The ages of the earliest sandy gyttjas are 11280 ± 80 BP (GrN-22004) near Groeningen (site 4) and 11260 ± 110 BP (GrN-22016) near Schafferden (13). Furthermore, the ^{14}C-age at site 4 matched the Late Allerød pollen age (Bohncke, pers.com.). Note that a large, more elongate channel near Smakt (14) cuts off these channels, in accordance with its younger date: 11170 ± 230 BP (GrN-22020). Other remnants of the Allerød floodplain level and related residual channels occur south of Meerlo and west of Lottum. A calcareous gyttja at the base of a coversand-buried residual channel at the Looiveld site (9) gave 10950 ± 280 BP (GrN-22010). Berendsen et al. (1995) obtained an age of 10720 ± 60 BP (GrN-15023) for a meandering Meuse channel about 20-km to the north-west in the Maas and Waal area. The oldest infilling of the residual channel west of Lottum (10c) has been dated 11640 ± 220 BP (GrN-23422), but probably suffers from ageing effects over 0.6-ka BP in view of its low δ^{13}C-value. Taking this into account an approximate age of ± 11-ka BP is found, which is in accordance with the Younger Dryas palynological age (Schuitwater pollen profile; Kasse et al., 1995).

Summarising, most Allerød channels were active meandering until 11.3-11.2 ka BP, but some shallower ones until the early part of the Younger Dryas Stadial. After abandonment of the floodplain, aeolian sand was deposited at some sites on this terrace level. Most sites have about 1.7-m aeolian cover, but at sites 9 and 14 up to 3-m is found. The aeolian cover (Younger Coversand II; Stiboka, 1975) is most probably time equivalent to the river dunes deposited on the Bølling floodplain on the east bank of the Meuse during the later part of the Younger Dryas (10.5-10.15 ka BP; Bohncke et al., 1993; Isarin, 1997). In the Heerenven channel (18), we dated an aeolian influx after 10.590 ± 70 BP (GrN-23424). At some places on this terrace level, prolonged anthropogenic manuring led to over 1-m thick reddish brown plaggensoils (Van de Westeringh, 1973).

Younger Dryas

The Younger Dryas floodplain can be found mainly south of sandpit Bosscherheide and as a small ribbon east of Lottum. Kasse et al. (1995) ascribe this floodplain with multiple shallow channels and mainly clastic infillings to a braided river system. At the site Kasteelweg (site 6), they palynologically dated the first infilling of a residual channel into the later part of the Younger Dryas and the early Preboreal. Near Linksstraat (8) and Heuloërbroek (5), the Geological Survey (NITG-TNO) dated the earliest gyttjas at 10850 ± 70 BP (GrN-18612) and 10600 ± 60 BP (GrN-18615) respectively. Teunissen (1990) found a Preboreal pollen age for the oldest peat in a driftsand-overblown residual channel (21), connected with a small floodplain remnant south of Gennep. In an Allerød residual channel near Smakt (site 14), we encountered a thin gravel layer intercalated between Late Allerød (11170 ± 230 BP; GrN-22020) and Preboreal (10350 ± 130 BP; GrN-22019) gyttjas, indicating that this channel has been re-used during the early part of the Younger Dryas.

Holocene

The Holocene floodplain can be traced as a continuous strip along the present Meuse. Northeast of Swolgen, we dated the earliest residual channel infilling at the Ooijen site (12) to 9170 ± 200 BP (GrN-22015). Kasse et al. (1992) palynologically demonstrated that infilling of this channel started shortly after the Preboreal/Boreal transition. Berendsen et al. (1995) found a comparable oldest age of 9240 ± 120 BP (GrN-18095) for an Early Holocene 6-m deep main channel in the Land van Maas en Waal, 20-km downstream of our study area. Likewise, Mingaars (1995) found a Late Boreal pollen age for the earliest infilling of an Early Holocene 7.2-m deep channel near Wanssum/Blitterswijck. Recoring this boring (site 2) we found a channel infilled with clayey gyttja and peat and the deepest part with sandy gyttja and gravel at 7.8-m. The peat interval was dated between 7320 ± 80 BP (GrN-23418) and 5340 ± 80 BP (GrN-23417). Unfortunately, there was too little bulk material to date the deepest sample. We expect that the extra 0.6-m represent the Late Preboreal and Early Boreal part of the infilling.

Floodplain levels

We identified three Late-glacial palaeofloodplains and one Holocene level of different fluvial generations based on topographical elevation, channel morphology and the ages of the oldest organic residual channel infillings. Due to the topographical low and wet positions of residual channels, registration of organic infillings will have started immediately after abandonment of the floodplain. E.g.: an oldest ^{14}C-age of 11780 BP indicates floodplain abandonment around 11780 BP. However, the channels on this floodplain carried water and migrated prior to 11780, i.e. the corresponding floodplain was in use during the Bølling. Our terrace ages thus relate to palaeofloodplain and channel activity. Both the spatial terrace map (Fig. 2.3) and the ideal "cross-section" (Fig. 2.5) show a Bølling level, an Allerød level, a Younger Dryas level and a Holocene level. For the 40-km Venlo-Boxmeer stretch, the gradients of the Bølling and Allerød floodplain levels are 15.0 and 12.3-cm/km respectively, while the Younger Dryas gradient is 14.3-cm/km and the Holocene gradient is 11.6-cm/km (Fig. 2.4). The pointbars of the Bølling age palaeomeander near Beugen are found at a lower elevation than was predicted by the Bølling palaeofloodplain gradient line. Probably Late-glacial Rhine influence or post-depositional subsidence is involved at this site, rather than a younger fluvial generation (see above).

With the help of the floodplain gradients, we calculated the elevations of the Bølling, Allerød, Younger Dryas and Holocene floodplains at 125-km downstream the Dutch-Belgian border at 17.43, 17.15, 14.88 and 14.05-m respectively. These values obviously differ from those at 100 and 150-km due to different floodplain gradients. The Bølling and Allerød floodplains converge near 135-km and so do the Younger Drays and Holocene floodplains near 150-km. The difference between the Bølling and Allerød floodplains demonstrates 0.3-m floodplain lowering (incision) at 125-km, but amounts 1-m and –0.4-m at 100 and 150-km respectively. Incision during the Allerød Interstadial results in a floodplain at a 2.6-m lower level during the Younger Dryas (2.7-m and 2.4-m. at 100 and 150-km respectively).

Incision from the Younger Dryas floodplain to the Holocene floodplain minimally amounts to 0.8-m. (1.5-m and 0.2-m). We refer to a minimal incision, because deforestation and agricultural practices have augmented clay sedimentation since the Late Atlantic (Teunissen, 1990). Consequently, the top of the present Holocene floodplain has been artificially raised with 1-2 m. The end of peat accumulation is dated 5340 BP and 3480 BP at sites 2 and 12 respectively (Fig. 2.5).

The Meuse floodplain thus has been lowered to successively lower levels due to Late-glacial incision. The difference between Bølling and Holocene floodplains suggests minimally 3.4-m incision at 125-km (4.2 and 2.5-m at 100 and 150-km respectively). Allowing for the Holocene artificial rise, Late-glacial incision at 125-km actually amounts to 4.4-5.4 m. Likewise, the base of a 7.8-m Preboreal residual main channel (site 2) is found some 5-m lower than the bases of the deepest Bølling residual main channels (Fig. 2.5: sites 1, 7 and 11a), after correcting for gradient of course. In contrast to the lower Holocene bases, the bases of most Allerød and Younger Dryas main channels (e.g. sites 6, 8, 10, 13, 14) are found higher within the ideal "cross-section" (Figures. 2.5 and 2.8b) and show constantly younger dates (Table 2.1). Because numerous borings were done at sites where highest channel depth could be expected (e.g. at the outer sides of meander bends) we interpret these higher base elevations as a systematic trend, rather than coincidental differences in local channelbase topography. In our view, this trend suggests the Late Allerød and Younger Dryas channels have become shallower in terms of bankfull discharge and show an increasing width/depth ratio.

In Figures 2.8A and 2.8B, a schematic floodplain plan view as well as the floodplain levels and corresponding bases of main channels are plotted versus time. As expected for a braided aggrading system, Late Pleniglacial channels are shallow (e.g. 1.2-m at site 3 (Bosscherheide) dated at 13280 BP). Pleniglacial channels of comparable depth were found in the sandpits Panheel and Grubbenvorst (Kasse et al., 1995; Van den Berg, 1996). Between 12.5-ka and 11.3-ka BP deep channels are found, which become gradually shallower again around 11.3-ka BP. Concurrently with the shallowing of the channels, a steeper river gradient is found for the Younger Dryas floodplain (Fig. 2.4). The floodplain width steadily decreases towards the Holocene.

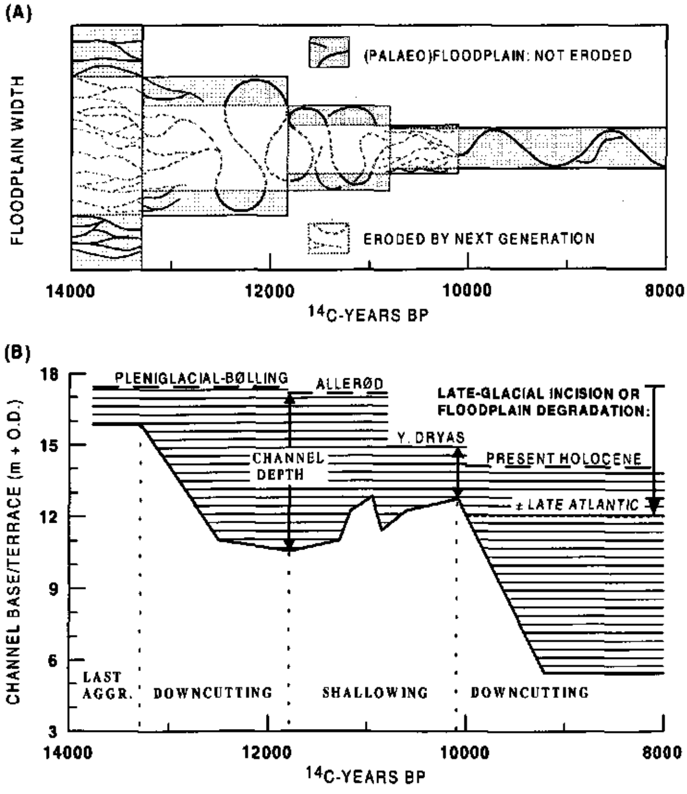

Figure 2.8 *(A)* Schematic representation of floodplain width and channel morphology in time. *(B)* Channelbase and terrace elevation (m above Ordnance Datum) versus time (uncalibrated ^{14}C-years) for an ideal cross-section at 125-km downstream of the Dutch-Belgian border. Original elevations have been recalculated using the floodplain gradients from the linear fits in Fig 2.4. The Late Atlantic floodplain is ~1-2 m lower, as augmented clay sedimentation due to human interference has raised the present Holocene floodplain.

2.5 DISCUSSION

River dynamics

After the last phase of Pleniglacial aggradation (ca. 13.3-ka BP), channels downcut rapidly at the beginning of the Late-glacial and soon follow a meandering course. The Beugen meander oldest infillings have been dated 12330 BP and palynologically as Early Allerød (Kasse et al., 1995). The deep, highly sinuous channels of Meerlo, Dubbroek and Keuter yielded 12500 BP, 11830 BP and 11780 BP. Overbank fine sands and silts from the meandering Bølling Meuse have been deposited on the floodplain east of the present river before 12.2-ka BP (e.g. Bosscherheide). Rapid channel downcutting and high-sinuosity meandering river activity should therefore be placed during the Bølling. Similarly, the small meandering river Mark in the southern Netherlands reached its deepest point of downcutting before 11850 BP (Bohncke and Vandenberghe, 1991). A major downcutting phase in the lower reach of the River Oise, which like the Meuse has its upper course in the northeast Paris Basin, ceases before 12400 BP (Pastre et al., 1997). The Warta and Vistula in Poland have a meandering course as from ± 13-ka BP (Starkel, 1991). Floodplain lowering during the Bølling is still of minor importance and amounts to only 0.3-m.

Channel downcutting ceases around 11.9-ka, just after the Older Dryas cooling event. River Meuse low-sinuosity meandering channels create a new floodplain between Boxmeer and Meerlo during the Allerød. The shallower and laterally eroding channels on this floodplain were in use until ± 11.3-ka BP. Kasse et al. (1995) and Huisink (1997) inferred a Bølling age (± 12.7-ka) for this same stretch based on terrace elevation, because they did not find suitable dating sites. Assuming this Bølling age, Huisink interprets this terrace level to be a transitional phase between the Late Pleniglacial braiding system and a high-sinuosity meandering system at the end of the Allerød. A transitional phase from the Younger Dryas braidplain to the low-sinuosity Holocene floodplain was not found, however. Solely using terrace elevation in this area to determine the age of floodplain activity can be disputed due to the presence of anthropogenic and aeolian covers. Dating of the two meander generations, especially of the transitional low-sinuosity phase is essential for a correct interpretation of Late-glacial river dynamics. We were able to date three main channels, which gave coherent dates around 11.3-ka BP. Since the channels have always been in a topographical low and wet position, we reject the possibility of a 1.4-ka hiatus (12.7-11.3) between channel abandonment and the start of organic formation. Combining the datings with floodplain level and channel patterns, we interpret this floodplain stretch to be of Allerød age. The ages of two meander generations near Dubbroek, verified by the Geological Survey, support this interpretation. The largest high-sinuosity meander was active prior to 11830 ± 180 BP (GrN-18605), whereas the small low-sinuosity meander (De Berckt) was abandoned before 11460 ± 80 BP (GrN-18607). Thus, in our view the Allerød floodplain represents the transition from a high-sinuosity meandering system with deep channels during the Bølling to the braided system with shallow channels during the Younger Dryas. The absence of a transitional phase from a Younger Dryas braiding system to the Holocene meandering system makes sense now. A transitional phase could have been expected if the disputed floodplain stretch were to be placed between a Pleniglacial braiding and Allerød meandering system.

The channels on the Allerød floodplain have been abandoned between 11.3 and 10.9-ka BP (Late Allerød) just before the onset of the Younger Dryas period. For the same period, Huisink (1997) envisages a major floodplain-wide incisional phase accounting for 8 to 10-m erosion of terrace material, followed by a phase of aggradation. If floodplain-wide erosion took place, one would expect to find one or more deep channel(s) responsible for this substantial lateral and vertical removal of 8-10 m terrace material. Various Late Allerød channels and Early Younger Dryas channels at sites Linksstraat (8, 10850 BP), Heuloërbroek (5, 10600 BP) and Kasteelweg (6), however are only 2-3m deep. Gravelly sediments were deposited at about the same level in the Smakt residual channel (14) between 11170 and 10350 BP. The absence of deep Early Younger Dryas channels, shallowing of the channels already during the Late Allerød and coarse clastic sediments testifying to higher energetic conditions in this period, make an 8-10 m incisional phase at the start of the Younger Dryas rather unlikely. We calculated the Younger Dryas floodplain at a 2.4-2.7 m lower level compared to the Bølling/Allerød levels. Kasse (1995) also mentions 2-3 m erosion, and assumes the erosion to result from higher peak discharges during the Late Allerød. Apparently, the Meuse lower reach first experienced rapid channel downcutting without substantial floodplain lowering before 11.9-ka BP and then lowered its floodplain by lateral erosion predominantly during the Allerød.

In general, sediments become finer as they are younger, except for the Younger Dryas period when a braiding multi-channel system is in use (Bohncke, 1993; Bohncke et al., 1993; Kasse, 1995). The coarser channel infillings as well as the shallower and more braided-type channels around the Allerød/Younger Dryas transition at 11-ka BP all point to higher energetic activity of the fluvial system. Both a floodloam in Bosscherheide (Kasse, 1995) and silty clay deposition in the Keuter (7) and Beugen (1) residual channels indicate renewed flooding of older floodplains and channels. At the same time, coarse sediments are deposited in the Smakt (14), Looiveld (9), Swolgen 3 (16b), Heuloërbroek (5) and Kasteelweg (6) residual channels. During the Younger Dryas, the supply of less weathered non-pedogenic clay minerals increases (Chapter 4, Tebbens et al., 1998), indicative of the erosion of immature soils, loess and supply of hillslope gelifluction debris (Pastre et al., 1997). A coarsening of sedimentation and a phase of deposition in the second half of the Late-glacial Interstadial and Younger Dryas has also been recognised for the River Oise (Pastre et al., 1997), several British rivers and the Warta in Poland (Collins et al., 1996).

Evidence for renewed channel downcutting during the Preboreal is provided by a residual main channel near Blitterswijck (2) having its base 7.5-m lower compared to the Younger Dryas channels. In several other residual channels the transition from aqueous gyttja to terrestrial peat accumulation suggests a lowering of the regional water level during the Early Preboreal. Figures 2.3 and 2.5 illustrate this by the following dates: sites 4 and 13: 10220 BP, site 14b: 9800 BP, site 11: 9590 BP and Dubbroek: 9740 BP. Both increasing (cold and) dry conditions during the later part of the Younger Dryas and downcutting of the deep Preboreal main channel could be responsible for water table lowering. The Ooijen channel (12) oldest infillings and the palynological data for the Blitterswijck channel (site 2; Mingaars, 1995) indicate that channel activity ended before 9170 ± 200 BP. An 8-m deep Holocene channel found in the Maas en Waal area was active before 9240 ± 120 BP at least, and comparable deep channels had been abandoned as early as 9800 ± 60 BP (Berendsen et al., 1995). The deep single meandering channels therefore were incised during the Early Preboreal (10.2-9.8 ka BP). Similarly, Pastre et al. (1997) found that the Rivers Oise, Marne and their tributaries downcutted very rapidly (within centuries) during the Early Preboreal. The earliest channel infillings in the Oise and Marne valleys dated ca. 9690 BP and those of the tributaries dated between 9500 and 9000 BP. The difference between the Meuse Younger Dryas and present-day Holocene floodplains suggests a minimal floodplain lowering of 0.8-m at 125-km (1.5-m and 0.2-m for 100 and 150-km respectively). Allowing for an anthropogenic raising of the present-day floodplain with 1-2 m, the Preboreal floodplain must have been 1.8-2.8 m lower. Thus, the Preboreal incision phase for this river stretch is of the same order of magnitude as the one during the Bølling-Allerød Interstadial.

Landscape changes and river response

The study of Late-glacial and Holocene Meuse palaeofloodplains revealed lowering and narrowing of the river floodplain (incision) during the Late Glacial and Preboreal, with the main phases of floodplain lowering during the Allerød Interstadial and Early Preboreal.

Bull (1991) studied the influence of Late Quaternary climate and landscape change for the temperate-zone Charwell River in New Zealand. He considers complex-response fill-cut terraces to reflect pauses in stream-channel downcutting or the degradation process. During glacials excessive sediment supply over river discharge causes floodplains to aggrade when long-term tectonic uplift cannot be compensated for by river downcutting. As soon as the climate ameliorates and excessive sediment supply fades, the river will catch up the downcutting trend and respond with degradation, ultimately resulting in floodplain lowering. Similarly, the Meuse fluvial system aggraded until the Late Pleniglacial and formed three fill-cut terraces during Late-glacial valley floor degradation. Van den Berg (1996) considers these fill-cut terraces as the erosional types of millennial-period fluvial cycles, driven by periodic changes in the strength and position of storm tracks over the North Atlantic.

The increase of temperature, precipitation and the proportion of precipitation falling as rain are the key climatic changes for the latest Weichselian (Bull, 1991). Isotope research on the Greenland ice cores has proven that interstadials began quite abruptly (within a few decades) and terminated gradually in the North Atlantic region and northwestern Europe (Dansgaard et al., 1989; Johnsen et al., 1992). Walker et al. (1994) and De Graaff and De Jong (1995) report rapid warming at the beginning of the Bølling Interstadial, gradual cooling during the Allerød and a cold phase between 11.5 and 11.3-ka BP. Guiot and Couteaux (1992) reconstructed a marked increase in mean annual temperature and mean annual precipitation for the period 15-8 ka BP (Fig. 2.9) for the Echternach area in Luxemburg (altitude 170-m). This region can be considered to be representative of the major part (Ardennes) of the Meuse drainage basin. The climatic amelioration is interrupted during both the Older Dryas and Younger Dryas Stadials. The Meuse responded to these major increases in temperature and precipitation by changing the width/depth ratio of its channels. The interaction between landscape stability, vegetation cover, soil formation, sediment supply and river discharge is crucial in this respect in causing complex-response lag effects in the channel downcutting process.

The Meuse lower reach demonstrates three phases of channel adjustments, namely rapid downcutting, slow aggradation and rapid downcutting again. Rapid channel downcutting occurred for the first time during the early part of the Bølling (± 12750-12500 BP). A high-sinuosity meandering system evolved, having a single channel with a small width/depth ratio during the (Late) Bølling. Permafrost completely disappeared during the Bølling (Hoek, 1997). The interplay between falling groundwater levels, increasing discharges and permafrost meltwater amounts and decreasing sediment supply owing to regionally increasing vegetation cover must have favoured this rapid channel downcutting (Pastre et al., 1997). Probably as a response to the Older Dryas cooling event and to rising groundwater levels during the Early Allerød (Hoek, 1997) Meuse channel downcutting ceases around 11.9-ka BP. The River Oise lower reach records a short sedimentation phase for this same period (Pastre et al., 1997) and similar observations are made in the River Somme (Antoine, 1997). Simultaneously, the Meuse adjusts to gradual Allerød cooling by transforming into a low-sinuosity meandering system characterised by shallower channels with higher width/depth ratios. By this time, lateral erosion starts to dominate over vertical erosion (Starkel, 1991), leading to floodplain lowering during the Allerød Interstadial. Renewed aggradation within the channels during the Late Allerød-Early Younger Dryas (± 11.3-10.9 ka) ultimately leads to a braided system with highest channel width/depth ratios during the Younger Dryas. Discontinuous permafrost is present again (Kasse, 1995). Floodplain lowering ceases and the Meuse aggrades again for a short while. A second phase of rapid channel downcutting occurs at the Younger Dryas-Preboreal transition (± 10200-9750) and results in the present Holocene low-sinuosity system. After channel downcutting has ceased, lateral erosion lowers the floodplain for the second time. If this follow-up proves to be correct, this would imply a quite direct response of the Meuse to Late-glacial climate change.

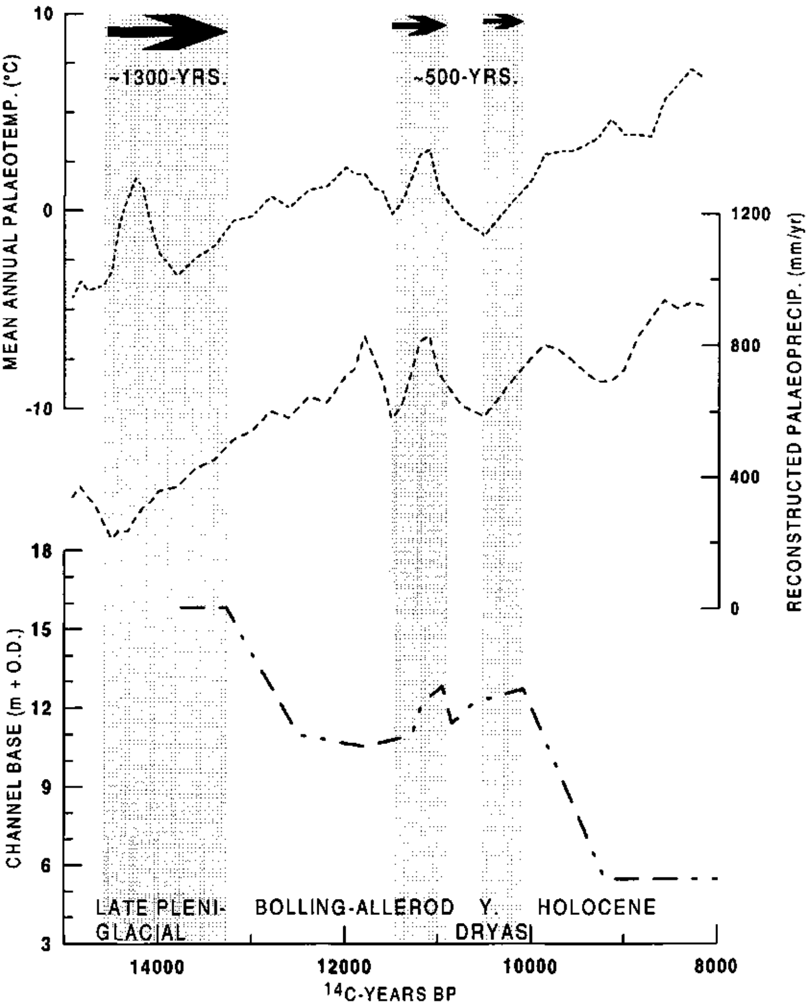

Figure 2.9 Lag time (uncalibrated ^{14}C-years) between climatic amelioration in the upstream Ardennes area (after Guiot and Couteaux, 1992) and onset of channel downcutting in the lower reach of the River Meuse. On the Y-axis, m + O.D. stands for m above Ordnance Datum. Note two main phases of channel downcutting during the early Bølling (~13-12.5 ka BP) and Preboreal (~10.2-9.2 ka BP) and shorter lag time for the Preboreal phase of channel downcutting.

In Figure 2.9, we plotted bankfull channel depth versus reconstructed palaeotemperature and palaeoprecipitation for the upstream Ardennes area (Guiot and Couteaux, 1992). Bankfull channel depth has increased 500-1300 years after a major temperature (and precipitation) rise, compared to channel depths from the preceding colder period. In our opinion, the different lag times between climatic amelioration and channel downcutting, which are in the order of centuries, suggest a major control of vegetation growth and soil formation on sediment supply. These lag periods probably differ due to the different initial conditions for a largely barren Pleniglacial landscape and a vegetated landscape during the Allerød and Younger Dryas.

Permafrost disappeared and a vegetation cover established in the Pleniglacial tundra landscape owing to rapid temperature rise since 14.5-ka BP (Guiot and Couteaux, 1992; Hoek, 1997). Decreasing gelifluction and hillslope processes will have diminished the sediment supply rapidly (Pastre et al., 1997). The combination of lower sediment supply and an increase in mean annual precipitation and permafrost/snowmelt peak discharges (due to increased January precipitation) favours an increasing river stream power. Thus, more stream power is available to entrain and transport sediment load in the channels and the ratio of stream power to resisting power is likely to exceed unity. The channels then start downcutting and vertical erosion will predominate.

A more gradual response can be expected and is observed for the Allerød to Younger Dryas gradual temperature and precipitation decrease. The break-up of soils needed to supply extra sediment in the whole drainage area obviously lags the cooling trend due to the Bølling-Allerød vegetation cover. To understand the full variability of river behaviour, one should include information for the whole drainage basin (e.g. Collins et al. 1996; Veldkamp and Van Dijke, 1998). Pastre et al. (1997) reconstructed an erosion and soil formation phase during the Bølling-Allerød Interstadial, followed by a main period of sedimentation during the Younger Dryas in the lower reach valleys of the Rivers Oise and Marne in the northeast Paris Basin. Lefèvre et al. (1993) were the first to investigate the Meuse upper reach before it enters the Ardennes. They conclude that wide channels, incised (during the Bølling?) in Late Weichselian gravels, have been progressively infilled under calm hydrodynamic conditions during the Younger Dryas. Information for the Bølling-Allerød Interstadial is not mentioned.

The break-up of interstadial vegetation cover and soils on a regional scale can be expected to occur earlier in the Ardennes area than in the French and Dutch parts of the fluvial system in view of the higher (colder) altitudes, steeper slopes and presence of coarser material. The Echternach pollen diagram (Guiot and Couteaux, 1992) indicates that the Bølling birch park landscape was replaced by an Early Allerød pine woodland, which gradually declined during the Middle and Late Allerød. Pine woods were established on the Meuse lower reaches in the Netherlands during the Late Allerød, but died off rapidly during the Early Younger Dryas (Hoek, 1997). Declining vegetation provoked renewed break-up of soils in at least the Ardennes catchments and probably also Dutch parts of the drainage area. Veldkamp and Van Dijke (1998) found that the Meuse lower reach is likely to register aggradation when the upper reach experiences erosion (break-up of soils) due to climate change. The associated increase in sediment load in lowland fluvial systems is reported for the period 11.3-10.9 ka BP (Bohncke, 1993) and is expressed by a coarsening of residual channel infillings around 11-ka BP. The gradual sediment-load increase results in a gradual increase of resisting power. Simultaneously, a 40-50% reduction of precipitation (Guiot and Couteaux, 1992) must have led to a considerable decrease in discharge and available stream power. From the Middle Allerød, the Meuse lower reach was characterised by slow aggradation at the channel level and higher energetic conditions that finally changed it into a braiding river system. Similarly, the short Older Dryas event may be reflected not by an individual floodplain level, but by a pause in channel downcutting (cf. Pastre et al., 1997).

In addition to hillslope and gelifluction processes, aeolian influxes have been important. During a modelling exercise, Van den Berg and Veldkamp (1996) could only simulate the last large Pleniglacial aggradation phase by assuming considerable cover sand and loess influxes into the Meuse fluvial system. River dunes were deposited after 10.5-ka BP (Bohncke, 1993) at the east side of the Meuse valley and at various sites aeolian relief was found on top of Bølling and Late Allerød floodplains. Ashley and Hamilton (1993) proposed the Kobuk River in Alaska as a modern analogue for Late-glacial north-west European river systems with a considerable aeolian supply. The Kobuk River aggrades due to aeolian influxes during cold and dry periods and downcuts during moist interstadials with vegetation growth. Like the Pleniglacial influx, the

Younger Dryas cooling event may have caused an extra aeolian influx in addition to the increase of gelifluction debris from the Ardennes area.

A rapid 7-°C warming and a 50% precipitation increase at the end of the Younger Dryas (Dansgaard et al., 1989; Guiot and Couteaux, 1992) induced a period of renewed rapid channel downcutting during the Early Preboreal. Palynological data (Van Geel, 1989; Teunissen, 1990; Guiot and Couteaux, 1992; Bohncke, 1993) indicate the presence of closed pine forest vegetation from the Preboreal onwards in both the Ardennes as well as in the Netherlands. Younger channels showed progressively finer and organic sediment infillings, with most Holocene channels recording clay, gyttja or peat (Fig. 2.5). At the same time, clay mineralogy and geochemistry indicate a relative increase of more weathered and pedogenic-type clay minerals like smectite and vermiculite (Chapter 4, Tebbens et al., 1998). All features indicate relative stable landscape conditions with decreasing clastic input during the Holocene (Pastre et al., 1997). Peat accumulates during the warm and wet climate of the Atlantic climatic optimum (e.g., 7320-5340 BP at site 2; 6800-3480 BP at site 12). Human-induced deforestation and agriculture interferes as from the Early Subboreal (Teunissen, 1990). Fluvial activity resumes from this time on with silty loam deposition under high sedimentation rates. The present Holocene floodplain is therefore at a higher elevation than it would have been without human interference.

2.6 CONCLUSIONS

The study and radiocarbon dating of Late-glacial and Early Holocene palaeofloodplains, residual channels and their infillings in the 40-km Venlo-Boxmeer lower reach of the River Meuse reveals a non-linear response to Late-glacial climate change. At a cold-to-relatively-warm transition, the Meuse responded very strongly to a major temperature and precipitation rise by rapid channel downcutting. In the reverse situation, a slow return to aggrading conditions is found. Apparently, the Meuse fluvial system is very sensitive to sediment supply and aggrades during excessive supply of gelifluction material or aeolian influxes in a cold climate. As soon as this supply of material fades away due to interstadial or interglacial vegetation growth and soil stabilisation, a phase of rapid channel downcutting occurs due to lower sediment supply and higher discharges. Channel downcutting phases are followed by periods of floodplain degradation, as appears from constant lowering and narrowing of the Meuse floodplain. The rapid downcutting of channels is not completely compatible with other interpretations, which suggested gradual adaptations of the fluvial system on 1-10 ka time-scales. Therefore, one should acknowledge the difference between channel downcutting and floodplain lowering (incision, degradation) and consider the entire Meuse drainage basin when evaluating climatic responses.

Following a 5-7 °C mean annual temperature rise and fourfold precipitation increase, two phases of rapid channel downcutting set in during the Bølling and the Preboreal within, respectively, ca. 1300 and ca. 500-years. The different lag times between climatic amelioration and onset of channel downcutting probably depend on initial landscape conditions, suggesting a major influence of vegetation and soil development on sediment supply. Climatic deterioration induces a gradual return to aggrading conditions. Floodplains were lowered (incision) mainly during the Late Allerød and the Preboreal when channel downcutting paused or ceased and lateral erosion occurred. Total incision during the Late Glacial and Early Holocene for a section 125-km downstream of the Dutch–Belgian border amounts to 4.4-5.4 m.

Late-glacial development of the lower reach of the Meuse is interpreted as follows. The last Pleniglacial aggradation took place after 13.3-ka BP. Following a 5-°C rise of mean annual temperature around 14.5-ka BP and a fourfold increase in precipitation in the Ardennes drainage area (Guiot and Couteaux, 1992), the Meuse main channels rapidly downcut between 13.3-ka BP and ca. 12.5-ka BP. A high-sinuosity meandering system with deep channels (low width/depth ratios) is present between 12.5-ka BP and 11.9-ka BP. The Older Dryas cooling event and rising groundwater levels possibly restrained further channel downcutting around 11.9-ka BP. The Bølling floodplain and channels were left and a low-sinuosity meandering system created a new floodplain between 11.9-ka BP and ca. 11.3-ka BP. Floodplain lowering was still of minor importance (0.3-m).

From 11.3 to 10.9-ka BP, the River Meuse lowered its floodplain in the Venlo-Boxmeer lower reach with 2.6-m. Late Allerød/Younger Dryas climate deterioration ultimately culminated in a renewed increase in sediment load and a decrease in discharge, leading to a coarsening of the sediment supply, shallowing of the river channels and abandonment of the Late Allerød floodplain. A braided shallow multichannel system with higher width/depth ratios occupied the Younger Dryas floodplain.

Finally, a 7-°C mean annual temperature rise and 50% annual precipitation increase during the Early Preboreal (Dansgaard et al., 1989; Guiot and Couteaux, 1992) causes renewed rapid downcutting by a single low-sinuosity channel. This downcutting period certainly ended before 9.2-ka BP, but maybe as early as 9.8-ka BP. A second phase of floodplain lowering followed and involved 1.8-2.8 m, depending on the amount of anthropogenically induced rise of the present Holocene floodplain surface.

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CHAPTER 3

NATURAL COMPOSITIONAL VARIATION OF RESIDUAL CHANNEL INFILLINGS IN THE RIVER MEUSE LOWER REACH

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3 NATURAL COMPOSITIONAL VARIATION OF RESIDUAL CHANNEL INFILLINGS IN THE RIVER MEUSE LOWER REACH

ABSTRACT

Unambiguously pristine and predominantly unpolluted sediments from the Late Weichselian and Holocene infillings of River Meuse (Maas) residual channels in northern Limburg (The Netherlands) have been sampled to determine the natural compositional variation of the River Meuse suspended load.

Bulk geochemical and granulometric analyses demonstrate that ~70% of variation can be ascribed to hydrodynamic mineral sorting. Clay- and fine silt-sized phyllosilicates are the most important deterministic variable, hosting the bulk of Al_2O_3, TiO_2, K_2O, MgO and trace element variability (notably Ba, Cr, Ga, Rb and V). Quartz is abundant in the fine and coarse sand fractions. Na_2O and the Zr, Nb, Nd, Y-quartet relate to albitic feldspars respectively heavy minerals in the coarse silt fraction. Therefore, the granulometry should be quantified if geochemical baseline data for a particular geological unit or region are drawn up to evaluate allegedly polluted sediments.

Early diagenetic siderite and vivianite formation in the anoxic gyttja environment causes relative accumulations of Fe_2O_3, MnO, P_2O_5, Co, Ni and notably Zn above the phyllosilicate background values. These accumulations are natural and demonstrate that sediments with elevated trace metal contents are not necessarily polluted. However, very early atmospheric pollution in relation to ore mining and smelting activities in the Roman era probably caused elevated Pb-contents in Subatlantic humic clays and peat samples, long before the historic pollution of the Industrial Revolution started.

The Al_2O_3-, Fe_2O_3- and CaO-contents can be used to predict the trace element values as a function of sample granulometry, siderite/vivianite- and lime-content respectively. As such, they can provide a sound basis for environmental researchers to determine baseline values for heavy metals in bulk samples of fine-grained fluvial sediments.

3.1 INTRODUCTION

Geochemical appraisal of soils and sediments has become increasingly important the last decade. In both the Netherlands and other densely populated areas, there is a growing need for geochemical baseline data to define the natural compositional variation of the grounds we live on (Darnley et al., 1995; Darnley, 1997). Geochemical baseline data account for regional and local geological variation and constitute the natural background values to which presumed polluted samples can be compared. As such, they are indispensable because they provide a sound and realistic basis for environmental legislation and policy making (Salminen and Tarvainen, 1997; Plant et al., 1997). The pollution of overbank and floodplain sediments of the River Meuse and its tributary Geul with Pb, Zn, Cu and Cd due to ore mining in Belgium certainly began more than 350-years ago (Leenaers, 1989; Rang and Schouten, 1989). During two major flooding events in the Netherlands in 1993 and 1995, public concern arose with respect to the degree of pollution of the present suspended load of the rivers Rhine and Meuse (Asselman and Middelkoop, 1996; Bleichrodt and Ensinck, 1994; Goudriaan, 1995; Wolterbeek et al., 1996). However, surprisingly little is known about the natural background values of major and trace elements of fine-grained sediments within the floodplains of these rivers.

Moura and Kroonenberg (1990) and Hakstege et al. (1993) were the first to describe the bulk geochemistry of Quaternary Rhine and Meuse sediments by sampling deposits of diverse granulometry and several geological formations in the south-east of the Netherlands. They found that differences in main and trace elements within and between Dutch Formations are mainly related to sorting processes and provenance, causing variation in the sediment mineralogy. Huisman (1997) sampled the Early and Middle Pleistocene Tegelen and Kedichem Formations (Rhine and Meuse deltaic deposits) of the southern Netherlands subsurface. He reconstructed provenance-related mineralogical changes and additionally found that syn- and post-depositional diagenetic processes can affect the contents of Fe_2O_3, MnO, P_2O_5 and associated trace elements like As, Ba, Ni, Pb and Zn. Van der Sluys et al. (1997), Swennen et al. (1997) and Swennen and Van der Sluys (1998a, b) completed the Meuse data set by sampling Holocene overbank sediments at the surface and at depth to assess geochemical baseline data for all major and minor Meuse tributaries in the Ardennes low mountain range in Belgium. They focused on provenance characteristics and the <125-μm fraction of lithologically homogenous overbank profiles. Therefore, they only sampled the tributaries and not the Meuse itself and they paid less attention to granulometric differences.

The tributary source catchments and their related sediment supplies show a diversity of provenance characteristics and obviously will differ geochemically from the sediments of the main river. For environmental research on Dutch River Meuse sediments, however, it is essential to have data on the main river "Meuse mixed composition" itself. It is the purpose of this paper to characterise the natural compositional variation of Late Weichselian and Early Holocene fine-grained Meuse River sediments in relation to its granulometry. We will report on the results of a bulk geochemical and granulometric research of some 640 samples taken from the infillings of River Meuse residual main channels. After abandonment due to river valley incision or morphodynamic cut-off processes, the inactive residual channels provide perfect low-energy environments to act as natural sediment traps for the suspended load passing by during a flood. During flooding events, a large number of sediment sources in a variety of tributary catchments will supply eroded material in the form of suspended load to the main river. Part of the suspended load will stay behind in the residual channel sediment traps. As such, the silty to clayey residual channel infillings can be regarded as the unconsolidated fluvial equivalents of floodplain overbank profiles (Ottesen et al., 1989). The samples taken from fine-grained residual channel infillings will therefore reflect the natural compositional variation of the suspended load of the River Meuse for a record of flooding events. Special attention will be given to granulometry as well as the interaction between gyttja/organic matter characteristics and trace elements. Finally, some time-composition relationships will be discussed.

3.2 GEOLOGICAL SETTING

The natural background values of major and trace elements of present-day River Meuse suspended load deposits should be based on unpolluted fine-grained sediments with a pure Meuse provenance, i.e. without admixture of Rhine sediments. Furthermore, the sampled sediments should have a composition comparable to the present-day suspension load, implying that post-depositional diagenesis or pedogenesis should not have modified their composition on a geological time-scale. The sediments of the Upper Kreftenheye Formation and Betuwe Formation (Doppert et al., 1975) in central and northern Limburg (The Netherlands) that are deposited in residual channels fulfil these requirements (Figs. 2.3, 2.4 and 2.5). They have a pure Meuse origin, and because they are still unconsolidated, they are unlikely to have suffered from post-depositional alteration. The Upper Kreftenheye Formation was deposited during the Late Weichselian and Early Holocene (Törnqvist et al., 1994). The drainage area of the River Meuse has remained essentially unchanged during the last 250-ka

BP (Bustamante, 1976, Van den Berg, 1996), excluding effects of large-scale provenance changes on bulk geochemistry as have been described for the Rhine system (Huisman, 1997).

At the climatic transition from the Late Weichselian to the Holocene (Lateglacial, ~13-9 ka BP), the Meuse progressively lowered its floodplain and changed from a braiding multi-channel into a meandering single-channel river (Chapter 2; Berendsen et al., 1995; Kasse et al., 1995; Huisink, 1997; Tebbens et al., 1999a). Consequently, fine-grained overbank sedimentation started in the Weichselian Late Glacial to cover the Pleniglacial gravels and coarse sands with a loam bed (Makaske and Nap, 1995). Törnqvist et al. (1994) thoroughly described these overbank deposits and their related clayey, silty or gyttja residual channel infillings and defined them as the Wijchen Member of the Kreftenheye Formation. The Holocene deposits should formally be attributed to the Betuwe Formation (Doppert et al., 1975).

3.3 SAMPLING

We took 636 bulk samples at 33 sites (Figs. 2.3 and 2.5) in the currently net-depositional 40-km stretch of the floodplain of the medium size River Meuse (Maas) that integrates the combined provenance and sorting effects of a drainage area of 33,000-km^2. Therefore, they are natural composite samples of the floodplain sample type and are significant at a continental or global scale (Darnley et al., 1995). The samples originate from largely reduced, unconsolidated sediments (unripened in soil science terminology) that have been deposited in several residual channels during the last 13,000-years. The residual channel infillings offered wide dating possibilities via intercalated peat and gyttjas (e.g. Figs. 2.5 and 3.7) and enabled sampling of indisputable pre-industrial unpolluted sediments. Details on the sampling sites, ^{14}C-dates and morphodynamic aspects of river incision and channel development can be found in Chapter 2 (Tebbens et al., 1999a).

The sampling strategy and number of samples virtually cover the whole range of possible textures and lateral and downstream variation within the Meuse sediments of the Upper Kreftenheye and Betuwe Formations, ensuring a very high regional representivity. For comparison with geochemical baseline data, we note that preferably unconsolidated residual channel infillings under reduced conditions were sampled, instead of regolith or alluvial overbank A- and C-soil horizon samples. We even avoided soil horizons particularly, to prevent post-depositional overprinting effects owing to recent soil formation or anthropogenic pollution to show up in the dataset. The minimum sample depth was 0.3-m, while the maximum depth was 7.7-m below ground surface. Fluvio-aeolian sands in the sandpits at Grubbenvorst and Panheel (not on map) could only be sampled above the groundwater table, however. Bulk clays and silts were sampled per lithological unit or every 10-cm in case of layers thicker than 10-cm, while sands were sampled per lithological unit to evaluate the influence of admixture of sand-sized minerals on the bulk geochemistry of the sediments.

3.4 BULK GEOCHEMICAL AND GRANULOMETRIC MEASUREMENTS

In the laboratory, the 640 samples were dried at 60-°C, gently disaggregated in a china mortar, dry sieved and finally homogenised. Utmost care was taken not to lose the finer or organic fractions. The size-fraction <2000-µm was analysed both for bulk geochemistry (performing X-Ray Fluorescence Spectroscopy, XRFS) and for granulometry (using the Laser-Diffraction technique). Bulk geochemical analysis (XRFS) enables rapid analysis of a large number of samples and yields extra information on rock fragments, fine fractions and diagenetic processes normally not studied in mineralogical studies (Kroonenberg, 1990, 1994; Moura and Kroonenberg, 1990). For the XRFS-measurements, a subsample of about 4-g was milled (and homogenised again) for 5-minutes in a Fritsch ball-mill apparatus with "Sialon" milling material (hardness ~9-10 on Mohs' scale). Subsequently, 0.600-g of milled sample was ignited with 2.400-g Li-tetraborate at 1800-°C to form a 23-mm diameter glass bead. The oxidation of organic matter and the volatilisation of: 1) crystal water from clay minerals, 2) CO_2 from carbonates and 3) SO_2 from sulphates at this high temperature yield a loss in sample weight. This loss represents and will be referred to in the following as the Loss on Ignition (LOI). A number of 636 samples have been measured using a Philips PW-1404 assembly and international geological reference standards. The contents of the major constituents SiO_2, TiO_2, Al_2O_3, Fe_2O_3, MgO, MnO, CaO, Na_2O, K_2O and P_2O_5 have been measured with a Sc-tube. The trace elements Ba, Ce, Co, Cr, Cs, Cu, Ga, La, Nb, Nd, Ni, Pb, Rb, Sr, Th, V, Y, Zn and Zr were measured with a Rh-tube.

Granulometric analyses were undertaken on a 1- to 20-g (depending on texture) sample after a 30% H_2O_2-pretreatment to oxidise the organic matter. The samples were not washed to retain the fine fractions. In case of samples exceptionally rich in organic matter (e.g. gyttjas), however, this yielded too high values for the 0-20 μm fraction due to incomplete destruction of fine organic particles. Sodium pyrophosphate was added to prevent flocculation of the clay fraction. Laser diffraction was performed at 10% sample obscuration using a Coulter LS230 Grain Sizer with software version 2.05 and measuring a range of 0.04-2000 μm (Buurman et al., 1997).

3.5 RESULTS

Bulk geochemical analyses

Covarying bulk geochemical and granulometric data clearly demonstrate that the elemental concentrations strongly depend on the sample granulometry. Sandy samples (defined as samples with <30% of the 0-20 μm fraction) typically contain >80% SiO_2 (Fig. 3.1), whereas silty and clayey samples (>30% of the 0-20 μm fraction) generally have <80% SiO_2 and >10% Al_2O_3. The clay and fine silt granulometric fractions concentrate Al-bearing phyllosilicates (clay minerals and micas) (Van den Broek and Van der Marel, 1980; Moura and Kroonenberg, 1990; Huisman, 1997). Consequently, Al_2O_3 strongly correlates ($r = 0.88$) to the 0-20 μm (clay and fine silt) granulometric fraction (Fig. 3.1), while SiO_2 is inversely correlated ($r = -0.82$). This enables us to infer a high clay and fine silt-content geochemically from a high Al_2O_3-percentage in a bulk sample (cf. Huisman and Kiden, 1998). Samples deviating from these linear trends appeared to be rich in organic matter (peaty and gyttja samples) and generally have >10% Loss On Ignition (LOI) owing to melting of the XRFS-beads at 1800 °C (Fig. 3.2).

Figure 3.3 shows the scatterplots of Al_2O_3-content (considered here to represent the clay and fine silt fraction) versus K_2O, MgO, TiO_2 and Fe_2O_3. Strongly organic peat and pure organic gyttja samples have been omitted, since they do not represent the detrital clastic suspension load of the main river. Silty and clayey bulk samples are systematically higher in K_2O, MgO, TiO_2, Fe_2O_3 and LOI (Figures 3.2, 3.3). Similarly, most trace elements show strong positive relationships to the Al_2O_3-content, such as Rb, V and Cr (Figs. 3.4A and B). However, note that particularly the samples in the clay and fine silt granulometrical range (with high Al_2O_3-contents) show considerable compositional variability. This phenomenon will be discussed in detail in Chapter 4. Clayey samples also contain more Fe_2O_3 (Fig. 3.3). Strongly organic samples like sideritic and/or vivianitic (clayey) gyttjas even contained 5-40% Fe_2O_3 (Fig. 3.5) due to the presence of siderite and vivianite (see below), 0.25-2% MnO, 0.5-5% P_2O_5 and in calcareous samples 1-20% CaO as well (the latter three are not shown for sake of brevity).

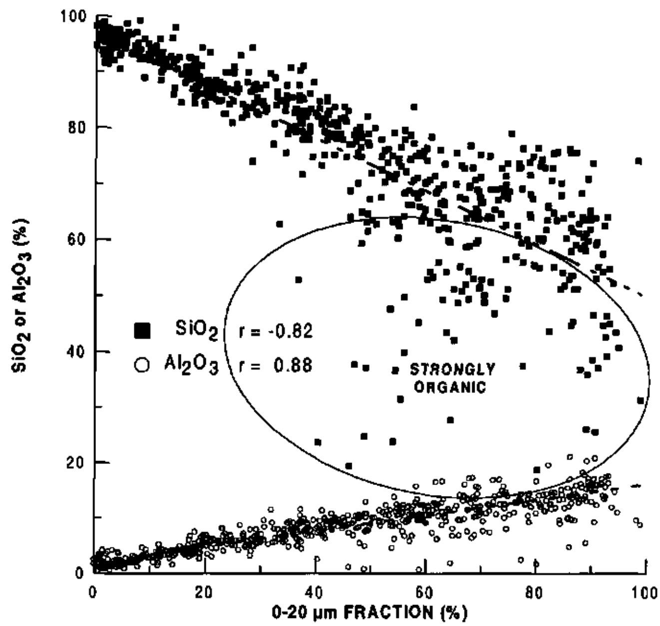

Figure 3.1 The relationship between the 0-20 µm fraction (determined by laser granulometry) and the contents of SiO_2 (squares) and Al_2O_3 (circles). Samples deviating from linear trends (e.g. squares within ellipse) have high organic matter contents.

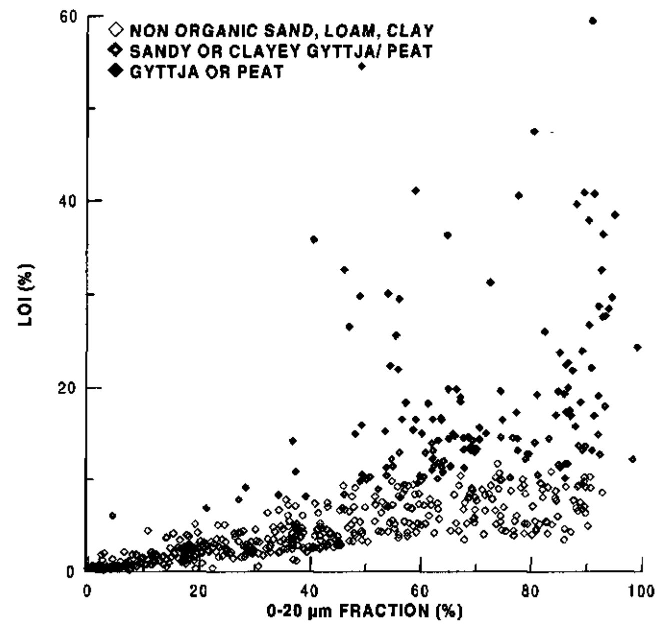

Figure 3.2 The relationship between the 0-20 µm fraction (determined by laser granulometry) and the Loss on Ignition (LOI). Note high LOI for gyttja and peat samples (rich in organic matter), but also for clayey samples (crystal water).

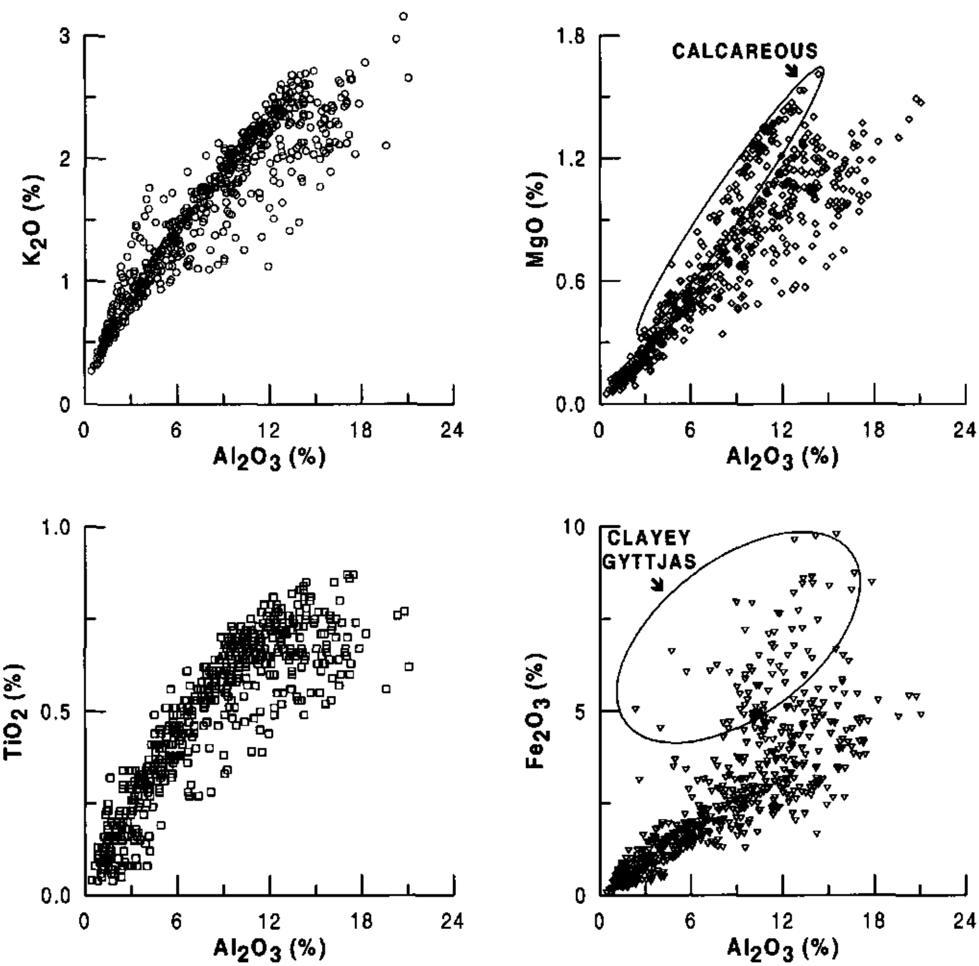

Figure 3.3 The relationship between the contents of the major constituents Al_2O_3 and K_2O, MgO, TiO_2 and Fe_2O_3 for 636 bulk sediment samples taken from Meuse lower reach residual channel infillings. Note the considerable variability of K_2O, MgO and TiO_2 with increasing Al_2O_3-contents. This phenomenon will be discussed in Chapter 4. Strongly organic gyttja samples are rich in Fe_2O_3 (see text). Extreme outliers have been omitted in the Fe_2O_3 scatterplot, but see also Figure 3.5.

Principal component analysis

As a first statistical reconnaissance study, a principal component analysis was performed on the dataset, containing both the bulk geochemical and the laser granulometric data (Table 3.1). We used the factor procedure of the statistical package SPSS-PC+, which extracts the principal components ("factors") within the dataset after Varimax rotation (Davis, 1986). The statistical analysis included the following four texture classes: 0-20 μm (clay and fine silt), 20-80 μm (coarse silt and very fine sand), 80-210 μm (fine sand) and 210-2000 μm (coarse sand). Bulk geochemical analyses below detection limits were labelled 0, and have been included in the statistical analysis. The trace elements Cs and Th were not included because too many samples were reported below detection limits. Cu has been left out too, for contamination during sieving over copper mesh was suspected. The analysis yielded four major factors (eigenvalues >1) that contained strongly covarying variables (with generally high communalities and factor loadings to the factors) and explained 82% of total variability in the dataset (Table 3.1). A mineralogical interpretation followed to account for the oxides or elements present in these extracted factors and to clarify the interrelationships between granulometric fractions, major constituents and trace elements. The mineralogical interpretation yielded a clay and fine silt/phyllosilicates factor (factor 1), a coarse silt/heavy minerals factor (factor 2), a gyttja/diagenetic minerals factor (factor 3) and a calcium carbonate factor (factor 4).

Clay and fine silt/phyllosilicates factor

The 0-20 μm fraction (CLFSILT in Table 3.1), the major constituents Al_2O_3, K_2O, MgO, TiO_2 and the trace metals V, Ga, Rb, Ba, Cr, Ce, Zn, Ni and La are strongly interrelated. They all have high positive factor loadings (>0.8) for the individual variables. SiO_2 and the 80-210 μm (FSAND) and 210-2000 μm (CSAND) granulometric fractions have negative factor loadings. The constituents with positive loadings (Al_2O_3 and others) occur in clay minerals and fine-silt micas hosted in the 0-20 μm fraction, while SiO_2 (negative loading) is enriched in the fine (80-210 μm) and coarse (210-2000 μm) sand fractions (Van den Broek and Van der Marel, 1964, 1980). The assemblage can be interpreted as a clay and fine silt factor and mainly indicates sorting of fine-grained phyllosilicates in the clay and fine silt size-fractions versus sand-sized quartz. This process explains 55.2% of total variation (Table 3.1). Nd and Co have a lower factor loading (>0.7) to the clay and fine silt fraction. Pb was not correlated to any of the factors due to some anomalous outliers in the dataset (see below). After removing these outliers for Pb, a relatively good correlation with the clay and fine silt fraction or Al_2O_3-content ($r = 0.71$) is found. Most trace elements, however, show strong linear relationships with Al_2O_3, permitting the conclusion that they are mainly built in into the phyllosilicate lattices or adsorbed on their reactive surfaces.

Coarse silt/heavy minerals factor

The 20-80 μm (coarse silt and very fine sand) fraction, the main constituent Na_2O and the trace elements Zr, Nb, Y, Nd show positive loadings for the second factor, versus negative loadings for the 210-2000 μm fraction (Table 3.1). Heavy minerals like zircon contain abundant Zr, Nb, Y and Nd. Nb substitutes for Zr in the zircon structure, while Y occurs as xenotime (YPO_4) inclusions in the zircon mineral (Deer et al., 1995). The factor indicates a concentration of heavy minerals like zircon in the silt fraction (20-80 μm) owing to their specific gravity and hydrodynamic properties. Therefore, we interpret factor 2 as a coarse silt/heavy minerals factor, which explains 14.5% of total variance. Swennen and Van der Sluys (1998a) additionally mention the concentration of Ti, Cr and Hf in heavy minerals as rutile, ilmenite and chromite. In this research however, TiO_2 and Cr correlated with the phyllosilicate factor (see above). In a mineralogical study of the Meuse alluvial sediments, Van den Broek and Van der Marel (1964)

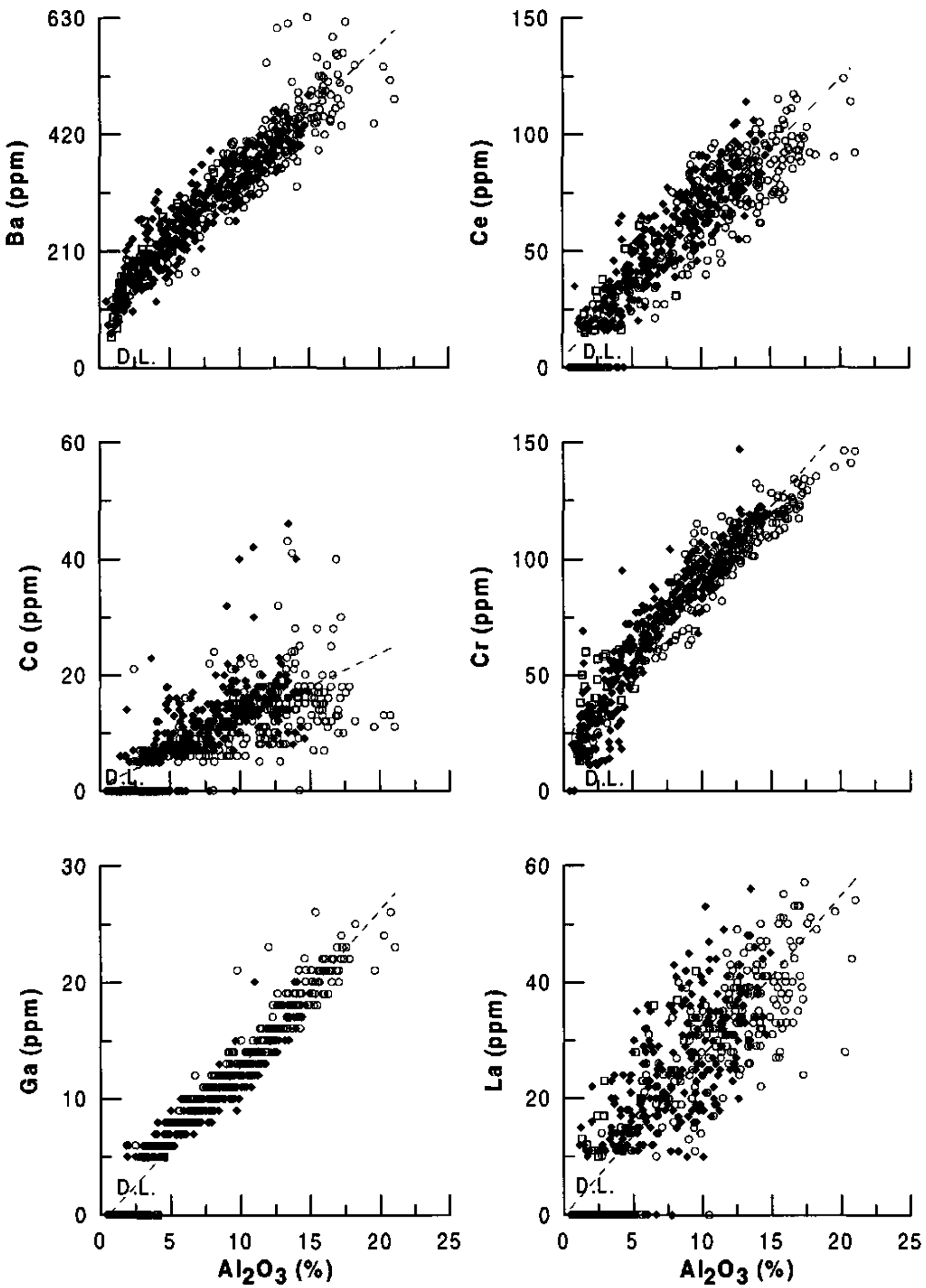

Figure 3.4A The relationship between the contents of Al_2O_3 and several trace elements for River Meuse bulk sediment samples. Data on the linear fits are listed in Table 3.2. D.L = detection limit. Legend: Pleniglacial samples: □; Late-glacial: ◆; Holocene: O.

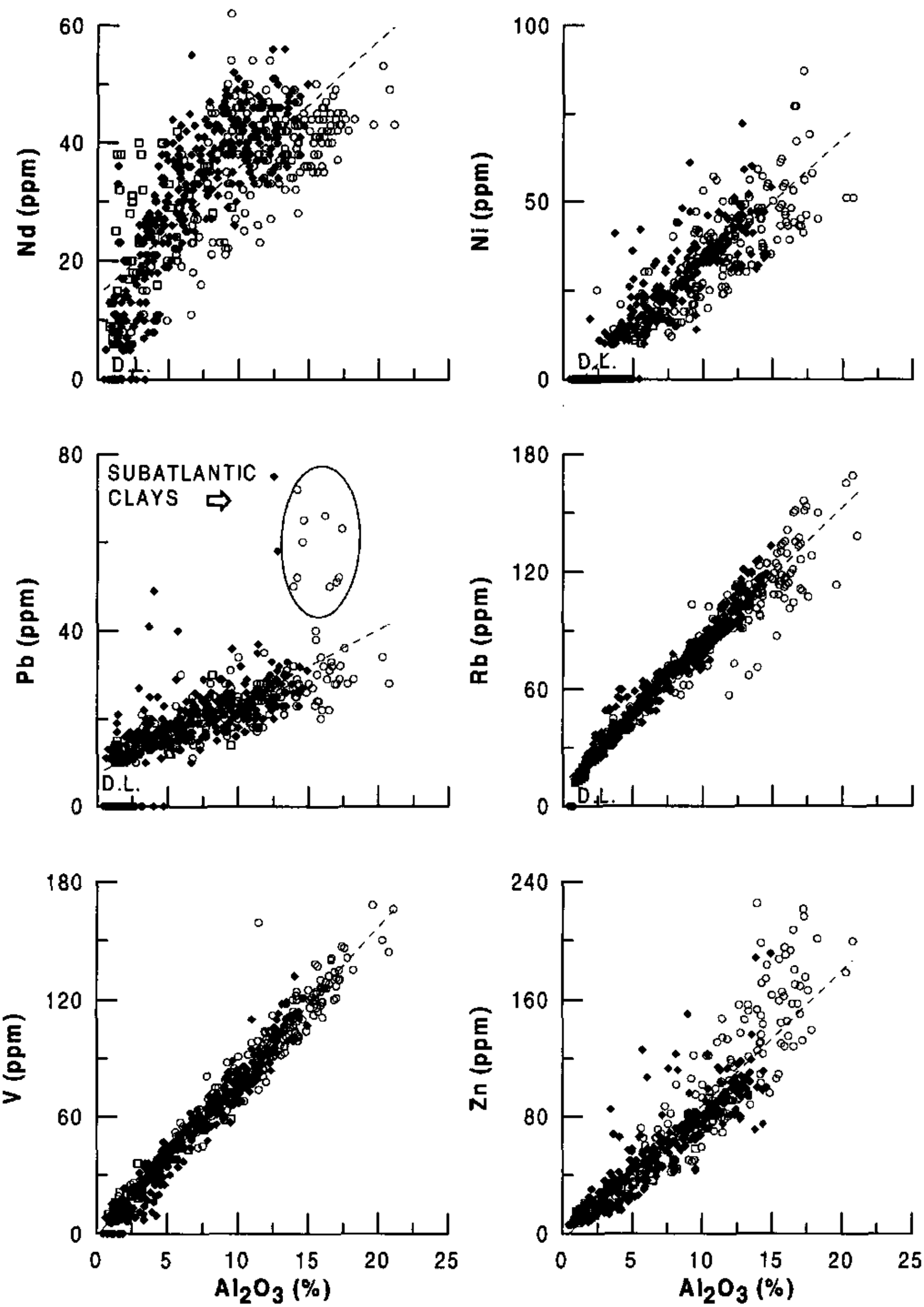

Figure 3.4B The relationship between the contents of Al_2O_3 and several trace elements for River Meuse bulk sediment samples. Data on the linear fits are listed in Table 3.2. D.L. = detection limit. Legend: see Fig. 3.4A. Some Pb- and Zn-outliers have been omitted (see text).

Table 3.1 Bulk sediment geochemical communalities and factor loadings (after Varimax rotation) for the elements of the first four factors with eigenvalues >1 (factor loadings <0.5 have been suppressed). The eigenvalues (EV) and percentages of trace (PT) are indicated per factor. The factor loadings can be considered as multiple correlation coefficients between the extracted factor and the involved variable (major constituent or trace element). CLFSILT = Clay and fine silt fraction (0-20 μm), CSILT = Coarse silt fraction (20-80 μm), FSAND = Fine sand fraction (80-210 μm) and CSAND = Coarse sand fraction (210-2000 μm), LOI = Loss on Ignition.

Variable	Communality	Factor 1 EV: 17.1 PT: 55.2%	Factor 2 EV: 4.5 PT: 14.5%	Factor 3 EV: 2.5 PT: 8.2%	Factor 4 EV: 1.2 PT: 3.8%
Al_2O_3	0.97	0.97			
V	0.96	0.97			
Ga	0.95	0.97			
Rb	0.95	0.96			
Ba	0.88	0.92			
Cr	0.96	0.92			
K_2O	0.93	0.90			
Ce	0.90	0.89			
Zn	0.85	0.88			
Ni	0.82	0.87			
CLFSILT	0.91	0.86			
TiO_2	0.96	0.84			
La	0.77	0.84			
MgO	0.92	0.83			
Nd	0.94	0.73			
Co	0.66	0.72			
FSAND	0.68	-0.67			
Pb	0.19	See text			
Zr	0.87		0.89		
CSILT	0.76		0.80		
Nb	0.81		0.73		
CSAND	0.86	-0.57	-0.62		
Y	0.65	0.53	0.60		
Na_2O	0.44		0.50		
Fe_2O_3	0.87			0.92	
P_2O_5	0.75			0.85	
MnO	0.73			0.77	
SiO_2	0.95	-0.59		-0.73	
LOI	0.68			0.66	
CaO	0.86				0.78
Sr	0.93	0.62			0.64

found 20-25% feldspars and 70-80% quartz in the 16-80 μm fraction. Probably owing to concentration of albitic (Na-) feldspars in the 20-80 μm fraction, Na_2O shows a weak correlation to this heavy mineral factor.

The 20-80 μm fraction demonstrates weak positive relationships with considerable scattering to the Zr-Nb-Nd-Y quartet, of which Zr is shown as an example (Fig. 3.6). The patterns in Figure 3.6 can be explained as follows: strongly organic (clayey) gyttja samples are low in detrital heavy minerals and thus are poor in Zr, Nb, Nd and Y. Fine sandy and silty samples are relatively rich in detrital heavy minerals and consequently have higher contents of trace elements. Silty clay samples are richest in Nd and Y because of their joint occurrence in coarse silt fraction heavy minerals and clay and fine silt-sized phyllosilicates. The main elements K_2O, MgO and TiO_2 are strongly positive linear related (data in Table 3.2) to the Al_2O_3-content.

Gyttja/diagenetic minerals factor

The third factor contains strong positive loadings from the major constituents Fe_2O_3, MnO, P_2O_5 and the variable Loss on Ignition. The SiO_2-content relates inversely to this factor. Polyvalent Fe, Mn and P are involved in microbially mediated redox conditions that are typical of the anoxic bicarbonate-rich freshwater gyttja environment of several residual channels in this research (Curtis, 1995). High contents of oxidisable organic matter in gyttja samples (Fig. 3.2) enable micro-organisms to reduce plenty of Fe(III)- and Mn(IV)-oxides. In the absence of dissolved sulphide, ferrous iron reacts with dissolved carbonate or phosphate to form post-depositional siderite ($FeCO_3$) and/or vivianite ($Fe_3(PO_4)_2 \cdot 8H_2O$). Mn is known to substitute for Fe and forms a solid-solution series with siderite and calcite. Therefore, siderite and vivianite are characteristic post-depositional early diagenetic minerals (Curtis, 1995) which are frequently reported in freshwater river fens and peat (Postma, 1977, 1982). Vivianite was easily recognised in several of our samples from the small white spots, which turned blue upon oxidation. The presence of both siderite and vivianite in our samples could also be established by SEM and XRD (Huisman, 1997: Chapter 4.1) and in samples from well-comparable residual Meuse channels (Westerhoff et al., 1990; Walraven et al., 1999). The third factor thus mainly represents post-depositional authigenic mineral formation and accounts for 8.2% variation.

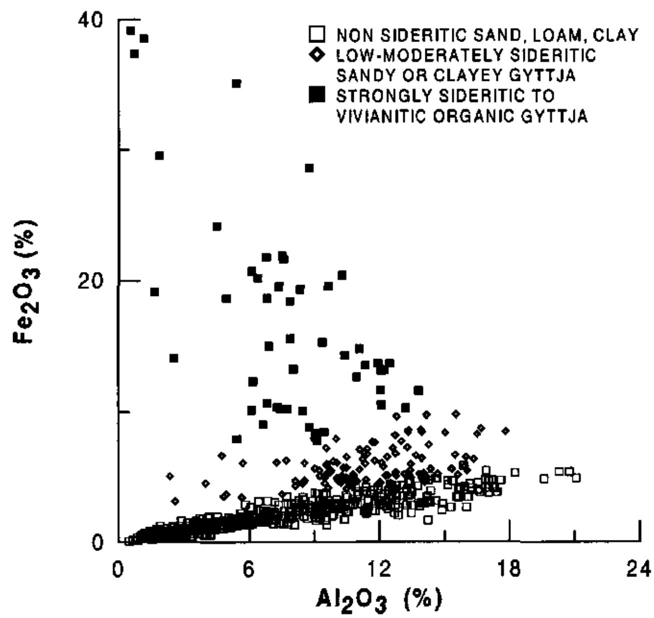

Figure 3.5 The relationship between the contents of Al_2O_3 (representing the clay-and-fine-silt content) and the contents of Fe_2O_3. Note Fe-enrichment due to authigenic mineral formation in moderately to strongly sideritic and vivianitic gyttja samples.

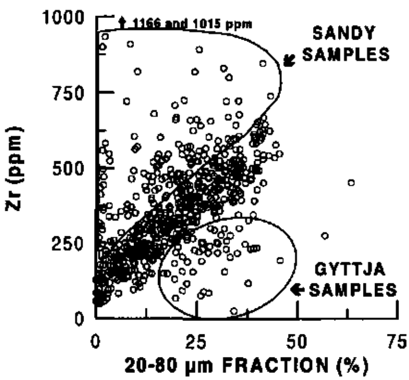

Figure 3.6 The relationship between the contents of the 20-80 μm fraction and Zr for bulk sediment samples. Note low values of Zr for gyttja samples and high values for sandy samples (<30% of the 0-20 μm fraction).

Accordingly, siderite and vivianite raise the contents of Fe_2O_3, MnO and P_2O_5 in pure organic and clayey gyttja samples (see the vertical profiles of Figures 3.7 and 3.8) and cause a weak correlation of these major constituents with the Al_2O_3-content (e.g. Fe_2O_3: Fig. 3.5). After omitting the most gyttja/siderite-rich samples, the bivariate correlation improves to r = 0.76 with the Fe_2O_3/Al_2O_3-ratio more resembling the one in phyllosilicates (Fig. 3.3). The gyttja/siderite factor did not incorporate the trace elements, probably because these are predominantly built in into the phyllosilicate lattices. However, individual gyttja samples do show a relative enrichment of Co, Ni and notably Zn. This only becomes clear after normalisation by Al_2O_3 (cf. Huisman, 1997) to account for their natural occurrence in phyllosilicates (Figs. 3.7 and 3.8). Note that the Holocene gyttja profile (B) shows the highest accumulation in absolute concentrations for P_2O_5, Co, Ni and Zn. Post-depositional accumulations of Co, Ni and Zn in gyttja samples causes scattering in the graphs (Figs. 3.4A and 3.4B) and consequently a less good correlation with the Al_2O_3-content. Similar enrichments for the other trace elements were expected in gyttja samples and have been checked, but they were not found.

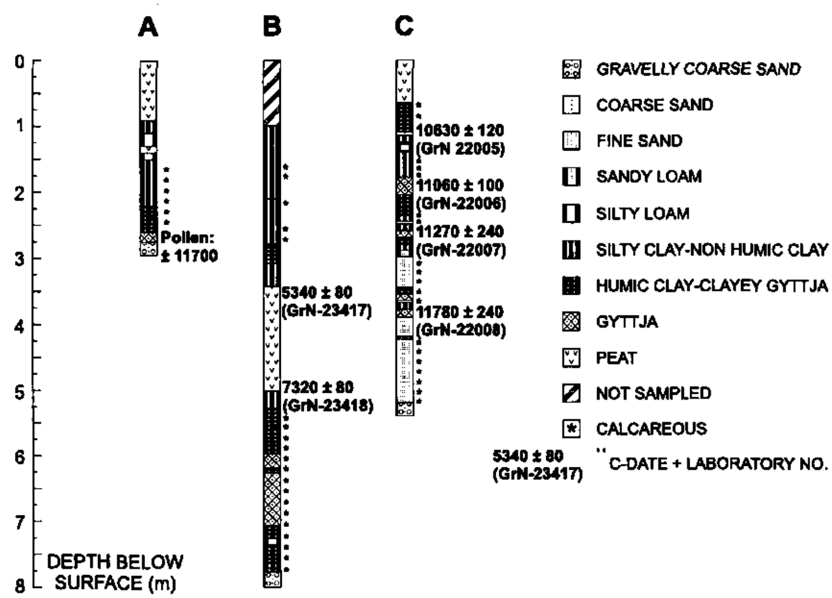

Figure 3.7 The lithological profiles of sites 1 (A, Beugen), 2 (B, Blitterswijck) and 7 (C, Keuter), showing several (clayey) gyttja intervals. The site locations are indicated on Figure 2.3, while data on a selection of main and trace elements can be found in Figure 3.8.

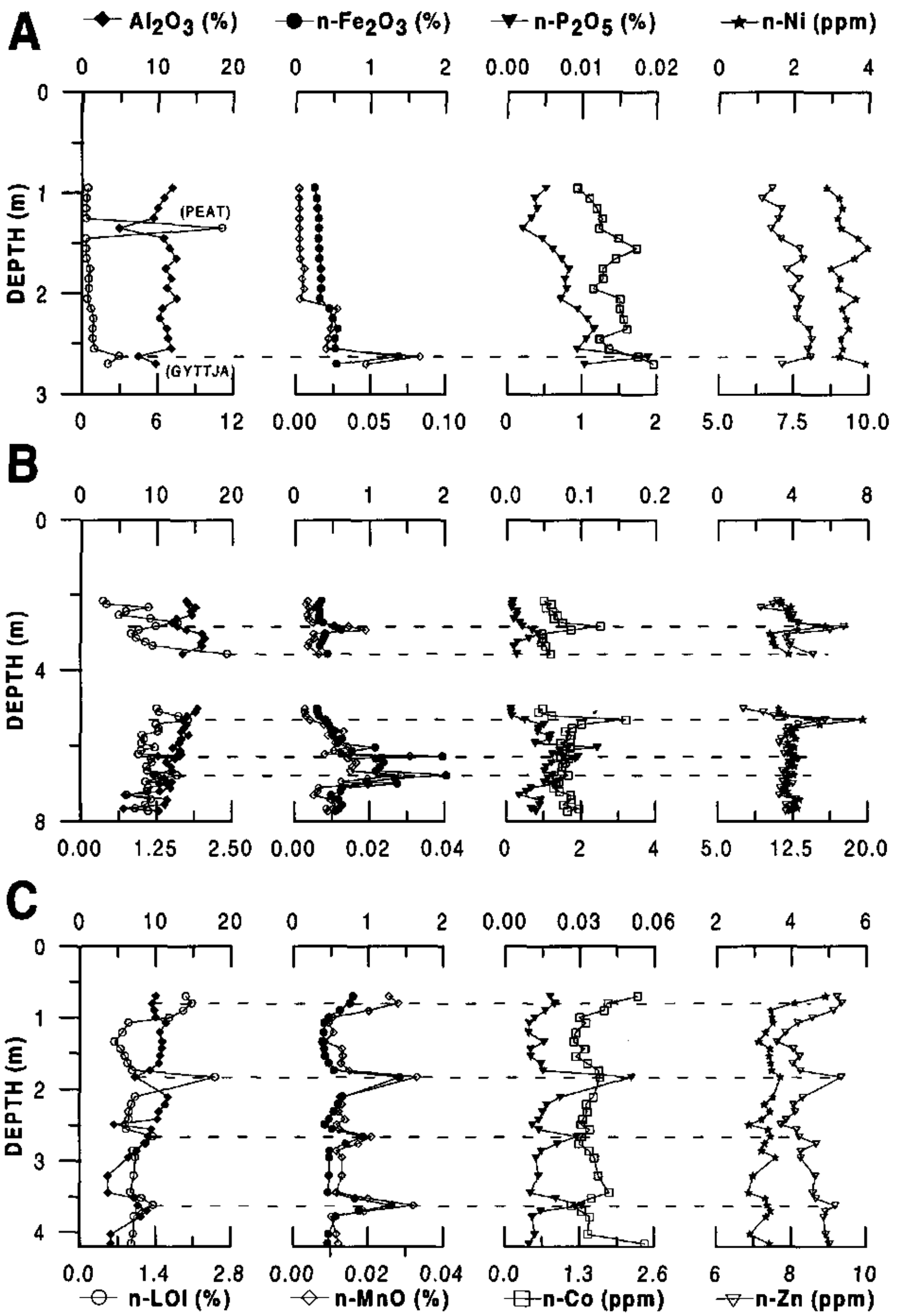

Figure 3.8 Al-normalised (n-...) main and trace elements for profiles A, B and C in Figure 3.7. Peat intervals (Tops of A, B, C and middle part of B) as well as the lowermost sands in C have not been sampled. Closed symbols correspond to upper x-axes and open symbols correspond to lower x-axes.

Calcium carbonate factor
CaO and the trace element Sr contribute positively to the fourth factor. A group of calcareous sandy and clayey samples (with roughly >1% CaO, effervescing with 1-M HCl) is present in the dataset. CaO is the main constituent in calcium carbonate and Sr isomorphically substitutes for Ca (Deer et al., 1995). Thus, the fourth factor with positive loadings for CaO and Sr and explaining 3.8% of variation can be interpreted as a calcium carbonate factor.

Strongly calcareous samples (samples effervescing with 1-M HCl and/or with roughly >1% CaO) contain about two to three times as much Sr as non-calcareous samples (not effervescing with 1-M HCl and/or <1% CaO) with comparable clay and fine silt-content (Fig. 3.9). Sr is not only related to the calcium carbonate-content, but increases with increasing Al_2O_3- or clay and fine silt-content as well. This suggests that Sr be built into phyllosilicates as well. Labelling further demonstrates elevated MgO-contents (Fig. 3.3; ellipse-shaped contour) and Fe_2O_3-contents in several calcareous samples (e.g. lower part of profiles A and B in Figures 3.7 and 3.8). The former is probably due to the substitution of Ca by Mg in the calcium carbonate crystal lattices (Van der Marel and Van der Broek, 1962). The latter is due to the presence of calcareous and clayey gyttja samples, which contain Fe-bearing siderite/vivianite as well.

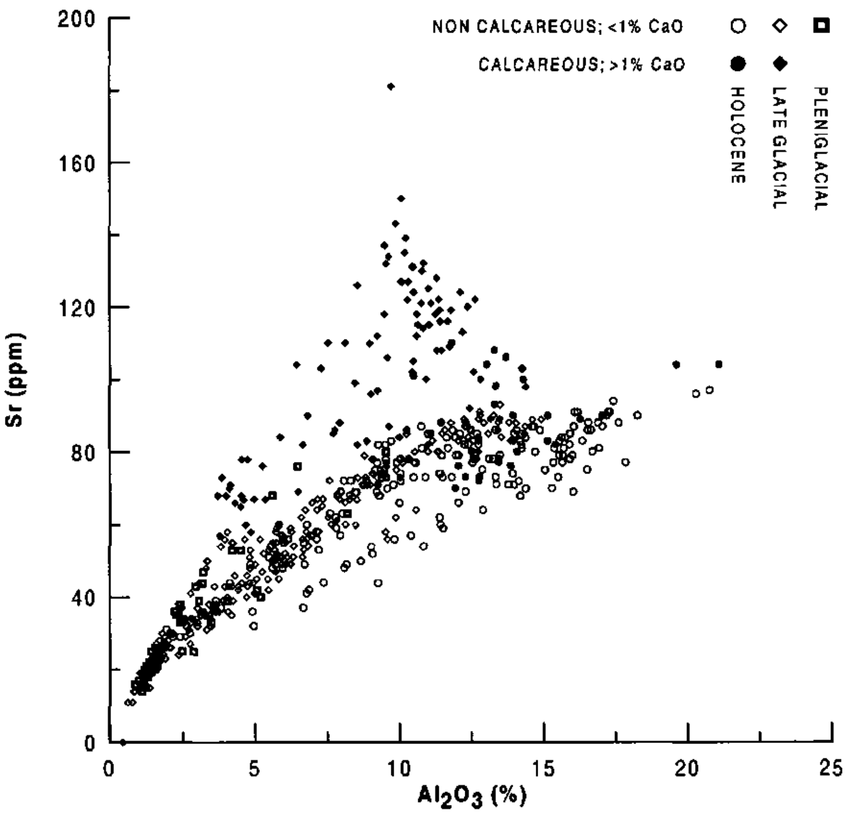

Figure 3.9 The relationship between the contents of Al_2O_3 and Sr. Open symbols indicate non-calcareous samples (<1% CaO) and closed symbols indicate calcareous samples (>1% CaO). Note Sr-enrichment in calcareous samples.

Anomalies

Six samples (of which five within one coring, but not successive) showed extreme values for Pb (375, 486, 188, 113, 76 and 75-ppm respectively), five other samples for Ni (104, 157, 138, 87, and 87-ppm respectively) and three for Zn (238, 173, 187-ppm). Recent anthropogenic pollution and laboratory contamination could be ruled out with absolute certainty after checking the position/chronology of these samples and their Zn-, Cu- and Cr-contents (showing normal values). The Pb-and Zn-accumulations were mainly found in humic clays and/or underneath peat layers. When these extreme values are excluded from the scatterplots, the Pb- and Zn-contents in some samples are still elevated (40-75 ppm). These samples appeared to be Subatlantic clays (Fig. 3.4B).

3.6 DISCUSSION

Natural compositional variability

The bulk geochemistry of samples taken from unconsolidated residual channel infilling sediments of the Upper Kreftenheye and Betuwe Formations varies with granulometry. The mixing of a variety of minerals (mainly phyllosilicates, feldspars and quartz) hosted in different size fractions, causes natural variation in the abundance of both main and trace elements and their ratios. Quartz grains in the fine and coarse sand fractions account for the bulk SiO_2-content. Phyllosilicates abound in the clay and fine silt size fractions and host varying contents of Al_2O_3, K_2O, MgO, TiO_2 and trace elements (Van den Broek and Van der Marel., 1980; Moura and Kroonenberg, 1990; Hakstege et al., 1993). K-feldspars, Na-feldspars and heavy minerals more frequently occur in the silt fraction (Van den Broek and Van der Marel, 1980). Accordingly, a coarse sandy sample only containing K-feldspar ($KAlSi_3O_8$) exhibits a higher K_2O/Al_2O_3-ratio relative to a clayey silt sample, in which the same K-feldspars are "diluted" by Al-rich clay minerals (e.g. illite: $KAl_5Si_7O_{20}(OH)_4$). Likewise, clayey overbank and stream sediments from the Meuse tributaries draining the Ardennes Palaeozoic shales were found to be richer in Al_2O_3, K_2O, several trace elements etc. compared to overbank sediments of streams and rivers draining the coarser Tertiary sands of northern Belgium (Van der Sluys et al., 1997). Therefore, any geochemical reference baseline for a particular region or geological unit should certainly include grainsize measurements that preferably cover the whole range of sedimentological niches in that region or unit. Apart from geochemical variation due to grainsize-sorting processes, the bulk and clay geochemistry of River Meuse sediments showed considerable compositional variability in samples with comparable high clay and fine-silt contents. We will elaborate on the causes of this compositional variability in the next Chapter (Ch. 4; Tebbens et al., 1998).

Post-depositional siderite and vivianite that are formed during early diagenesis raise the contents of Fe_2O_3, MnO and P_2O_5 in (clayey) gyttja samples (cf. Postma, 1982; Curtis, 1995; Huisman, 1997; Walraven et al., 1999). The at-site prevailing groundwater and redox conditions strongly determine authigenic siderite and vivianite formation. Therefore, these minerals might be strong indicators for the local or regional palaeo-environment, but they seem less suitable to use them as palaeoclimatic indicators for an entire upstream drainage basin. E.g., based on detailed cm-sampling of two Late-glacial gyttja deposits (from sites 1c and 6 in this study, Fig. 2.3), Walraven et al. (1999) infer two diagenetic stages of siderite formation, which they relate to regional climate change. According to them, the Fe- and Mn-(oxy)hydroxides have been deposited during the cold phases, while siderite preferentially precipitated during the warm phases. Poulton and Raiswell (1996) performed sequential extractions on suspended loads, to demonstrate that glacial river systems are low in highly-reactive Fe and that poorly-reactive Fe dominates their total Fe-contents. The reverse was found for present interglacial European rivers.

They consider the larger fraction of highly-reactive Fe in the present river basins to arise from increased (chemical) weathering as opposed to dominant physical weathering processes in glacial river systems.

The contents of the trace elements Co, Ni and Zn in clayey gyttjas are higher than their natural occurrence in phyllosilicate minerals predicts. Historic mining of ore deposits since the Industrial Revolution is known to have raised the contents of Pb, Zn, Cu and As in recent Meuse and Geul overbank sediments (Leenaers, 1989; Hindel et al., 1996; Swennen et al., 1997; Swennen and Van der Sluys, 1998a,b). In this research, the samples were of unambiguously pristine nature, excluding effects of ore mining or coal slags on trace element geochemistry. Hakstege et al. (1993) mention Pb- and Zn-accumulation in former vegetation horizons of old floodplain deposits. Huisman (1997) found similar accumulations of the same elements and describes Ni-enrichment in Early and Middle Pleistocene organic subsurface samples. He suspects Ni is released from Fe-hydroxides during glacial-interglacial groundwater fluctuations. Hill and Aplin (1996) sampled unpolluted rivers with diverse contents of dissolved organic matter to find that 20-40% of total Zn and 60-80% of Ni and Pb were associated with fine colloidal organic material via humic and fulvic acid complexation. Our samples with Ni-accumulations were situated in or below gyttja intervals. Apparently, the organic matter complexes these trace metals to elevate the Co-, Ni-, Pb- and Zn-contents after subsequent flocculation with clay in the high-organic matter gyttja or peat environment. Biological accumulation due to vegetation growth may have caused the Holocene organic-matter rich profile to contain more trace elements in absolute sense (Korobova et al., 1997). Finally, Shotyk et al. (1998) reconstructed enhanced atmospheric Pb-influxes in a Swiss peat bog. They attribute this to increased soil dust inputs that originated from soil erosion, following two climate change events at around 10590 (Younger Dryas) and 8230 ^{14}C-years BP.

The elevated Pb-contents in Subatlantic clays (younger than 2900 ^{14}C-years BP) probably reflect anthropogenic deforestation practices leading to the removal of the upper organic soil horizons rich in bio-accumulated Pb or Zn (e.g. Macklin et al., 1994; Korobova et al., 1997). Indeed, Shotyk et al. (1998) provide evidence for increased atmospheric lead deposition after 5320 ^{14}C-years BP, related to soil erosion caused by forest clearing and agricultural tillage. They also reconstructed the beginning of increased lead pollution due to mining and smelting at ~3000 ^{14}C-years BP. Likewise, Greenland ice-core analyses indicate an increase of atmospheric lead concentrations to four times the natural background values during ~2500-1700 years BP, the time of mining and smelting activities by the Greek and Roman civilisations (Hong et al., 1994). Therefore, direct human atmospheric pollution might have elevated the Pb-contents in soils and sediments as early as 3000 BP. So, as far as Pb is concerned, this implies that Subatlantic clayey sediments can strictly seen not be considered unpolluted.

Prediction of natural contents

The contents and ratios of main and trace elements in pristine samples of Late-glacial and Early Holocene River Meuse sediments depend for the greater part (70% of explained variation) on their granulometry. Climatic change and early diagenesis provide some variations on this theme. Pre-erosional weathering in interstadial or interglacial soil environments depletes clay minerals of K_2O and MgO and causes relative Al_2O_3-enrichment (see Chapter 4). Siderite/vivianite accumulation and organic complexation in strongly organic gyttjas and peat elevate the contents of Fe_2O_3, P_2O_5, MnO, Co, Ni, Pb and Zn. MgO and Sr are elevated in calcareous samples. Heavy mineral concentration in fine sands elevates their Zr-, Nb-, Nd- and Y-contents.

Now the causes of the most important factors in natural variation are clear, it is possible to predict the contents of major and trace elements as a function of sample granulometry and to allow for siderite/vivianite accumulation and lime content. The Al_2O_3-content showed the strongest factor loading to the clay and fine silt factor (Table 3.1). The Al_2O_3-content is to be preferred as a proxy for the clay and fine silt-content instead of laser granulometry, because Al_2O_3 and other constituents can be measured simultaneously in the same sample (cf. Huisman and Kiden, 1998). Moreover, the XRFS-measurements do not suffer from incomplete oxidised organic matter particles in strongly organic samples (unlike the grainsize measurements). Likewise, the Fe_2O_3- and CaO-content show strongly positive factor loadings to the gyttja/siderite and carbonate factors respectively and can serve as a proxy for elevated siderite/vivianite- or lime-contents. In case of K_2O, MgO and TiO_2, it makes sense to distinguish between Holocene and Late-glacial/Pleniglacial samples, to allow for the relative enrichment with Al_2O_3 in Holocene clayey samples (see Chapter 4). Table 3.2 gives data on the multiple linear regression models, with the Al_2O_3, Fe_2O_3 and CaO-contents as independent variables. Most models are very good fits and have multiple correlation coefficients of >0.85 (or explain >75% of variation). Despite a smaller number of samples, the Holocene fits show lower r^2–values compared to the Late-glacial ones, stressing their greater variability. The lower ratios of K_2O, MgO and TiO_2 to Al_2O_3 for Holocene samples are reflected in lower multiple regression coefficients (a-values in Table 3.2).

Table 3.2 Multiple linear regression data on a selection of major and trace elements for the entire dataset. For calculation of Co, Ni, Pb and Zn, 13 samples showing anomalies or extreme outliers have been omitted, see text.

Multiple linear regression model: Y = **a** • Al_2O_3-% + **b** • Fe_2O_3-% + **c** • CaO-% + **d**
(Y = Predicted major constituent in %, trace elements in ppm)

Constituent Y	a	b	c	d	Fit (r^2)	# Obs.(n)
K_2O (all)	0.1300	-0.0113	0.0207	0.5454	**0.86**	636
K_2O (Pleniglacial)	0.2145	-0.1329	0.7669	0.2893	**0.97**	38
K_2O (Late Glacial)	0.1668	-0.0101	-0.0054	0.4053	**0.95**	372
K_2O (Holocene)	0.1095	-0.0074	0.0598	0.6187	**0.77**	226
MgO (all)	0.0728	0.0020	0.0401	0.0472	**0.86**	636
MgO (Pleniglacial)	0.0832	0.0045	-0.2831	-0.0101	**0.97**	38
MgO (Late Glacial)	0.0975	0.0032	0.0326	-0.0500	**0.95**	372
MgO (Holocene)	0.0601	0.0041	0.0620	0.0998	**0.80**	226
TiO_2 (all)	0.0426	-0.0037	0.0065	0.1376	**0.79**	636
TiO_2 (Pleniglacial)	-0.0379	0.2077	1.0347	0.0262	**0.88**	38
TiO_2 (Late Glacial)	0.0564	-0.0029	-0.0004	0.0718	**0.91**	372
TiO_2 (Holocene)	0.0325	0.0017	-0.0055	0.2070	0.60	226
Ba	24.0455	-0.5776	-3.4005	110.57	**0.89**	636
Ce	5.9904	0.0662	-0.1190	4.04	**0.83**	636
Co	1.0492	0.2152	0.1684	0.71	0.53	623
Cr	6.7234	0.2730	-0.8947	21.5	**0.90**	636
Ga	1.3432	0.0736	-0.2014	-0.76	**0.94**	636
La	2.7435	0.1159	-0.1406	-0.29	**0.71**	636
Nd	2.2392	-0.1732	-0.1081	14.05	0.59	636
Ni	3.5159	0.4110	0.2717	-4.68	**0.81**	623
Pb	1.6862	-0.2942	-0.3670	7.78	0.54	623
Rb	7.2146	-0.4260	-0.2230	11.71	**0.83**	636
Sr	4.3123	-0.8737	8.8090	23.59	**0.88**	636
V	7.7766	0.5764	-0.8111	-0.95	**0.96**	636
Zn	1.7405	-1.699	0.6546	8.44	**0.79**	623

3.7 CONCLUSIONS

Late-glacial and Early Holocene Meuse sediments deposited since 13000 years BP and before 3000 ^{14}C-years BP are unaffected by anthropogenic pollution and are of unambiguously pre-industrial or pristine nature. Therefore, well-dated fine-grained Meuse residual channel infillings from this period offer wide possibilities for environmental research to determine the natural sedimentary compositional variation of major and trace elements.

Unconsolidated, reduced sediments show considerable compositional variation owing to hydrodynamic sorting processes during fluvial transport and sedimentation, affecting the abundance of phyllosilicates in the clay and fine silt size fractions and feldspar and quartz minerals mainly in the coarser silt and sand fractions. The clay- and fine-silt fraction is predominantly characterised by the elevated and covariant contents of Al_2O_3, K_2O, MgO, TiO_2 and trace elements like Ba, Ce, Cr, Ga, La, Ni, Pb, Rb, V and Zn. Post-depositional formation of siderite and vivianite in gyttjas strongly influences the Fe_2O_3-, MnO- and P_2O_5-contents. Co, Ni and Zn occur as natural accumulations in these samples, which are therefore not necessarily polluted. Natural processes of organic matter complexation suggest the same applies to Pb in strongly organic samples. However, a small group of Subatlantic clays with elevated Pb-contents probably reflect increased atmospheric Pb-input resulting from deforestation and Roman era mining activities. In addition, heavy mineral-content and lime-content are able to influence the background values of Zr, Nb, Nd, Y and MgO and Sr respectively. The composition of fine-grained Meuse sediments thus systematically varies on a time-scale of 1000 to 10,000 years and on a local to regional spatial scale. Any environmental legislation should certainly account for this variation to be realistic.

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CHAPTER 4

THE IMPACT OF CLIMATE CHANGE ON THE BULK AND CLAY GEOCHEMISTRY OF FLUVIAL RESIDUAL CHANNEL INFILLINGS

Based on: Tebbens, L.A., Veldkamp, A. and Kroonenberg, S.B. 1998.

The impact of climate change on the bulk and clay geochemistry of fluvial residual channel infillings: The Late Weichselian and Early Holocene River Meuse sediments (The Netherlands).

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4 THE IMPACT OF CLIMATE CHANGE ON THE BULK AND CLAY GEOCHEMISTRY OF FLUVIAL RESIDUAL CHANNEL INFILLINGS

ABSTRACT

Fine-grained fluvial residual channel infillings are likely to reflect systematic compositional changes in response to climate change due to changing weathering and geomorphological conditions in the upstream drainage basin. This research focuses on the bulk sediment and clay geochemistry, laser granulometry and clay mineralogy of Late-glacial and Early Holocene Meuse (Maas) unconsolidated residual channel infillings in northern Limburg (the Netherlands). We demonstrate that although sediment provenance has not changed considerably, residual channel infillings register a systematic bulk and clay compositional change related to climate change. This indicates that the composition of River Meuse sediments can not be considered constant on a 1-10 ka time-scale.

Late-glacial and Holocene climatic amelioration stabilised the landscape and facilitated prolonged and intense chemical weathering of phyllosilicates and clay minerals due to soil formation. Clay translocation and subsequent erosion of topsoils on Palaeozoic bedrock and loess deposits increased the supply of smectite and vermiculite within River Meuse sediments. Smectite plus vermiculite contents rose from 30-40% in the Pleniglacial to 60% in the Late Allerød and to 70-80% in the Holocene. Younger Dryas cooling and landscape instability caused almost immediate regression to low smectite and vermiculite contents. Following an Early Holocene rise, within about 5000 years steady state supply is reached before 5-ka BP (Mid Holocene). Holocene sediments therefore contain higher amounts of clay that are richer in high-Al, low-K and low-Mg vermiculites and smectites compared to Late (Pleni-)Glacial sediments. The importance of clay mineral provenance and loess-admixture in the Meuse fluvial sediments is discussed.

4.1 INTRODUCTION

Palaeohydrological studies indicate that fluvial systems react strongly to climate change through associated complex response dynamics in their drainage basins (Starkel et al., 1991; Kasse et al., 1995; Vandenberghe, 1995). These studies have mostly concentrated on the morphological behaviour of rivers, but little attention has been paid to systematic compositional changes of fluvial sediments in relation to climate change. The use of sedimentary palaeoclimatic indicators in fluvial sediments therefore still requires rigorous calibration of the climatic controls on the formation of those indicators (Parrish et al., 1993).

Fluvial sediments are derived from soils and rocks, which have been exposed to physical and chemical weathering processes in the upstream drainage basin. Provenance determines the bulk geochemical characteristics of fluvial sediments via the mineralogical composition of the source rocks before weathering modifies them. Weathering can be described in terms of weathering intensity and duration. Temperature and precipitation determine weathering intensity and thus climate has significant influence on the mineralogical composition of soils and fluvial sediments when erosion of parent material is slow (Irion, 1991). The tectonic regime influences weathering duration via its control on local relief. Local relief in the drainage basin determines the residence time of parent material in a catchment and intervenes in the erosion, transport and sedimentation pathways of weathering products (Curtis, 1990; Johnsson, 1993). Soil formation lengthens chemical weathering duration and intensifies hydrolysing processes. Only in a transport-limited situation, fluvial sediments will largely originate from soils and reflect the climatic conditions under which the soils have formed (Singer, 1984). Climate will then be reflected in soil clay mineralogy and in its bulk geochemical composition (Curtis, 1990).

Erosion, transport, sedimentation and burial of characteristic soil clay assemblages can obscure the climatic signal within sediments, by mixing of material of different soil horizons with different weathering intensities (Curtis, 1990; Johnsson, 1993). Therefore, careful consideration should be given to reworking, sorting and diagenetic overprinting processes. In spite of these potential pitfalls, we made a study of the composition of fine-grained residual channel infillings for evidence of the climatic record as information in this field is still scarce (Chamley, 1989). To this purpose, we investigated the geochemistry of Dutch River Meuse sediments deposited during the transitional period of the Weichselian Late-glacial to the Early Holocene (13-9 ka BP). A general warming trend characterises this period, interrupted by two cooling phases during the Late-glacial Older and Younger Dryas Stadials (Van Geel et al., 1989; Bohncke, 1993; Hoek, 1997).

4.2 LATE-GLACIAL GEOLOGICAL SETTING

The Meuse headwaters drain Jurassic marls and limestones (7007-km^2) in the northeastern Paris Basin. Several tributaries draining Palaeozoic low-grade metapelites and quartzites (13,672-km^2) in the Ardennes low mountain range join the Meuse middle reach. Via a small area of Upper Cretaceous limestones, the Meuse enters the loess-covered Quaternary fluvial terrace and delta landscape (8691-km^2) in the Netherlands (Berger and Mugie, 1994). The loess cover is thickest on the terraces, but less on the limestones. This drainage area has not changed since the Moselle captured the Upper Meuse Vosges basin during the Early Saalian and made it part of the Rhine fluvial system (Fig. 1.1A; Bustamante, 1976; Bosch, 1992). Thus, sediment provenance changes can be excluded for the last 250,000 years.

The Late Pleniglacial climate was extremely cold and dry. A large influx of cover sands and loess in the Meuse drainage basin caused intermittent fluvio-aeolian sedimentation (Schwan, 1988; Mol et al., 1993) and the aggradation of a well-expressed Late Pleniglacial terrace (Van den Berg and Veldkamp, 1996). Permafrost conditions prevailed in northwestern Europe (Vandenberghe and Pissart, 1993), reducing water mobility and penetration within soil profiles and thus favouring slope erosion and gelifluction processes via runoff. Physical weathering dominated and residence time of weathering products within the catchment was presumably short.

The morphodynamics of the Meuse in the study area (Fig. 2.2, 2.3) changed considerably as a response to Late-glacial climatic change (Fig. 2.8, 2.9). Increasing temperatures, precipitation and vegetation cover during the Late Glacial are well documented in northwestern Europe (cf. Van Geel et al., 1989; Bohncke, 1993; Hoek, 1997). Vegetation stabilised the slopes and plateaux leading to favourable preconditions for soil formation. Sea-level effects on sedimentation can be excluded, because sea-level rise did not influence the downstream Maas and Waal area before the Late Atlantic (Berendsen et al., 1995). Moreover, the study area has always been located upstream of the terrace intersection. Van den Broek and Maarleveld (1963), Kasse et al. (1995), Huisink (1997) and Tebbens et al. (Chapter 2; 1999a) reconstructed the following Late-glacial fluvial morphodynamics. The Pleniglacial Meuse was a braided system, evolving to a meandering system during the Bølling-Allerød Interstadial and back to braiding conditions during the Younger Dryas Stadial. River dunes were formed along the east bank of the Meuse during the dry second half of the Younger Dryas (between 10500 and 10150 BP; Bohncke et al., 1993). The Holocene Meuse saw a single meandering channel again. Van den

Broek and Van der Marel (1964, 1980) described soil formation, soil properties and sediment characteristics in Meuse alluvia. Miedema (1987) and Schröder (1983) studied the soils on nearby Rhine sediments. Anthropogenic agriculture and deforestation practices have augmented soil erosion and subsequent suspended load sedimentation in the Meuse valley since the Early Subboreal (Havinga & Van den Berg van Saparoea, 1980; Teunissen, 1990; Pastre et al., 1997).

4.3 METHODS

We studied Late (Pleni-) Glacial to Early Holocene infillings of Meuse residual channels, a construction pit in Lottum and the sand pits at Bosscherheide, Grubbenvorst and Panheel in the southeastern Netherlands (Fig. 2.3). Clayey and silty sediments have been extensively cored to provide as complete a sedimentary record as possible (Chapter 2, Fig. 2.5; Tebbens et al., 1999a). The sampling strategy aimed at preferably unconsolidated and unpolluted samples to perform granulometric measurements and bulk geochemical analyses on major constituents and trace elements. Details on the sampling sites, sampling strategy and analysing techniques are given in Chapter 2 and Chapter 3 (sections 3.3 and 3.4). Chapter 3 also provides information on the variables and processes determining the lions' share of bulk geochemical variation in the samples.

In addition to the bulk geochemistry of the samples, the mineralogy and geochemistry of about 60 fractionated clay samples has been determined with X-Ray Diffraction (XRD) and XRFS (see 3.3). Clay fractions were made according to standard methods and clay samples were oriented on ceramic platelets. The oriented clay samples were saturated with 1-M $MgCl_2$, glycerol and 1-M KCl. The KCl-saturated samples were additionally heated at 150-°C. The swelling characteristics and clay mineralogy were determined via XRD-measurements of the 001-reflections. Semi-quantitative peak surfaces were obtained by multiplying the 0.7, 1.0, 1.4 and 1.8-nm peak heights above background with the peak widths at half peak heights. Within one dataset, reliable relative clay mineral percentages can then be calculated by dividing the peak surface for an individual clay mineral by the sum of all peak surfaces. The clay geochemistry was analysed by XRFS (see section 3.4) after the clay fractions had been separated, saturated and settled with 1-M $BaCl_2$ and then freeze-dried.

The (approximate) time of deposition of the sediments was determined to relate bulk geochemical variation to Late Glacial climatic changes. For this purpose, the start and end of channel infillings were conventionally ^{14}C-dated on strongly organic gyttjas and peat (see Chapter 2). As it is practically impossible to date every clayey sample, the age of clayey and silty layers that were intercalated between these organic deposits was determined by linear interpolation between successive ^{14}C dates. Sandy units were excluded from these calculations as they are assumed to have been deposited instantaneously or can result from aeolian influxes. Dating standard errors lie in the range of 50 to 250 years. Fine-grained clastic residual channel infillings will originate from several flooding events rather than from continuous sedimentation. Therefore, depositional hiatuses can be present in the clastic interval between dated organic layers. In those cases, linear interpolation introduces errors, because one supposes constant sedimentation rates for the interval. In this research, clastic intervals between successive organic layers in Late-glacial residual channels lasted 210 to 820 ^{14}C-years, so any depositional hiatus in these intervals is certainly shorter than 820 years (see Fig. 2.5). Four clayey gyttja intervals without any lithological signs of hiatuses lasted 1040 to 1730 ^{14}C-years. Three Holocene clayey intervals (Sites 1e, 2 and 12 in Figures 2.3 and 2.5) lasted 1920, 2370, and 3510 years, but the coarser textures in some samples hint to changing depositional environments and possible hiatuses. Clay sample ages can thus differ some hundreds of years from real age of deposition.

To minimise these errors, we used additional palynological information on the age of both gyttja and clayey units (Teunissen, 1990; Bohncke et al., 1993; Hoek 1997). Nevertheless, since we are interested in bulk sediment and clay compositional changes on a time-scale of 1-10 ka, we accept possible age uncertainties in the order of centuries.

4.4 RESULTS

Bulk geochemical scatterplots

Before a climatic signal is to be distinguished within the large dataset, first the effects of various sorting and diagenetic overprinting processes on bulk geochemistry have to be understood and quantified. Grainsize-sorting (factor 1), diagenetic mineral formation in anoxic gyttja environments (factor 3), and variations in heavy mineral (factor 2) and carbonate content (factor 4) appeared to cause the main bulk geochemical variation in the dataset (Chapter 3). Chamley (1989) and Curtis (1990) anticipated that climatic signals are likely to survive best in bulk compositional data of fine-grained Quaternary fluvial sediments. Climatic signals are thus most likely to be found in the finest size fractions. Therefore, we will concentrate on the geochemistry of the clay and fine silt size-fraction in this section.

The principal component analysis (section 3.5) and Figure 3.3 indicated that the major constituents Al_2O_3, K_2O, MgO and TiO_2 are mainly hosted within the 0-20 μm size-fraction and therefore they were strongly interrelated in factor 1. Because the former constituents did not show up in the other factors, they are the most likely candidates to register a climatic signal that is not disturbed by overprinting processes. On the other hand, major constituents like Fe_2O_3 and P_2O_5 were hosted in factor 3 and are likely to have been overprinted by early diagenetic siderite and vivianite formation under local redox conditions in gyttja environments. It will be clear that local overprinting processes severely hamper a reconstruction of climatic fluctuations in the upstream drainage basin from the detrital sedimentary record. Thus, the constituents in factor 3 are unsuitable as sedimentary palaeoclimatic indicators. Likewise, post-depositional precipitation or dissolution of lime can blur climatic reconstructions based on CaO-content (factor 4).

Figure 3.2 demonstrated a strongly positive relationship ($r = 0.88$) between the 0-20 μm granulometric fraction (clay and fine silt) and Al_2O_3. Thus, the Al_2O_3-percentage is a very suitable geochemical proxy for the clay and fine silt content (Chapter 3, cf. Moura and Kroonenberg, 1990; Huisman and Kiden, 1998). Figure 4.1 shows that the Holocene samples appear to have higher Al_2O_3-contents than Late-glacial samples with a comparable granulometry. The K_2O-Al_2O_3- and the MgO-Al_2O_3-scatterplots (Figs. 4.2 and 4.3) demonstrate strong clay-mineralogical control on both the K_2O-contents ($r = 0.93$) and MgO-contents ($r = 0.88$) of the samples, as was expected from their covarying behaviour in principal component factor 1 (Table 3.1). Chronological labelling reveals that Holocene bulk sediment samples and separated clay fractions have distinctly lower K_2O- and MgO-contents than Late-glacial samples with comparable Al_2O_3-contents or granulometry. Furthermore, calcareous samples have elevated MgO-contents compared to non-calcareous samples, but Holocene calcareous samples still show lower MgO-contents than Late-glacial calcareous samples (Fig. 4.3).

The grainsize factor (1) also hosted several trace elements that are typically built into clay mineral lattices, like Ba, Cr, Rb and V (Table 3.1). Correspondingly, the Al_2O_3-Rb plot (Fig. 3.4B) and the plots of K_2O versus Ba, Cr, Rb and V (Fig. 4.4) reveal similar differences between Pleni- and Late-glacial and Holocene samples as found in the Al/K- and Al/Mg-plots. The Rb-contents are lower with respect to Al_2O_3 in the silty to clayey granulometric range, but the Ba-, Cr-, Rb-, and V-contents are higher compared to K_2O in Holocene samples.

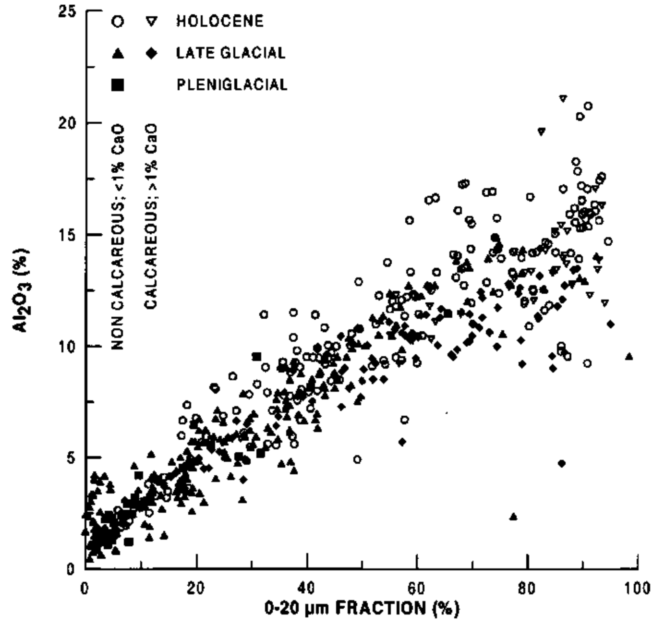

Figure 4.1 The relationship between the 0-20 µm fraction (determined by laser granulometry) and the contents of Al_2O_3 for samples of Pleniglacial and Late-glacial (filled symbols) and Holocene (open symbols) age. Note higher Al-contents for both clayey and Holocene samples.

Clay mineralogy and geochemistry

A fine-grained Late Pleniglacial residual channel infilling, dated at 13280 ± 70 BP (GrN-21882), in the Bosscherheide sandpit contained many small mica flakes (muscovite), small leaves (a.o. *Betula nana*, dwarf birch) and moss remnants, indicating a low-energy setting. An important observation was that clay-size material was virtually absent, although deposition of clay-sized particles could have been expected here. Furthermore, most Holocene residual infillings generally showed higher clay contents than Late-glacial residual infillings did (compare Late-glacial and Holocene samples in Fig. 4.1), suggesting higher supply of clay-size material to younger sediments. The higher Al_2O_3-contents and lower K_2O and MgO-contents in Holocene samples might be related to grainsize differences only (supply of clay-sized particles). To check whether the compositional changes could be attributed to a true clay-compositional trend as well, we separated clay fractions for a selection of samples and investigated them for their geochemistry and clay mineralogy.

Clay-fraction geochemistry indicates high Al_2O_3 and high K_2O and MgO for Late-glacial samples, but significantly lower K_2O- and MgO-contents in the same Al_2O_3-range for Holocene samples (Figs. 4.2 and 4.3). Considerable overlap exists for the clay fractions in the Al_2O_3/K_2O plot. From labelling these overlapping samples it turned out that Holocene samples high in K_2O have been deposited during the transitional Early Holocene and that Late-glacial samples with low K_2O-contents (Bosscherheide sandpit) were deposited as a 0.4-m overbank loam. Although the overbank loam was reduced when sampled, it may have went through a period of oxidation and soil formation in the past.

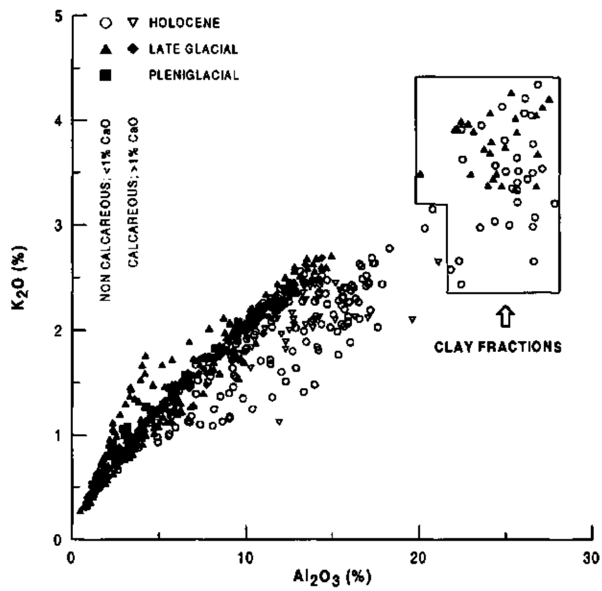

Figure 4.2 The relationship between Al_2O_3 and K_2O for bulk sediment and clay fraction samples (within "rectangle") of Pleniglacial and Late-glacial (filled symbols) and Holocene (open symbols) age. Note distinctly lower K_2O-contents in Holocene samples.

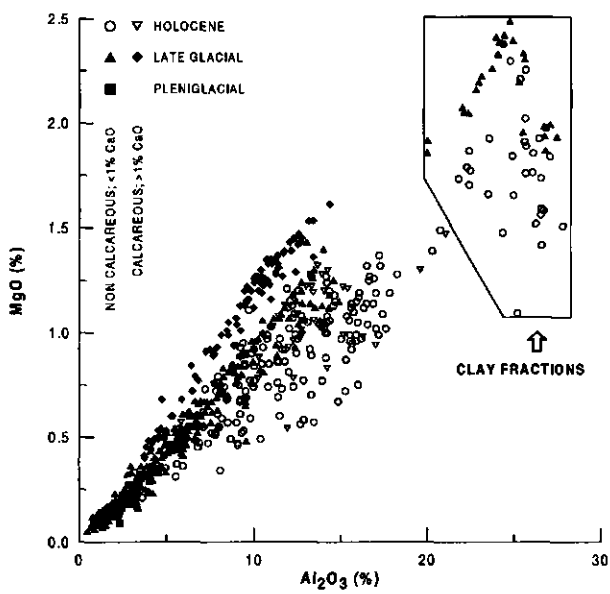

Figure 4.3 The relationship between Al_2O_3 and MgO for bulk sediment and clay-fraction samples (within "rectangle") of Pleniglacial and Late-glacial (filled symbols) and Holocene (open symbols) age. Note relatively low MgO-contents in Holocene samples and relatively high MgO-contents in calcareous samples.

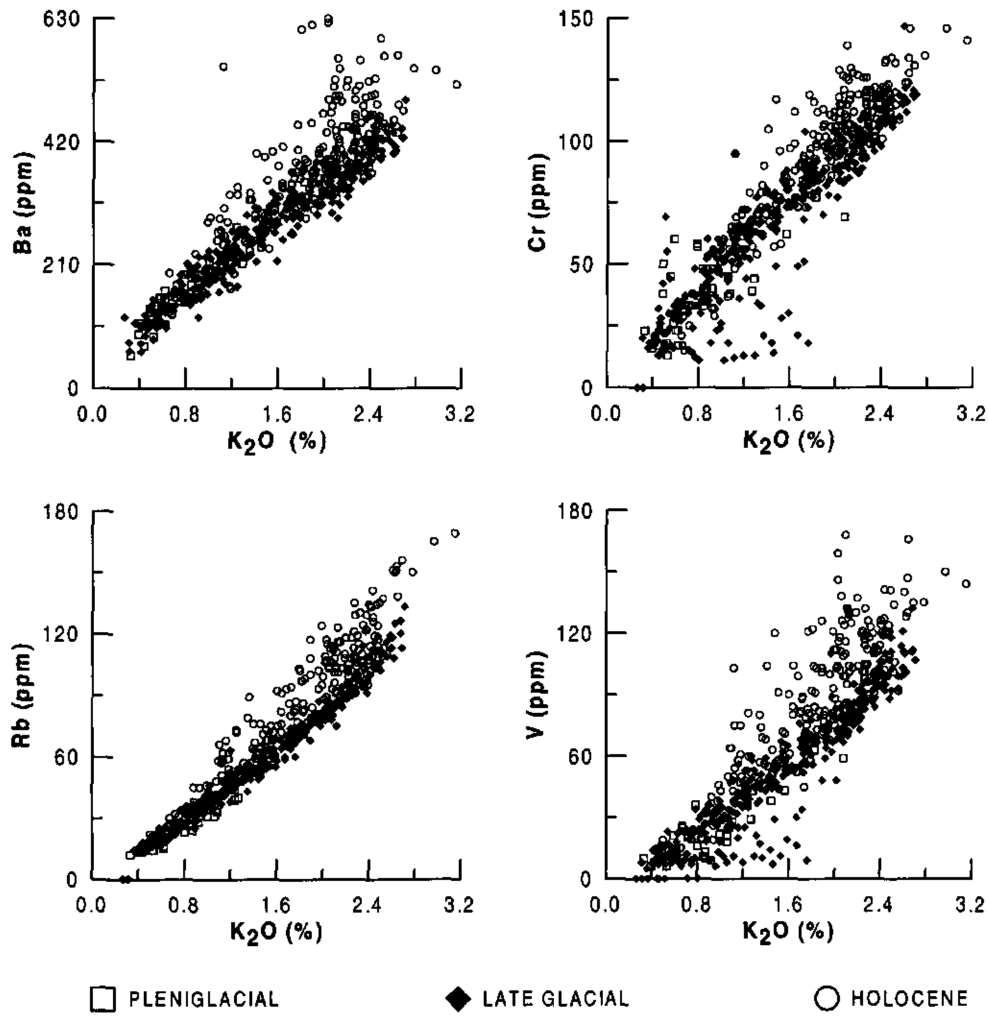

Figure 4.4 The relationship between the contents of K_2O and the trace elements Ba, Cr, Rb and V for 636 bulk sediment samples of Pleniglacial, Late-glacial and Holocene age. Note the systematic shift towards relatively higher values for Ba, Cr, Rb and V with respect to K_2O for Holocene samples.

The XRD-analysis and semi-quantitative peak surface measurements indicate the presence of the following clay minerals (indicative chemical composition between brackets): kaolinite ($Al_2Si_2O_5(OH)_4$), illite ($KAl_5Si_7O_{20}(OH)_4$), little chlorite ($Mg_{10}Al_4Si_6O_{20}(OH)_{16}$), smectite ($Na_{0.7}(Al_{3.3}Mg_{0.7})_4Si_8O_{20}(OH)_4$), vermiculite ($(Mg,Ca)_{0.7}Mg_6(Al,Si)_8O_{20}(OH)_4$) and 1-1.4 nm mixed-layer clay minerals. Kaolinite, illite and chlorite dominate Late-glacial clays, whereas Holocene clays contain 30-40% more smectite and vermiculite, but no chlorite. The time-clay mineralogy plot (Fig. 4.5) clearly indicates a rise in smectite plus vermiculite content during the course of the Holocene. A maximum is reached around the climatic optimum at 5-ka BP, follow-

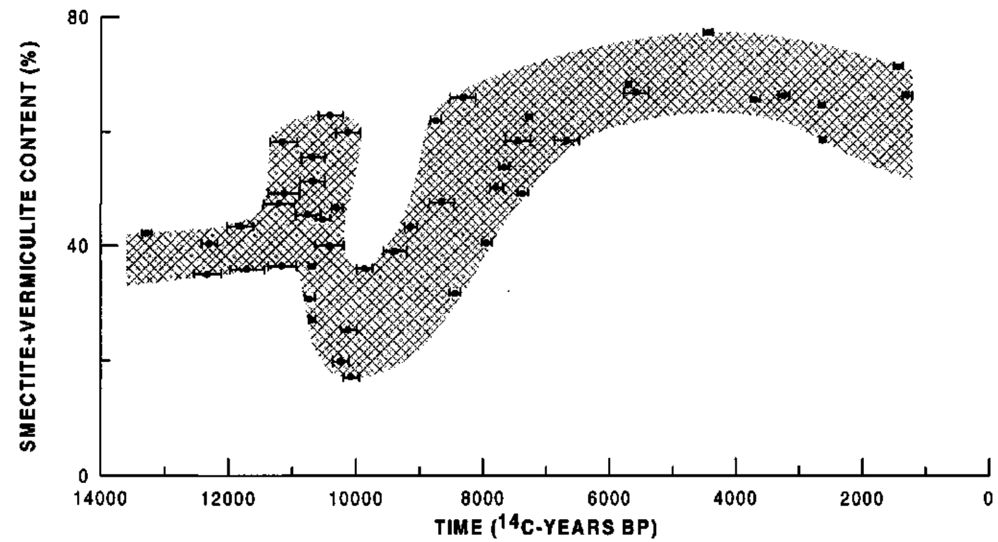

Figure 4.5 Semi-quantitative contents (XRD) of smectite plus vermiculite in clay fraction samples versus their approximate age of deposition (see text). The error bars indicate ^{14}C-dating standard errors (2σ).

ed by somewhat lower smectite and vermiculite contents in the Subboreal and Subatlantic periods. Note a scarcity of samples in the 7-4 ka time-frame because of large-scale non-clastic peat accumulation during the warm and humid Atlantic period. When the Late-glacial period is included (14-10 ka BP), the picture is somewhat blurred due to the broad range in smectite plus vermiculite contents. The same rise however, seems to be present for the Bølling-Allerød Interstadial (13-10.8 ka BP), followed by a sharp drop at the Younger Dryas Stadial (10.8-10 ka BP).

A principal component analysis (combining semi-quantitative clay mineralogy with clay geochemistry) yielded that K_2O was positively related to the illite content (r = 0.79) and MgO slightly positively related to the chlorite content (r = 0.47). Both however, were negatively related to the vermiculite content with multiple correlation coefficients of -0.59 and -0.64 respectively. Likewise, it extracted a factor with high positive loadings for the interlayer cations K_2O, its substitute Rb, and MgO, and strong negative loadings for vermiculite content and LOI.

4.5 MEUSE CLAY MINERALOGY: TIME AND PROVENANCE

Age labelling in scatterplots shows that the bulk composition of Meuse River sediments can not be considered constant on a time-scale of 1000-10,000 years. The contents of K_2O, MgO and TiO_2 were found lower in Holocene samples relative to Late-glacial and Pleniglacial samples with comparable Al_2O_3-contents. The contents and ratios of Ba, Cr, Rb and V relative to Al_2O_3 are only slightly lower in Holocene sediments. However, when Ba, Cr and especially Rb and V are plotted versus the K_2O- (Fig. 4.4) and MgO-contents (not shown), a shift to considerably higher values relative to K_2O and MgO in Holocene samples emerges. This suggests that the climatic scattering be mainly due to preferential depletion of K_2O and MgO in Holocene samples. Late-glacial samples contain relatively more illite and chlorite and are relatively rich in K_2O and MgO. Calcareous samples incorporate more MgO than non-calcareous samples. Holocene samples contain more Al-rich smectite and vermiculite and consequently have lower K_2O- and MgO-contents. Calcareous Holocene samples showed lower MgO-contents than Late-glacial calcareous samples, suggesting a clay-mineralogical effect.

Likewise, Van den Broek and Van der Marel (1980) found large amounts of smectites (they use the term "swelling illites") in young Holocene Meuse alluvial *soils* and delta deposits. They ascribed this to post-depositional weathering of illites to smectites in a swampy delta environment, but also mention the clays could have been deposited this way. In a research of Late Weichselian and Holocene Rhine alluvial soils, Schröder (1983) too found larger amounts of smectites in soils on younger alluvia compared to older alluvia. He assumed that the smectites in the older soils had been transformed to illites. This research demonstrates that in most Holocene *sediment* samples the amount of vermiculites and smectites has increased at the expense of kaolinite, illite and chlorite contents. The question now arising is how to explain the abundance of Al-rich and K- and Mg-poor smectite and vermiculite in Holocene sediments compared to Late-glacial sediments. Is the change in clay mineralogy with time due to post-depositional alteration, to provenance effects or to climatic change?

Post-depositional alteration

As was already briefly mentioned above, post-depositional alteration has been invoked to explain differences in clay mineralogy in literature on alluvial *soils*. Most of our samples however, originated from 2 to 8-m deep residual channels with no burial diagenetic overprint and have been taken from *unconsolidated (unripened) reduced sediments*. Due to reaction of Fe(II) with groundwater rich in bicarbonate and phosphate some diagenetic siderite and vivianite has been formed in (clayey) gyttja environments (Chapter 3). Neither in this research nor in literature (Chamley, 1989) we found evidence that high organic matter contents and/or redox reactions influence the original clay mineralogy and bulk composition on a time-scale of 1-10 ka. Post-depositional alteration of clay minerals can therefore be expected to be minimal and the clay mineral assemblage thus very likely reflects the original suite of clay minerals at the time of deposition. Therefore, we argue that during the Holocene the supply of smectite, vermiculite and 1-1.4 nm mixed-layer clay minerals to the Meuse (and apparently the Rhine sediments too) has increased relative to the detrital chlorite-illite-kaolinite background suite. Because post-depositional alteration can be excluded, the causal question of the higher smectite and vermiculite contents reduces to their source areas and weathering regimes: where do these higher amounts of smectite and vermiculite originate?

Provenance

In this section, we will assess the contribution of different source areas and whether the clay minerals reflect pedogenetic and climatic conditions in the drainage area or not. Clay minerals in unaltered fluvial sediments are always from detrital origin, i.e. eroded from soil environments and thus fingerprint the weathering conditions in the upstream drainage basin (Singer, 1984; Chamley, 1989; Velde, 1995). In this way, large sedimentary basins integrate the clay mineral response to climate change (Hillier, 1995). Clay mineral provenance is very important because regions with different bedrock and climatic conditions will supply a wide diversity of clay minerals. Since the drainage area essentially remained unchanged during the glacial-interglacial transitional period, source rock provenance changes that are related to the gain or loss of tributary catchments can not be held responsible. The clay mineral suites possibly originate from several weathering cycles. However, we expect that the Pleistocene glacial conditions, including the latest Weichselian, will have cleared most pre-existing Tertiary and Pleistocene interglacial weathering mantles and have rejuvenated the landscape to great extent. The South Limburg Pleistocene flight of 31 aggradational gravelly terraces with decreasing quartz contents for the younger terraces corroborates this (Van den Berg, 1996). The clay mineral suites thus reflect source rocks and overlying soils, which for the greater part have been developed under Latest Weichselian and Holocene weathering conditions.

Three major clay-mineral source areas have to be considered, namely *1)* Carboniferous, Jurassic and Upper Cretaceous limestones in the headwaters of the northeastern Paris Basin, *2)* Palaeozoic slates, phyllites and quartzites of the Ardennes low mountain range and *3)* Weichselian loess and fluvial deposits in the southern Netherlands, northern Belgium and Germany (Fig. 4.6).

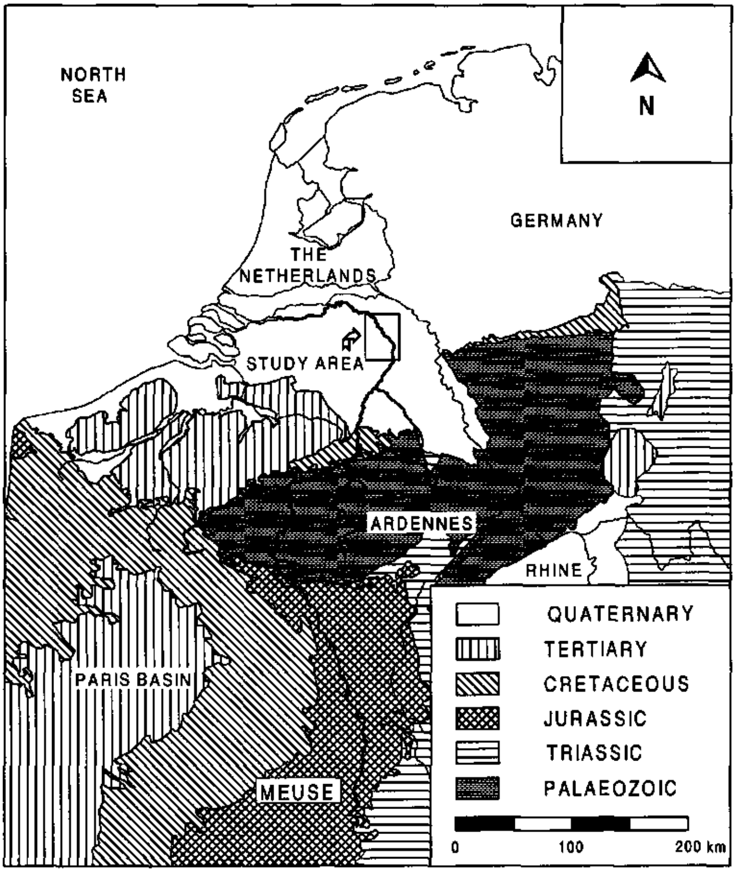

Figure 4.6 The River Meuse drainage area with the major chronostratigraphical units.

Carboniferous, Jurassic and Upper Cretaceous limestones

Clay originating from weathered Carboniferous limestones and calcareous slates contains illite, kaolinite and small amounts of illite-smectite interstratified minerals (Thorez and Bourguignon, 1971). The dissolution residues of Jurassic and Upper Cretaceous limestones drained by the

Upper Meuse mainly contain kaolinite and smectite (Pomerol and Riveline-Bauer, 1967; Chamley, 1989). We calculated mean annual suspended loads with the help of present day discharges and mean suspended load concentrations measured at St. Mihiel (pre-Ardennes, France), Liège (including Ardennes, Belgium) and Eijsden (including loess area, Netherlands). This yielded a 1:5:1-ratio for contributions of the French, Belgian and Dutch drainage areas to the annual suspended load of the Meuse respectively. The dissolution residues at maximum constituting 1-2% of the limestones, their contribution to the Meuse clay mineral assemblage is considered negligible in view of the large amounts that are contributed by weathering of Palaeozoic shales and phyllites or erosion of loess deposits. Testifying to this, Van den Broek and Van der Marel (1980) found smectites in only minor amounts in the upstream Ardennes part of the River Meuse.

Palaeozoic source area

Erosion of weathering mantles and soils of the low-grade metamorphic Palaeozoic Ardennes (including Famenne and Condroz regions) forms a second major clay mineral source. Bedrock of diverse geological age and soil horizons (with some loess admixture) in the Ardennes demonstrate similar mineralogy: quartz, feldspars, micas (biotite and illite) and tri-octahedral chlorite in both silt and clay fractions, with additional amphibole in the silt fraction. Smectites are absent (De Coninck et al., 1979). Contemporary soils of the Ardennes, Famenne and Condroz regions supply weathered muscovite and illite, vermiculite ("expanded illite"), chlorite and kaolinite to the Meuse tributaries and trunk stream. Smectites ("swelling illites") on the other hand were found in a very limited amount only (Van den Broek and Van der Marel, 1980). Dufey and Lambert (1987) report illite-smectite interstratified clay minerals in soils on Jurassic rocks and illite, vermiculite and kaolinite in the Palaeozoic Ardennes and Famenne soils. Soils on Palaeozoic bedrock thus are a main source for vermiculite, but lack smectite.

Erosion of loess deposits and/or their top luvisoils

The following evidence indicates a loess provenance for at least part of the Meuse clay minerals:

a) On granulometrical grounds, Bohncke et al. (1993) and Törnqvist et al. (1994) state that loess admixture has taken place in the Wijchen Member of the Kreftenheye Formation. The Wijchen Member is the current lithostratigraphical name for Late Weichselian and Early Holocene silty to sandy clay overbank deposits and residual channel deposits from which our samples largely originated. Loess deposits are dominated by the 16-80 μm fraction and contain calcium carbonate particles with some substitution of Ca by Mg (Van der Marel and Van den Broek, 1962). In this research, many Late-glacial granulometrical curves show a top in the 20-80 μm range. Calcareous silty sediments with high percentages of the 20-80 μm fraction contained more CaO and MgO. Loess admixture within Meuse sediments therefore seems to be very important. Substitution of dolomitic carbonate in the calcium carbonate crystal lattices is probably also responsible for the elevated MgO-percentages in calcareous samples.

b) According to Van den Broek and Van der Marel (1980) young Meuse alluvial soils contain 30-40% smectites ("swelling illites"). As "swelling illites" were not encountered in the soils of the Palaeozoic drainage area, neither in the main Ardennes Meuse tributaries, they are probably derived from the Geul and Roer tributaries. These tributaries drain the loess belt in Southern Limburg and adjacent Germany. Their sediments contain up to 30-40% "expanding or swelling" illites. The clay fraction from weathered and unweathered loess profiles is dominated by illite but also contains 5-15% smectite and vermiculite (Mücher, 1973). Considering the large amounts present in the clay fraction of sediments of the Geul and Roer tributaries that drain the loess belt, the "swelling illites" probably largely originate from fine suspended clay eroded from loess deposits and associated luvisoils (Tavernier and Louis, 1986).

c) Paepe (1971) did not find Late-glacial paleosols in Belgian loess areas with a pronounced relief and concludes that the time for their formation was too short. In our opinion, this strongly suggests that erosion has been very important in sloping loess areas, as Late-glacial soils do exist in the coversand area. In a forest-covered loess area, which has been stable in the last 20,000 years, Langohr and Sanders (1985) found evidence for one relatively short fluviatile erosional cycle at the end of the Weichselian, post-dating a period of decalcification and clay illuviation. Strong erosion in loess blanketed Upper Cretaceous hills, drained by the Meuse tributary Geul, was not reported before forest clearance in the Subboreal (Havinga and Van den Berg van Saparoea, 1980). However, the deepest part of their Etenaken profile shows at least 0.5-m "fluvially reworked" loess deposited in Preboreal and Boreal times, suggesting that pre-anthropogenic loess erosion did take place in the Late-glacial or Early Holocene.

With the most important source areas in mind, namely the Palaeozoic rocks and the Weichselian loess belt, one can begin to reconstruct the effect of climate on clay mineral supply. Changes in the respective contributions of these areas or prolonged soil formation in the drainage basin are the most logic explanations for the observed change in Meuse clay mineralogy with time. These effects will be discussed in the following section.

4.6 LANDSCAPE STABILITY, SOIL FORMATION AND CHEMICAL WEATHERING

During preceding glacials and the Weichselian Early Glacial and Pleniglacial, gelifluction processes have led to drainage area wide soil stripping, leading to short residence times for weathering material in the sloping Ardennes tributary catchments. Permafrost, water shortage, cold temperatures and sparse vegetation must have led to strongly reduced hydrolysing processes and chemical weathering rates in Late Pleniglacial periglacial soil environments. Immature frozen soils predominated (Bohncke, 1993) suggesting a poverty of erodible clayey soil horizons before the onset of the Late Glacial. An extreme Antarctic example shows that even after 4-Ma soil formation almost no clay is formed and, if any, consists of source rock inherited illite and chlorite (Matsuoka, 1995). The low clay content of Late-glacial Meuse sediments - even in low-energy settings - suggests low clay availability and input during the early Late-glacial.

Chemical weathering duration and intensity during the Pleniglacial must have been minimal within the whole Meuse drainage basin. Fresh loess deposits covered older interglacial soils and were still unweathered as they contain 10-15% lime (Langohr and Sanders, 1985). Late-glacial erosion of the Pleniglacial weathering mantles and calcareous loess deposits within the Meuse drainage basin (Kasse et al., 1995, Van den Berg, 1996) gave rise to a small amount of physically weathered clays. Detrital kaolinite, illite, chlorite and relatively low amounts of pedogenetic smectite, vermiculite and 1-1.4 nm mixed-layer clay minerals thus dominate the Late-glacial Meuse clay mineralogy.

The Late-glacial climatic amelioration induced higher landscape stability and stronger chemical weathering processes in the upstream drainage basin. The disappearance of permafrost favoured water percolation in soil profiles and successive leaching of basic cations to the groundwater. In the southern Netherlands and Belgium lowlands, permafrost had completely disappeared before 12.4-ka BP (Hoek, 1997), but the higher-elevated Ardennes will have followed later. Leaching of cations to the groundwater deplete the soil solution of K and Mg and so does nutrient uptake by vegetation. The interstadial or interglacial climatic conditions enabled the re-establishment of a protecting vegetation cover (Guiot and Couteaux, 1992; Lefèvre et al, 1993; Hoek, 1997; Pastre et al., 1997), favouring soil formation and concurrent chemical weathering. Nutrient uptake by this vegetation cover may account for the withdrawal of K and

Mg from soil mineral environments on a landscape-wide scale. (e.g. Bain et al., 1995; Berner et al., 1996; Korobova et al., 1997). Paleosols indicate Bølling-Allerød soil formation in the upstream Paris Basin (Pastre et al., 1997), in a coversand area in Belgium (Paepe, 1971) and at the Bosscherheide and Grubbenvorst sites in our study area. Late-glacial clay illuviation is reported for soils in the Ardennes area (De Coninck et al., 1979; Langohr and Van Vliet-Lanoë, 1979) and in the loess belt (Langohr and Sanders, 1985, Sanders et al., 1986). Intense soil formation in Rhine alluvia included decalcification, clay illuviation and subsequent pseudogleying in higher topographical positions and gleying in low topographies (Miedema, 1987). Drainage basin wide clay translocation may rapidly have increased the supply of fine suspended pedogenetic clays (smectites and vermiculites) to streams and tributaries.

Higher temperatures along with abundant and percolating water due to higher precipitation and the disappearance of permafrost speed up the hydrolysis of substrate minerals. The intense chemical weathering resulting from Temperate Zone pedogenetic processes involves extraction of interlayered potassium from muscovite, biotite and illite and dissolution of Mg-bearing brucite layers from chlorite. The mineralogical evolution of the Ardennes soils (De Coninck et al., 1979) shows that silt-size micas and chlorite in the topsoils are physically broken down into clay particles, leading to an absolute accumulation of clay. Clay fraction phyllosilicates undergo a partial vermiculitisation and are transformed into chlorite-vermiculite and mica-vermiculite interstratified clay minerals. Part of the chlorite is broken down completely, without formation of smectites. Bronger et al. (1976, 1998) showed that phyllosilicate and feldspar minerals in the silt fractions from B- and Bt-horizons in loess soils weather to illites, vermiculites and sometimes to smectites. Mica weathering in acidic soils involves two steps (Aoudjit et al., 1996): first the micas transform into 1-1.4 nm mixed-layer minerals and then into hydroxy-Al interlayered vermiculite. In this way, weathering products like vermiculite, smectite, mica/vermiculite, illite/smectite and chlorite/vermiculite mixed-layer clay minerals are formed which are enriched in Al and impoverished in K and Mg (Righi and Meunier, 1995; Dixon and Weed, 1989). Miedema (1987) found smectites and strongly Al-interlayered vermiculites in topographically higher soils, but no Al-interlayering in topographically lower and recent Holocene soils. Both weathering duration and weathering intensity thus increase as a consequence of postglacial soil formation.

To produce a characteristic pedogenetic clay mineral assemblage requires time. A higher supply of typically soil derived clay minerals as vermiculite, smectite and 1-1.4 nm mixed-layer clay minerals can thus be expected to appear after an appropriate period of soil formation. These conditions existed at the end of the Bølling-Allerød Interstadial and during the course of the Holocene. Figure 4.5 illustrates that the smectite/vermiculite content rose between ± 12 and 11-ka BP (Late-glacial Interstadial) as well as between 10 and 6-ka BP. The sharp drop from 10.8 to 10-ka BP possibly reflects reworking and input of gelifluction material into the Meuse due to Younger Dryas discontinuous permafrost, forest dieback and braided river conditions (Kasse, 1995). The Holocene trend in Figure 4.5 suggests a rapid rise in smectite and vermiculite content from the Younger Dryas minimum of 20-30% at about 10-ka BP to a mid-Holocene maximum at approximately 5-ka BP of 70-80%. After 5-ka BP, the smectite and vermiculite contents stabilise at a high level and hardly change anymore. We interpret this to be the result of an extensive vegetation cover and the lack of fluvial incision or erosion in the Meuse drainage basin. These relatively calm and stable landscape equilibrium conditions (Pastre et al., 1997) must have led to a constant or steady state supply of smectite plus vermiculite clay minerals to the Meuse river. From these data, we suggest that a minimum period of approximately 5000 years is needed to reach a steady state in Holocene smectite and vermiculite supply to the River Meuse.

Vermiculitic clays are derived mainly from erosion of topsoils from the Ardennes region (see above), and vermiculitic and smectitic clays from erosion of loess deposits and their topsoils. Ultimately, interstadial and interglacial weathering in soil environments thus have increased the supply of detrital soil-derived clay minerals to the fluvial system. Higher amounts of K- and Mg-depleted smectite, vermiculite and their interstratifications to illite lowered the contents of K_2O, MgO and TiO_2 and raised the Al_2O_3-contents of Holocene sediments. Consequently, Holocene sediment samples have lower K/Al, Mg/Al and Ti/Al-ratios and higher ratios of Rb, V, Cr and Ba to K_2O and MgO compared to Late-glacial samples. Therefore, Late-glacial and Holocene soil formation is of key importance in controlling the bulk geochemical composition of Late-glacial and Holocene Meuse clays and clay fractions.

4.7 CONCLUSIONS

The composition of unconsolidated, diagenetically unaltered fine-grained River Meuse sediments deposited in Late-glacial residual channels has systematically changed on a 1-10 ka time-scale. Therefore, the bulk and clay geochemistry of fine-grained sediments may provide useful clues to reconstruct long-term climate-controlled weathering processes in the upstream drainage area. Changing relative source area contributions and prolonged, intensive weathering of phyllosilicates in both interstadial and interglacial soil environments have most likely caused a shift in the detrital clay mineral supply towards higher contents of pedogenetic clay minerals. This led to increasing Al_2O_3-contents and decreasing K_2O-, MgO- and TiO_2-contents in Holocene Meuse sediments. Consequently, the ratios of a variety of trace elements that are typically related to clay minerals - like Ba, Cr, Rb and V - to these major constituents changed too. Thus, the composition of the 0-20 μm fraction cannot be considered constant over a period of 1000-10,000 years. Geochemists and soil scientists should bear this in mind during sampling of a sedimentary record covering one large temporal interval or when they compare samples from overbank sediments that differ more than 1000 years in age, even when these overbanks show no signs of recent soil formation.

Periglacial climate conditions and widespread landscape instability before 14-ka BP resulted in low residence times in tributary catchments, low chemical weathering rates, minor clay production and therefore low clay supply to the fluvial system. Late-glacial silty residual infillings therefore have low bulk clay contents. The Ardennes Palaeozoic rocks and/or still unweathered calcareous deposits in the loess belt of South Limburg, Belgium and Germany supplied detrital kaolinite, chlorite and illite and minor amounts of smectite, vermiculite or 1-1.4 nm mixed-layer clay minerals.

Late-glacial climatic amelioration (13-10 ka BP) facilitated the melting of permafrost and onset of postglacial soil formation. After decalcification, clay was translocated on a regional scale in soils on both Palaeozoic rocks and in the loess belt. Prolonged and more intense chemical weathering caused higher bulk clay contents and clay mineralogical differentiation. Smectite, vermiculite and 1-1.4 nm mixed-layer clay minerals are the ultimate weathering products during pedogenesis in a humid temperate climate. Separated clay fractions from sediments deposited between 12 and 11-ka BP and between 10 and 5-ka BP show increasing contents of pedogenetic Al-rich smectite and vermiculite. A steady-state supply during the Holocene was reached within a period of approximately 5000 years. Late-glacial topsoils from the Ardennes Palaeozoic rocks and loess deposits are the main sources for vermiculites, while smectites are predominantly derived from eroded topsoils of loess deposits. Aeolian influxes within fluvial drainage basins thus can influence the bulk mineralogy considerably.

Smectites, vermiculites and 1-1.4 nm mixed-layer clay minerals dominate the bulk and clay fractions of Holocene samples. These clay minerals are impoverished in K_2O and MgO and have higher Al_2O_3-contents relative to illite and chlorite. Consequently, the K/Al- and Mg/Al-ratios in Holocene sediments are lower relative to Late-glacial sediments. The K/Al- and Mg/Al-ratios of fine-grained fluvial sediments thus provide information on the abundance of purely detrital clay minerals and altered clay minerals derived from weathering processes in soil environments, provided that attention is paid to reworking and post-depositional processes. The post-depositional diagenesis of siderite and vivianite in anoxic gyttja environments affects the contents of Fe_2O_3, MnO and P_2O_5 and is site-specific. In the same manner, deposition and dissolution of lime strongly influences the CaO-contents. These constituents are therefore unsuitable to fingerprint weathering conditions in the upstream drainage basin, as they suffer from local diagenetic overprinting processes.

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PART II

LONG-TERM FLUVIAL DYNAMICS
AND SEDIMENT COMPOSITION

A 250-0 ka BP FORWARD-MODELLING CASE STUDY
OF THE RIVER MEUSE FLUVIAL SYSTEM

CHAPTER 5

MODELLING FLUVIAL LONGITUDINAL PROFILE DEVELOPMENT IN RESPONSE TO LATE QUATERNARY TECTONIC UPLIFT, SEA-LEVEL AND CLIMATE CHANGES

Based on: Tebbens, L.A., Veldkamp, A., Van Dijke, J.J. and Schoorl, J.M. 1999.

Modelling fluvial longitudinal profile development in response to Late Quaternary tectonic uplift, sea-level and climate changes: the River Meuse

Submitted to Global and Planetary Change

5 MODELLING FLUVIAL LONGITUDINAL PROFILE DEVELOPMENT IN RESPONSE TO LATE QUATERNARY TECTONIC UPLIFT, SEA-LEVEL AND CLIMATE CHANGES

ABSTRACT

The erosional and depositional dynamics of fluvial systems on Quaternary time-spans have gained the interest of many fluvial geomorphologists and sequence-stratigraphers. In quantifying the relative importance of long-term external and internal controls, they are confronted with rivers responding to the combined effects of tectonics, climate and sea-level change. In this paper, we intend to elucidate the fluvial response to long-term external forcing on a basin-wide scale. We present a semi 3-D forward-modelling case study of longitudinal profile development of a basin-marginal fluvial system (River Meuse) responding to Late Quaternary tectonic uplift, sea-level and climate changes.

Modelling results predict climatic control to be expressed most clearly in the uplifting Meuse upper and middle reaches upstream of the Roer Valley Graben. The alternation of stadial depositional and interstadial erosional phases causes net aggradation of the river valley during glacials, while the river valley is degraded during interglacials. Sea level strongly influences the subsiding lower reaches in the southern part of the North Sea Basin. Continuous deposition during the glacial periods fills in the subsiding graben areas and the southern North Sea Basin, suggesting that interstadial erosion is subdued. However, fading sediment supply at the start of interglacials combined with increasing discharges leads to discharge-controlled (kinetic) incision in these net depositional areas. The subsequent sea-level highstands generate gradient-backfilling events that migrate upstream with time, causing the terrace intersection to migrate upstream too. The Meuse case study indicates a possible shift over some 100-140 km. Increasing sediment supply shifts the coast line position seawards (delta progradation). It also increases the difference between the coast line and terrace intersection positions, which is a measure of the lateral extent of the depositional wedge or coastal prism. Increasing sediment travel distances (transport capacity) at constant sediment supply shifts the positions of both the coast line and terrace intersection upstream, but does not affect the difference between them.

5.1 INTRODUCTION

Basin-marginal fluvial systems in slowly uplifting tectonic settings respond dynamically to the combined effects of upstream control (uplift and climate change) and downstream control (glacio-eustatically driven relative sea-level changes) on Quaternary time-spans (Schumm, 1977; Van den Berg, 1996; Törnqvist, 1998). Fluvial systems archive these responses in their morphological patterns and sedimentary records. Classic research approaches involve the reconstruction of fluvial morphodynamics and erosional/depositional phases from sedimentary records. However, the interpretations from these studies are likely to be incomplete or unjustified, if they are based on the preserved deposits for a certain river reach or short time-spans only. The internal complex response of the fluvial system dictates that the processes of erosion and deposition in the upper, middle and lower reaches are strongly interrelated (Schumm, 1977). This clearly complicates the extrapolation of fluvial response from single river reaches to the scale of the whole fluvial system. Moreover, sedimentary records often show hiatuses rather than preserving a deposit correlative to every single climatic event or time-slice. Finally, both external and internal controls are known to interact on different time-scales, generating non-linear fluvial behaviour (Vandenberghe, 1995). Consequently, fluvial response varies both in temporal and 3-D spatial domains and has to be seen within the total fluvio-systematic context of external forcing and internal self-(re)organisation. For this reason, a sedimentary record taken out of this fluvio-systematic context will not suffice to reconstruct fluvial behaviour on larger spatial or temporal scales. Therefore, it is essential to understand the system dynamics before interpreting individual sedimentary records or quantifying the relative effects of the individual controlling variables.

To understand and quantify the long-term non-linear and complex-response behaviour of Late Quaternary fluvial systems, one has to account for processes affecting the whole drainage basin on a time-scale of 10^3-10^6 years (Veldkamp and Van Dijke, 1998). These processes involve external forcing of discharge and sediment load resulting from tectonic uplift, climate and sea-level change (cf. Merritts et al. 1994; Maddy, 1997). Long-term tectonic uplift generates relief and in this way provides the potential energy for a river to enable lowering of its longitudinal profile (Bull, 1991; Maddy, 1997). Climatic forcing of fluvial behaviour is very important in the slowly uplifting northwest European tectonic setting (Kiden and Törnqvist, 1998). Climate directly controls river discharge and sediment supply, which on their turn determine the delicate balance between stream power and resisting power in the fluvial system. These opposing forces determine whether the fluvial system will aggrade, degrade or maintain equilibrium.

Long-term changes in the relative contributions of external controls forces a river to readjust its longitudinal profile to the new environmental conditions (Bull, 1991). In the fluvial system, phases of net river valley aggradation alternate with phases of net river valley degradation in time and in space. Tectonic uplift and base level lowering do not necessarily initiate river valley degradation over the whole length of the longitudinal profile. Instead, a surplus of sediment supply under favourable cold climatic conditions might delay or even outpace long-term degradation (Bull, 1991). Furthermore, the fluvial system might respond to external disturbances without degradation or aggradation at all, solely by internally readjusting gradient, sinuosity and channel depth or width to a new morphological pattern within individual reaches, e.g. from braiding to meandering (Ouchi, 1985; Schumm, 1993). Finally, upstream degradation in incised channels might provoke downstream aggradation, so that different river reaches will respond out of phase. Thus, different river reaches respond in a complex manner to differential uplift, climate and sea-level changes, depending on their position along the longitudinal profile and the rates of change (cf. Merritts et al., 1994; Veldkamp and Van Dijke, 1999). This phenomenon is often referred to as complex-response dynamics (Schumm, 1977). Apart from spatial non-linearity, the fluvial response to external forcing is temporally non-linear as well (Bull, 1991; Vandenberghe, 1995).

This paper intends to explore the effects of the interplay between tectonic uplift, climate and sea-level change for river systems on a time-scale of 10^3-10^6 years. We chose the Meuse fluvial system to perform a semi 3-D (4-D including time) forward-modelling case study of longitudinal profile development for the Late Quaternary, i.e. 250-0 ka BP. The Quaternary climo-tectonic framework and the morphodynamics of the River Meuse are well-known (cf. Kasse et al., 1995; Van den Berg, 1996). Since the capture of the Vosges headwaters ~250-ka BP (Fig. 1.1-A), the Meuse drainage basin has been non-glaciated and is entirely rain-fed, creating excellent conditions to model discharges as a function of long-term climate change. We followed a robust downscaling approach to avoid problems in the upscaling of short-term hydrological processes on the channel level and chicken-egg discussions on e.g. fluvial morphodynamics in relation to river gradient. These problems are inherent to many large-scale modelling exercises (e.g. Howard et al., 1994). They mostly arise from uncertainties and missing field data concerning palaeodischarge and sediment-specific properties like erodibility and grain size.

Of course, one could reason that a robust downscaling approach grossly simplifies "real" fluvial behaviour. Given the many inferences and assumptions underlying the transfer and extrapolation of field measurements or observations to different temporal and spatial scales, one could question if such verity should be aimed at in the first place or can be reached at all. In this respect, we would like to cite the view of Oreskes et al. (1994), who state that the primary value of numerical models is mainly heuristic: "They are representations at best, useful for guiding further research and to challenge existing formulations, rather than pretending to be an exact, verified copy of the real world". Nevertheless, numerical modelling of highly simplified stream systems is able to demonstrate various types of non-linear behaviour (Snow and Slingerland, 1990).

5.2 MODELLING APPROACH AND INPUT DATA

We simulated Late Quaternary dynamics of the River Meuse using an elaborated version of the FLUVER2 model. The FLUVER2 model describes fluvial dynamics on a basin-wide scale with equations that relate the sediment transport capacity of the fluvial system to discharge, topography and slope processes (Veldkamp and Van Dijke, 1998). The model is semi three-dimensional and allows for the different sizes of tributary subcatchments. We assigned a tributary catchment-dependent width to every 1-km segment to increase the main river discharge proportionally over the stretch between the mouths of two tributaries. However, channel-specific parameters like width, depth and morphological pattern are not included and neither are lateral channel belt shifts.

The stream power of the fluvial system (valley gradient · upstream drainage area discharge) is linked to sediment production from the same drainage area to calculate a sediment flux through the fluvial system. On a time-scale of 10^3 to 10^5-years, the variations in the external forcing parameters (tectonic uplift rate, sea level and climate) perturb the balance between the sediment supply to the fluvial system and its sediment transport capacity. Assuming mass continuity and dynamic equilibrium, the model quantifies the fluvial response to these perturbations by calculating the sediment flux resulting from the difference in sediment detachment rate and settlement rate from node to node for 1-km segments along the longitudinal river profile. A positive sediment flux at a given position on the profile raises the height of the inter-nodal profile segment and thus mimics a sedimentation event. A negative sediment flux lowers the height of the inter-nodal profile segment and represents the erosion process.

We chose to simulate River Meuse fluvial dynamics for the time-span 250-0 ka BP for three reasons. Firstly, the Meuse has intermittently been a tributary of the Rhine until the Mid-Cromer period (Bustamante, 1976; Van Dijke and Van den Berg, 1999). From then on, the river can be considered to flow autonomously. Secondly, a tributary of the Moselle captured the Upper Meuse headwaters draining the crystalline Vosges area (Fig. 1.1-A) around 250-ka BP (Bustamante, 1976). The loss of this catchment to the Rhine drainage system decreased the Meuse discharge to an unknown, but probably considerable extent. From the moment of capture, the influence of the Vosges glaciers (e.g. Salomé, 1968) on the Meuse discharge regime can be ignored. The total Meuse drainage area has not changed considerably since (Hantke, 1993). Thirdly, the 250-ka time-span covers two glacial-interglacial cycles for which climate and sea-level data are relatively well known. High-quality input data are essential to obtain as realistic model outputs as possible.

The model input data required to simulate fluvial dynamics on a time-scale of 10^3 to 10^5-years involve information on the initial longitudinal profile of the Meuse, tectonic uplift or subsidence rates, sea-level changes and climatic data.

Initial longitudinal profile

Information on the 250-ka BP longitudinal profile of the River Meuse is not available. Given the position of the Saalian terraces and sediment bodies, we took the present-day longitudinal profile (Fig. 5.1) as a best approximation of initial conditions. This profile can be subdivided into three different stretches coinciding with the major litho-stratigraphical units in the drainage basin (Berger and Mugie, 1994). The upper reach or Meuse Lorraine (0-339 km) drains Jurassic limestones of the north-eastern fringe of the Paris Basin, the Chiers being the only major tributary. The middle reach or Ardennes Meuse (339-597 km) drains impermeable Palaeozoic metapelites and quartzites in the Ardennes low mountain range. Several major tributaries (Semois, Lesse, Sambre and Ourthe) increase the main river discharge considerably in this area (Figs. 5.2 and 6.6). The Meuse abandons the Ardennes near Liège (597-km) to enter the depositional Quaternary delta landscape of the Rhine and Dutch Meuse (Maas) via loess-covered Late Cretaceous limestones. Minor tributaries (Geul, Roer and Niers) drain the Late Pleistocene loess and coversand deposits in the lower part of the basin (Fig. 5.2 and 6.6).

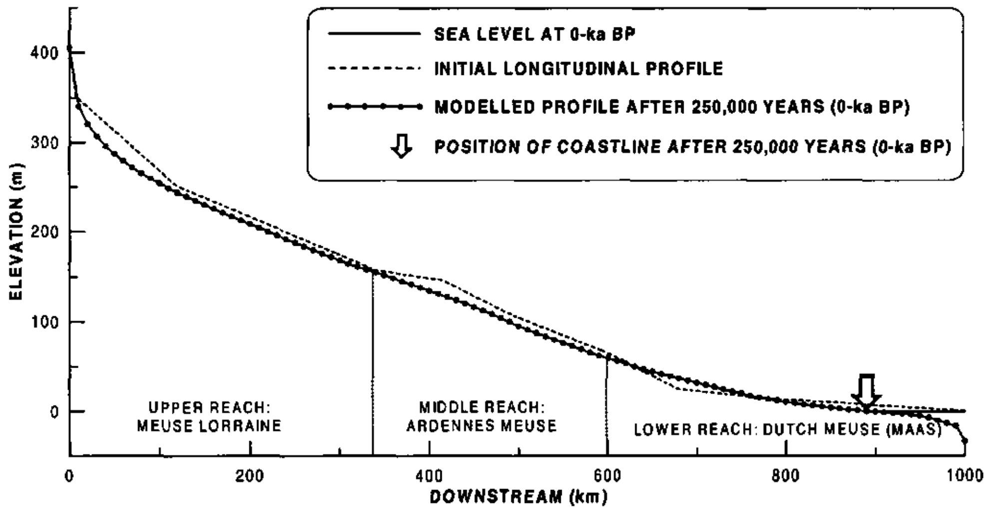

Figure 5.1 The initial present-day River Meuse longitudinal profile and the position of the modelled longitudinal profile and coastline after a model run of 250,000 years.

Tectonic setting

The Meuse crosses various structural tectonic units, each with their own uplift or subsidence rates, which constrain the modelled river stretches. During the past 250-ka, the northeastern Paris Basin and Ardennes region have been broadly uplifting (Demoulin, 1989, 1998). Rift-related lithospheric extension and crustal thinning caused the Rhenish Shield and especially the northern Ardennes to be uplifted more rapidly than the north-eastern Paris Basin, resulting in a dome-shaped area (Fuchs et al., 1983; Ziegler, 1994; Demoulin, 1998). Accordingly, the longitudinal profile in Figure 5.1 exhibits a convex part and knickpoints along the Ardennes reach (339-597 km). The Meuse kept pace to this epeirogenetic uplift by incision of the exposed Palaeozoic rocks and formed a macro-scale alluvial fan at the point where it leaves the Ardennes region (Berger and Mugie, 1994). Continuous Late Cenozoic uplift of this mainly glacially aggraded fan and interglacial dissection resulted in a flight of 31 terraces (Van den Berg, 1996). Veldkamp and Van den Berg (1993) found that these terraces arising from glacial aggradation and interglacial degradation in the Meuse fluvial system are preserved best at a tectonic uplift rate of $5{\cdot}10^{-5}$-$1{\cdot}10^{-4}$-m/yr.

Periods with relatively slow as well as rapid uplift reflect the waxing and waning of the regional tectonic stress field in the Meuse drainage basin. Van den Berg (1994) used the surface altitudes from uplifted terraces in the Maastricht region at 627-km to calculate an average Pleistocene uplift rate of ~$6{\cdot}10^{-5}$-m/yr. At present, however, the Maastricht region raises at a tenfold higher rate of $8{\cdot}10^{-4}$-m/yr (Van den Berg, 1996). Contrary, Juvigné and Renard (1992) estimated the incision rate from the bases of terrace remnants near Liège at less than ~$2{\cdot}10^{-5}$-m/yr for the Latest Pleistocene. They attribute this lower incision rate to the capture of the Upper Meuse by the Moselle engendering a decreasing discharge and stream power. It is unclear, however, which of their terraces is correlated to the age-referenced Caberg (270-ka BP) and Eisden-Lanklaar (175-125 ka BP) terraces. If we correlate the Jupille, Herstal and Vivegnis glacial aggradational terraces with the Caberg 3 (Saalian I), Eisden-Lanklaar (Saalian II) and Maasmechelen (Weichselian) levels (Paulissen, 1973; Van den Berg, 1996) respectively, then we reconstruct an average incision rate of ~$5.5{\cdot}10^{-5}$-m/yr at Liège.

Because of its dome-shaped appearance, the uplift rate for the Central Ardennes area should be larger than that of the flanking Maastricht region (Ziegler, 1994; Fuchs et al., 1983). Although we are aware that the Rhenish Shield/Ardennes region is not uplifted *en bloc* (Müller, 1983; Demoulin, 1989, 1998), we simplified the tectonic modelling by using a symmetric sine-shaped uplift curve for the Ardennes-Feldbiss stretch (370-661 km; Fig. 5.2). The wavelength and maximum amplitude of the sine curve have been calculated to reproduce the Maastricht uplift rate (*sensu* Van den Berg, 1996) at the appropriate distance in the longitudinal profile, i.e. $6{\cdot}10^{-5}$-m/yr at 627-km. We assumed broadly equal uplift rates on both flanks of the Ardennes and thus the uplift for the north-eastern Paris Basin was modelled with a constant Maastricht value of $6{\cdot}10^{-5}$-m/yr over the entire 0-370 km stretch (Fig. 5.2). The calculated sine-function constrained the maximum uplift rate in the central Ardennes part (500-km) at ~$2{\cdot}10^{-4}$-m/yr, equivalent to ~50-m uplift in 250-ka.

The area of net uplift extends to 661-km downstream (Fig. 5.2). At this point, the South Limburg Feldbiss fault forms the break to the Roer Valley Graben in the southern Netherlands. The Roer Valley Graben consists of the subsiding Central Graben, the uplifting Peel-Venlo Block and the subsiding Venlo Graben. The Venlo Graben is situated at the upstream fringe of the hinge line of the continuously subsiding North Sea basin (Van den Berg et al., 1994). Several faults border these tectonic units. To simplify the modelling in the Roer Valley Graben (661-773 km), we specified the faults as uplift/downwarp inflection points with zero vertical displacement. A sine-shaped curve was used to simulate an increasing subsidence from zero at the Feldbiss fault to $1.36{\cdot}10^{-4}$-m/yr in the centre of the Central Graben (661-706 km). Subsidence decreases to zero at the Peelrand Fault, to change into increasing uplift of $1.36{\cdot}10^{-4}$-m/yr for the centre of Peel-Venlo Block (706-751 km). From there on, subsidence decreases to zero again at the Velden fault and to $1.36{\cdot}10^{-4}$-m/yr at 773-km (Venlo Graben; 751-773 km).

Stage-5e (Eemian) buried marine and brackish-lacustrine sediments, corrected for compaction, indicate that the reconstructed peak sea levels between the "presumably tectonically stable" Channel area and the subsiding southern North Sea Basin differ some 17-m (Zagwijn, 1983). If an absolute (U/Th-series) age of 124-ka BP (Chappell and Shackleton, 1986) is assigned to stage-5e, then a mean subsidence rate of $1.36{\cdot}10^{-4}$-m/yr can be calculated for the Dutch coast relative to the Channel area. Accordingly, a constant subsidence rate of $1.36{\cdot}10^{-4}$-m/yr has been assumed for the North Sea basin (Fig. 5.2).

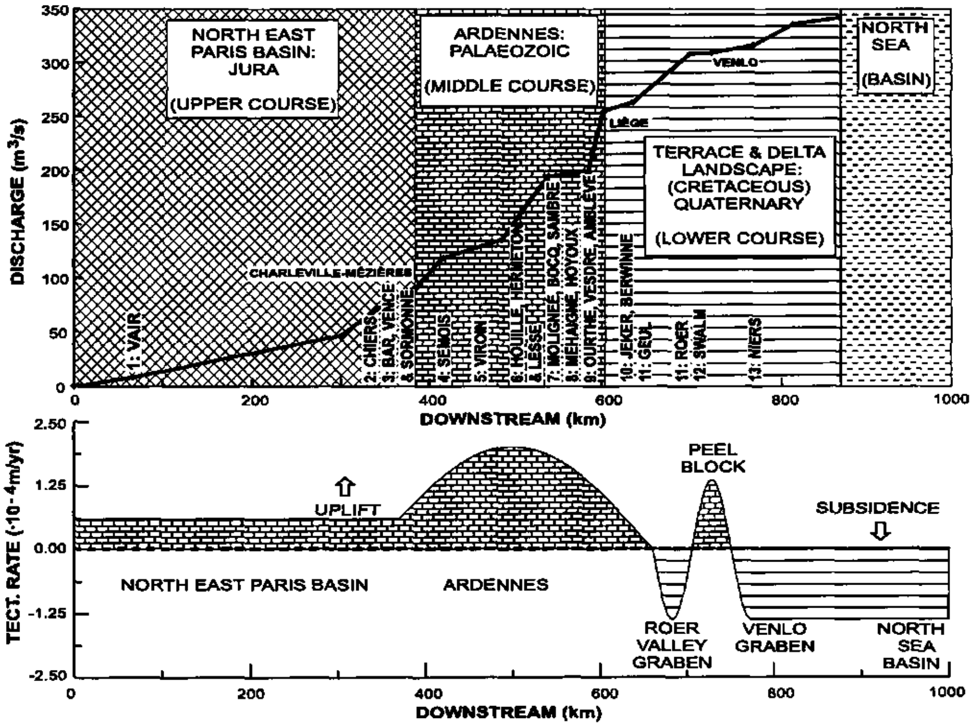

Figure 5.2 Discharge and tributaries of the present-day River Meuse (top). Schematic overview of model input tectonic rates versus downstream distance to source (bottom).

Sea-level changes

Variations in global ice volume on a time-scale of glacial-interglacial periods and millennia caused Late Pleistocene sea level to fluctuate. Extensive U-Th series dating of tectonically uplifted coral reef terraces has allowed a reconstruction of the eustatic sea-level record for 130-0 ka BP. Most authors agree the Eemian Interglacial caused a highstand of +6-m above the present sea level at around 124-ka BP (Broecker et al., 1968; Ku et al., 1974; Chappell and Shackleton, 1986). Fairbanks (1989) found that sea level sunk to 121-m below current datum during the coldest part of the Late Weichselian at 18-ka BP. The following rapid glacio-eustatic sea-level rise (18-0 ka BP) slowed down during the Younger Dryas climatic cooling event. In between the 124-ka highstand and 18-ka lowstand extremes, the coral terraces of the Huon Peninsula (Papua New Guinea) provide the most complete sea-level record (Chappell and Shackleton, 1986; Richards et al., 1994; Linsley, 1996). Chappell et al. (1996) recently improved the performance of this sea-level curve to obtain a better agreement with isotopic sea levels for the period of 70 to 30-ka BP.

We based our sea-level input (Fig. 5.3-A) for 124-18 ka BP on the adapted Huon data of Chappell and Shackleton (1986) and Chappell et al. (1996), and derived the more detailed record for 18-0 ka BP from Fairbanks (1989). We linearly interpolated the sea levels for the time-intervals in between the coral reef terraces. Sea-level highstand data for the 250-124 ka period are very scarce due to erosion of coral reef terraces and sea-level lowstand data for this period are very insecure (Chappell and Shackleton, 1986). However, both the sea-level record

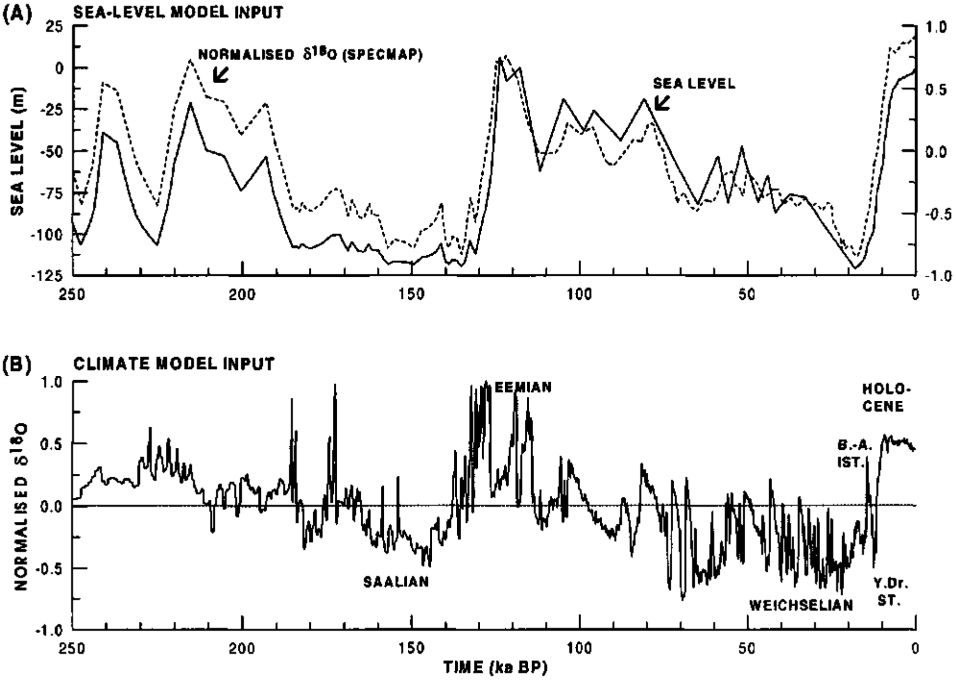

Figure 5.3 *(A)* Sea-level and SPECMAP-data for 250-0 ka BP (Martinson et al., 1987; Chappell and Shackleton, 1986 and Chappell et al., 1996). ***(B)*** Normalised δ^{18}O-data from the GRIP ice-core (Dansgaard et al., 1993), used to simulate discharge and hillslope supply for the period 250-0 ka BP. "B.-A. IST" = Bølling-Allerød Interstadial; "Y.Dr. St." = Younger Dryas Stadial.

and the normalised δ^{18}O-record respond in a similar way to changes in the global ice volume (e.g. Linsley, 1996). Therefore, we fitted the detailed sea-level record of Fairbanks (covering a range of 121-m sea-level rise) to the astronomically tuned and well-dated SPECMAP normalised δ^{18}O-isotope stratigraphy of Martinson et al. (1987). Next, we predicted the sea-levels for the period 250-124 ka BP from the curve fit formula (r^2 = 0.97) and the available normalised δ^{18}O-data from Martinson et al. (Fig. 5.3-A). This yielded four Early Saalian sea-level highstands at -39, -21, -53 and -100-m respectively, whereas Chappell and Shackleton (1986) found them at -14, -12, -9 and -58-m. Thus, one should keep in mind that our 250-124 ka curve derived from SPECMAP δ^{18}O-data yields lower sea-levels during the Saalian period.

The Huon sea-level record matches other sea-level records, which are based on raised coral reef terraces measured in a variety of tectonic settings. This supports global use of the Huon curve (Chappell and Shackleton, 1986). However, continuing isostatic readjustments of the land caused considerable regional sea-level variations since the latest deglaciation and might have controlled relative sea-level change in NW Europe too (Emery and Aubrey, 1985; Lambeck et al., 1990). E.g. former icecap centres in the Gulf of Bothnia and Scotland exhibit uplifts of >100-m and 6-m respectively during the last 9000 years. The southern North Sea basin is part of a collapsing peripheral bulge and consequently subsided at least 45-m during this same time, corresponding to a subsidence rate of ~$5 \cdot 10^{-3}$-m/yr. This value closely matches the present subsidence rate of $2\text{-}7 \cdot 10^{-3}$-m/yr obtained by recent precision levelling and tide-gauge stations (Emery and Aubrey, 1985). Within this subsiding area, Kiden (1995) mentions differential glacio-isostatic crustal uplift rates for Belgium relative to the subsiding Netherlands to slow down from $2 \cdot 10^{-3}$-m/yr for the period 9000-5500 cal. years BP to $2.5 \cdot 10^{-4}$ m/yr since 5500- cal. years BP.

We chose not to correct the eustatic sea levels for ice-cap loading or unloading in the model, since we only have data for the last 10-ka. Moreover, the isostatic uplift rates themselves have been derived from model calculations having their own specific uncertainties. Multiple problems in dating the raised beaches of Southern England and determining (glacio-isostatic) uplift rates lead Keen (1995) to conclude that altitudinal values for sea-level highstands can neither be accurately determined for any of the Late Quaternary interglacial stages. Finally, Emery and Aubrey (1985) state that it is still impossible to separate eustatic from isostatic sea-level changes. Thus, we had to assume that relative sea-level movement followed global eustatic sea-level change and applied a constant subsidence rate for the southern North Sea Basin of $1.36 \cdot 10^{-4}$-m/yr, based on the data of Zagwijn (1983).

Climate changes

High-resolution (200-yr averages) $\delta^{18}O$-isotope data from the GRIP ice core at Summit, Greenland have revealed marked climatic instability in the North Atlantic region during the Late Pleistocene (Fig. 5.3-B; Johnsen et al., 1992; Dansgaard et al., 1993). Because the Northern Hemisphere depressions track from W to E, the reconstructed climate changes from the GRIP core closely match climate changes in Northwest Europe. Bond et al. (1993), Thouveny et al. (1994) and Fronval et al. (1995) provided evidence that Late Pleistocene climatic instability is indeed reflected in North Atlantic sediments and in maar lake deposits in France. However, the Eemian data (~128-115 ka BP) have to be treated with care, for marine records suggest a more stable climate than the GRIP ice core data do (McManus et al., 1994). Johnsen et al. (1995) still consider the rapid climatic fluctuations in the Eemian part of the GRIP core to reflect interglacial climatic instability. Peel (1995) and Chappellaz et al. (1997) mention final conclusive evidence that the lower part of the GRIP core has suffered stratigraphic disturbance to produce cold spells in the Eemian ice that are actually ice-flow artefacts.

Although we alternatively could have chosen the SPECMAP deep-sea chronostratigraphy of Martinson et al. (1987), we preferred to benefit from the independent higher-resolution GRIP ice-core data (200-yr sample interval) and the direct link between the depression tracks and Northwest-European climate. We normalised the GRIP-core isotope data (Dansgaard et al., 1993) and linked the youngest (Holocene) normalised $\delta^{18}O$-value to measured present-day discharges (Berger and Mugie, 1994). In this way, the River Meuse discharges were simulated to respond to Northwest-European climate change, with maximum discharges for the Eemian being 25% higher and minimal discharges for full-glacial conditions being 55% lower compared to present-day conditions. Hillslope-processes (see below) were simulated the same way, and set to be maximal during glacial or stadial stages in areas with high local relief (the Ardennes low mountain range) and minimal during interglacial or interstadial stages in areas with low relief (the Netherlands).

5.3 CALIBRATION

Three main variables in the model can be tuned to stabilise the computational procedure and/or to determine the performance of the simulations. They are the bedrock/alluvium erodibility (*k_sed*), the sediment travel distance (*d_tra*) and a hillslope-processes rate variable (*difco*). K_sed quantifies the ease with which the rocks or sediments in a longitudinal profile segment are eroded. When multiplied with the stream power, it directly determines the sediment detachment rate, which can be considered equivalent to the potential energy of the river to erode the bedrock or alluvium. D_tra or travel distance is the downstream distance over which a certain amount of sediment that is already in transport is dispersed along the longitudinal profile. Dividing the incoming sediment flux by this sediment travel distance yields a settlement rate for the sediment load in any inter-nodal segment. The hillslope process variable only becomes operative in places with local relief, i.e. there where the height of the watershed differs from that of the river valley. Local relief is only generated in the 0-661 km section, corresponding to the north-east Paris Basin and Ardennes area. Multiplication of the hillslope variable with the amount of local relief generates a hillslope process rate and simulates an additional sediment flux into the fluvial system in uplifting areas. The hillslope process rate on its turn is proportionally scaled to climate via the link with the GRIP ice core data: the colder it is, the higher the rate.

To tune the *k_sed* and *d_tra* variables, we first kept hillslope process dynamics constant and ran the model with different settings for *k_sed* and *d_tra*. Then, results including the timing of erosional and depositional events in the fluvial system, the total amount of net river degradation (profile lowering) in the terrace flight area (627-km) and the modelled position of the Holocene coastline were compared to real-world data. Figure 5.4 shows the resulting calibration stability fields for 250-ka model runs at several combinations of *k_sed* and *d_tra*. The zero-field (0) represents numerical model instability, occurring at values of *k_sed* $>36\cdot10^{-10}$ m^{-2}. Below this value, the model is numerically stable. The model should at least simulate an incisional phase in the Venlo Graben area at the start of the Holocene Interglacial, because in this area channel downcutting and valley degradation was the most prominent fluvial response during the Weichselian Late-glacial and Early Holocene (Kasse et al., 1995; Tebbens et al., 1999a). Results in the minus-field (–) represent model runs without this incisional phase and indicate net profile aggradation (too little erosion) in the Ardennes area. The runs in the plus-field (+) yielded too much fluvial erosion in both the Ardennes and terrace flight area. This leaves the outcome-field (=) to represent those combinations of *k_sed* and *d_tra* that yield the most realistic simulations of longitudinal profile development.

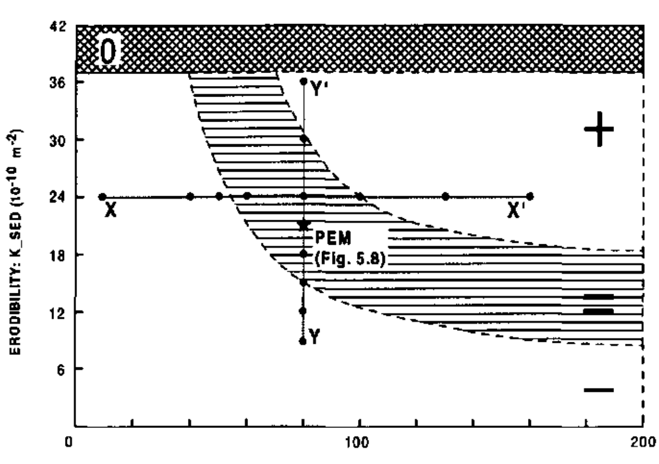

Figure 5.4 Stability plot (see text) for 250-0 ka BP simulations of the River Meuse longitudinal profile, performed at different erodibility and sediment travel distance settings. The X-X' and Y-Y' transects represent model simulations at different settings indicated in Figure 5.5. The star depicts the settings of the Profile Evolution Map in Figure 5.8.

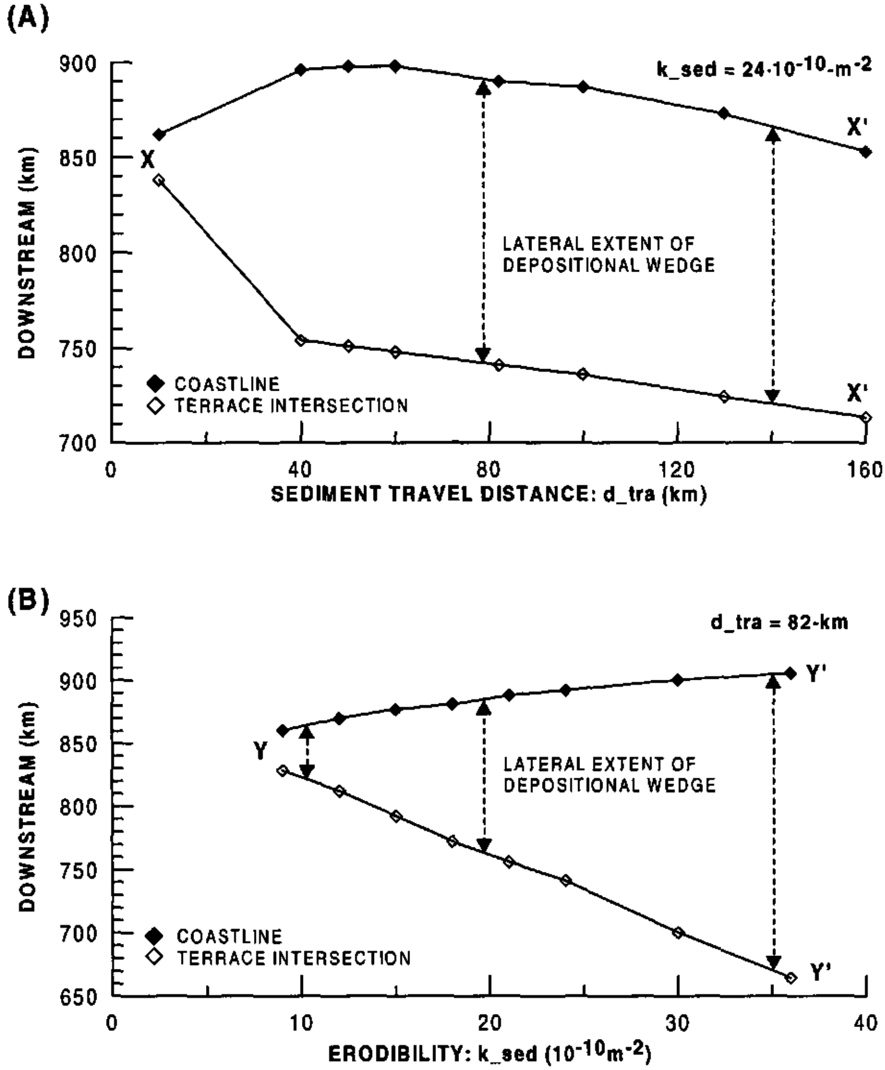

Figure 5.5 The positions of the modelled coastline and the terrace intersection after 250-ka model runs at different settings for *(A)*: the sediment travel distance (at constant *k_sed* of $24{\cdot}10^{-10}$ m^{-2}) and *(B)*: erodibility (at constant d_tra of 82-km). Transects X-X' and Y-Y' are indicated in Figure 5.4.

Transect X-X' (Figs. 5.4 and 5.5) illustrates the impact of different sediment travel distance values (*d_tra*) on the modelled coastline and the location of the terrace intersection point at a constant erodibility (*k_sed*) value of $24{\cdot}10^{-10}$ m^{-2}. E.g. at a sediment travel distance of 10-km, too little sediment is exported from the supplying areas, leading to overall profile aggradation and a coastline at 860-km downstream of the source. The modelled coastline extends maximally (~895-km) at a sediment travel distance of ~50-km. At higher values, so much sediment is exported to the North Sea, that the river profile is lowered in the subsiding area and sea level comes inland again. Interglacial sea-level highstands cause aggradational effects related to gradient backfilling reaching up to hundreds of kilometres inland (Schumm, 1993; Törnqvist, 1998). Figure 5.5 indicates the downstream distance where fluvial degradation changes into backfilling-related aggradation (the terrace intersection). The spatial difference between the downstream positions of the terrace intersection and the modelled coastline corresponds to the lateral extension of the so-called coastal prism or depositional wedge (see below). Figure 5.5 shows that the lateral extension of the depositional wedge for the Meuse amounts ~140-km, and is not dependent on sediment travel distance (at a constant erodibility factor of $24 \cdot 10^{-10}$ m^{-2} and for sediment travel distances >40-km).

Consider now a situation in which the sediment travel distance is kept constant at 82-km and the erodibility factor is increased (Transect Y-Y' in Figures 5.4 and 5.5). Figure 5.5 shows that in this situation the modelled coastline steadily extends in seaward direction. High erodibility values correspond to high sediment supply and lead to profile aggradation if fluvial transport capacity does not suffice to transport the extra amount of sediment. Apparently, the resulting aggradation is able to outpace long-term subsidence in the North Sea Basin area, which obviously must lead to delta progradation. Consequently, the river gradient is flattened progressively and hence the effects of an interglacial sea-level highstand on sedimentation (gradient backfilling) should be noticed further inland. Indeed, the lateral extension of the depositional wedge increases from only 32-km for the lowest *k_sed* value to 240-km for the highest *k_sed* value (Fig. 5.5).

Given the stability fields, we then tried to determine a most plausible value for the sediment travel distance. The model calculates a sediment settlement rate for each 1-km internodal segment by dispersing its incoming river sediment load over the sediment travel distance. Thus, high values for the sediment travel distance will induce low settlement rates to delay sediment fall out, equivalent to an increasing transport distance of the eroded sediment. We propose the natural process of large-scale sediment grading to enable tuning of the sediment travel distance as follows. The current river gradient of the Meuse (Fig. 5.1) is highest in the Ardennes (0.4-0.5 m/km) but decreases rapidly from ~0.5-m/km before the city of Maasbracht (679-km, Fig. 6.6) to ~0.11-m/km between Maasbracht and Ottersum (765-km). From then on, the gradient is only ~0.06-m/km to the river mouth in the North Sea at 874-km. The gradient decrease causes substantial grading of sediment deposition in the Dutch Meuse alluvial fan from the point where the Meuse leaves the Ardennes (597-km). Large-scale grainsize sorting causes gravel to dominate the fluvial deposits from 597-679 km, whereas sands grading into silts and clay are mainly found from 679-874 km. From this large-scale sorting process in the Meuse alluvial fan, we suggest a value of 679-597 = 82-km for the sediment travel distance.

The Holocene sea-level rise already influenced the present longitudinal profile and glacial river gradients have been quite different (Törnqvist, 1998). However, these effects are obliterated during the first 25-ka of simulation time, during which the longitudinal profile rapidly tends to a dynamic equilibrium with climatic and sea-level input data and reaches a concave-shaped profile. Thus, as a first approximation, we chose to base the sediment travel distance on the present-day longitudinal profile (for which the discharge characteristics are available) and kept this value constant during the whole simulation. Within the stability field of most realistic model runs (=-field, horizontally hatched), a set of plausible *k_sed*-values can now be obtained at *d_tra* = 82-km. With these combinations of settings, the Meuse roughly maintains its longitudinal profile at the same vertical position/elevation, indicating that fluvial erosion is able to keep pace with regional or epeirogenetic uplift. Thus, the settings correspond to a scenario of consequent fluvial degradation.

As a final test, all aforementioned factors were kept constant and hillslope rates were reduced with 50% to assess their influence on fluvial development. The resulting simulations (not shown) were compared with the model runs presented above. Notably, aggradation events in the Central Ardennes area were found less severe and because less sediment was supplied to the system, coastline and delta progradation were less too. In the stability plot, this produced a roughly parallel shift of boundary lines to the left, but the performance of the profile evolution maps essentially remained the same. Only the intensity and therefore the net amount of degradation and aggradation decreased.

5.4 RESULTS

This section presents the results of a model run with the *k_sed* and *d_tra* parameters set at $21 \cdot 10^{-10}$ m^{-2} and 82-km respectively (Fig. 5.4, ☆ in the =-field). Results for the first 25-ka are not shown, since they represent model initialisation artefacts. Figure 5.1 gives the initial longitudinal profile, the final concave-shaped longitudinal profile and the 0-ka BP (Holocene) sea level. Note that the Ardennes reach (339-597 km) has a convex part and two major knickpoints in the initial longitudinal profile. Sea level and final longitudinal profile cross each other at ~888-km, which roughly coincides with the present coastline at 874-km (Hollands Diep).

Figures 5.6-A to 5.6-I illustrate the evolution of the river profile topography versus time for three river stretches, namely the central Ardennes region (500-km), the terrace flight area near Maastricht (627-km) and the Venlo Graben (762-km). The time-relief plot of Figure 5.6-A shows that the ***Ardennes*** river valley has been uplifted ~5-m in 225-ka relative to the geoid. Apparently, fluvial valley erosion in this part is not able to compensate the total amount of uplift. However, if we want to know the net amount of degradation or aggradation in time with reference to a virtual non-eroded position outside the river valley, the time-cumulative amount of uplift has to be subtracted from the modelled actual heights. Figure 5.6-B demonstrates that after this reduction, the Ardennes river valley is degraded progressively (~40-m) with reference to a virtual uplifted position outside the river valley where no fluvial erosion would take place. This illustrates the long-term consequent incision into the Ardennes Palaeozoic rocks. Finally, the plot of net river valley profile changes (Fig. 5.6-C) shows that individual erosion and sedimentation events can be recognised, superimposed on the long-term uplift and degradation of the Ardennes river valley.

Although the ***Terrace flight area*** is a net uplifting area, the plot (Fig. 5.6-D) for the terrace flight area near Maastricht (627-km) shows ~3-m degradation of the river valley on the long term. Thus, fluvial erosion outpaces the amount of uplift in the terrace flight area, and is actively lowering the river valley elevation relative to the geoid. The process of terrace formation can be inferred from Figure 5.6-E. The river valley slightly maintains its altitude during a glacial, but is rapidly degraded (~5-m) during both the Eemian and Holocene Interglacials. The total amount of river valley degradation relative to a virtual non-eroded oldest terrace level is ~17-m. The observed difference in real world terraces will be less, since the 17-m are calculated with respect to the Holocene degradational valley surface and not to the aggradational surface of the next glacial. Plot 5.6-F shows that although the river valley is actively degrading its profile, several clustered Weichselian sedimentation events can still be distinguished.

The time-relief plot (Fig. 5.6-G) for the subsiding ***Venlo Graben area*** (762-km) shows that fluvial aggradation or infilling of the graben does not completely compensate long-term subsidence: the river valley and graben surface are lowered with respect to the geoid. Consequently, Figures 5.6-H and 5.6-I demonstrate that the Venlo Graben is a net depositional area.

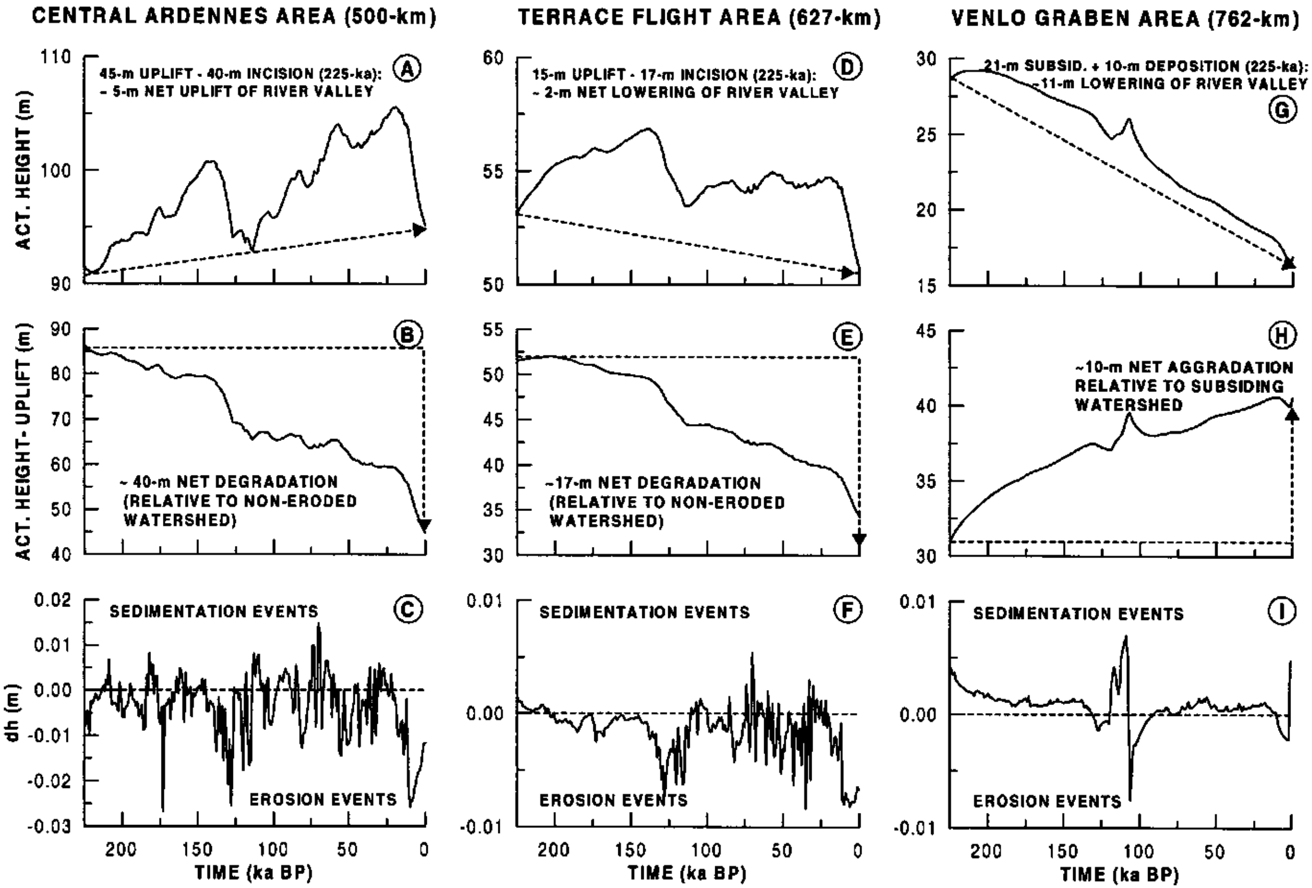

Figure 5.6 Modelled fluvial dynamics for three reaches along the River Meuse longitudinal profile. Plots A, B, D, E, G and H show the evolution of river valley elevation, while plots C, F and I show the net depositional and erosional events for the last 225-ka BP.

In spite of that, Figure 5.6-I shows that two erosional phases interrupt the depositional sequence at the start of the Eemian and Holocene Interglacials. A phase of increased sedimentation succeeds the Eemian erosional phase due to gradient backfilling (see below), which is related to the Eemian sea-level highstand of 124-ka BP. The model output predicts these deposits are eroded again during the start of the next glacial.

The allocation or prediction of these net aggradational and degradational phases in the fluvial system with time is most interesting in view of understanding complex-response fluvial behaviour. Where does the longitudinal profile aggrade or degrade and when? The model calculates the vertical profile adjustments per 1-km segment (dh, in m), owing to positive sediment fluxes (deposition) or negative sediment fluxes (erosion events) along the longitudinal profile. Figure 5.7 gives these results for one time-slice, namely the longitudinal profile during full-glacial conditions at 21.9-ka BP. It demonstrates fluvial erosion (negative dh-values) only in the upstream (0-100 km) stretch of the longitudinal profile, aggradation (positive dh-values) in the Ardennes and Roer Valley Graben (100-710 km), erosion on the Peel Block (710-750 km) and aggradation again for the 750-1000 km reach.

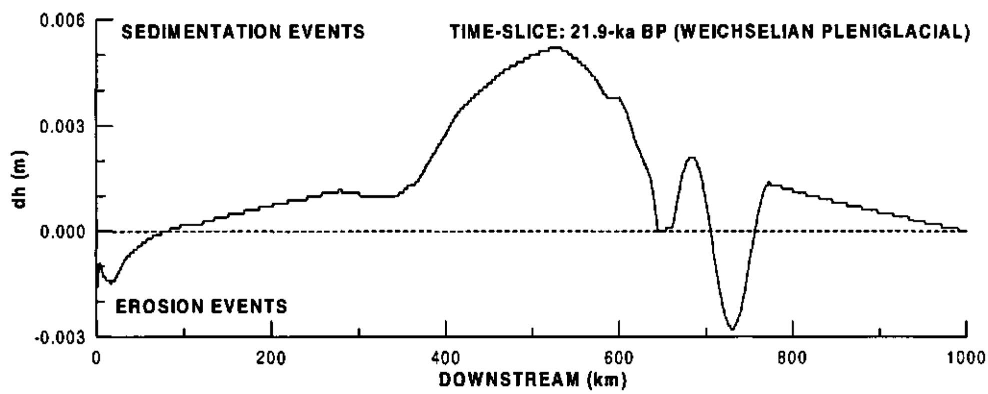

Figure 5.7 The occurrence of depositional and erosional events along the Meuse longitudinal profile for the time-slice 21.9-ka BP (Weichselian Pleniglacial).

When similar vertical profile changes of all individual time-slices are successively plotted with time, a Profile Evolution Map (PEM, Veldkamp and Van Dijke, 1998) can be drawn. This PEM shows where (x-axis) and when (y-axis) deposition or erosion occurs along the longitudinal profile. Figure 5.8 gives the PEM for the simulation depicted with a star in the stability field of Figure 5.4. The upstream and middle reaches of the longitudinal profile, representing the Meuse Lorraine (0-339 km) and the Ardennes Meuse (339-597 km) are considerably eroded during most of the simulated time (indicated with 'A' in Fig. 5.8). In this region, deposition events occur only during full-glacial conditions (Fig. 5.8: left-hand 'B'), but deposition is omnipresent during both glacial and interglacial conditions in the subsiding parts (Fig. 5.8: right-hand 'B') of the Roer Valley Graben (661-706, 751-773 km) and the subsiding coastal North Sea Basin area (>773-km). However, two incisional phases (Fig. 5.8: 'C') accompany the beginning of the Eemian and Holocene Interglacials in this same reach (see also Fig. 5.6-H and 5.6-I). They migrate downwards from the lower Ardennes area (~600-km) towards the subsiding basin. The uplifting model Peel Block (700-740 km) on the other hand is incised most of the time (Fig. 5.8: 'D'), which perfectly matches the field situa-

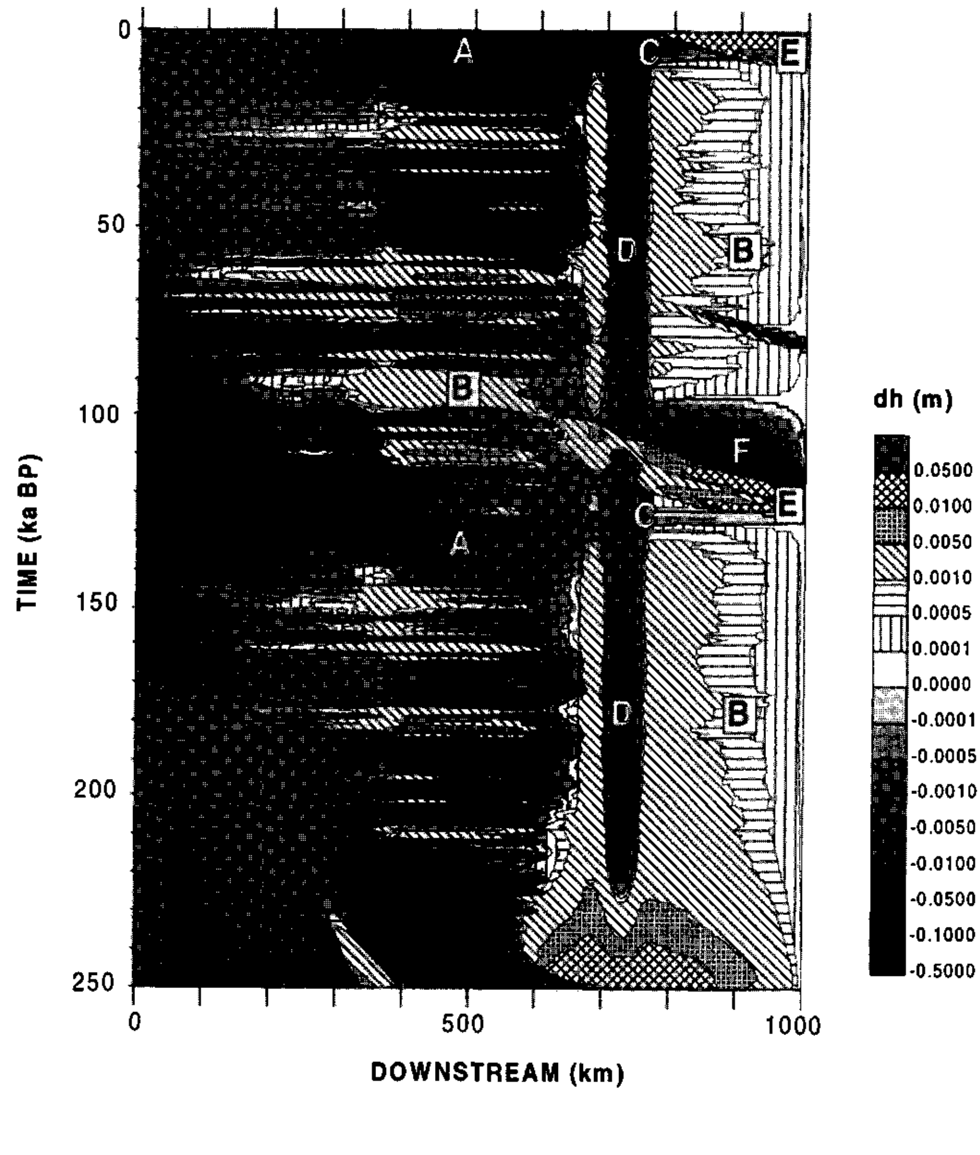

Figure 5.8 Profile Evolution Map showing the occurrence of depositional and erosional events along the River Meuse longitudinal profile for the period 250 to 0-ka BP. Prominent features are (see text): **A**: Area where erosion dominates (dh<0), **B**: Area where deposition dominates (dh>0), **C**: Post-glacial kinetic incision, **D**: Incision of the Peel Block, **E**: Interglacial depositional wedge and **F**: Post-interglacial potential incision of depositional wedge. North-east Paris Basin: 0-370 km, Ardennes: ~370-661 km, Roer Valley Central Graben: 661-706 km, Peel Block: 706-751 km, Venlo Graben: 751-773 km and North Sea Basin: 773-1000 km. The modelled 0-ka BP coastline is found at 888-km (see also Fig. 5.1).

ation of a presently small and confined river valley at the section where the Meuse crosses the real-world Peel Block (Paulissen, 1973).

The influence of sea-level highstands emerges from two depositional wedges (Fig. 5.8: 'E'), representing coastal prisms (Talling, 1998) or coastal onlap in the River Meuse lowermost reaches (800-1000 km) during the Eemian and Holocene Interglacials. The zone of increased sedimentation related to the Eemian sea-level highstand appears to migrate upstream during the period ranging from ~128 to 115-ka, to finally "connect" with glacial aggradational events (~100-ka) in the lower Ardennes region. Simultaneously, the subsequent sea-level fall of 115-100 ka BP provokes incision (Fig. 5.8: 'F') of the previously deposited coastal onlap section. This elegantly demonstrates non-linear fluvial response during the time-frame 115-100 ka BP: whereas the 800-1000 km river section is already being incised due to sea-level fall, the 600-800 km section is still responding to the 124-ka sea-level highstand by backfilling and aggradation of the river profile. The most recent Holocene depositional wedge extends to 756-km inland at 0-ka BP.

5.5 DISCUSSION

Within the constraints of the input data, the model provides the first well-calibrated view on the River Meuse longitudinal profile development of the last 250-ka. It shows complex response and non-linear behaviour for the Meuse system on a time-scale of 10^4-10^5 years due to the interplay of tectonic uplift, climate and sea-level change. The stability plot indicates that, given specific combinations of erodibility, sediment travel distance and hillslope rate settings, one can find a domain of most plausible outputs. Thus, one should not consider the results presented above as the absolute truth, but merely as a guide to comprehend fluvial behaviour on larger temporal and spatial scales in the following discussion.

Tectonic uplift causes the Meuse to degrade its longitudinal profile in the Lorraine and Ardennes reaches (0-597 km) during most of the time. However, the Ardennes longitudinal profile raises relative to the geoid, because net fluvial degradation is not capable to compensate all the uplift. This is consistent with the convex part in the present-day longitudinal profile. Should the present-day 105-m actual height of the 500-km river valley have risen with an equivalent mean 5-m uplift in 225-ka, then the river valley would have been at sea level (0-m) at around 4.7-Ma BP. This corresponds fairly well with Late Pliocene low-relief peneplain conditions. Fluvial degradation intensifies during the Eemian and Holocene Interglacials to migrate downstream at the start of these interglacials. Leeder and Stewart (1996) refer to this type of downstream-migrating incision as discharge-controlled incision or "kinetic incision", as it is clearly the change in hinterland conditions (discharge versus sediment supply) that causes the incision. Similarly, Merritts et al. (1994) label the climate-influenced supply of sediment and water as "upstream control", because the upstream river reaches respond most directly to relative changes in these variables. They conclude that the upper and middle reaches are vertically incised at about the same rate of uplift, as long as these river reaches remain located upstream of the depositional wedge related to a base level rise (see below). Evidently, the balance between river discharge and sediment load in the upper reaches of the Meuse Lorraine is such, that the modelled transport capacity or stream power outweighs sediment supply under both glacial and interglacial climatic conditions.
In contrast, periglacial hillslope processes acting during full-glacial periods induce excess sediment supply over river transport capacity, favouring well-expressed aggradation events in the middle reaches (500-661 km) of the Ardennes region and in the Maastricht terrace flight section (Veldkamp and Van den Berg, 1993; Veldkamp and Van Dijke, 1999). Glacial aggradation in north-eastern France is predicted only during very severe glacial conditions

(e.g. ~69 and 20-ka). Lefèvre et al. (1993) corroborate the presence of Weichselian Pleniglacial gravels in the upstream northeast Paris Basin. Clustered sedimentation events are predicted during the harshest stadials (Fig. 5.6-C, F), but slight degradation occurs during relatively warm glacial interstadials (e.g. Figs. 5.6-B, C and 5.6-E, F). The alternation of clustered sedimentation events with erosive degradational surfaces might find its real-world counterpart in the internal architecture of glacial terraces. These often show stacking of multiple (3-4) gravel units or take the form of several sandy fining upward sequences, separated from each other by bounding surfaces. Van den Berg (1996) referred to these individual units as parasequences and correlated them to fluvial activity on a millennial time-scale. The gravelly unit in Figures 2.6 and 2.7 (Chapter 2) supports this view, as it was deposited between 13280 and 12110 ^{14}C-years BP.

The semi 3-D model simulates vertical profile changes, but does not consider lateral channel erosion or channel belt shifts. Therefore, in a vertical sense, thickest aggradational deposits arising from the most severe and prolonged stadials are the ones most likely to be preserved, while others are eroded. Since the model does not allow for lateral movements and the actual Meuse is sliding off to the west due to long-term tectonic tilting, one can not compare every single aggradational unit directly with a preserved terrace in the Maastricht terrace flight. However, like Veldkamp and Van den Berg (1993), we find that interglacial degradation is far more extensive than interstadial degradation (Fig. 5.6-E). Continuous valley degradation in the Maastricht region is consistent with parallel terrace surfaces (Juvigné and Renard, 1992; Van den Berg, 1996) and will ultimately transform the cold-stage aggradational units into the preserved alluvial fill-type terraces as found by Paulissen (1973) and Van den Berg (1996). Other north-west European rivers like the Thames (Gibbard, 1997; Maddy, 1997), the Somme (Antoine, 1994), the Loire (Veldkamp, 1991; Veldkamp and Van Dijke, 1998) and the Rhine show equivalent behaviour.

Despite similar climatic inputs as for the upper reaches and the Ardennes, the model does not predict a stadial/interstadial alternation of aggradation and degradation for the Meuse lower reaches. Instead, aggradation is predicted to occur over the whole length of the glacial periods in the net depositional subsiding Roer Valley Graben, Venlo Graben and North Sea basin areas (Figs. 5.2; 5.6-H, I and 5.8). These aggradational units are only incised (kinetic incision) at the start of the subsequent Eemian and Holocene Interglacials. The Late Weichselian to Early Holocene degradation is consistent with two periods of channel downcutting (12.5-11.9 ka BP and 11.3-9.8 ka BP) and the subsequent formation of factual Late-glacial cut terraces in the Venlo Graben area. Late-glacial (Bølling/Allerød) and Early Holocene downcutting in the Weichselian aggradational surface began at ~13-ka BP and is reported in many other north-west European rivers as well (Starkel et al., 1991). Several adjustments of the channel and floodplain morphodynamics accompanied this phase of cut-terrace formation (Kasse et al., 1995; Chapter 2: Tebbens et al., 1999a).

An interglacial sea-level highstand often initiates gradient backfilling (Schumm, 1993) to produce a zone of increased sedimentation that migrates upstream with time and which builds up a depositional wedge (Merritts et al., 1994) or coastal prism (Talling, 1998). Likewise, relative sea-level rise in the North Sea Basin must have forced the Rhine-Meuse fluvial systems to become graded to a higher base level (Törnqvist, 1998). The model predicts the Holocene sea-level highstand to influence fluvial dynamics in the downstream Venlo Graben and North Sea Basin area (750-870 km) as from 8-5 ka BP. This prediction is corroborated by Törnqvist (1993) and Weerts and Berendsen (1995), who reported high avulsion frequencies for this same period and related these to the influence of rapid sea-level rise for the downstream part of the Rhine-Meuse delta. Törnqvist (1998) suggested that glacio-eustatic control on fluvial longitudinal profile development of the Rhine-Meuse system might extend as much as hundreds of kilometres inland. Our model results clearly demonstrate that sea-level rise causes backfilling of the longitudinal profile in the most downstream river reaches. The modelled 0-ka BP Holocene coastline was found at 888-km, while the fringe of the depositional wedge reaches much further inland, namely to 756-km. Thus, given the external forcing input and aforementioned settings, our model suggests that a Holocene backfilling of the longitudinal profile is possible to a distance of 132-km upstream from the modelled coastline. Interestingly, the position of the modelled 0-ka BP terrace intersection practically coincides with the position of the actual present-day terrace intersection at ~765-km (cf. Törnqvist, 1995), which has not been included in the model calibration. Holocene clays are predicted to overlie the Late Weichselian aggradational deposits of the Meuse downstream of this point. Indeed, detailed reconstruction of the Late-glacial gradient lines (Fig. 2.4, p. 20) showed the upstream Younger Dryas erosional cut-terrace to disappear under an aggradational Holocene clay cover at ~765-km (corresponding to the village of Gennep; Tebbens et al., 1999a). The downstream low-gradient part of the Rhine-Meuse delta demonstrates ~15-m Holocene fine-grained sediments deposited on a Weichselian higher-gradient and gravelly subsurface (Törnqvist, 1995).

Because Holocene sea level is still rising, the Holocene depositional wedge may not have reached its maximum extension yet. This can be inferred from the 6-m higher Eemian sea-level highstand that influenced fluvial dynamics even further upstream than the Holocene sea level. The model predicts that a similar gradient-backfilling event in the 800-1000 km river section occurred around ~124-ka BP. This event migrated upstream with time and has induced sedimentation at ~110-ka BP even on the uplifting Peel Block, which normally is constantly being incised (Fig. 5.8: 'D'). The Peel Block river section thus shows a 14-ka delayed response to the preceding high sea level of 124-ka BP. Unfortunately, River Meuse fine-grained deposits of the Eemian depositional wedge have hardly been found, but they very likely once existed (Törnqvist, 1995). This leaves us no possibility to confirm a zone of increased sedimentation migrating to over the Peel Block during the Eemian. Depositional wedges are subsequently incised as sea level falls again (cf. Talling, 1998), to demonstrate the type of base-level controlled or potential incision *sensu* Leeder and Stewart (1996). Indeed, the model predicts a prolonged and considerable post-interglacial erosional phase (Fig. 5.8: 'F') in the Eemian depositional wedge at the start of the Weichselian Glacial. This erosional phase might very well explain the absence of the Eemian deposits.

5.6 MODEL SENSITIVITY

Since FLUVER2 is a process-based model, we expect that most large fluvial systems will respond in a similar way to Late Quaternary climate and sea-level changes. As such, it can be used to test the response of other river systems in comparable tectonic and climatic settings as well. Of course, any change in the timing of the model-input climatic or sea-level data will change the timing of fluvial response too. Changes in tectonic uplift or subsidence rates and hillslope parameters will influence the spatial allocation and the intensity of depositional and erosional events.

Figures 5.4 and 5.5 already showed that higher values of erodibility (*k_sed*) and sediment travel distance (*d_tra*) set the model in a "higher gear": more sediment is eroded (higher supply) and/or sediment in transport is carried further away (higher transport capacity). Some uncertainty in our approach arises from assuming a constant erodibility (*k_sed*-value) along the whole length of the longitudinal profile. However, the diversity in more or less resistant rocks in the Upper Meuse Lorraine and Ardennes sections and the downstream transition to the Dutch alluvium-dominated river stretch would actually call for varying *k_sed* values along the profile. Similarly, Ashley and Hamilton (1993) and Van den Berg and Veldkamp (1996:

Ch. 5) argued that the aeolian influx of extra-basin cover sands and loess provides indirect climatic control, a factor that is not addressed in our model approach. At present, we do not know the input quantities of cover sand and loess, yet the additional sediment input is likely to have had an impact on Meuse fluvial dynamics, as several tributaries have their origin in the loess or coversand-overblown areas.

The semi 3-D model does not allow for lateral channel (belt) shifts or changes in the morphological pattern of the fluvial system. Therefore, the model is able to predict where and when erosion and sedimentation events will occur along the longitudinal profile, but it is not possible to infer firm conclusions on the preservation potential of aggradational units. Although it is true that additional parameters or dimensions would probably yield more quantitative results, we seriously doubt if scientific insight would be enhanced in a qualitative sense. For the moment, we considered it premature to introduce more parameters creating more degrees of freedom, without having the possibility to test their plausibility or validity in this calibrational approach.

5.7 CONCLUSIONS

The well-studied non-glaciated drainage basin of the rain-fed River Meuse provides excellent opportunities to study the response of a basin-marginal fluvial system to Late Quaternary tectonic uplift, sea-level and climate changes. Complex-response fluvial behaviour accomplishes that the allocation of depositional and erosional events not only depends on the interaction between external controls, but on the position of a certain reach along the longitudinal profile as well. We have demonstrated that only by modelling the development of a 1000-km longitudinal profile of the Meuse on a basin-wide scale, one can begin to reconstruct the sensitive long-term (10^3-10^6 years) fluvial records of basin-marginal systems as stored in terraces and fluvial sediments.

The results from our forward-modelling case study are able to reproduce and confirm qualitative inferences (Merritts et al., 1994) that climate and tectonics are the main forcing factors of fluvial dynamics in the middle and upper reaches. Thus, climatic variability can be expected to be expressed most clearly in the uplifting middle and upper reaches of the fluvial system (cf. Veldkamp and Tebbens, 1999). The stadial/interstadial and glacial/interglacial variability of discharge and sediment supply lead to an alternation of depositional and erosional events, superimposed on a long-term uplift-driven degradation of the longitudinal profile.

In a similar way, climate and sea level are the main controlling factors of fluvial dynamics in the lower reaches. High sediment supply combined with low river discharges during cold stages favour long-term aggradation in the subsiding lower fluvial reaches. Climatic warming and increasing discharges at the dawn of the next interglacial interrupt the long-term aggradation by provoking a short phase of kinetic incision (Leeder and Stewart, 1996), which migrates downstream. The subsequent interglacial sea-level highstand induces a zone of increased fluvial sedimentation in the lower reaches. This zone of gradient backfilling (Schumm, 1993) migrates upstream with time and finally results in the building up of a coastal prism (Talling, 1998) or depositional wedge. The modelling exercise suggests that the extension of the depositional wedge strongly increases with higher sediment supply to the fluvial system, but is independent of an increasing transport distance of the sediment.

The point where the fringe of the interglacial depositional wedge meets the upstream zone of kinetic incision corresponds with the position of the terrace intersection. Upstream from this point, sea-level rise is thus unlikely to have effected long-term longitudinal profile aggradation. The upstream migration of the zone of increased sedimentation (or: gradient backfilling) implies that the terrace intersection migrates upstream too (Törnqvist, 1998), in case of the River Meuse to some 130-km upstream from the modelled coastline. The modelling exercise suggests that the higher Eemian sea level has shifted the position of the terrace intersection even further upstream. Simultaneous and subsequent interglacial kinetic incision in the upstream reaches that are beyond the reach of the depositional wedge (above the terrace intersection), is predicted to produce the type of 100-ka alluvial fill-terraces in the terrace flight area near Maastricht (Fig. 1.1-A) and millennial-period cut-terraces in North Limburg.

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CHAPTER 6

MODELLING THE DOWNSTREAM AND TEMPORAL EVOLUTION OF RIVER MEUSE SEDIMENT COMPOSITION

6 MODELLING THE DOWNSTREAM AND TEMPORAL EVOLUTION OF RIVER MEUSE SEDIMENT COMPOSITION

ABSTRACT

The processes of source rock weathering and erosion, transport and sedimentation will change fluvial sediment composition on different temporal and spatial scales. The residence time of weathering material in a variety of drainage basin catchments determines the influence of weathering processes on sediment composition. Cold-stage climates induce conditions of weathering-limited fluvial sediment supply leading to short weathering duration. Warm-stage climates will favour conditions of transport-limited sediment supply leading to long weathering duration. The interaction of spatial complex response and temporal non-linear behaviour in the fluvial system require thorough quantification of fluvial dynamics to assess the long-term climatic impact on sediment composition.

This paper addresses the influence of external forcing (changes in tectonics, sea level and climate) on the downstream and long-term (10^3-10^5 years) evolution of sediment composition along a fluvial longitudinal profile. The River Meuse (southern Netherlands) served as a case study for a semi 3-D forward-modelling approach to quantify the downstream and 250-0 ka BP sediment fluxes. These have been related to the bulk geochemical properties of the tributary catchments to calculate the composition of the main river sediment flux.

The incorporation of GRIP ice-core climatic change during the simulation of long-term fluvial dynamics clearly yielded systematic changes in fluvial sediment composition. A scenario comprising minimal discharges and maximum hillslope process dynamics during cold glacial periods alternating with maximal discharges and minimal hillslope processes during prolonged interstadials or interglacials largely succeeded to reproduce long-term measured bulk geochemical trends registered in 15-0 ka BP fine-grained sediments of the River Meuse lower reach.

6.1 INTRODUCTION

The geochemical evolution of sedimentary sequences has intrigued many Quaternary earth scientists. In order to make correct interpretations of compositional changes it is essential to understand the natural variability of sediment composition (or bulk mineralogy) on the appropriate spatial and temporal scale. Most fine-grained fluvial sediments are ultimately derived from the erosion of pre-weathered saprolites and/or the truncation of clay-rich horizons of soil profiles (Singer, 1984; Curtis, 1990). Consequently, fluvial sediment composition will vary due to source rock diversity in the drainage basin (provenance), due to physical and chemical weathering of these source rocks under varying climatic conditions and as a result of subsequent erosion and transport pathways within the fluvial system (Johnsson, 1993). Post-depositional diagenetic processes may result in additional authigenic mineral formation in (bio-)chemically reactive environments (Curtis, 1995).

During transport, the sediments of various point sources and a wide range of tributary catchments are intimately mixed to contribute to the suspension load in the main river. In this way, the suspension load integrates the provenance and weathering effects over an entire drainage basin area (Ottesen et al., 1989). During frequently occurring flooding events, the suspension load will be deposited as fine-grained overbank sediments on distal river floodplains and in residual channels acting as natural sediment traps. Bulk geochemical analyses on composite samples of these sediments are therefore highly representative to describe fluvial sediment composition on a basin-wide scale (Ottesen et al., 1989). In Western society, 19^{th}-century industrialisation and other anthropogenic influences often blur natural compositional variation and fluvial sedimentation processes. Therefore, one often has to rely on Quaternary fine-grained fluvial archives to characterise true natural variability. Only in these records pristine and undisturbed terrestrial sediments are still preserved to provide clues to the relative importance of long-term provenance changes and climatic variability (Chamley, 1989).

Bulk geochemical analysis has proven a powerful tool to rapidly quantify the natural compositional variability of fluvial sediments (Veldkamp and Kroonenberg, 1993; Huisman, 1997). Many inventories concentrated on characterisation of the granulometrical controls on bulk geochemistry, on the spatial variation between different tributary catchments or on short-term compositional variation related to man-induced pollution. Moura and Kroonenberg (1990), Hakstege et al. (1993), Veldkamp and Kroonenberg (1993) and Kroonenberg (1994) showed that provenance characteristics and grainsize-sorting during transport may explain >70% of fluvial sediment compositional variation. Huisman (1997) and Walraven et al. (1999) reconstructed post-depositional diagenetic processes in Early Quaternary Rhine and Late-glacial Meuse sediments respectively. Several Western European countries performed systematic geochemical mapping to describe spatial and temporal natural compositional variation and to obtain baseline reference data for environmental pollution (Darnley et al., 1995; Plant et al., 1997). Macklin et al. (1994), De Vos et al. (1996), Swennen et al. (1997) and Van der Sluys et al. (1997) studied sediment composition on a relatively short time-scale (10^0-10^2 years) to assess pollution related to historic ore exploration. Long-term studies (10^3-10^6 years) mainly focus on the reconstruction of sediment routing within fluvial drainage basins (Veldkamp, 1991; Veldkamp and Kroonenberg, 1993; Weltje, 1994; Huisman, 1997) or on reconstructing past climatic variability (Curtis, 1990). Climatic changes in the drainage basin might directly or indirectly change the long-term composition of fluvial sediments (Veldkamp, 1991; Chapters 3 and 4 in this thesis: Tebbens et al., 1998, 1999b).

To enable linking of sediment composition data (bulk geochemical analyses) with internal fluvial dynamics requires quantitative information on the sediment flux of the main river on a time-scale of 10^3-10^5 years. The sediment flux on its turn directly relates to the varying contributions of tributary catchments and the timing, frequency and intensity of erosional and depositional phases in the fluvial system. The fact that erosional and depositional phases are often reconstructed from field data introduces at least three problems, namely that the reconstruction *1)* relies by definition on the preserved sedimentary record, *2)* is often a qualitative one and *3)* is specific for a certain river reach only (e.g. Huisink, 1997). Complex-response behaviour of fluvial systems (Schumm, 1993) complicates the translation of these reconstructed sediment fluxes to larger temporal and spatial scales. Indeed, Merritts et al. (1994), Rose (1995) and Veldkamp and Van Dijke (1998) demonstrated that it is unlikely that different river reaches will respond simultaneously and in the same way to changes in the external controlling factors (tectonics, climate and sea-level change). Thus, one cannot simply extrapolate the reconstructed sediment fluxes for a given river reach directly to another reach. We performed a semi 3-D forward-modelling approach to find a solution to this problem and to quantify the compositional response of a fluvial system to the external controls of tectonic uplift, sea level and climatic change on a basin-wide scale (Chapter 5: Tebbens et al., 1999c).

6.2 THE RIVER MEUSE CASE STUDY

The well-known fluvio-systematic context of the River Meuse provides an excellent starting-point to characterise long-term compositional variation in the sedimentary records. The River Meuse (Chapter 1, Fig. 1.1) is well investigated for its response to external forcing (tectonics, climate and sea level change) on a time-scale of 10^3-10^6 years (Vandenberghe, 1995; Van den Berg, 1996). Berendsen et al. (1995), Kasse et al. (1995), Huisink (1997) and Tebbens et al. (1999a) described the morphodynamic expression of fluvial response to Late-glacial climate change. The Meuse lower reach preserved a record of fine-grained sediments correlative to the period 15-0 ka BP (Chapter 2-4). Late Weichselian fine-grained residual channel fills of the River Meuse contained less Al_2O_3 and more K_2O and MgO than Holocene ones (Chapter 4: Tebbens et al., 1998). The trace element geochemistry yielded similar differences that appeared to be related to higher smectite and vermiculite contents in Holocene samples compared to Late Weichselian samples (Chapter 3 and 4: Tebbens et al., 1998, 1999b). Large-scale provenance changes and post-depositional alteration could not be held responsible for the long-term geochemical variation, raising the fundamental question whether climate change could have changed the detrital clay mineralogy.

Climate influences the intensity (rate) and duration of source-rock weathering. In the Temperate Zone, smectite and vermiculite are typically formed in interglacial soil environments as they are the K- and Mg-depleted weathering products of illite and chlorite (Chamley, 1989; Bronger et al., 1998; Chapter 4: Tebbens et al., 1998). Thus, interglacial climatic conditions might favour a higher supply of pedogenetic clay minerals in Holocene sediments, but this will only happen if the residence time of the weathering material in the source areas (weathering duration) is long enough for chemical weathering to affect the source rocks. Climate not only directly controls the residence time of weathered materials in fluvial drainage basins (e.g. via vegetation cover), but also indirectly via the intensity of hillslope processes and fluvial erosion. Fluvial sediment supply can be weathering-limited or transport-limited (Johnsson, 1993). In a weathering-limited system, the residence time of the weathering products in the upstream catchments is too short for weathering to leave clear climatic imprints on sediment composition. In a transport-limited system, residence time is long enough for climate to overprint provenance characteristics. Relating the observed clay mineralogical differences to the exclusive effect of chemical weathering implies the hypothesis that sediment supply in the Holocene fluvial system was transport-limited (long weathering duration). Alternatively, climate-controlled varying intensities of hillslope supply and fluvial erosion might have changed the bulk (and clay) mineralogy in time, solely via changes in the relative contributions of tributary catchments. This hypothesis implies a short weathering duration and is equivalent to weathering-limited sediment supply. It emphasises the initial provenance characteristics of the different tributary catchments and stresses the role of internal fluvial transport.

On a time-scale of 10^3-10^5 years, the climatic control interacts with tectonics and sea level to generate erosional and depositional phases in the fluvial system. These erosional and depositional phases cause changes in the transported sediment flux quantities. In this way, they influence the basic mineralogy of the sediments. Therefore, it is interesting to study the long-term effects of tectonics, climate and sea-level change on sediment composition. The present paper will focus on the climatic forcing of fluvial dynamics and hillslope supply. We will simulate effects of changing weathering duration on the long-term evolution of sediment composition. We will try to falsify the hypothesis that internal fluvial dynamics and external hillslope supply do not influence the long-term evolution of sediment composition.

6.3 FORWARD MODELLING

Longitudinal profile development

Variations in the external forcing parameters tectonics, sea level and climate perturb the mass balance between the sediment supply to and sediment transport capacity of the fluvial system on a time-scale of 10^3 to 10^5 years. The FLUVER2 model has been developed to describe this long-term fluvial dynamics on a basin-wide scale (Veldkamp and Van Dijke, 1998). The longitudinal profile model includes the two spatial dimensions of length and height (and the third dimension of time), but does not incorporate palaeohydrological channel width, lateral channel(belt) shifts or other morphodynamic features. However, the model accounts for the different sizes of tributary catchments to simulate a downstream increase of discharge (Fig. 5.2) and to cope with the influence of their hillslope sediment supply on sediment composition. The model was made semi 3-D (4-D including time) by assigning a virtual width to each 1-km segment along the longitudinal profile. Suppose the length of the Meuse between two tributaries A and B is 50-km and the catchment area of the downstream tributary B is 500-km^2. One can increase the cumulative catchment area of the Meuse over the stretch A to B by multiplying each 1-km segment in this 50-km stretch with its virtual width, which is 500/50 = 10-km. These virtual widths are solely dependent on the distance between two tributaries and the catchment area of the downstream tributary. With the cumulative catchment area known and the tributary-dependent amount of present-day run off in m/yr, one can now simulate the effects of increasing downstream discharges and stream power on sediment load and transport capacity.

In Chapter 5 we elaborated, calibrated and tested a model version for the River Meuse using tectonic uplift rates, GRIP ice-core climate data and sea-level data reconstructed from coral reef terraces for the last 250-ka. The 250-ka period was chosen to exclude side effects of a capture of the Upper Meuse near Toul (Fig. 1.1-A) on sediment provenance and river discharge and allowed the use of high-resolution input data. A climate and relief feedback mechanism allowed varying hillslope supply in time and along the river profile. The semi 3-D model calculates highest hillslope supply in sloping areas and during relatively cold periods (stadials, full glacial conditions) to simulate short residence times of weathering material in the sediment production source areas. Lowest hillslope supply is calculated in low relief areas during relatively warm periods (interglacials, prolonged interstadials) to represent long residence times in the source catchments. Thorough assessment of the modelling results (and ignoring bulk geochemical data) ended in a most plausible combination of the substrate erodibility, sediment travel distance and hillslope-supply parameters (Ch. 5, Tebbens et al., 1999c). Running the model with this combination yielded a simulation output that largely approximates actual erosional/depositional features for different sections along the Meuse longitudinal profile. A Profile Evolution Map (PEM, Fig. 5.8) visualises these morphological features for the last 250-ka BP. The PEM indicates where and when phases of erosion (grey-shading) and deposition (hatching or pattern fill) are predicted along the Meuse longitudinal profile. The most prominent features in the PEM are an alternation of erosional and depositional phases in the uplifting Ardennes (339-597 km) and its flanking Maastricht terrace flight area (627-km), superimposed on a general degradation of the longitudinal river profile. Gradient backfilling accelerates sedimentation during interglacial sea-level highstands in the subsiding and therefore net depositional Roer Valley Graben, Venlo Graben and North Sea Basin (661-1000 km). Two incisional phases interrupt this depositional record at the start and end of the Eemian (~125-ka BP) and the start of the Holocene interglacial (~10-ka BP). Chapter 5 discusses these phenomena into more detail (Tebbens et al., 1999c).

Sediment composition

We will try to model the geochemical evolution of River Meuse sediments using bulk geochemical data from tributary overbank sediments in the upstream drainage area. The bulk geochemistry of samples taken from a clastic sedimentary record simply reflects the bulk mineralogy of those sediments (Chapter 3). Weathering processes are recorded best in fine-grained clayey and silty deposits that are for the greater part derived from the suspension load of the river (Chamley, 1989). The sediment composition in the River Meuse was quantified using the ratio of K_2O to Al_2O_3 (K/Al-ratio). The major constituents K_2O and Al_2O_3 are strongly positively correlated to phyllosilicates in the clay and fine silt size-fractions and therefore, the K/Al-ratio serves as a good representative for the suspension load composition. Moreover, these constituents are not influenced by lime content, heavy mineral enrichments or local post-depositional overprinting processes in gyttja redox environments (Chapters 3 and 4: Tebbens et al., 1998). The model input K/Al-ratios (Table 6.1) for all major Meuse tributaries in the Ardennes and southern Netherlands were obtained from data relating to the L-samples in Hindel et al. (1996), Swennen et al. (1997) and Van der Sluys et al. (1997). These authors systematically sampled and analysed the <125-μm size-fractions of the upper and lower parts of Holocene overbank loams. The L-samples were taken from the lower parts and represent a preserved record of non-polluted suspension loads resulting from several pre-industrial flooding events. These lower overbank sediments were sampled some distance upstream from the confluence with the main river. Therefore, they integrate the natural composition of a variety of weathered parent materials and soils in the tributary catchments. The K/Al-ratios derive from Holocene overbanks and hence may already bear a Holocene climatic signature. This will be evaluated in the discussion of the results.

We used an extended version of the FLUVER2 model to simulate the spatial and temporal evolution of sediment composition. The erosional and depositional processes within 1-km segments of the longitudinal river profile form the basis to calculate the volume of the sediment flux that is in transport within the main river. Next, the modelled sediment fluxes resulting from river valley erosion and hillslope processes have been linked to the K/Al-ratios of the tributary catchments to quantify downstream and long-term compositional changes of the main river sediment flux. This calculation comprised the following basic assumptions (Figures 6.1-A to C). Consider the sediment flux leaving a segment (Qs[out]) is larger than the incoming sediment flux (Qs[in]). The elevation of the river valley in this segment is lowered only ($dh<0$, erosion) if a negative sediment flux related to internal river valley erosion is not counterbalanced by a positive sediment flux related to external hillslope supply (Fig. 6.1-A). In this situation, both the "fresh" material eroded within the segment and the external hillslope supply will have a local source area composition, namely "Y". These sediment inputs will change the composition "X" of the incoming sediment flux to the new composition X' (Fig. 6.2). It follows that the sediment flux that leaves the segment will bear an increasing local signature if either the valley erosion sediment input or the simulated hillslope supply is large relative to the sediment flux coming in from the drainage area upstream of the segment.

Likewise, the river valley must rise ($dh>0$, deposition) in a segment (Figures 6.1-B and 6.1-C), if the sediment flux leaving the segment is smaller than the incoming sediment flux. This situation corresponds to a river starting to flow over its own valley alluvium. Two possibilities are conceivable: *1)* A positive sediment flux related to external hillslope supply counterbalances a negative flux related to internal valley erosion (Fig. 6.1-B). Again, both the erosion in the river valley and the hillslope supply will contribute material with a local source area composition (Y) to change the composition of the incoming sediment flux (X) to the new composition X'(Fig. 6.2). *2)* In the second situation, both the hillslope supply and the river sediment flux are positive (Fig. 6.1-C). The sediment that is deposited within the segment will

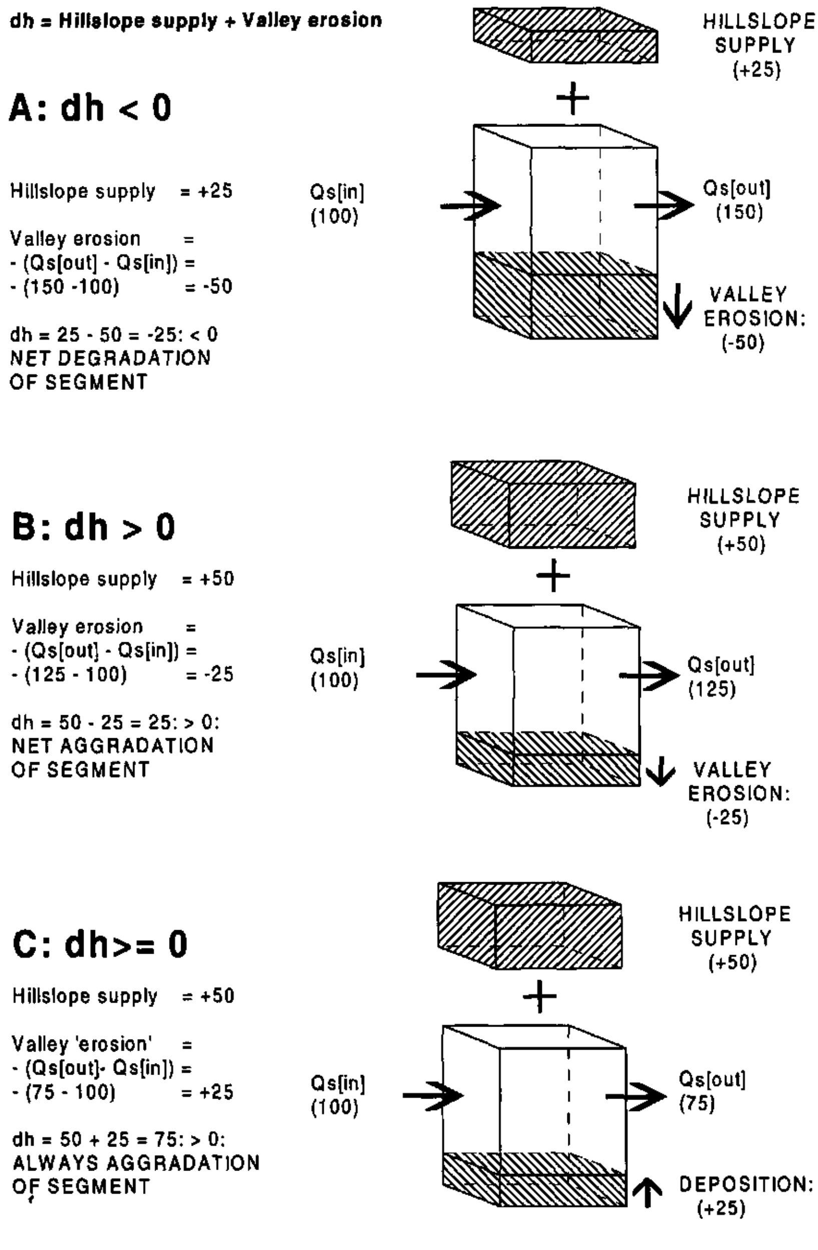

Figure 6.1 Different conditions leading to degradation or aggradation along a 1-km long segment of the Meuse longitudinal profile. The segment elevation is updated with hillslope supply originating from outside the river valley and with a contribution from internal river valley erosion or deposition. The latter is calculated from the difference of Qs[in] and Qs[out] (respectively the volume of the sediment flux entering [in] and leaving [out] the segment) The hillslope supply and river valley erosion processes on their turn depend on external forcing (especially climate) and varying tributary catchment contributions (see Chapter 5).

have the composition of the incoming sediment flux (X). However, the sediment originating from hillslope supply has a local composition (Y) and will start to influence the composition (X) of the incoming sediment flux downstream of the segment (Fig. 6.3). This assumption is justified by the observation that complete sediment mixing at a main river-tributary confluence can show a delay of one to several kilometres (Veldkamp, 1991: Ch. 3.3).

Figures 6.2 and 6.3 show that simple linear equations were used to update the downstream composition of the main river sediment flux. The composition of an incoming sediment flux in a given segment is recalculated by multiplying the fluvial erosion sediment input and hillslope input with their local compositions, adding these to the incoming sediment flux and then dividing the total proportionally by the cumulative amount of sediment that leaves the segment. The underlying assumption for proportionality is that immediate and ideal linear mixing (Weltje, 1994) takes place. Common sense and sediment mixing delay (see above) indicate this necessarily is a robust simplification of 3-D reality.

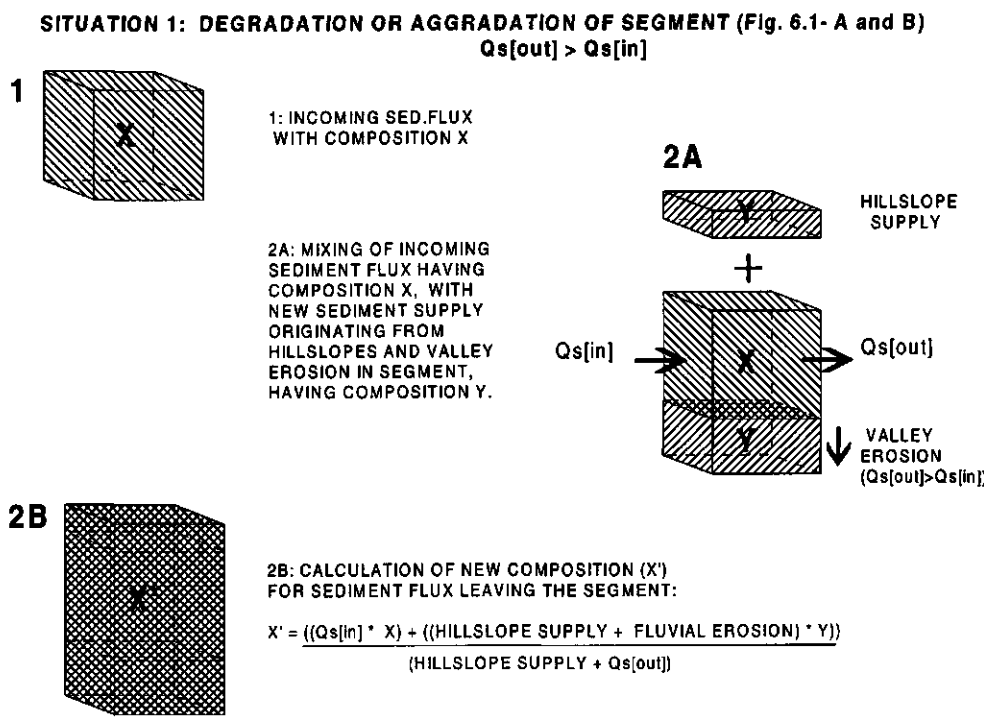

Figure 6.2 The calculation of the main river sediment flux composition for situation A and B in Figure 6.1. X is the composition of the incoming sediment flux (e.g. the K/Al-ratio), Y is the composition of the local hillslope supply originating from a tributary catchment (the input K/Al-ratios) and X' is the updated composition after mixing of the different contributions. The input K/Al-ratios (the Y-values) are listed in Table 6.1. Tributaries are shown in Figure 6.6.

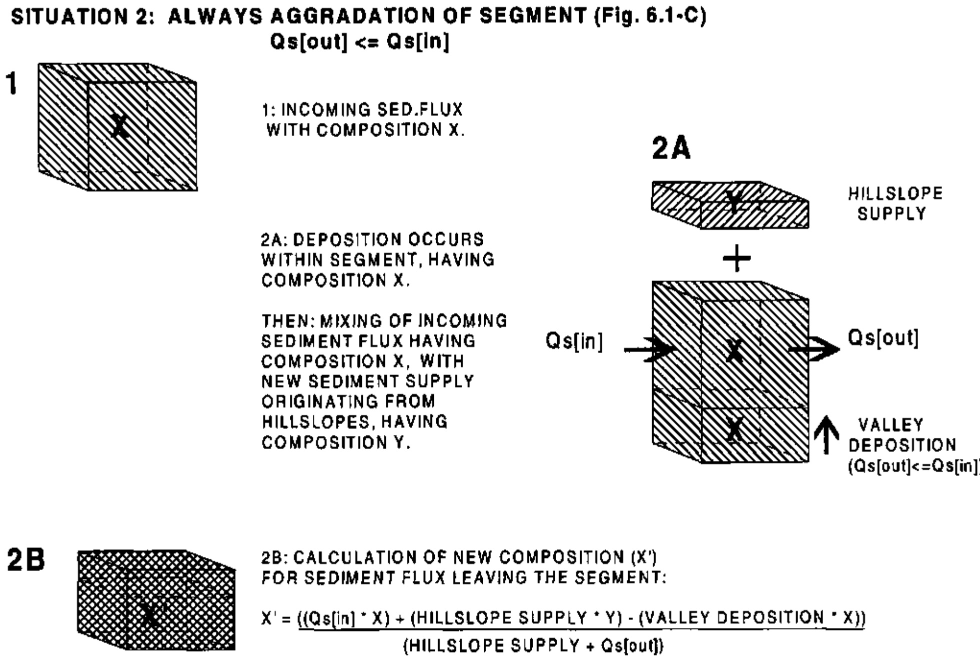

Figure 6.3 The calculation of the main river sediment flux composition for situation C in Figure 6.1. X is the composition of the incoming sediment flux (e.g. the K/Al-ratio), Y is the composition of the local hillslope supply originating from a tributary catchment (the input K/Al-ratios) and X' is the updated composition after mixing of the different contributions. The input K/Al-ratios (the Y-values) listed in Table 6.1. Tributaries are shown in Figure 6.6.

6.4 RESULTS AND DISCUSSION

Forward modelling of the development of the River Meuse longitudinal profile enabled the quantification of the main river sediment flux and its composition for the last 250-ka BP. The Profile Evolution Map (PEM, Fig. 5.8) shows where and when erosional and depositional phases are predicted to provide insight into the long-term evolution of the longitudinal profile on a basin-wide scale. In the same manner, the geochemical composition of the sediment flux in every 1-km segment along the longitudinal profile is plotted versus time to generate a Geochemical Evolution Map (GEM, Fig. 6.4). By reading the map from left to right (one time-slice), one can trace the effects of downstream sediment mixing on sediment composition along the longitudinal profile. By reading the map from bottom to top (one site-slice along the profile), one can trace the effects of long-term external forcing and more specifically of climate changes on sediment composition as time proceeds. Thus, the GEM shows the combined effects of downstream variation as well as temporal variation.

Downstream variability

Linear mixing of the sediment contributions of successive downstream tributary catchments, each having their own local catchment composition, causes considerable compositional varia-

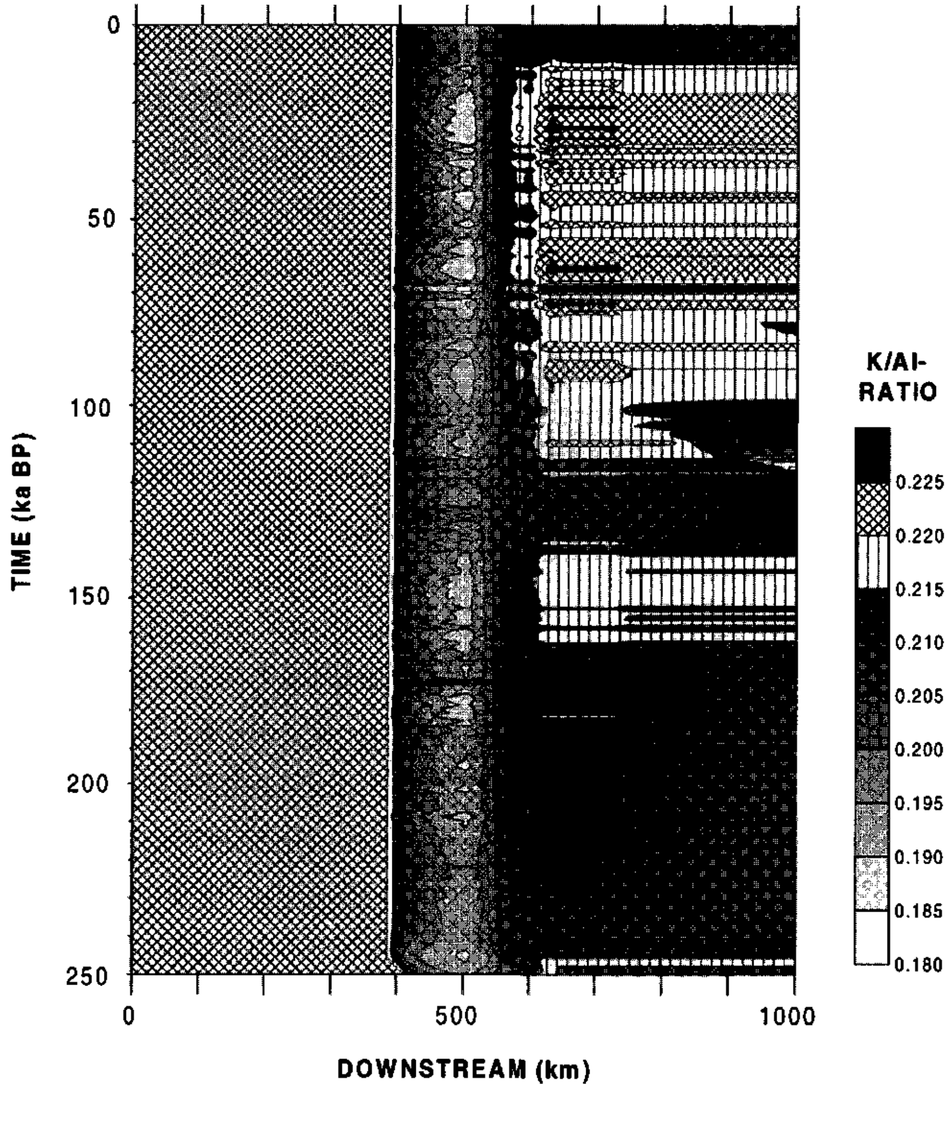

Figure 6.4 Geochemical Evolution Map showing the modelled downstream and temporal evolution of the composition of the River Meuse sediment flux for the period of 250 to 0-ka BP. Initial settings and modelled fluvial dynamics are identical to those for the PEM of Figure 5.8. Figure 6.5 shows the impact of tributaries on the downstream evolution of sediment composition and Figure 6.6 presents their exact positions along the longitudinal profile. Initial K/Al-ratios for the tributary catchments are listed in Table 6.1. Figure 6.9 presents the temporal evolution for the 500-, 627- and 762-km sites along the longitudinal profile.

bility along the longitudinal profile. In Fig. 6.5, the modelled sediment flux K/Al-ratios have been plotted versus the downstream position of every segment along the longitudinal profile and for two climatically contrasting time-slices of the GEM, namely 21.9-ka BP and 6-ka BP. The 21.9-ka BP time-slice represents the Late Pleniglacial coldest part in the GRIP-core, for which the model simulates a relatively low discharge input and a high hillslope supply. The 6-ka BP time-slice represents the mid-Holocene climatic optimum, for which the model simulates a relatively high discharge input and low hillslope supply. In the following section, we will first discuss downstream variability in K/Al-ratios on the basis of the results for the 21.9-ka BP time-slice.

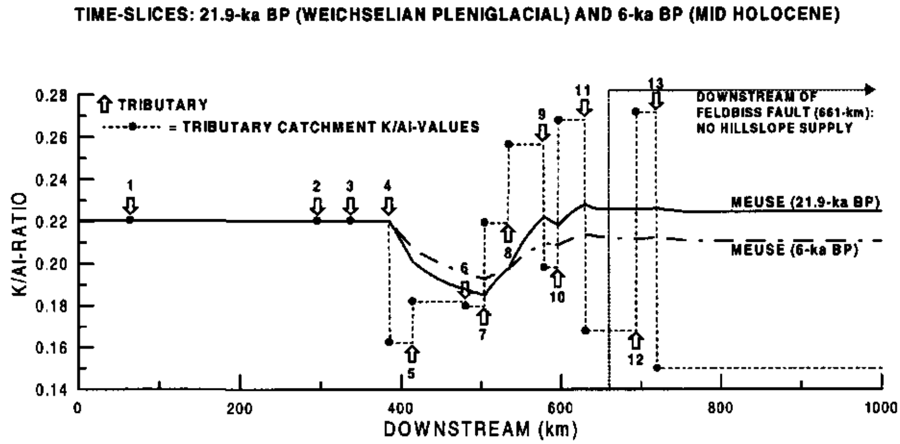

Figure 6.5 The evolution of downstream sediment composition (represented by the K/Al-ratio) for two climatically contrasting time-slices, namely 21.9-ka BP (Weichselian Pleniglcial) and 6-ka BP (Mid Holocene). The arrows and dashed line indicate the initial K/Al-ratio input values of the following tributary catchments (see Fig. 6.6): 1 = Vair, 2 = Chiers, 3 = Bar, Vence, Sormonne, 4 = Semois, 5 = Viroin, 6 = Houille, Hermeton, Lesse, 7 = Molignée, Bocq, Sambre, 8 = Mehaigne, Hoyoux, 9 = Ourthe, Amblève, Vesdre, 10 = Jeker, Berwinne, 11 = Geul, Roer, 12 = Swalm, 13 = Niers.

The Meuse, the major tributary Chiers and five minor tributaries all drain the same lithostratigraphical unit along the entire upstream section (0-385 km), namely Jurassic marls and limestones in the north-east Paris Basin (Fig. 6.6 and 6.7). Local relief is still unimportant here. The tributary catchments in this area have been assigned identical K/Al-ratios (River Chiers: 0.22) and consequently, the calculated K/Al-ratios do not change for the main river segments in the 0-385 km upstream section. Downstream of ~339-km, local relief markedly increases as the Meuse enters the uplifting Ardennes low mountain range in Belgium. Inputs from the second largest tributary Semois and the major tributaries Viroin and Lesse enlarge the main river discharge and stream power considerably from 385-km to 505-km (Figs. 5.2, 6.6 and 6.7). The high local relief causes maximum hillslope process activity in the Ardennes area (Fig. 6.8). This generates abundant sediment supply, having the tributary catchment composition of local Palaeozoic shales and phyllites. With K/Al-ratios ranging from ~0.16 to 0.18, the high Ardennes supplies effectively dilute the K/Al-ratio of the Jurassic sediment flux coming in from the upstream Paris Basin. The K/Al-ratio of the main river sediment flux......

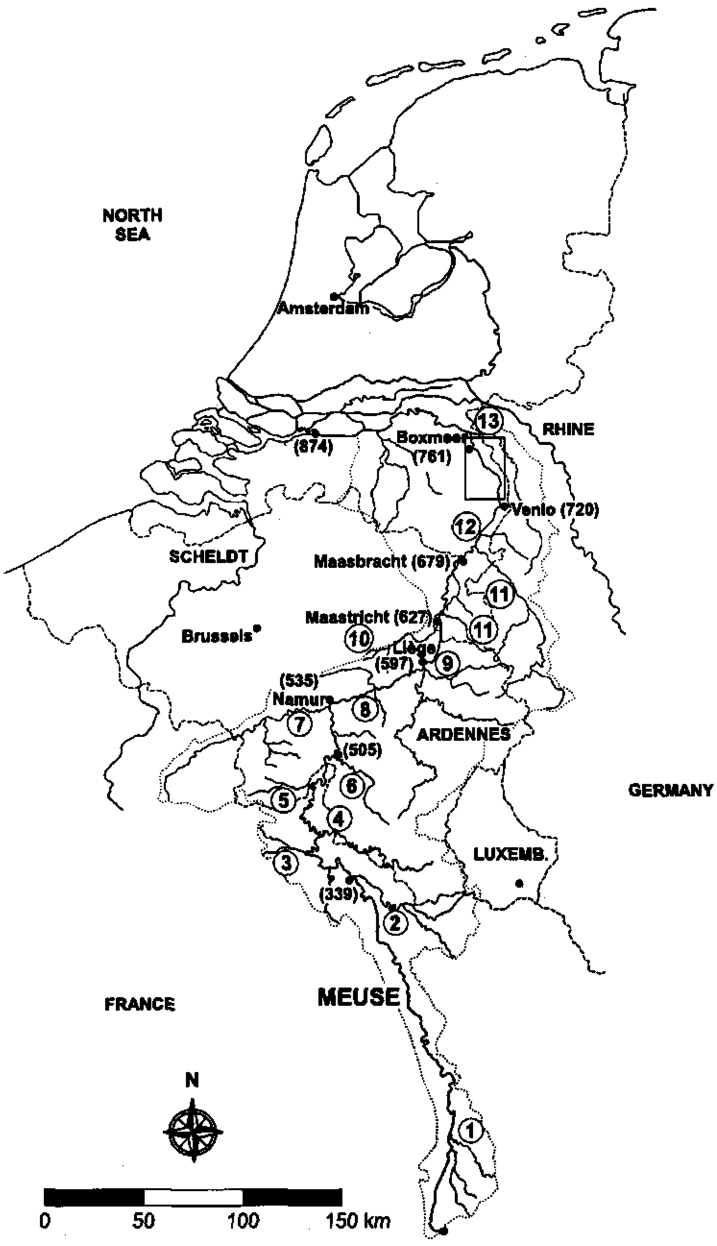

Figure 6.6 Tributaries in the River Meuse drainage area (adapted from Berger and Mugie, 1994): 1 = Vair, 2 = Chiers, 3 = Bar, Vence, Sormonne, 4 = Semois, 5 = Viroin, 6 = Houille, Hermeton, Lesse, 7 = Molignée, Bocq, Sambre, 8 = Mehaigne, Hoyoux, 9 = Ourthe, Amblève, Vesdre, 10 = Jeker, Berwinne, 11 = Geul, Roer, 12 = Swalm, 13 = Niers. Numbers between brackets indicate the distance in km downstream from the source. Table 6.1 lists the input K/Al-ratios for these tributaries. The rectangle indicates the study area (Chapters 2 to 4).

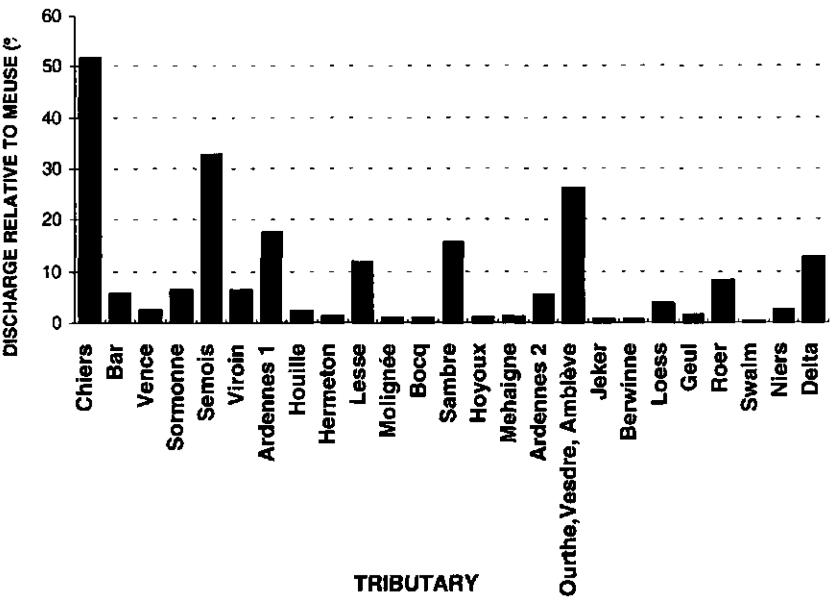

Figure 6.7 The discharges of the tributaries of the River Meuse drainage area expressed as a percentage of the River Meuse discharge at the point of confluence. E.g., the discharge of the tributary Lesse is 11% of the Meuse discharge before their discharges are summed. Discharge data were obtained from Berger and Mugie (1994). The Ardennes 1 and 2, Loess and Delta discharges are restgroups and can be considered as baseflow discharges from these areas.

approximates the tributary catchment values near 500-km (Central Ardennes). This implies that at this point, the admixture from the southern Ardennes tributaries has completely overshadowed the influx from the upper reaches draining the Jurassic rocks.

At 535-km (Namur), the fourth major tributary Sambre joins the Meuse (Figs. 6.6 and 6.7). This river and four minor Ardennes tributaries supply sediments with relatively high K/Al-ratios, whilst hillslope supply is at a maximum in this area (Fig. 6.8). Accordingly, the K/Al-ratio of the main river sediment flux rises considerably again to ~0.22. Near 597-km (Liège), the fifth major tributary Ourthe (including the major Ardennes Amblève and Vesdre tributaries) joins the Meuse, to lower the K/Al-ratio once more. Past this confluence, the contributions of downstream tributaries become less and less important relative to the discharge and sediment flux of the main river, which can be considered almost mature now. Additionally, local relief and hillslope activity decrease from 500-km onwards (Fig. 6.8). Both features accomplish that the local K/Al-ratios of tributary catchments downstream of 597-km have progressively less influence on the K/Al-ratio of the large main river sediment flux. The slight downstream changes in the GEM of Figure 6.4 and the time-slices of Figure 6.5 corroborate this.

From ~630-km onwards, the hillslope supply fades to zero near the Feldbiss fault at 661-km. Simultaneously, the PEM (Fig. 5.8) indicates that the Meuse starts to flow over its own alluvium in the depositional area of the Rhine/Meuse delta in the Netherlands. Here, the supply of material with a local composition originating from hillslope processes is nullified and because fluvial erosion has changed into deposition, a local erosive contribution from the

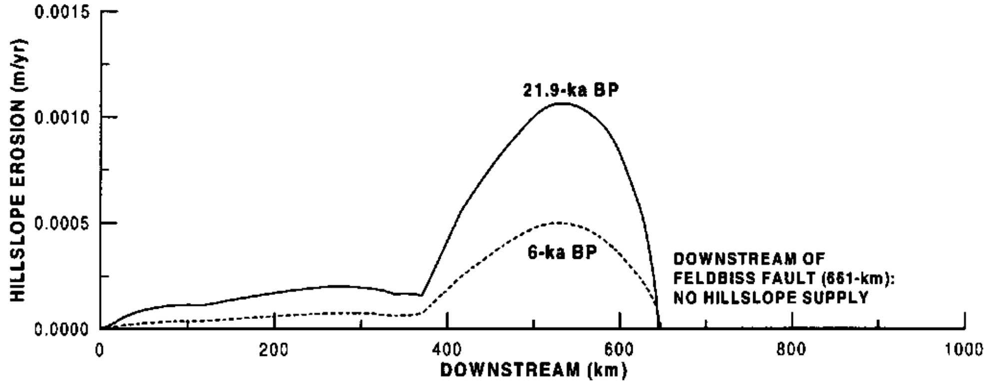

Figure 6.8 The increase of modelled hillslope erosion (= hillslope supply to the river) owing to the downstream increasing local relief along the River Meuse longitudinal profile. The time-slices correspond to the cold and dry Weichselian Pleniglacial (21.9-ka BP) and the warm and humid mid Holocene (6-ka BP).

main river can be expected neither. Therefore, the K/Al-ratio of the main river sediment flux hardly changes as from ~630-km downstream.

Figure 6.5 shows that linear mixing of the main river sediment flux with the sediment supplies from tributary catchments causes comparable downstream changes in sediment composition for the time-slice 6-ka BP (Mid Holocene). However, the amplitude of the changes is smaller compared to the 21.9-ka time-slice, suggesting that temporal differences play a role as well.

Temporal variation

The GEM (Fig. 6.4) shows a compilation of all time-slices and enables us to assess the geochemical evolution in time for one site (or segment) along the longitudinal profile. At this moment, it is important to realise that the input values for the K/Al-ratios of all tributary catchments have been kept constant in time during the whole simulation. Thus, the simulation did not incorporate effects of changing weathering rates (intensity) in the tributary catchments, enabling us to assess the pure effects of varying hillslope activity and internal river dynamics on the evolution of sediment composition.

Figure 6.4 indicates that the composition of the main river sediment flux for segments in the 0-385 km section does not change with time. This is because both the Meuse and involved tributaries drain the same geological unit and the composition of the respective catchments has been considered constant in time. However, the K/Al-ratios do change with time downstream of 385-km, with changes being most clearly expressed in the Meuse lower reaches (>597-km). Figures 6.9-A, 6.9-D, and 6.9-G testify to this in showing the long-term evolution of sediment composition for three sites at 500-km (Central Ardennes), 627-km (terrace flight area, Maastricht) and 762-km (net depositional area, Venlo Graben). The linear regression fits (plotted for indication only) point to a long-term trend of slightly decreasing K/Al-ratios for the Central Ardennes area as time proceeds and simulated sediment fluxes are younger. The terrace flight area and the Venlo Graben area on the other hand show long-term trends of increasing K/Al-ratios, although external forcing input data have been the same as for the Central Ardennes area. The long-term compositional changes remind of the well-esta-

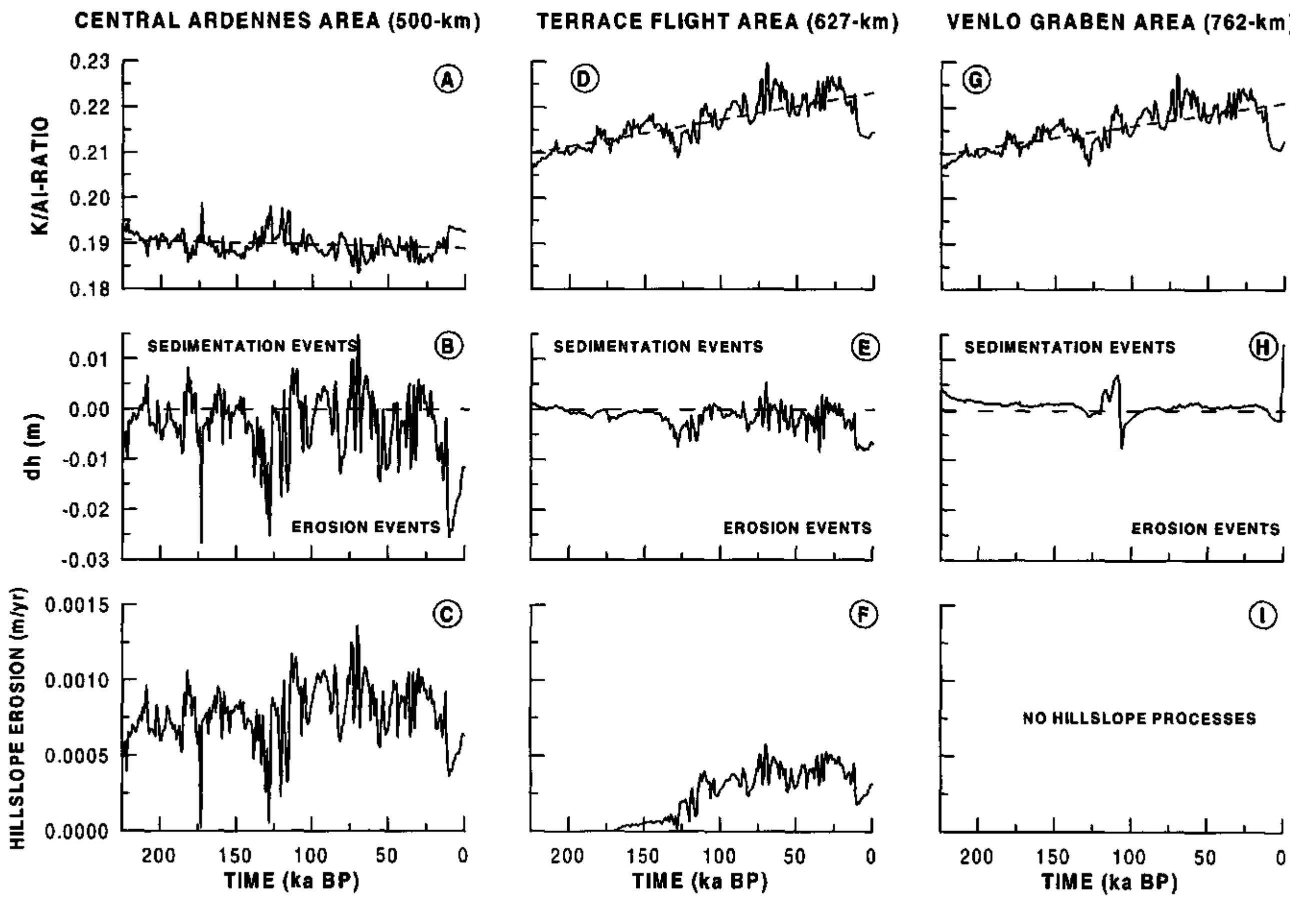

Figure 6.9 The temporal evolution of *sediment composition* (A, D and G), the temporal evolution of fluvial dynamics (B, E and H) and the temporal evolution of hillslope erosion (C, F and I) for three different reaches along the Meuse longitudinal profile.

blished trend of increasing slate and phyllite contents and decreasing quartz contents in the Quaternary terrace gravels for the terrace flight area near Maastricht (Felder and Bosch, 1989; Van den Berg, 1996). Excursions from these long-term trends at ~125-ka and from 10 to 0-ka predict that the interglacial K/Al-ratio will increase in the Central Ardennes area, but will clearly decrease in the Maastricht and Venlo Graben areas. Prolonged interstadials demonstrate this same phenomenon, although to a lesser extent. The PEM (Fig. 5.8) and Figures 6.9-B, 6.9-E and 6.9-H indicate that the excursions coincide with erosion events in the river system. Figures 6.9-B to H are "site-slices" of the PEM, showing erosional and depositional phases versus time for sites at 500-km, 627-km and 762-km.

The different behaviour of upstream and downstream river reaches that nevertheless simultaneously experience comparable climatic conditions once again (Chapter 5) stresses the importance of the downstream position of the river section. The spatial variation in Figure 6.8 and Figures 6.9-C, 6.9-F and 6.9-I shows that the intensity of hillslope processes is considerably less for the sections at 627-km (Maastricht) and 762-km (Venlo Graben) compared to the one at 500-km (Central Ardennes). This is a direct consequence of fading local relief in downstream direction of the uplifting Ardennes low mountain range. Moreover, the model simulates that the intensity of hillslope processes is even less during interglacial or prolonged interstadial periods, because it incorporates a negative feedback-mechanism between climatic amelioration and hillslope processes (see above). Figure 6.8 elegantly demonstrates this difference between the full-glacial 21.9-ka BP time-slice and the interglacial 6-ka BP time-slice. A decreasing contribution of hillslope erosion processes to the sediment supply relative to increasing river discharges will generate interglacial/interstadial erosion events (Chapter 5: Tebbens et al., 1999c). Furthermore, a decreasing hillslope supply from the tributary catchments will decrease the relevance of local K/Al-ratios for the composition of the main river sediment flux too. Thus, it becomes clear that the influence of local hillslope supply on the K/Al-ratio of the main river sediment flux in sloping areas is dampened during interglacial periods, while the influence of local supply is strengthened during glacial periods.

Indeed, Figure 6.5 shows that the Pleniglacial (21.9-ka BP) K/Al-ratios are distinctively lower compared to the interglacial (6-ka BP) values for the 385-535 km section. The combination of high hillslope material admixture with very low local K/Al-ratios in this section causes the 21.9-ka sediment flux to reach lower K/Al-ratios than the 6-ka BP sediment flux. On the other hand, this same high hillslope supply and exactly higher K/Al-ratios of tributary catchments in the section >535-km will lead to higher K/Al-ratios during the 21.9-ka BP time-slice compared to the 6-ka BP time-slice. These results indicate that without having incorporated effects of increasing weathering rates, climatic forcing of hillslope supply and internal river dynamics is able to cause changes in long-term sediment composition via indirect provenance effects and complex-response processes. Now the main downstream and temporal compositional trends have been established, it is interesting to check whether the modelled K/Al-ratios reflect actual trends in measured data.

Modelled K/Al-ratios versus measured K/Al-ratios

The modelled K/Al-ratios can be compared to measured granulometrical, clay mineralogical and bulk geochemical data from Chapter 4 (Tebbens et al., 1998) by zooming in on the Venlo Graben area (~762-km) and on the time-frame 15-0 ka BP (note: model/GRIP-core years, not ^{14}C-years). First, the modelled temporal trends of the K/Al-ratios will be compared to the measured data and then the absolute values of the data and the amplitude of the changes will be discussed.

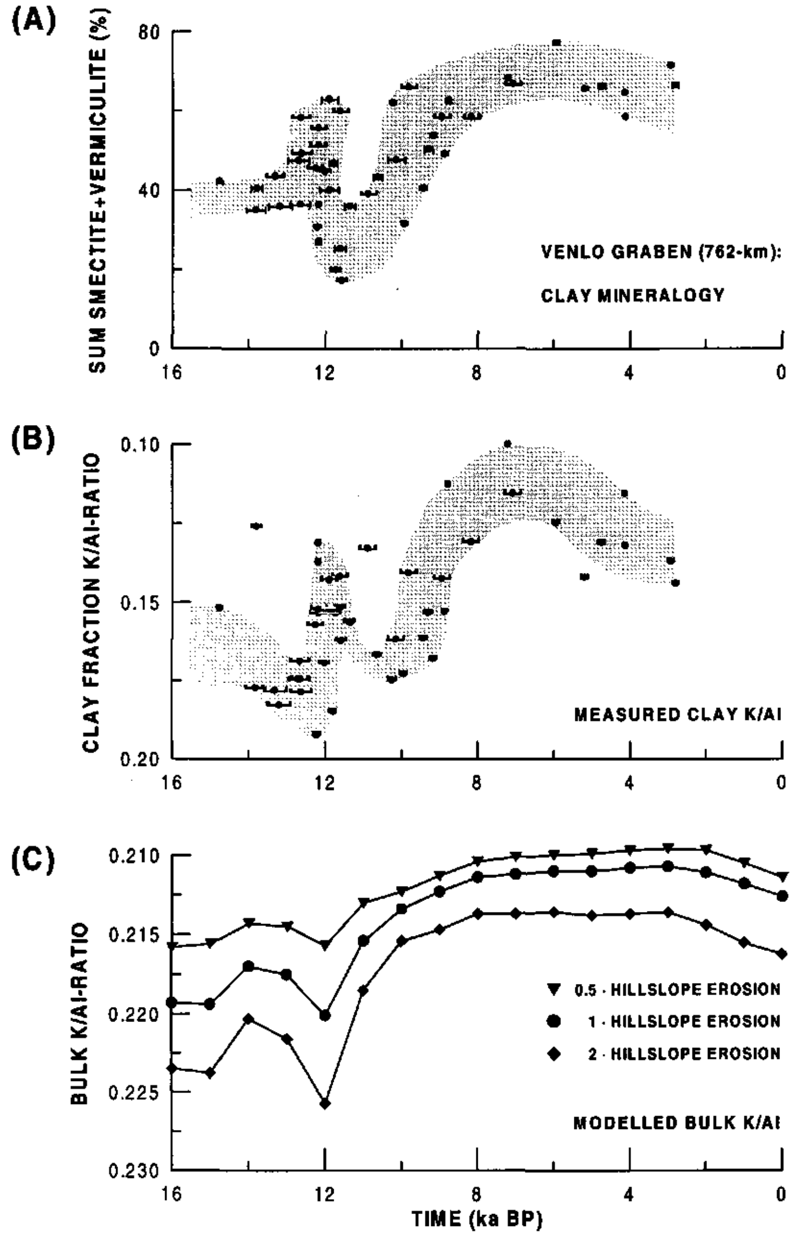

Figure 6.10 The evolution of *(A)* measured clay mineralogy, *(B)* clay fraction K/Al-ratios and *(C)* modelled bulk fraction K/Al-ratios during the Late-glacial time-frame for the Venlo Graben river section (at ~762-km downstream of the source). Figure C also demonstrates the impact of three different intensities of hillslope erosion processes on the modelled bulk K/Al-ratios (see text).

Temporal trend

In Chapter 4 (Tebbens et al., 1998) we reconstructed a bandwidth (Fig. 6.10-A) for the summed smectite and vermiculite contents in separated clay fraction samples versus their approximate time of deposition. The bandwidth suggests increasing smectite and vermiculite contents during the relatively warm Lateglacial Interstadial (~14.7-12.3 ka BP in the GRIP ice-core). A sharp drop registers the cold Younger Dryas Stadial (~12.3-11.5 ka BP), followed by a second distinct rise during the Early Holocene. Figure 6.10-B demonstrates that the measured K/Al-ratios of these same clay fraction samples reflect the major trend in this bandwidth. Increasing smectite and vermiculite contents correspond to decreasing K/Al-ratios (note reverse y-axis). Comparing the measured clay fraction data (Fig. 6.10-A and -B) with the modelled bulk K/Al-ratios for the Venlo Graben area (Fig. 6.10-C) shows that both the timing and the direction of changes match quite good. This qualitatively good temporal correspondence suggests that climatic forcing of hillslope supply and internal fluvial dynamics during modelling is able to simulate the measured evolution of K/Al-ratios in the Meuse fluvial system to great extent. Therefore, the model simulation corroborates the hypothesis that changes in hillslope processes and internal river dynamics are able to change the composition of fluvial sediments on a time-scale of 10^3-10^5 years.

Absolute values

The K/Al-ratios of the measured bulk samples in Chapter 2 to 4 were obtained from different sieving fractions than the model input K/Al-ratios. Phyllosilicates in the clay and fine silt size-fractions (having low K/Al-ratios) are known to dilute feldspar minerals (having higher K/Al-ratios) in the sand fractions (cf. Chapter, 3; Van den Broek and Van der Marel, 1980; Huisman, 1997). For this reason, the K/Al-ratios obtained from clay fractions (no particles >8-μm) are in absolute sense lower (Fig 6.10-B) than the modelled values, which were based on bulk data. So, in order to compare the K/Al-ratio magnitudes of measured samples with the modelled bulk K/Al-ratios, it is important to correct for granulometric differences. Comparing samples that contain similar Al_2O_3-percentages can do this, because these samples will have a roughly comparable granulometry (Chapter 3).

Table 6.1 The K/Al-ratio input values for all major Meuse tributaries. Data for tributary 1 to 12 are calculated from K_2O- and Al_2O_3-percentages in Hindel et al. (1996); Swennen et al. (1997) and Van der Sluys et al. (1997). The tributaries Swalm and Niers were not sampled. Their ratios were based on the Bosbeek stream data and Holocene Meuse sediments with >12% Al_2O_3 respectively (see Fig. 6.11).

Number and tributary catchment (Fig. 6.6)	K/Al-ratio
1: Vair	0.2206
2: Chiers	0.2206
3: Bar, Vence, Sormonne	0.2206
4: Semois	0.1621
5: Viroin	0.1819
6: Houille, Hermeton, Lesse	0.1796
7: Molignée, Bocq, Sambre	0.2194
8: Hoyoux, Mehaigne	0.2565
9: Ourthe, Vesdre, Amblève	0.1982
10: Geer/Jeker, Berwinne	0.2682
11: Geul, Rur/Roer	0.1679
12: Swalm (data from Bosbeek-stream in Van der Sluys et al.)	0.2717
13: Niers (data from lower reach Meuse sediments)	0.15

The model input K/Al-ratios for the tributary catchments (Table 6.1) stemmed from analyses on the <125-µm size-fractions of bulk samples (Hindel et al., 1996; Swennen et al., 1997; Van der Sluys et al., 1997). Our previous bulk data originated from the <2000-µm size-fractions (Chapter 3: Section 3.4). To correct for granulometry, we plotted our measured K/Al-ratios versus the Al_2O_3-percentages of the samples in Figure 6.11. Silty and clayey bulk samples contain high Al_2O_3-percentages (Chapter 3) and will obviously have low K/Al-ratios. In the Meuse drainage area, the K/Al-ratios measured on the <2000-µm fraction of bulk samples will thus always be higher than those obtained from the <125-µm fractions.

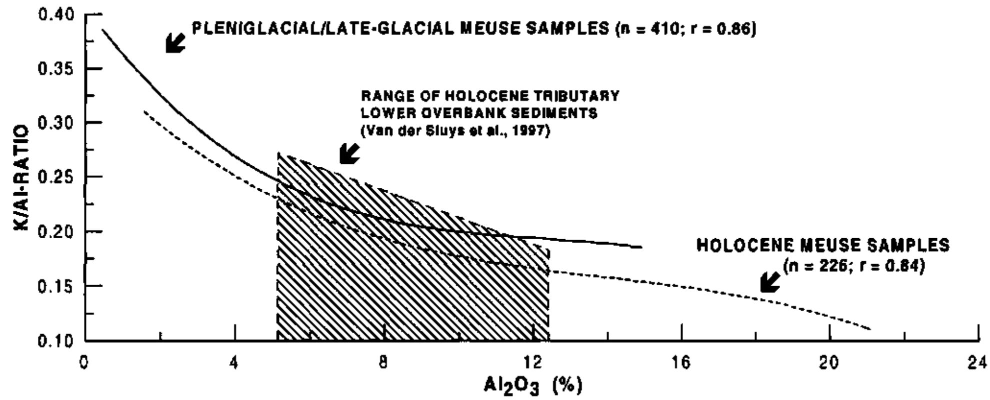

Figure 6.11 The relation between the measured K/Al-ratios and the Al_2O_3-percentages of Pleniglacial/Late-glacial and Holocene fine-grained Meuse samples.

The lower samples of pristine Holocene tributary overbank sediments in Van der Sluys et al. (1997) contained 5.1 to 12.4% Al_2O_3 (range indicated in Fig. 6.11). The fits in Figure 6.11 indicate that in this Al_2O_3-range, the measured K/Al-ratios may be expected to range from ~0.15 to 0.175 for Holocene silty and clayey bulk samples and from ~0.18 to 0.21 for Late-glacial samples. The model input values were based on measurements of Holocene tributary overbanks, so one would expect the modelled K/Al-ratios of the main river sediment flux to be nearer the measured Holocene values than the Late-glacial values. However, the modelled values (~0.21-0.22) appear to approximate the measured Late-glacial sediment values most. Table 6.1 and Figure 6.5 show that most major tributary catchments already have high Holocene K/Al-ratios (the model input values) of >0.18. Because the model calculations assume ideal linear mixing, it will be clear that these high input values will never be able to yield the low range (~0.15-0.175) of measured Holocene values. Therefore, the modelled K/Al-ratios will in absolute sense be higher than the measured ones, but the temporal trends remain the same.

Amplitude

Chapter 4 discussed the phenomenon that the measured K/Al-ratios of Holocene bulk samples are lower with respect to the Pleniglacial and Late-glacial samples in comparable granulometric ranges. Figure 6.11 suggests that these ratios can be expected to differ 0.015-0.03 for samples containing 5.1 to 12.4% Al_2O_3. However, forward modelling of the K/Al-ratios in the Venlo Graben indicates that the modelled 12-ka BP value and the 6-ka BP value differ only ~0.01 (Fig. 6.10-C, ●-line). Thus, the measured difference between Late-glacial and Holocene samples (= the amplitude of the temporal compositional change) is 1.5 to 3 times higher compared to the modelled amplitude. To assess a possible influence of the magnitude of hillslope supply on the compositional difference between Late-glacial and Holocene samples (see Fig. 6.5 and 6.8), two scenarios were tested in which the hillslope supply was halved and doubled respectively (Fig. 6.10-C). The "halved"-scenario decreased the difference between the modelled 12-ka BP and the 6-ka BP K/Al-ratios with 37% (from 0.0091 to 0.0057). The "doubled"-scenario increased the difference between the modelled 12-ka BP and the 6-ka BP K/Al-ratios with 33% (from 0.0091 to 0.0121), but this was still not sufficient to increase the modelled difference to the measured difference of 0.015-0.03 (+ 65-230%). One could try to incorporate even higher hillslope supply to allow for this difference, but this would yield unrealistic results in view of longitudinal profile development. Thus, with the current model settings, it is not yet possible to simulate the exact magnitude of change in the bulk K/Al-ratios.

6.5 CLIMATIC INFLUENCE ON SEDIMENT COMPOSITION

Climatic change not necessarily directly controls fluvial sediment composition, but can exert its influence indirectly via the hillslope contributions of tributary catchments in a high-relief or sloping area. The forward-modelling exercise shows that climatic forcing of the intensity of hillslope processes lowers the sediment supply of tributary catchments in the high-relief Ardennes area during interglacials and prolonged interstadials, generating erosional phases in the fluvial system and weakening the influence of local catchment K/Al-ratios. In this way, the sediment supply is simulated to shift from glacial weathering-limited to interglacial transport-limited conditions. This scenario performed well to predict the temporal dimensions of compositional change as well as the direction of change for sediments in the Meuse lower reach. Given the current model settings, we therefore suggest that internal fluvial dynamics and provenance changes are able to alter the sediment composition of sediments in the lower reach of the fluvial system and that our initial hypothesis is not falsified. However, the absolute K/Al-ratios and the amplitude of change did not fully match with the measured data, suggesting that other factors must interfere, but which have not yet been included in the model scenario.

We suggest that the direct effects of climatic change may be one of those factors. Especially in an interglacial transport-limited situation, the residence time in the catchment will be long enough to change the composition of parent rocks via chemical weathering. Chemical weathering processes will not only depend on the residence time of weathering material in the catchment, but on the intensity or rate of the process as well. Rising temperatures and greater availability of hydrolising agents like water and organic acids (from vegetation) will strongly speed up chemical weathering rates of minerals. However, the model did not incorporate increasing weathering rates, because it assumed constant weathering-intensity conditions during sediment production in the source areas: the model input K/Al-ratios for different tributary catchments have been kept constant in time during the whole simulation. The validity of this assumption may be questioned, as follows. Both temperature and precipitation have fluctuated substantially during the Weichselian (Dansgaard et al., 1993). Temperatures, precipitation and water discharges in the important upstream Ardennes area increased dramatically at the start of the Late-glacial and during the Early Holocene (Guiot and Couteaux, 1992; Walker, 1995) and presumably also during earlier interstadials or interglacials. Permafrost disappeared to accelerate percolation (Hoek, 1997), needed to dispose dissolved basic cations via base flow to the groundwater. Ultimately, the growth and maturation of closed forest vegetation favoured stabilisation of the substrate and interstadial/interglacial soil formation (cf. Guiot and Couteaux, 1992; Bohncke, 1993; Hoek, 1997). In this way, the interglacial intensity (the rate) of hydrolysis of minerals or chemical weathering processes must have been considerably higher in the whole drainage basin to generate higher clay contents and a higher amount of pedogenetic clay minerals that are chemically more stable. Thus, one might expect that interglacial chemical weathering in the tributary catchments lowers the K/Al-ratios of the sediments owing to the formation and ultimately the supply of a larger amount of clay minerals that are richer in Al_2O_3 and poorer in K_2O (Chapter 4: Tebbens et al., 1998). Such a situation would create temporal compositional changes in the upstream 0-385 km section as well, whereas the current model does not predict them there. Additional chemical weathering processes will certainly amplify the compositional changes related to downstream sediment mixing and fluvial dynamics as described above. This suggests that the alternation between cold-stage weathering-limited sediment supply and warm-stage transport-limited sediment supply via climatic forcing of hillslope processes and internal river dynamics can explain at least part of the observed temporal variation, but not all of it.

Another mechanism that may interfere in the sediment composition of fine-grained sediments is the large-scale aeolian influx of loess in the Meuse drainage basin. Van den Berg and Veldkamp (1996) had to assume considerable inputs of aeolian coversands to explain the last Pleniglacial aggradation event. In Chapter 4 (Tebbens et al., 1998) we and Törnqvist et al. (1994) already pointed out that a substantial part of the fine-grained sediments in the River Meuse drainage area might originate from aeolian input. Several tributaries (Sambre, Berwinne, Jeker, Geul, Rur/Roer) drain an extensive loess-covered area in northern Belgium, the southern Netherlands and Germany (Mücher, 1973; Langohr and Sanders, 1985). Additionally, loess is admixed in many present-day soils on Palaeozoic rocks in the Ardennes (cf. De Coninck et al., 1979; Langohr and Van Vliet-Lanoë, 1979). The anthropogenic introduction of agricultural and deforestation practices has augmented erosion in the loess belt since the Late Atlantic (cf. Langohr and Sanders, 1985; Pastre et al., 1997). Many present-day interglacial overbanks of tributaries draining the loess belt might therefore originate from the erosion of previously stabilised loess deposits. This additional loess component has likely influenced the bulk geochemical data and model-input values for our lower reach tributaries. Indeed, bulk geochemical analyses of the sediments of loess-dominated tributary catchments (e.g. Mehaigne, Jeker and Geul) indicate higher K/Al-ratios (Hindel et al., 1996; Van der Sluys et al., 1997). Likewise, it may be expected that an influx of unweathered loess increases the K/Al-ratios of fine-grained sediments in the fluvial system.

6.6 CONCLUSIONS

By linking fluvial overbank bulk geochemical data of tributary catchments to a modelling approach of Late Quaternary fluvial dynamics on a basin-wide scale, it is possible to simulate long-term (10^3-10^5) compositional changes of the sediment flux along the longitudinal profile of the main river. This opens new perspectives in the reconstruction of sediment fluxes and sediment routing in Quaternary fluvial systems. Measured bulk geochemical data of well-dated fine-grained sediments deposited in the lower reaches of the fluvial system can be used to validate model output data, provided that they have been obtained from comparable grainsize-classes and that they have not been affected by post-depositional processes.

We used the K/Al-ratios of fine-grained tributary overbanks to simulate long-term compositional changes along the longitudinal profile of the River Meuse in a 250-0 ka BP case study. Mixing of the main river sediment flux with the supplies of the tributary catchments causes considerable downstream variation in composition. The high sediment supply from tributary catchments in the high-relief Ardennes middle reach overshadows the sediment flux of the upper reaches within 100-km downstream of the upper reach source areas. However, as the main river discharge and sediment flux grow larger, the tributaries in the lower reaches are predicted to have progressively less impact on sediment composition.

Modelled temporal variation is pronounced most clearly in the lower reaches and consistently reflects the measured trends in Late-glacial fine-grained residual channel infillings. Considerable temporal variation in sediment composition was predicted for the lower reaches, although the bulk geochemical input data (K/Al-ratios) for the tributary catchments have been assumed constant over the modelled time-span. Therefore, weathering-limited sediment supply is able to change the composition of fluvial sediments via internal river dynamics and indirect provenance changes. The timing of long-term compositional change and the direction of changes provided a good qualitative match with measured data. This indicates that changes in weathering duration can explain the greater part of natural compositional variation. Thus, the modelling results strongly suggest that the shifts from glacial weathering-limited sediment supply to interstadial or interglacial transport-limited sediment supply cause the greater part of natural compositional variations of sediments from the River Meuse. The additional influx of loess has very likely interfered in sediment composition. Because the quantities involved have yet to be determined, this influx could not be incorporated in model simulations.

CHAPTER 7

SYNTHESIS

7 SYNTHESIS

Fluvial clastic sedimentary records constitute the natural terrestrial archives of a variety of geomorphological and environmental changes in the upstream drainage basins. The concept of extrinsic and intrinsic geomorphological thresholds (Bull, 1991) implies that fluvial response to external forcing is non-linear and fundamentally dependent on time-scale (Vandenberghe, 1995). The fluvial response to external forcing commonly involves the adjustment of the river longitudinal profile to reach or to maintain a graded, dynamic equilibrium at the base level of erosion (Bull, 1991), which is usually sea level (Schumm, 1993). Adjustment of the longitudinal profile results in the redistribution of sediment fluxes within the drainage basin, leading to compositional changes as well. The spatial dimension of the fluvial system entails that the response to external perturbations may include internal adjustments, because the lower, middle and upper reaches mutually influence their respective responses. This is defined as complex response (Schumm, 1977; Bull, 1991). Together, the concepts of non-linear fluvial behaviour and complex response demonstrate that researches performed on single time-slices, single river reaches or single sedimentary sequences do not suffice if one aims at the long-term reconstruction of fluvial response at the spatial scale of a whole landscape or fluvial drainage basin. Thus, one should preferably apply an integrated 4-D approach to reconstruct climatic responses in all Quaternary river systems for which the basic information concerning provenance (cf. source rock mineralogy), base-level (sea-level) changes, as well as tectonic and climatic setting is available. The 4-D approach involves the three-dimensional landscape and the fourth dimension of time. In this way it allows for the fluvio-systematic context, i.e. the context of fluvial response to long-term (10^3-10^5 years) external and internal forcing at the spatial scale of the entire drainage basin.

The main aim of this PhD-project was to determine whether fluvial sediments register measurable effects of climatic change on sediment composition on a time-scale of 10^3-10^5 years. To this purpose, a case study was made for the rain-fed River Meuse (Maas) draining into the southern part of the North Sea basin. The Northwest-European climate frequently changed from cold and dry periglacial conditions during glacials to humid and warm, temperate conditions during prolonged interstadials and interglacials like our present-day Holocene climate. Particularly the synergetic combination of bulk geochemical analyses and forward modelling of longitudinal profile development proved to be a powerful tool to quantify fluvial response to these climate changes. A field study of Late-glacial and Holocene residual channel infillings in the Meuse lower reach enabled the collection of empirical evidence. In this way, the composition of fluvial sediment fluxes could be reconstructed for the 15-0 ka BP time-frame. A semi 3-D forward modelling study (4-D if one includes time) provided important process-based information on the timing, frequency and magnitude of fluvial sediment fluxes. These model calculations addressed theoretical considerations to place the fieldwork study area into its fluvio-systematic context. In this way, we were able to allow for complex- response and non-linear fluvial dynamics on larger time-scales. The integration of geomorphological fieldwork with granulometrical, bulk geochemical, clay mineralogical and forward-modelling analysing techniques provided valuable insights on the downstream and long-term evolution of sediment fluxes and sediment composition in fluvial systems.

The choice to study the Late Quaternary evolution of sediment composition of the River Meuse was largely based on the fact that large-scale provenance changes could be eliminated for the last 250-ka BP. Moreover, the river drainage basin was non-glaciated and experienced a diversity of climatic conditions during the Late Quaternary. The alternation of relatively cold and dry stadial-glacial and relatively warm and humid interstadial-interglacial stages thoroughly influenced landscape development and in this way, the pathways of discharge and sediment supply to the river system. The results of the forward-modelling study confirmed many qualitative theoretical considerations (Chamley, 1989; Bull, 1991; Leeder, 1997) that long-term variations in sediment flux and composition may largely originate from the shift of glacial weathering-limited sediment supply to interglacial transport-limited sediment supply.

The presence or absence of a low to high mountain range may influence the spatial distribution of precipitation and discharge and strongly increase the activity of hillslope processes, leading to additional sediment supply in sloping areas. This will have considerable consequences for the response of the lower reaches downstream, stressing the fact that direct correlations of external (climatic) forcing and fluvial response of single river reaches on a one-to-one basis are highly unlikely (Veldkamp and Tebbens, 1999). It could be demonstrated that the tectonically induced downcutting in the Meuse upper and middle reaches (Ardennes low mountain range and terrace flight area) reverses or pauses during cold-stage climatic aggradation. The increase in discharge and stream power relative to decreasing hillslope rates and sediment supply both in space (as the river 'matures' downstream) and in time (during prolonged interstadials and interglacials), determine the major periods of episodic erosion (Schumm, 1977) in the Meuse fluvial system. Interglacial degradation very likely transforms the glacially aggraded middle reach river valleys into major climatic terraces. The same process of degradation or kinetic incision engenders the minor complex-response fill-cut terraces (*sensu* Bull, 1991) in the alluvial lower reach of the river Meuse. The current semi 4-D forward model did not allow for lateral erosion, which is necessary to provide shifts of individual channels or entire channel belts. Thus, at present the preservation potential of the major climatic terraces and minor terraces can not be evaluated yet. This problem might be elucidated in the near future by integrating the results of the longitudinal profile model (river valley gradient, eroded volume and erosional/depositional dynamics) into a 3-D finite state model describing local valley development (Veldkamp and Vermeulen, 1989, Veldkamp and Van den Berg, 1993; Tebbens et al., 1999c).

The direct and indirect climatic influence on spatial and temporal sediment redistribution described above largely determines the downstream and long-term evolution of sediment composition. Linear mixing combined with high sediment yield in the sloping Ardennes low mountain range wipes out any signal from the uppermost reaches and causes downstream tributary catchments to have progressively less impact on sediment composition. Furthermore, measured compositional data strongly suggest that enhanced weathering intensity during interglacial transport-limited sediment supply will amplify the spatio-temporal palaeoclimatic compositional signal. Therefore, it may be expected that medium to large river basins in comparable tectonic and (palaeo)climatic settings will show similar compositional variations in their lower reaches as found for the Meuse lower reach. Several reservations have to be made, however: climatic responses are unlikely to be found in the reaches of fluvial systems that are both in time and/or in space completely dominated by weathering-limited sediment supply. This especially applies to the eroding upper reaches of fluvial systems in a variety of climatic zones. Other examples are fluvial systems that are either completely situated in high relief/high erosion mountainous areas (e.g. rivers at an active plate margin, Ibbeken and Schleyer, 1991) or in very cold climatic regions in which chemical weathering processes are insignificant (cf. Matsuoka, 1995). In these situations, mechanical weathering and linear mixing of tributary catchments will dominate and might produce compositional changes, but these differences will be solely provenance-related. Likewise, the process of river capture is likely to subdue climatic signals in fluvial systems via the gain or loss of important river reaches, i.e. those reaches that drain tributary catchments with high sediment yields. Again, provenance changes will prevail in the recorded compositional signal. Other exceptions are fluvial reaches in which post-depositional overprinting and/or weathering processes have wiped out possible palaeoclimatic signals, for example in tropical river basins where the processes of chemical weathering are so intense that they will rapidly overwhelm any other compositional differentiation.

Finally, we have anticipated on the influence of large-scale aeolian loess influxes in the River Meuse drainage basin. Strong indications (although of the circumstantial-evidence type) were found that the loess component constituted an additional supply of fine-grained sediment to the river sediment load originating from weathering and subsequent saprolite and soil erosion in the drainage basin area. Aeolian influxes originating from sources outside the fluvial drainage basins have been common throughout the Weichselian period in the periglacial areas of Europe and North America (Ashley and Hamilton, 1993). The specific properties of loess (high initial clay and fine-silt content, well-sorted homogeneous lime-rich sediment, easily erodible material, easily weatherable minerals) may have provided a strong control on river (morpho)dynamics (cf. Fried, 1993) as well as on sediment composition. Thus, aeolian influxes probably have been very important in former periglacial areas, but their influence on the development of river systems has hardly been quantified yet. Therefore, it was not possible to include the loess dynamics in the present forward-modelling study. Maybe, this final remark might encourage earth scientists to hunt for the influence of aeolian influxes in the fluvial systems of (former) periglacial areas.

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SUMMARY

SUMMARY

All fluvial systems ultimately drain into alluvial basins, where the weathering products of their upstream drainage areas accumulate over a time-span varying from 10^0 to 10^6 years. Most silted-up alluvial basins are low-gradient deltas that are densely populated, because their high fertility maintains a high agricultural potential. Global warming due to increased concentrations of greenhouse gases in the Earth's atmosphere will very likely affect the quantity and quality of both water discharges and fine-grained sediment fluxes to these alluvial basins. The long-term interplay between tectonics, climate and sea level determines the frequency, the timing, the allocation and magnitude of erosional and depositional events in fluvial systems. Erosional or depositional events are the direct consequence of changes in the quantity of discharge and sediment fluxes and thus will influence the sediment composition too. Therefore, a thorough fundamental study of the response of fluvial systems to climatic change is indispensable to understand the natural erosional and depositional dynamics and to assess possible future changes in the bulk geochemical composition of sedimentary sequences. In this respect, the expression of past climate changes in the composition of fluvial sedimentary records may serve to predict future fluvial response to climate change: the past as a key to the future.

This thesis presents the research results for a case study of the River Meuse that combines geomorphological, bulk geochemical and forward modelling techniques. The study focuses on the impact of climate change on the natural composition of clastic river sediments, *in casu* variations in the bulk geochemical composition of fine-grained residual channel infillings on a temporal scale of 10^3-10^5 years. The main topic was split up into two main parts, each dealing with individual queries. **Part I** addresses the first goal of the research, namely to determine which fluvial sediments are likely to register a palaeoclimatic signal, where these sediments are found and how the climatic signal is expressed in these sediments. Earlier work directed the focus towards fine-grained sediments, because the clay and fine-silt size fractions are the most likely candidates to register evidence of changing weathering and transport pathways. Chapter 2, 3 and 4 describe the results of geomorphological fieldwork and subsequent laboratory analyses by zooming in on fine-grained deposits in the Meuse lower reach, which date to the 15-0 ka BP time-frame.

Chapter 2 describes the general geological setting of the study area that forms part of the Meuse lower reach in North Limburg (southern Netherlands). The response of the River Meuse to Late Glacial climate change was reconstructed to provide insight into the origin and age of the residual channels and their infillings. Ice-cores drilled within the framework of the Greenland Ice Core Project (GRIP) indicate frequent, profound and abrupt climatic change over the last 250,000 years Before Present (250-ka BP) in the North Atlantic region and Northwest Europe. The Late Glacial part of the GRIP ice-core demonstrated that mean annual temperature rose dramatically at the beginning of the Late Glacial (~14.5-ka BP; 5-°C) and in the Early Holocene (~10.2-ka BP; 7-°C). Quantitative climate reconstructions based on palynological and coleopteran data for the upstream Ardennes (where the major part of the River Meuse discharge and sediment load originates) and in the Netherlands corroborated these temperature rises. The Ardennes data also indicated a fourfold increase in mean annual precipitation. The rain-fed River Meuse is very sensitive to temperature and precipitation changes and accordingly responded by lowering its Late-glacial floodplain: glacially aggraded deposits were incised and several previously active channels were abandoned and turned into low-energy depositional environments. Subsequent flooding events left their traces in the form of a discontinuous clayey and silty sedimentary record within the residual channels. Periods of sediment by-pass led to gyttja and peat accumulation. This offered excellent

opportunities to date the residual channels and the intercalated clastic sediments, because the organic material has been formed amply within the dateable ^{14}C-range.

Extensive conventional ^{14}C-dating of strongly organic gyttja and peat intervals in the residual channels permitted the reconstruction of two Late-glacial phases of rapid vertical channel downcutting, preceding lateral river valley degradation. The first phase dated between ~13.3 and 12.5-ka BP (Early Bølling) and the second one between ~10.2 and 9.8-ka BP (Early Preboreal). The onset of channel downcutting lagged the climatic amelioration in the Ardennes some 500-1300 years, depending on initial landscape conditions. This suggests a major influence of interstadial and interglacial vegetation growth on landscape fixation and stabilisation and soil development, resulting in a decrease of the sediment supply relative to simultaneously increasing river discharges and hence channel downcutting. Following the downcutting phases, meandering river patterns suggest low-energy river dynamics during the Late-glacial Interstadial (Late Bølling, Allerød) and the Late Preboreal. Simultaneously, increasing landscape stability contributed to a gradual fining of the sediments. Renewed climatic deterioration around 11.3-ka ^{14}C BP (Late Allerød) and during the severe Younger Dryas cooling event (10.8-10.2 ka BP) caused a short return to high-energetic braided river conditions. All in all, the results of Chapter 2 show that the Late-glacial geological setting of the Meuse lower reach in North Limburg provided the ideal environment for deposition and preservation of fine-grained clastic sediments.

Chapter 3 zooms in on a set of ~640 samples taken from non-polluted, largely unconsolidated, fine-grained residual channel infillings deposited in the Meuse lower reach. The <2000-µm fractions have been analysed for their granulometry (determined with a Laser Grainsizer), bulk geochemistry (using X-Ray Fluorescence Spectroscopy, XRFS) and part of the samples for their clay geochemistry and clay mineralogy (X-Ray Diffraction, XRD). The natural compositional variation in these fine-grained sediments was quantified using multivariate statistical analyses. This enabled distinguishing the palaeoclimatic signal from variation that is not directly related to climate, like post-depositional overprinting effects. Hydrodynamic sorting of minerals in different size fractions explained over 70% of the variation in sediment composition. Phyllosilicates in the clay and fine-silt size fractions hosted the major part of the major constituents Al_2O_3, TiO_2, K_2O, MgO and the trace elements Ba, Co, Ce, Cr, Ga, La, Nd, Ni, Pb, Rb, V and Zn. Early-diagenetic formation of siderite and vivianite in the strongly organic and clayey anoxic gyttja environment appeared to cause relative natural accumulations of Fe_2O_3, MnO, P_2O_5, Co, Ni and notably Zn above the phyllosilicate background values. High lime contents caused elevated contents of CaO and Sr, while Na_2O and Zr, Nb, Y and Nd were more or less strongly related to the occurrence of albitic feldspars and heavy minerals respectively in the coarse silt fraction. Only 13 samples out of 636 showed strong anomalies or accumulations of the trace elements Pb, Zn, Ni and Co, which in high concentrations can pose a potential threat to the environment. This confirmed the assumption that most samples did not suffer from post-depositional anthropogenic pollution. However, bio-accumulation and very early atmospheric pollution due to small-scale ore-mining and smelting in the Roman era might explain the elevated Pb-contents found in Subatlantic clays.

Chapter 4 deals with the same set of samples and confirms theoretical considerations that the composition of fine-grained clastic sediments does not remain constant over a period of 10^3-10^4 years. The ^{14}C-dating of frequent-occurring organic intervals and additional palynological information enabled time labelling of samples taken from intercalated clastic layers. Bivariate scatterplots showed that the Pleniglacial, Late Glacial and Holocene sample groups differ considerably in their clay contents and in their contents of several main and trace constituents. Firstly, Holocene samples were found to have significantly higher clay contents, suggesting higher clay mineral supply. Secondly, Holocene samples contained more

Al_2O_3 and less K_2O, MgO and TiO_2 relative to Pleniglacial and Late-glacial specimens within a comparable granulometrical range. The typically clay-related trace elements Ba, Cr, Rb and V showed similar chronological differentiation as for the main constituents, namely lower Holocene values relative to Al_2O_3. However, these trace elements have higher ratios in Holocene samples relative to K_2O and MgO, owing to relative depletion of the latter constituents. Detailed clay mineralogical analysis of separated clay fractions and clay mineral weathering literature strongly suggested that this systematic shift in sediment composition could be ascribed to both an absolute and relative increase of the smectite and vermiculite contents and interstratifications of these minerals with illite in Late-glacial Interstadial and Holocene sediments. Because overprinting effects owing to post-depositional soil formation and anthropogenic effects could be excluded, the changes in detrital clay mineralogy have been interpreted as a systematic sedimentary palaeoclimatic signal.

Climatic amelioration and increasing landscape stability during prolonged interstadials and interglacials increased the weathering intensity (rate) of phyllosilicates and lengthened weathering duration. Widespread soil formation on Palaeozoic metapelitic rocks in the Ardennes low mountain range as well as on loess deposits has most likely caused the clays were progressively depleted of the main constituents K_2O, MgO and TiO_2 relative to Al_2O_3 and Ba, Cr, Ga, Rb and V (Chapter 4). The resulting higher supply of the typically pedogenetic high-Al, low-K and low-Mg smectites and vermiculites ultimately constituted a palaeoclimatic signal in the clay and fine silt size fractions of Meuse sediments. Several early diagenetic post-depositional processes favour the formation of authigenic minerals in the gyttja-redox environment (siderite and vivianite). They exclude the use of the following major and minor constituents for reconstructing long-term detrital compositional changes: Fe_2O_3, MnO and P_2O_5. In the same manner, variations in heavy mineral and lime content exclude the use of Nb, Y, Zr and the CaO-Sr pair respectively.

Part II encompasses the second subgoal of the research, namely to answer the questions why the sediments one is interested in have been laid down there in the first place, how they relate to long-term and large-scale fluvial dynamics and how sediment compositional changes relate to internal fluvial dynamics. Sediment composition is namely directly related to the sediment flux, which itself depends on the evolution of the fluvial system. Therefore, the long-term river dynamics had to be quantified at the spatial scale of the whole drainage basin to account for other factors than climate and to get a grip on long-term compositional changes. A forward modelling study in Chapter 5 serves to understand the 15-0 ka BP lower reach results of Chapter 2-4 within its fluvio-systematic context. This context concerns the influence of external forcing on the long-term evolution of the Meuse longitudinal profile on the spatial scale of the whole drainage basin. Chapter 6 attempts to give a finishing touch by combining sediment flux calculations with bulk geochemical data to provide insight in the long-term evolution of fluvial sediment composition.

Chapter 5 contains the interpretations concerning long-term fluvial dynamics resulting from sediment flux calculations in a well-calibrated semi 3-D forward modelling study. The longitudinal profile development of the River Meuse was simulated in response to changes in the external forcing variables tectonics, climate and sea level for the time-span 250-0 ka BP. The modelling results showed that a scenario of climate-controlled discharge and hillslope sediment supply (interstadial or interglacial increasing discharges and simultaneously decreasing hillslope sediment supply) is able to reproduce a phase of river valley degradation at the start of interglacial periods. The incisional phase is identical to the observed phase of Late-glacial incision in the Meuse lower reach, which set the favourable conditions for preservation of fine-grained sediments in the resulting residual channels (Ch. 2). This suggests that the followed forward modelling approach adequately simulates the long-term evolution of the Meuse fluvial system.

The Profile Evolution Map (Fig. 5.8, page 93) visualises the long-term evolution of sediment fluxes and demonstrates the timing and allocation of erosional and depositional phases along the longitudinal profile. The downstream positions of sections along the longitudinal profile strongly determine how they respond to time-equivalent changes in the external forcing variables. The fact that the upstream and downstream sections appeared to mutually influence each other clearly indicated complex-response in the fluvial system. Tectonic uplift and climate change could be demonstrated to dominate fluvial response in the North French and Ardennes upper and middle reaches. Here, the model predicts erosional phases to be most pronounced during prolonged interstadials and interglacials, leading to climate-controlled river valley incision and degradation of the longitudinal profile. On the other hand, continuing tectonic subsidence within the Roer Valley Graben and the southern North Sea Basin and eustatic sea-level changes dominated the response in the lowermost reaches. Here, continuous deposition takes place, interrupted by an incisional phase at the beginning of interglacials. The subsequent rising Eemian and Holocene sea levels caused increased sediment aggradation, leading to gradient backfilling. This process generated a depositional wedge that protruded progressively upstream and shifted the terrace intersection land-inward over some 100-150 km.

In **Chapter 6**, the bulk geochemical data of various upstream Ardennes tributary catchments have been coupled to the sediment supply arising from climate-controlled hillslope processes and from internal valley-erosion. The integration of bulk geochemical data with calculated sediment fluxes originating from the forward modelling study allowed simulating the effects of changes in the external forcing variables on sediment composition in the fluvial system. The Geochemical Evolution Map (Fig. 6.4, page 109) visualises the long-term evolution of sediment composition along the longitudinal profile. We modelled a scenario of changing weathering duration, namely alternating glacial weathering-limited (short weathering duration) and interglacial transport-limited sediment supply (long weathering duration. This scenario performed well in simulating the timing and direction of changes in the long-term sediment composition. Furthermore, the simulated changes in sediment composition were of the same order of magnitude as the measured changes recorded in the fine-grained sediments of the Meuse lower reach. However, a discrepancy with absolute values indicated that the direct effects of increased weathering intensity could not be excluded, especially in prolonged interstadial or interglacial periods with transport-limited sediment supply. Furthermore, circumstantial evidence indicated that loess influxes might play an important role as well, but these have yet to be quantified.

SAMENVATTING

SAMENVATTING

Ieder riviersysteem mondt uiteindelijk uit in een sedimentair bekken, waar de verweringsproducten van het bovenstroomse drainage-gebied accumuleren gedurende een tijdspanne variërende van 10^0 tot 10^6 jaar. De meeste opgeslibde sedimentaire bekkens zijn delta's met een laag verhang, die vanwege hun hoge vruchtbaarheid een hoog landbouwkundig potentieel hebben en daarom dichtbevolkt zijn. Het is zeer waarschijnlijk dat de globale opwarming - vanwege de toegenomen concentraties aan broeikasgassen in de atmosfeer van de Aarde - de kwantiteit en kwaliteit zal beïnvloeden van zowel de rivierafvoer als de fijnkorrelige sedimentlast naar deze sedimentaire bekkens. De lange-termijn interactie tussen tektoniek, klimaat en zeespiegel bepaalt hoe vaak, wanneer en waar de erosie- en sedimentatiefasen zich voordoen in het riviersysteem en tevens hoe groot zij zijn. Deze erosie en sedimentatiefasen zijn het directe gevolg van veranderingen in de rivierafvoeren en de sedimentlast en zullen dus ook de sedimentsamenstelling beïnvloeden. Daarom is een grondige, fundamentele studie van de reactie van riviersystemen op klimaatveranderingen onmisbaar om de natuurlijke erosie en sedimentatie dynamiek te begrijpen en om het belang van eventuele toekomstige veranderingen in de bulk geochemische samenstelling van rivierafzettingen te kunnen inschatten. In dit opzicht kan de manier waarop klimaatveranderingen uit het verleden tot uiting komen in de samenstelling van oude rivierafzettingen een goed begin zijn om het riviergedrag bij toekomstige klimaatveranderingen te voorspellen: het verleden als sleutel naar de toekomst.

Dit proefschrift behandelt de onderzoeksresultaten voor een gedetailleerde studie van de rivier de Maas, waarin resultaten afkomstig van geomorfologische, granulometrische- en bulkgeochemische technieken alsmede van *forward*-modellering worden geïntegreerd. De studie betreft vooral de invloed van klimaatverandering op de samenstelling van klastische rivierafzettingen en legt de nadruk op natuurlijke variaties in de bulk geochemische samenstelling van fijnkorrelige restgeulopvullingen op een tijdschaal van 10^3-10^5 jaar. Dit hoofonderwerp is opgesplitst in twee delen, die ieder aparte deelvragen behandelen.

Deel I heeft betrekking op het eerste hoofddoel van het onderzoek, namelijk om te bepalen in welke rivierafzettingen een klimaatsignaal het best bewaard is gebleven, waar deze rivierafzettingen voorkomen en hoe dat klimaatsignaal tot uiting komt in de rivierafzettingen. Eerder onderzoek vestigde de aandacht op fijnkorrelige rivierafzettingen, omdat een eventueel bewijs voor veranderende verwerings- en transportdynamiek het meest waarschijnlijk bewaard zou zijn gebleven in de klei en fijn-silt (slib) korrelgroottefracties. Hoofdstuk 2, 3 en 4 beschrijven derhalve de resultaten van het geomorfologisch veldwerk en de daaropvolgende laboratoriumanalyses. Daarbij wordt ingezoomd op het stroomafwaartse deel van de Maas en de daar voorkomende kleiïge en siltige afzettingen, die dateren van de laatste 15.000 jaar.

Hoofdstuk 2 beschrijft het algemene geologische raamwerk van het studiegebied, dat deel uitmaakt van het benedenstroomse deel van de Maas in Noord-Limburg. De reactie van de Maas op Laat-glaciale klimaatveranderingen is gereconstrueerd om inzicht te verkrijgen in de ontstaanswijze en ouderdom van de restgeulen en hun opvullingen. IJskernen die geboord werden in het kader van het Groenland IJskern Project (GRIP) lieten veelvuldige, aanzienlijke en abrupte klimaatveranderingen zien in de Noord-Atlantische regio en Noord-West Europa gedurende de laatste 250.000 jaar (250 ka BP). Het laat-glaciale deel van de GRIP-ijskern openbaarde dat de gemiddelde jaarlijkse temperatuur aanmerkelijk steeg aan het begin van het Laat-glaciaal (~14,5 ka BP; 5-°C) en tijdens het Vroeg-Holoceen (~10,2 ka BP; 7-°C). Kwantitatieve klimaatreconstructies gebaseerd op stuifmeelonderzoek in Nederland en in de bovenstroomse Ardennen (waar het grootste deel van de afvoer en sedimentlast van de Maas vandaan komt) bevestigden deze temperatuurstijgingen. De gegevens voor de Ardennen gaven ook een viervoudige toename van de gemiddelde jaarlijkse neerslag te zien. De uitsluitend door regen gevoede Maas is zeer gevoelig voor temperatuur- en neerslagveranderingen en reageerde dienovereenkomstig door haar Laat-glaciale rivierdalvlakte te verlagen. De Maas sneed zich in in haar glaciaal geaggradeerde afzettingen en verliet diverse voormalig actieve geulen, die vervolgens veranderden in laag-energetische afzettingsmilieu's. De daaropvolgende hoogwaterperioden lieten vervolgens hun sporen achter in de vorm van discontinue, kleiïge en siltige rivierafzettingen in de restgeulen. Perioden waarin nauwelijks of niets werd afgezet leidden tot gyttja en veenvorming. De organische restgeulopvullingen konden uitstekend gedateerd worden met de conventionele ^{14}C-methode, omdat het materiaal ruim binnen het dateerbare ^{14}C-bereik werd gevormd.

Door de veelvuldig voorkomende sterk organische gyttja- en veen-intervallen in de restgeulen te dateren, konden er twee Laat-glaciale fasen van relatief snelle verticale geulinsnijding gereconstrueerd worden. De eerste fase dateerde van ~13,3 tot 12,5 ka BP (Vroeg-Bølling) en de tweede van ~10,2 tot 9,8 ka BP (Vroeg Preboreaal). Deze fasen gingen vooraf aan laterale verlagingen van de rivierdalvlakte. Het begin van de geulinsnijding liep met respectievelijk 1300 tot 500 jaar achter op de klimaatverbeteringen in de Ardennen, afhankelijk van initiële landschapscondities. Dit suggereert dat de interstadiale en interglaciale vegetatiegroei een grote invloed had op de vastlegging en stabilisering van het landschap en op de bodemvorming. Deze landschapsstabilisatie moet geresulteerd hebben in geulinsnijding, vanwege een afname van de sedimentaanvoer bij een gelijktijdige toename van de rivierafvoeren. Een meer meanderend rivierpatroon volgend op deze insnijdingsfasen suggereert een laag-energetische rivierdynamiek gedurende het Laat-glaciale Interstadiaal (Laat-Bølling, Allerød) en het Preboreaal. Tegelijkertijd droeg de toenemende landschapsstabiliteit bij aan een alsmaar fijner wordende sedimentlast. Een hernieuwde klimaatverslechtering rond 11,3 ka BP (Laat-Allerød) en tijdens de sterke afkoelingsfase van de Late Dryas (10,8-10,15 ka BP) veroorzaakte een korte terugkeer naar een hoog-energetisch verwilderd rivierpatroon. Samenvattend laten de resultaten in Hoofdstuk 2 zien dat de Laat-glaciale rivierdynamiek in het benedenstroomse deel van de Maas in Noord-Limburg het ideale milieu creëerde voor de afzetting en het behoud van fijnkorrelige klastische sedimenten.

Hoofstuk 3 gaat in op een set van 636 monsters, die genomen werden van niet vervuilde, grotendeels ongerijpte, fijngetextureerde restgeulopvullingen die werden afgezet in het benedenstroomse deel van de Maas. De <2000 μm fracties werden onderzocht op hun korrelgrootteverdeling (bepaald met de *Laser Grainsizer*) en op hun bulk geochemie (met Röntgen Fluorescentie Spectroscopie, XRFS). Van een deel van de monsters werd tevens de kleigeochemie en kleimineralogie bepaald (met Röntgendiffractie, XRD). De natuurlijke variatie in de samenstelling van fijnkorrelige afzettingen werd met behulp van multivariate statistiek gekwantificeerd. Dit gebeurde om een eventueel paleoklimaatsignaal te kunnen onderscheiden van variatie die niet direct aan klimaat gerelateerd is, zoals processen die de samenstelling van het sediment ná afzetting beïnvloeden. Hydrodynamische sortering van mineralen in verschillende korrelgroottefracties kon ongeveer 70% van de natuurlijke variatie in sedimentsamenstelling verklaren. Phyllosilicaten in de klei en fijn-silt korrelgroottefracties bevatten het grootste deel van de hoofdcomponenten Al_2O_3, TiO_2, K_2O, MgO en de spore-elementen Ba, Co, Ce, Cr, Ga, La, Nd, Ni, Pb, Rb, V and Zn. Vroeg-diagenetische sideriet- en vivianietvorming in het sterk organische en soms kleiïge anoxische gyttja milieu bleken natuurlijke relatieve aanrijkingen te veroorzaken van vooral Fe_2O_3, MnO, P_2O_5, Co, Ni en Zn boven de phyllosilicaat-achtergrondwaarden. Hoge kalkgehalten veroorzaakten verhoogde gehalten aan CaO en Sr, terwijl Na_2O en Zr, Nb, Y en Nd min of meer sterk gerelateerd waren aan respectievelijk albitische veldspaten en zware mineralen in de grof-silt fractie. Slechts 13 van de 636 monsters toonden sterke afwijkingen of aanrijkingen van de spore-elementen Pb, Zn, Ni en Co, die in hoge concentraties potentieel toxisch kunnen zijn voor het milieu. Dit bevestigde de aanname dat de meeste monsters na hun afzetting niet door menselijke invloed waren vervuild. Echter, bio-accumulatie en zeer vroege atmosferische vervuiling tijdens de mijnbouw en het smelten van ertsen door de Romeinen zou de oorzaak kunnen zijn van verhoogde loodgehalten in Subatlantische kleien.

Hoofdstuk 4 behandelt dezelfde set monsters en bevestigt theoretische aannames dat de samenstelling van fijnkorrelige sedimenten niet constant blijft over een periode van 10^3-10^4 jaar. De ^{14}C-datering van veelvuldig voorkomende organische intervallen en additionele palynologische informatie maakte het mogelijk om de monsters die genomen waren van tussengeschakelde klastische lagen in de tijd te plaatsen. Bivariate *scatterplots* toonden aan dat de Pleniglaciale, Laat-glaciale en Holocene monstergroepen aanmerkelijk verschilden in hun kleigehalten en in hun gehalten aan hoofdcomponenten en spore-elementen. Allereerst bleken de Holocene monsters significant meer klei te bevatten, hetgeen suggereert dat het aanbod aan klei in deze monsters groter was. Ten tweede bleken Holocene monsters meer Al_2O_3 en minder K_2O, MgO en TiO_2 te bevatten ten opzichte van Pleniglaciale en Laat-glaciale monsters met een vergelijkbare korrelgrootteverdeling. De typische klei-gerelateerde spore-elementen Ba, Cr, Rb en V lieten een vergelijkbare chronologische differentiatie zien als de hoofdcomponenten, namelijk lagere waarden ten opzichte van Al_2O_3 in Holocene monsters. Echter, deze spore-elementen hadden hogere ratio's ten opzichte van K_2O en MgO, ten gevolge van relatieve depletie van de laatstgenoemde hoofdcomponenten. Gedetailleerde kleimineralogische en geochemische analyses van afgescheiden kleifracties suggereerden dat deze systematische verschuiving in sedimentsamenstelling in sterke mate kan worden toegeschreven aan zowel een absolute als relatieve toename van de smectiet- en vermiculiet-gehalten in Laat-glaciale Interstadiale en Holocene sedimenten. Omdat de sedimenten na hun afzetting zeker niet door bodemvorming of anthropogene vervuiling zijn beïnvloed, zijn deze veranderingen in de detritische kleimineralogie geïnterpreteerd als een systematisch sedimentair paleoklimaatsignaal.

Klimaatverbetering en toenemende landschapsstabiliteit gedurende langdurige inter-stadialen en interglacialen verhogen de verweringsintensiteit (snelheid) van phyllosilicaten en verlengt de verweringsduur. Uitgebreide bodemvorming in Paleozoïsche metapelitische gesteenten (Ardennen) en in de lössafzettingen tijdens het Bølling/Allerød-Interstadiaal en het Holoceen heeft zeer waarschijnlijk veroorzaakt dat de kleien in de loop der tijd relatief verarmd raakten aan de hoofdcomponenten K_2O, MgO en TiO_2 ten opzichte van Al_2O_3 en Ba, Cr, Ga, Rb en V. Het hogere aanbod van typisch pedogenetische Al-rijke, K-arme en Mg-arme smectieten en vermiculieten dat daar het gevolg van was, creëerde zo een paleoklimaatsignaal in de klei en fijn-silt korrelgroottefracties in de Maassedimenten. Verschillende post-sedimentaire vroeg-diagenetische processen begunstigen de vorming van authigene mineralen (sideriet en vivianiet) in het door redoxprocessen beïnvloede gyttja-milieu. Deze sluiten het gebruik uit van de componenten Fe_2O_3, MnO en P_2O_5, als men lange-termijn veranderingen in de detritische sedimentsamenstelling wil reconstrueren. Op dezelfde manier sluiten variaties in zware mineralen en kalkgehalte het gebruik uit van respectievelijk Zr, Nb, Y, Nd en het CaO-Sr paar.

Deel II omvat het tweede doel van het onderzoek, namelijk om te achterhalen waarom de fijnkorrelige sedimenten juist in Noord Limburg zijn afgezet, hoe dat in verhouding staat tot lange-termijn en grootschalige rivierdynamiek en hoe veranderingen in de sedimentsamen-stelling samenhangen met interne rivierdynamiek. De sedimentsamenstelling hangt namelijk direct samen met de sedimentflux, die op zijn beurt weer afhangt van de ontwikkeling van het riviersysteem. Om rekening te kunnen houden met andere factoren dan klimaat en om greep te krijgen op lange-termijn veranderingen in de sedimentsamenstelling moest de lange-termijn rivierdynamiek gekwantificeerd worden op de spatiële schaal van het hele drainagegebied.

De *forward*-modelleringsstudie in Hoofdstuk 5 dient om de resultaten van de periode 15-0 ka BP en het benedenstroomse deel van de Maas te begrijpen binnen de fluviatiel-systematische context. Deze behelst de invloed van externe sturing op de lange-termijn en grootschalige ontwikkeling van het longitudinale profiel van de Maas. In Hoofdstuk 6 worden de berekende sedimentfluxen gekoppeld aan bulk geochemische gegevens, teneinde inzicht te verschaffen in de lange-termijn ontwikkeling van de riviersedimentsamenstelling.

Hoofdstuk 5 bevat de interpretaties betreffende lange-termijn rivierdynamiek, gebaseerd op berekeningen van de sedimentflux in een uitgebreid gecalibreerde semi 3-D *forward*-modelleringsstudie. De ontwikkeling van het longitudinale profiel van de Maas werd gesimuleerd als resultante van veranderingen in de externe sturende variabelen tektoniek, klimaat en zeespiegel voor de tijdsspanne 250-0 ka BP. De "profielontwikkelings-kaart" (Fig. 5.8, pagina 91) visualiseert de lange-termijn ontwikkeling van de sedimentfluxen en laat zien waar en wanneer erosie- en sedimentatiefasen plaatsvinden langs het longitudinale profiel. De modelleringsresultaten toonden aan dat een scenario waarin zowel de rivierafvoer als het aanbod van hellingmateriaal klimaatgecontroleerd was (met interstadiaal of interglaciaal toenemende afvoeren en gelijktijdig afnemend aanbod van hellingmateriaal), in staat was om een degradatie ofwel verlaging van de rivierdalvlakte te reproduceren aan het begin van interglaciale perioden. Deze insnijdingsfase komt overeen met de geobserveerde Laat-glaciale insnijdingsfase in de benedenloop van de Maas. Deze vormde de restgeulen en creëerde daarmee gunstige omstandigheden voor het bewaard blijven van fijnkorrelige sedimenten (Hfst. 2). Dit suggereert dat de gevolgde benadering van *forward*-modellering de lange termijn ontwikkeling van het Maas-riviersysteem adequaat simuleert.

De stroomafwaartse posities van secties langs het longitudinale profiel bepalen in sterke mate hoe zij reageren op gelijktijdige veranderingen in de externe sturende variabelen. Het feit dat de stroomopwaartse en stroomafwaartse delen elkaar wederzijds bleken te beïnvloeden, was een duidelijke aanwijzing voor het optreden van *complex-response* in het riviersysteem. Aangetoond kon worden dat tektonische opheffing van Noord Frankrijk en de Ardennen samen met klimaatveranderingen het riviergedrag domineren in de bovenstroomse en middenstroomse delen van de Maas. Hier voorspelt het model dat de erosiefasen het meest intensief zijn tijdens langdurende interstadialen en interglacialen, hetgeen leidt tot een klimaatgecontroleerde verlaging van het longitudinale profiel en dus versnijding van de rivierdalvlakte. Daarentegen domineerden de continue daling van de Roerdalslenk en het zuidelijke Noordzeebekken samen met eustatische zeespiegelbewegingen het riviergedrag in de meest stroomafwaartse delen. Hier vindt continue sedimentatie plaats, onderbroken door een klimaatgestuurde insnijdingsfase aan het begin van interglacialen. De daaropvolgende hoge zeespiegels tijdens het Eemien en het Holoceen veroorzaken versnelde aggradatie en leiden tot *gradient-backfilling*. Dit proces genereert een sedimentaire wig die naarmate de zeespiegel stijgt, zijn invloed steeds verder stroomopwaarts doet gelden en daarbij de terrassenkruising eveneens zo'n 100 tot 150 km landinwaarts verschuift.

In **Hoofstuk 6** zijn de bulk geochemische gegevens van verschillende stroomgebieden van de zijrivieren in de Ardennen gekoppeld aan het sediment-aanbod afkomstig van klimaatgestuurde hellingprocessen en van interne rivierdal-erosie. De integratie van bulkgeochemische gegevens met de berekende sedimentfluxen in de *forward*-modelleringsstudie maakte het mogelijk om de effecten van veranderingen in de externe sturende variabelen op de sedimentsamenstelling te simuleren. De geochemische ontwikkelingskaart (Fig. 6.4, pagina 107) visualiseert deze lange-termijn ontwikkeling van de sedimentsamenstelling langs het longitudinale profiel. De modelsimulaties kwamen overeen met grootschalige en lange-termijn klimaatgestuurde veranderingen in de verweringsduur, namelijk een afwisselend glaciaal verwering-gelimiteerd sediment aanbod (korte verweringsduur) en interglaciaal transport-gelimiteerd sediment aanbod (lange verweringsduur). Dit scenario voldeed goed in de simulatie van de *timing* en richting van de veranderingen in de lange-termijn sedimentsamenstelling. De gesimuleerde veranderingen in de sedimentsamenstelling waren verder van dezelfde ordegrootte als de gemeten veranderingen in de fijnkorrelige sedimenten van de benedenloop van de Maas. Echter, een discrepantie met de absolute waarden gaf aan dat de invloed van een toenemende verweringsintensiteit niet kon worden uitgesloten, vooral tijdens langdurende interstadialen en interglacialen met een transport-gelimiteerd sediment-aanbod. Daarnaast gaf indirect bewijs aan dat de invloed van löss influxen in het riviersysteem wel eens zeer belangrijk zouden kunnen zijn, maar deze moeten vooralsnog worden gekwantificeerd.

CURRICULUM VITAE

Leonardus Antonius Tebbens was born on May 11, 1967 in Heemskerk, The Netherlands. After having completed secondary education (VWO) at the Bonhoeffer College in Castricum, he started his study Soil Science in August 1986 at the Wageningen Agricultural University. In the mid-eighties, the research from the Soil Science and Ecopedology group on the topic of soil acidification was at its hayday. Therefore, his first MSc-research was an investigation to the influence of acid atmospheric deposition on the behaviour and potential buffering capacity of Fe-oxides in the soil environment.

Triggered by the contents and enthusiastic teaching of subjects on geology and landscape development, his interest shifted more and more towards soil formation within its physical-geographic context. From May to September 1990, he had a very pleasant time in the magnificent landscape of southern Ticino, Switzerland. Here, together with a fellow-student, he investigated how the site factors bedrock, vegetation, climate and soil profile exposition determined the formation of so-called cryptopodzolic soils and how these soils fitted into the mainstream definition of standard podzolic soils. The results of the research were published in the European Journal of Soil Science. While thinking "saving the best for last", his last MSc-research was a major subject on Geology and Mineralogy. He took great pleasure in patiently handpicking, quantifying and interpreting the compositional variation of some 1600 detrital garnets in sediments of the Rhine, Meuse and Baltic River fluvial systems.

Following graduation in August 1992, he was employed as a junior researcher from October until December 1992 at the Department of Soil Science and Geology to publish the encouraging results on the detrital garnets. In April 1993, he accepted the offer to carry out the PhD-research project: "Variability of fluvial sediment composition in different time scales: Assessing the impact of climatic change" for the Dutch Organisation of Scientific Research (NWO). After some delay of bureaucratic nature, he could start the very interesting research at the Laboratory of Soil Science and Geology at the Wageningen Agricultural University in November 1993. The efforts of the subsequent period have resulted in the present thesis.

In the near future, he will be employed part-time at the Department of Physical Geography of Utrecht University on the post-doc project: "High-resolution measurement of vertical accretion rates in mangrove ecosystems (Usumacinta-Grijalva delta, Mexico) by means of AMS ^{14}C wiggle matching". If all things work out fine, he will combine this job for one year with upgrading the education in geomorphology and landscape processes at the Laboratory of Soil Science and Geology at Wageningen Agricultural University.