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Late Quaternary evolution of rivers, lakes and peatlands in northeast Germany reflecting past climatic and human impact – an overview

Knut Kaiser, Sebastian Lorenz, Sonja Germer, Olaf Juschus, Mathias Küster, Judy Libra, Oliver Bens, Reinhard F. Hüttl

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Abstract: Knowledge of regional palaeohydrology is essential for understanding current environmental issues, such as the causes of recent hydrologic changes, impacts of land use strategies and effectiveness of wetland restoration measures. Even the interpretation of model results on future impacts of climatic and land-cover changes may be improved using (pre-)historic analogies. An overview of palaeohydrologic findings of the last c. 20,000 years is given for northeast Germany with its glacial landscapes of different age. River development is examined with a focus on valley(-floor) formation and depositional changes, river course and channel changes, and palaeodischarge/-floods. Major genetic differences exist among 'old morainic' (Elsterian, Saalian) and 'young morainic' (Weichselian) areas, and among topographically high- and low-lying valleys, the latter of which are strongly influenced by water-level changes in the North and Baltic Seas. Lake development was analysed with respect to lake formation, which was predominantly driven by late Pleistocene to early Holocene dead-ice dynamics, and with respect to depositional changes. Furthermore, lake-level changes have been in the focus, showing highly variable local records with some conformity. The overview on peatland development concentrated on phases of mire formation and on long-term groundwater dynamics. Close relationships between the development of rivers, lakes and peatlands existed particularly during the late Holocene by complex paludification processes in large river valleys. Until the late Holocene, regional hydrology was predominantly driven by climatic, geomorphic and non-anthropogenic biotic factors. Since the late Medieval times, human activities have strongly influenced the drainage pattern and the water cycle, for instance, by damming of rivers and lakes, construction of channels and dikes, and peatland cultivation. Indeed, the natural changes caused by long-term climatic and geomorphic processes have been exceeded by impacts resulting from short-term human actions in the last c. 50 years as discharge regulation, hydromelioration and formation of artificial lakes.

Die spätquartäre Entwicklung von Flüssen, Seen und Mooren in Nordostdeutschland als Spiegel klimatischer und anthropogener Einflüsse – eine Übersicht

Kurzfassung: Die Kenntnis der regionalen Paläohydrologie ist eine wesentliche Grundlage für das Verständnis aktueller Umweltfragen, wie zum Beispiel nach den Gründen von hydrologischen Veränderungen, dem Einfluss von Landnutzungsstrategien und der Wirksamkeit von Renaturierungsvorhaben in Feuchtgebieten. Auch die Interpretation von Modellierungsergebnissen zu den künftigen Einflüssen des Klima- und Landnutzungswandels auf das Gewässersystem kann durch die Einbeziehung (prä-) historischer Analogien verbessert werden. Für das glazial geprägte nordostdeutsche Tiefland wurde eine Übersicht der vorliegenden paläohydrologischen Befunde für den Zeitraum der letzten etwa 20.000 Jahre erarbeitet. Die Entwicklung der Flüsse wurde mit Blick auf die Tal-/Auengenese und das Ablagerungsmilieu, die Veränderung des Tal- und Gerinneverlaufs sowie den Paläoabfluss bzw. das Paläohochwasser betrachtet. Wesentliche genetische Unterschiede bestehen zwischen Alt- (Elster- und Saalekaltzeit) und Jungmoränengebieten (Weichselkaltzeit) sowie zwischen hoch und tief gelegenen Tälern. Letztere sind stark durch Wasserspiegelveränderungen in der Nord- und Ostsee beeinflusst worden. Die Entwicklung der Seen wurde hinsichtlich der Seebildung, die überwiegend eine Folge der spätpleistozänen bis frühholozänen Toteistieftau-Dynamik ist, und der Veränderungen im Ablagerungsmilieu analysiert. Weiterhin standen Seespiegelveränderungen im Fokus, wobei sich hoch variable lokale Befunde mit einigen Übereinstimmungen zeigten. Der Überblick zur Moorentwicklung konzentrierte sich auf hydrogenetische Moorentwicklungsphasen und auf die langfristige Entwicklung des Grundwasserspiegels. Enge Beziehungen zwischen der Entwicklung der Flüsse, Seen und Moore bestanden insbesondere im Spätholozän durch komplexe Vermoorungsprozesse in den großen Flusstälern. Bis in das Spätholozän wurde die regionale Hydrologie überwiegend durch klimatische, geomorphologische und nicht-anthropogene biologische Faktoren gesteuert. Seit dem Spätmittelalter wurde in der Region das Gewässernetz und der Wasserkreislauf im starken Maß durch anthropogene Interventionen beeinflusst (z.B. Aufstau von Flüssen und Seen, Bau von Kanälen und Deichen, Moorkultivierung). In den letzten etwa 50 Jahren haben dann sogar die kurzfristigen anthropogenen Eingriffe, z.B. in Form von Abflussregulierung, Hydromelioration und künstlicher Seebildung, die Wirksamkeit langfristiger klimatischer und geomorphologischer Prozesse übertroffen.

Keywords: *palaeohydrology, valley formation, depositional change, lake- and groundwater-level fluctuation, mire, late Pleistocene, Holocene*

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Contents

1 Introduction

Global climate change causes regional and local variations in the terrestrial water balance (e.g. Tao et al. 2003, IPCC 2007, Bates et al. 2008, Gerten et al. 2008, Kundzewicz et al. 2008, Huang et al. 2010), influencing the hydrologic, geomorphic and ecologic properties of the regional drainage system comprised of flowing (rivers, streams) and stagnant waters (lakes, ponds) as well as peatlands of varying dimension. An aridification trend, for example, will inevitably cause a reduction (1) in the discharge of rivers by diminishing supply, (2) in the size of lakes by level lowering and (3) in the extension of peatlands by groundwater lowering.

As hydrologic and climatic research in Europe shows, there are currently distinct changes in water balances with regionally differing trends (e.g. Lehner et al. 2006, BACC Author Team 2008, EEA 2009, Merz et al. 2012). In northeast Germany widely a 'drying' trend prevails, resulting in decreasing groundwater and lake levels as well as river discharges (e.g. Gerstengarbe et al. 2003, Kaiser et al. 2010, 2012a, Germer et al. 2011). If this trend continues, a negative influence on ecosystem services, such as the provision of water for human use and wetland conservation, is to be feared.

Undoubtedly, the knowledge of both historic hydrologic (last c. 1000 years) and palaeohydrologic developments can help us to understand the hydrologic system dynamics at present and even in the future (e.g. Branson et al. 1996, Gregory & Benito 2003, Brázdil et al., 2006, Gregory et al. 2006, Czymzik et al. 2010). In particular, the frequency and magnitude of short-term events, such as river floods and droughts, as well as long-term processes, such as lake-level fluctuations, changes in the river's mean annual discharge and its hydromorphologic status can be detected retrospectively (e.g. Petts et al. 1989, Berglund et al. 1996a, Harrison et al. 1998, Brown 2002, Starkel 2005, Baker 2008, Battarbee 2010). Insights gained through such historic analogies can be used to improve the interpretation of modelled future impacts of climatic and land-cover changes and, hence, to develop and optimise adaptation strategies. Furthermore, information on the pre-modern ecologic status of aquatic landscapes is a precondition for developing restoration measures in accordance with the European Union Water Framework Directive (CEC 2000, Bennion & Battarbee 2007, Zerbe & Wiegleb 2009).

In theory, palaeohydrology is concerned with all components of the hydrologic cycle. But in practice most research focuses on specific compartments, such as river channels and discharge, lake- and groundwater-level fluctuations, isotope chemistry, or on proxy indicators of past precipitation characteristics (Anthony & Wohl 1998, Gregory & Benito 2003). Such knowledge on the palaeohydrology of temperate regions in the world is well-established. Particularly western and central Europe have a long-standing research tradition (e.g. Starkel et al. 1991, Gregory 1995, Hagedorn 1995, Vandenberghe 1995a, Starkel 2003, Macklin et al. 2006, Hoffmann et al. 2008). However, stronger integration between the regional findings as well as with related disciplines is necessary.

In northeast Germany, there are well-structured scientific communities dealing with both present-day and future hydrologic changes (investigated by hydrologists and climate impact researchers) as well as with palaeohydrology (investigated by geoscientists and palaeoecologists). Unfortunately joint investigations by both communities are lacking. In addition, existing palaeohydrologic knowledge is not sufficiently being considered in the interpretation of (pre-)recent hydrologic trends and prospective (modelling) purposes. Obstacles to the exploitation of hydrologic palaeo-data are the multitude of local case studies, and their prevailing publication in German periodicals and monographs with a regional or national focus. Publications synthesising regional paleohydrologic results are rare.

This overview offers access on the results of regional palaeohydrologic research over the last c. two decades. The consolidation of findings into one paper will hopefully foster the consideration of (pre-)historic hydrologic changes into the respective discussions, increasing the interpretational power for modelling results. This paper primarily focuses on the evolution of drainage systems during the last c. 20,000 years, spanning the late Pleistocene and the Holocene epochs. The long-term and partly interdependent development of the region's main aquatic *inland* environments – rivers, lakes and peatlands – will be outlined. For several specific issues (e.g. river valley formation, palaeodischarge characteristics, dead-ice dynamics, lake- and groundwater-level changes, peatland formation), the state-of-the-art will be reported.

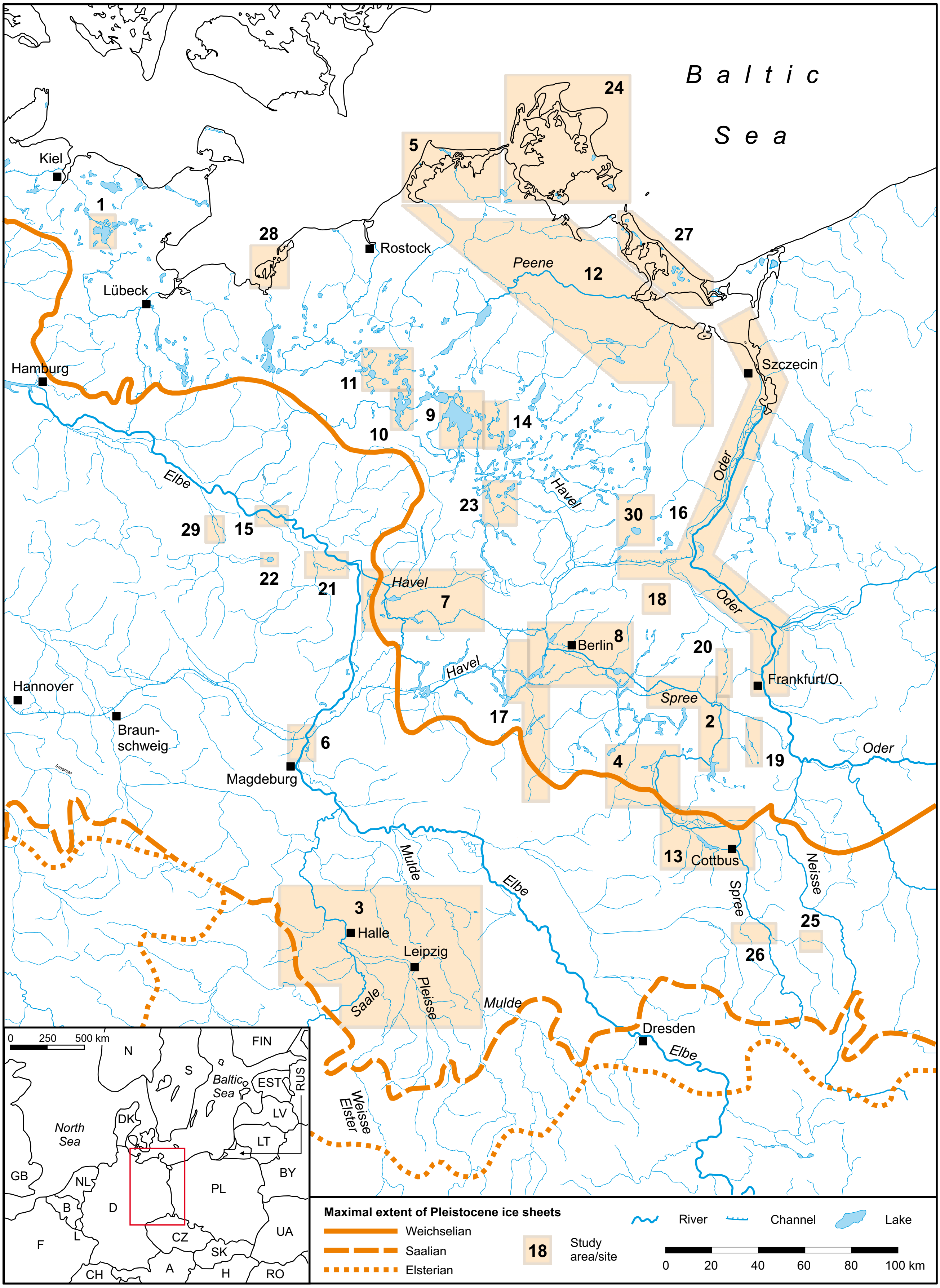

Fig. 1: Hydrography, main glacial structures and study areas/sites with palaeohydrologic findings in northeast Germany (map after BMUNR 2003, adapted). The numbers refer to the study areas/sites presented (see Tab. 1).

Abb. 1: Hydrografie, glaziale Hauptstrukturen (Marginalzonen) und Arbeitsgebiete/-orte mit paläohydrologischen Befunden in Nordostdeutschland (Karte nach BMUNR 2003, verändert). Die Zahlen beziehen sich auf die vorgestellten Arbeitsgebiete/-orte (siehe Tab. 1).

Tab. 1: Study areas and sites with palaeohydrologic findings in northeast Germany (see Fig. 1).

Tab. 1: Arbeitsgebiete und -orte mit paläohydrologischen Befunden in Nordostdeutschland (siehe Abb. 1).

No.	Study area / site	Research field[1]	References
1	Lake Plöner See	LB, LL, NT, GA, PL, PE	Sirocko et al. 2002, Dörfler 2009
2	Lower Spree River	FM, PD, PE	Schulz & Strahl 1997, Schulz 2000, Schönfelder & Steinberg 2004, Hilt et al. 2008
3	Leipzig-Halle area	LB, LL, GA, PL, FM, PE, GG, HI	Hiller et al. 1991, Mania et al. 1993, Wolf et al. 1994, Mol 1995, Böttger et al. 1998, Fuhrmann 1999, Tinapp et al. 2000, 2008, Eissmann 2002, Wennrich et al. 2005, Czegka et al. 2008
4	Lower Spree River, lower Spreewald area and Dahme River	FM, LB, PE, GG	Böttner 1999, Juschus 2002, 2003
5	Darss peninsula, Barthe River and Endinger Bruch basin	LB, LL, GA, PL, FM, PT, CE, PE	Kaiser 2001, 2004a, de Klerk 2002, Kaiser et al. 2000, 2006, 2007, Lampe 2002, Lane et al. 2012
6	Elbe River N of Magdeburg	FM, HI	Rommel 1998
7	Lower Havel River, Elb-Havel-Winkel and Rhinluch/Havelländisches Luch areas	PT, GG, PE, GA, LL, GW, PL, HI	Mundel et al. 1983, Kloss 1987a, 1987b, Mundel 1995, 1996, 2002, Schelski 1997, Küster & Pötsch 1998, Rowinsky & Rutter 1999, Gudermann 2000, Mathews 2000, Zeitz 2001, Gramsch 2000, Kaffke 2002, Weisse 2003, Schönfelder & Steinberg 2004
8	Berlin area	LB, LL, GW, GA, PL, FM, PE, GG, PT, HI	Böse & Brande 1986, 2009, Pachur & Röper 1987, Brande 1986, 1988, 1996, Gärtner 1993, Schich 1994, Uhlemann 1994, Varlemann 2002, Grünert 2003, Kossler 2010, Neugebauer et al. 2012
9	Lake Müritz	LL, PL, PE, HI	Kaiser 1998, Kaiser et al. 2002, Ruchhöft 2002, Lampe et al. 2009
10	Lake Plauer See	LL, GA, HI	Ruchhöft 2002, Bleile et al. 2006, Bleile 2008
11	Nossentiner/Schwinzer Heide area	LB, LL, PE, PL, FM, HI	Schmidtchen et al. 2003, Lorenz 2003, Rother 2003, Hübener & Dörfler 2004, Lorenz & Schult 2004, Kaiser et al. 2007, Lorenz 2007, 2008, Lorenz et al. 2010
12	Low-lying river valleys of Vorpommern (e.g. Recknitz, Peene and Uecker River)	FM, LB, LL, GW, PE, GG, CE, PT, GA, HI	Kaiser & Janke 1998, Helbig 1999, Kaiser et al. 2000, 2003, Michaelis 2000, Schatz 2000, Helbig & de Klerk 2002, Janke 2002, 2004, de Klerk 2004, Kaiser 2004b, Berg 2005, Krienke et al. 2006, Michaelis & Joosten 2010, Jantzen et al. 2011, Küster et al. 2011
13	Upper Spreewald and Cottbus areas	FM, GA, GG, PE, PT, HI	Kühner et al. 1999, Neubauer-Saurer 1999, Rolland & Arnold 2002, Woithe 2003, Poppschötz & Strahl 2004, Brande et al. 2007
14	Headwaters of Havel River	LB, LL, PE, PL, FM, HI	Kaiser & Zimmermann 1994, Küster 2009, Küster & Kaiser 2010, Küster et al. 2012
15	Lower Elbe River at Lenzen	FM, HI, GA	Schwartz 1999, Schatz 2011
16	Lower Oder River, Oderbruch area, Stettiner Haff (Szczecin Lagoon), Eberswalder Urstromtal (spillway)	FM, GG, PL, PE, CE, PT, PD, HI	Dobracka 1983, Brose 1994, 1998, Schlaak et al. 2003, Borówka et al. 2005, Carls 2005, Dalchow & Kiesel 2005, Schlaak 2005, Lutze et al. 2006, Börner 2007
17	Potsdam area, Havel and Nuthe Rivers	LB, GW, PL, FM, PE, GG, PT, HI	Rowinsky 1995, Weisse et al. 2001, Wolters 2002, 2005, Hickisch 2004, Hickisch & Päzolt 2005, Lüder et al. 2006, Kirilova et al. 2009, Enters et al. 2010
18	Biesenthal Basin, upper Finow Stream	LB, PE, GG	Chrobok & Nitz 1987, 1995, Nitz et al. 1995
19	Schlaube Stream	PL, LB, PE	Schönfelder et al. 1999, Brose 2000, Giesecke 2000
20	Kersdorfer Rinne (tunnel valley)	LB, GG, PE	Schulz & Brose 2000, Schulz & Strahl 2001
21	Wische area (lower Elbe River)	FM	Caspers 2000
22	Lake Arendsee	PL, PE, HI	Scharf 1998, Scharf et al. 2009
23	Lake Stechlinsee, Upper Rhin River	LB, FM, PL, PE	Gärtner 2007, Brande 2003, Kaiser et al. 2007
24	Rügen Island and adjacent coastal and land areas	LB, GW, NT, GA, PL, PE, GG, PE, GG, CE, PT	Kliewe 1989, Strahl & Keding 1996, Helbig 1999, de Klerk et al. 2001, Krienke 2003, Verse 2003, Hoffmann & Barnasch 2005, Hoffmann et al. 2005, de Klerk et al. 2008a, 2008b, Kossler & Strahl 2011
25	Weisser Schöps River (Reichwalde area)	FM, PT, GW, GA, PE	Friedrich et al. 2001, van der Kroft et al. 2002
26	Upper Spree River (Nochten/Scheibe area)	FM, GG	Mol 1997, Mol et al. 2000, Hiller et al. 2004
27	Usedom Island	LB, NT, PL, PE, GG, CE	Helbig 1999, Hoffmann et al. 2005
28	Poel Island and adjacent coastal and land areas	CE, NT, PE, GA, CE	Lampe et al. 2005, 2010
29	Jeetzel River	FM, PE, GA	Turner 2012
30	Schorfheide area	LB, PE, PT	Schlaak 1997, Stegmann 2005, van der Linden et al. 2008

[1]LB = Lake-basin formation, LL = Lake level, GW = Groundwater level, NT = Neotectonic, GA = Geoarchaeology, PL = Palaeolimnology, FM = Fluvial geomorphology / valley formation, PD = Palaeodischarge, PE = Palaeoecology, GG = Glacial geomorphology / geology, CE = Coastal evolution, PT = Peatland evolution, HI = Human impact on inland waters

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Tab. 2: Conceptual models of late Quaternary river valley and lake basin development in northeast Germany. Adapted and modified from Marcinek & Brose (1972) and Nitz (1984).

Tab. 2: Konzeptionelle Modelle der spätquartären Flusstal- und Seebeckenentwicklung in Nordostdeutschland (nach Marcinek & Brose 1972 und Brose 1984, verändert).

Chronology	**Phases of river valley genesis** (Marcinek & Brose 1972)	**Phases of (lake-) basin genesis** (Nitz 1984)
Late Holocene (0–4 kyrs BP)	**Holocene phase influenced by man** *('Anthropogen beeinflusste, holozäne Phase')* • strong human influence on the drainage system by channels, weirs, hydro amelioration and agriculture	**Colluvial phase** *('Kolluviumsphase')* • man-induced filling up of smaller depressions by colluvial sediments (hillwash)
Mid-Holocene (4–8 kyrs BP)	**Natural Holocene phase** *('Natürlich holozäne Phase')* • weak fluvial erosion and accumulation	**Aggradation phase** *('Verlandungsphase')* • filling up of lake basins by sedimentation of gyttja and peat
Early Holocene, Lateglacial (8–13 kyrs BP)	**Lateglacial-Early Holocene transitional phase** *('Spätglazial-altholozäne Übergangsphase')* • reversals of flow direction • partly formation of interior drainage • melting of stagnant ice / lake formation • decay of permafrost	**Deep melting phase** *('Tieftauphase')* • melting of stagnant ice, formation of (lake) basins • decay of permafrost
Late Pleniglacial (20–13 kyrs BP)	**Fluvial periglacial phase** *('Fluvioperiglaziäre Phase')* • formation of a hierarchic river system on permafrost	**Conservation phase** *('Konservierungsphase')* • conservation of stagnant ice by permafrost • sedimentation of periglacial lacustrine, fluvial and aeolian deposits
Late Pleniglacial (>20–14 kyrs BP)	**Fluvioglacial Phase** *('Fluvioglaziäre Phase')* • initial ice-marginal drainage, later ice-radial drainage • outwash plain formation	**Ice-melting phase** *('Niedertauphase')* • inclusion / burial of stagnant ice by sediments
		Formation phase *('Anlagephase')* • formation of depressions by ice exaration and glaciofluvial erosion

2 Regional settings

The region northeast Germany is part of the North European Plain, which is bounded by the coasts of the North Sea and Baltic Sea to the north and the German Central Uplands to the south. The surface relief (<200 m a.s.l.) varies from flat to undulating. Several Quaternary glaciations of Scandinavian ice sheets, subsequent periglacial shaping and interglacial processes have formed this area. A multitude of ice terminal zones of the Saalian and Weichselian glaciations traverse the region and reflect the glaciation/deglaciation (Fig. 1). The complex glacial and interglacial processes produced a mosaic of glacial, fluvial, lacustrine, colluvial, marine and aeolian landforms and sediments.

The Weichselian glacial belt ('young morainic area') covers the northern area and comprises landscapes with an immature *river* system that developed following the last deglaciation (c. 24,000–17,000 cal yrs BP; Böse 2005, Lüthgens & Böse 2011). River valleys in that belt are characterised by frequently alternating degradational (erosion) and aggradational (accumulation) river stretches, by frequent shifts in direction, by the common presence of lake basins (partly within the valley floors) and by frequent areas with interior drainage. By contrast, the river system of the Elsterian (c. >330 kyrs) and Saalian belts (c. >125 kyrs; 'old morainic areas') is maturated. Major rivers in the region are the Elbe and Oder which drain northeast Germany into the North Sea and the Baltic Sea, respectively. These rivers are character-

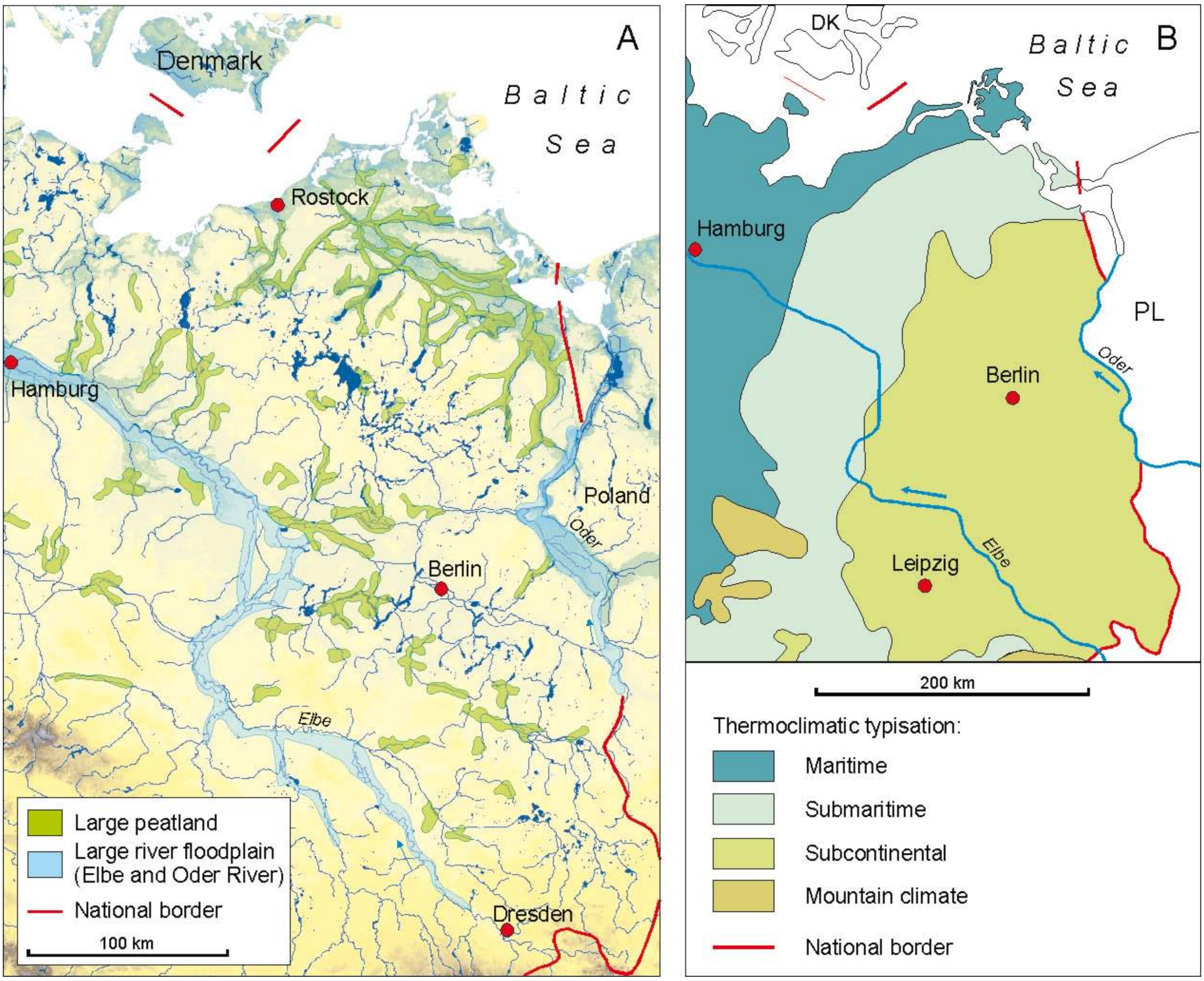

Fig. 2: Distribution of large peatlands and large river floodplains (A) as well of thermoclimatic zones (B) in northeast Germany (after BGR 2007, IfL 2008, adapted).

Abb. 2: Verbreitung großer Moorgebiete und großer Flussauen (A) sowie thermoklimatischer Zonen (B) in Nordostdeutschland (nach BGR 2007, IfL 2008, verändert).

ised by present-day mean annual discharges in a range of 500–700 $m^3 s^{-1}$. Several tributaries exist; the most important are the Saale, Havel, Mulde, Spree and Peene (20–120 $m^3 s^{-1}$; BMUNR 2003). A mainly east-west-oriented network of canals used for inland navigation connects the rivers.

The Weichselian belt is characterised by the occurrence of numerous *lakes* of different size and of different genetic, hydrologic and ecologic type. According to estimations from the adjacent Polish young morainic area, only one-third to half of the former lakes from the late Pleistocene to early Holocene have remained due to aggradation caused by natural and anthropogenic processes (Starkel 2003). By contrast, only a few natural lakes in the Saalian belt occur, but several artificial lakes originating from river damming and lignite opencast mining exist. In northeast Germany the total area of natural lakes amounts to c. 1300 km^2 (Korczynski et al. 2005). In general, the region's natural lakes largely represent 'hollows' located in the first unconfined aquifer. Thus groundwater and lake hydrology are closely connected.

In addition, a large area of the region (c. 5800 km^2) is covered by *peatlands*. This term refers to all kinds of drained or undrained areas with a minimal thickness of peat of at least several decimetres (Joosten 2008). Peatlands primarily occur in river valleys and large basins in the Federal States of Mecklenburg-Vorpommern (2930 km^2) and Brandenburg/Berlin (2220 km^2; Fig. 2A). Smaller areas are distributed in the lowland parts of Sachsen-Anhalt (580 km^2) and Sachsen (70 km^2). Groundwater-fed peatlands dominate with c. 99 % versus only 1 % rain-fed peatlands (Couwenberg & Joosten 2001).

The present-day climate of the region (Hendl 1994) is classified as temperate humid with mean annual air temperatures around 8–9 °C and mean annual precipitation ranging from 773 mm a^{-1} (Hamburg) to 565 mm a^{-1} (Cottbus). A distinct thermoclimatic gradient exists from northwest to southeast, dividing the region into maritime, sub-maritime and sub-continental parts with decreasing precipitation (Fig. 2B). The driest sites are located at the Saale (Halle/S.) and Oder Rivers (Frankfurt/O.) with a mean annual precipitation of about 450 mm a^{-1}.

3 Principle research questions, concepts and methods used in regional studies

The main disciplines providing regional palaeohydrologic knowledge (Fig. 1, Tab. 1) are geomorphology, Quaternary geology, palaeobotany and historical sciences. The prin-

Tab. 3: Facies areas of Holocene river valley development in northeast Germany considering geographic location, river valley dimension and valley history (Brose & Präger 1983, adapted).

Tab. 3: Faziesgebiete der holozänen Flusstalentwicklung in Nordostdeutschland unter Berücksichtigung der Lage, der Flusstaldimension und der Talgeschichte (nach Brose & Präger 1983, verändert).

<table>
<tr><th>Zone</th><th>Facies area</th><th>Example (river)</th><th>Selected genetic properties</th><th>Comparing conclusions (cross-zonal)</th></tr>
<tr><td>I</td><td>periglacial valley bottoms in the German Uplands</td><td>Saale, Mulde (middle reaches)</td><td>▪ state of equilibrium between erosion and aggradation in the early Holocene
▪ deposition of gravel during Atlantic frequently burying oak stems
▪ late Holocene deposition of flood loams and/or erosion</td><td rowspan="7">▪ as most river valleys (facies zones) are only initially investigated, comparing conclusions are partly of preliminary status
▪ after the retreat of the Weichselian ice sheet an erosional phase took place (Lateglacial) affecting the large river valleys up to the uplands
▪ erosion / aggradation in northern valleys is mainly controlled by water-level changes in the Baltic Sea and North Sea basins, whereas southern valleys are controlled by climatic and (in the late Holocene) by human impact
▪ widespread deposition of organic sediments (peat, gyttja) and soil formation characterises the Atlantic and Subboreal
▪ areal deposition of human-induced flood loams is a characteristic of the late Holocene except in low-lying valleys of zone IVb</td></tr>
<tr><td>II</td><td>valley bottoms in the loess belt</td><td>Elster, Unstrut</td><td>▪ erosional phase in the early Holocene with subsequent deposition of gravel, sand and topping overbank fines
▪ mid-Holocene hiatus (soil formation)
▪ late Holocene deposition mainly of flood loams</td></tr>
<tr><td>IIIa</td><td>valley bottoms in the old morainic area between Weichselian maximum and loess belt</td><td>Spree, Neiße (middle reaches)</td><td>▪ similar depositional history as in zone II</td></tr>
<tr><td>IIIb</td><td>valley bottoms in the young morainic area between Weichselian maximum and Pomeranian stage</td><td>Havel, Dosse, Spree (lower reaches)</td><td>▪ frequent occurrence of fluvial connections of basins (river-lake-structures)
▪ erosion / aggradation depending from river bed changes of Elbe and Oder (zone IVa)</td></tr>
<tr><td>IVa</td><td>valley bottoms of large transzonal rivers occupying several facies areas</td><td>Elbe, Oder</td><td>▪ erosional phases during (Pre-?)Bølling (lower Oder) and early Holocene (Elbe)
▪ early to mid-Holocene deposition of gravels and sands (Elbe) and mainly of peat (lower Oder)
▪ late Holocene deposition of overbank fines</td></tr>
<tr><td>IVb</td><td>valley bottoms of tributaries of the Baltic Sea north of the Weichselian Pomeranian stage</td><td>Peene, Warnow</td><td>▪ erosional phases during (Pre-?)Bølling, Preboreal and late Boreal
▪ flattening of the river bed gradient by organic sedimentation in the Atlantic/ Subboreal caused by marine influence (Littorina transgression)
▪ dominating deposition of peat and gyttja instead of overbank fines in the late Holocene</td></tr>
</table>

ciple research questions – some of which have been posed periodically for more than 100 years (e.g. Woldstedt 1956, Marcinek 1987, Kaiser 2002) – concern (1) the structure and formation of the natural drainage system, (2) its anthropogenic use and historic reshaping, and (3) the (palaeo-) ecologic status and change. More specific research questions in relation to the single aquatic environments investigated – rivers, lakes and peatlands – are given in chapters 4.1, 4.2 and 4.3.

Corresponding to different thematic approaches, the research concepts used come from different disciplines. Both geosciences and palaeoecology use climatologic- and biostratigraphic concepts and units. They are defined for dividing and explaining stages of deposition, relief formation and biotic changes, respectively. More specifically, the general model for the regional late Quaternary relief formation with emphasis on fluvial geomorphology, proposed by Marcinek & Brose (1972) and extended to incorporate the development of lake basins (Nitz 1984), has been adapted for use in this overview (Tab. 2). Additionally, the conceptualised regional facies areas of Holocene river development by Brose & Präger (1983) will be outlined (Tab. 3). These models and schemes provided the thematic framework for most of the later geomorphic and palaeohydrologic research. However, they base on relatively few local field studies only and generally lack sufficient numeric age control.

Archaeology, as a discipline of the historic sciences, has concentrated on the settlement and human use of aquatic landscapes in pre-Medieval (i.e. 'pre-German') times, thought to be a period with little human impact on the aquatic environment (e.g. Bleile 2012). History and historic geography have dealt with strong human impact on the regional drainage system since Medieval times (e.g. Schich 1994, Driescher 2003, Blackbourn 2006).

Corresponding to the disciplines involved, the results presented are based upon a broad range of geoscientific (including geochronology), biological (palaeoecology) and historic methods. The basic geoscientific methods used include the analysis of thousands of sedimentary profiles from corings as well as open sections, geomorphic mapping of fluvial and lacustrine structures, sedimentologic analyses and geophysics. Geochronology provides absolute chronologic control

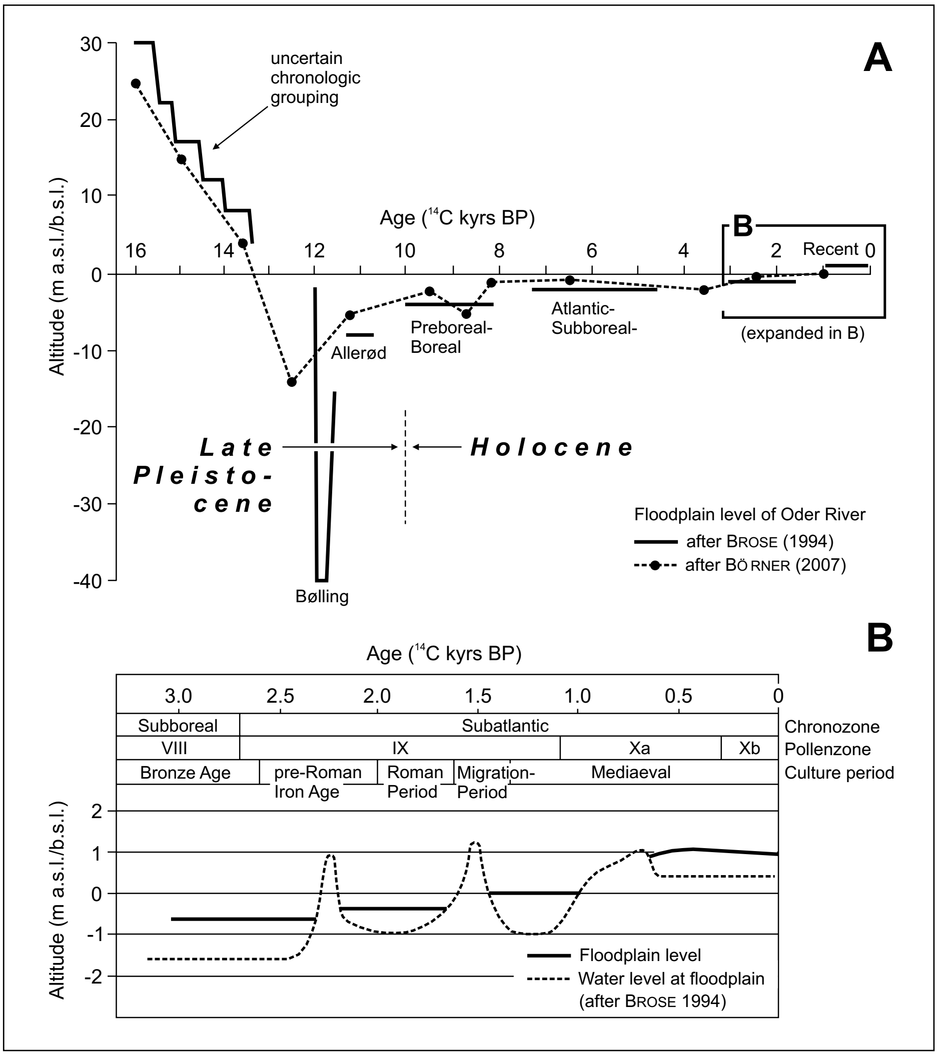

Fig. 3: Changes in the floodplain level of the lower Oder River. A: General development during the late Pleistocene and Holocene (after Brose 1994, Börner 2007, adapted). B: Detailled development during the late Holocene (after Brose 1994, adapted).

Abb. 3: Veränderungen des Auenniveaus der unteren Oder. A: Generelle Entwicklung während des Spätpleistozäns und Holozäns (nach Brose 1994, Börner 2007, verändert). B: Detaillierte Entwicklung während des Spätholozäns (nach Brose 1994, verändert).

comprising radiocarbon dating and, at a progressive rate, luminescence dating (mostly OSL). Normally, the chronology in this overview is based on calibrated radiocarbon ages (cal yrs BP). But, depending on the context, some other chronologic systems were also used (e.g. yrs BC, yrs AD, ^{14}C yrs BP, varve yrs BP). The most important biological method applied is pollen analysis providing both stratigraphic (thus to a certain degree even chronologic) information and palaeoecologic data (e.g. on vegetation structure, groundwater situation, human impact). Regional knowledge of the historic sciences is mainly based on archaeological excavations including find matter analysis, and interpretation of historic public records (documents) and maps. The latter are not available earlier than the 16th century AD.

4 Results and discussion
4.1 Rivers

In general, subjects of research on regional river evolution have been mainly (glacio-) fluvial geology and geomorphology (e.g. change of river course, river bottom incision/aggradation, valley mire formation), and palaeoecology, particularly analysing sedimentary archives in river valleys for vegetation and water trophic level reconstruction. It is only in recent years that quantitative estimations of palaeodischarge were attempted for some rivers (Elbe, Oder and Spree), using palaeoecologic, climatic and hydraulic data. The following overview on river and valley development concentrates on the aspects (1) river valley formation and depositional changes, (2) changes in the river courses and channels, and (3) palaeodischarge and palaeofloods.

4.1.1 River valley formation and depositional changes

The backbone of the regional river network has been a system of glacial spillways (ice-marginal valleys). These spillways worked as southeast-northwest oriented drainage following the retreat of the Weichselian ice sheet, except for the southernmost spillways, which originated from the previous Saalian glaciation. The valleys were operating from c. 26,000 to 17,000 cal yrs BP, partly initiated by the glacier blocking of northwards, i.e. to the North Sea and Baltic Sea basins, flowing rivers (Marcinek & Seifert 1995). The general subglacial and subaerial drainage of the ice sheet to the south led to connections of these spillways via lower-scale valleys. After glaciers decay, unblocking of the terrain often has initiated flow reversals (e.g. Kaiser et al. 2007, Lorenz 2008). In parallel, several short-lived ice-dammed (proglacial) lakes of different dimension developed; some of them of vast extent (see chapter 4.2.1.2)

A striking geomorphic property of the young morainic area is the existence of numerous so-called (*open*) ‘tunnel valleys' (glacial channels), containing rivers and streams as well as lakes and peatlands. Additionally, *buried* tunnel valleys of similar dimension occur both in the young and old morainic area (Eissmann 2002). The valleys were mainly eroded by meltwater supposed to have drained from subglacial lakes. Their water was most likely released in repeated outburst floods (so-called 'jökulhlaups') and flowed in relatively small channels on the floors of the tunnel valleys (Piotrowski 1997, Jørgensen & Sandersen 2006).

Knowledge on late Quaternary river development is very irregularly available in the region (Fig. 1, Tab. 1). The region's main river, the Elbe, has been recently only marginally in the (geo-) historic focus (e.g. Rommel 1998, Caspers 2000, Thieke, 2002, Turner 2012), in further contrast to other large central European rivers, such as Vistula and Rhine (Schirmer et al. 2005, Starkel et al. 2006).

A characteristic of low-lying valleys in the northern part of the study area, comprising the lower sections of the Elbe and Oder Rivers as well as the Vorpommern rivers (e.g. Uecker, Peene, Trebel, Recknitz; Fig. 1), is the hydraulic dependency of valley bottom processes from water-level changes in the North Sea and Baltic Sea basins and from isostatic movements. In general, a rise in the water level in the sea basins causes a lower hydraulic river bed gradient, whereas a water level fall leads to the opposite. This strongly influences several processes in the river and its floodplain (e.g. transport, flooding, sedimentation/erosion, vegetation). The Oder River and some Vorpommern rivers were extensively investigated in this respect. In the Lateglacial and early Holocene marked valley bottom changes were caused by lake-level changes of ice-dammed lakes in the Baltic Sea basin (Fig. 3). The mid- to late Holocene sea-level rise (Lampe 2005, Behre 2007, Lampe et al. 2010) triggered a large-scale formation of peatlands (mostly of percolation mires), temporally even the drowning of lower valley sections (e.g. Brose 1994, Janke 2002, Börner 2007, Michaelis & Joosten 2010). Thus, in contrast to river valleys of the higher-lying glacial landscape and the German Uplands, which are mainly filled by minerogenic deposits (gravels, sands, flood loams), peat widely fills the present valleys (Fig. 4).

Most regional studies have noticed that Holocene river bottom development up to the late Atlantic/early Subboreal is exclusively controlled by climatic and (natural-) geomorphic as well as biotic processes, such as fluvial erosion/aggradation and beaver damming. By contrast, Neolithic and subsequent economies, regionally starting in the south c. 7300 cal yrs BP (Tinapp et al. 2008) and in the north c. 6100 cal yrs BP (Latałowa 1992), considerably changed the vegetation structure, water budget and geomorphic processes of the catchments. Erosional processes, following forest clearing and accompanying agricultural use, increased the suspended load of rivers causing deposition of flood loams (overbank fines, 'Auelehm' in German) during flood events. Accordingly, a larger number of flood loams date from the late Atlantic (e.g. Hiller et al. 1991, Mundel 1996, Caspers 2000). Moreover, there is a multitude of flood loam records dating from the Subboreal and Subatlantic (e.g. Fuhrmann 1999, Börner 2007, Brande et al. 2007, Kaiser et al. 2007, Tinapp et al. 2008).

As shown by palaeo-flood indicators, human-induced changes in the catchment hydrology led to an increase in the frequency and magnitude of floods in the late Holocene (see chapter 4.1.3). The river valley bottoms shifted from quasi-stable to unstable conditions (Schirmer 1995, Kalicki 1996, Starkel et al. 2006, Hoffmann et al. 2008). More frequent and heavy floods caused both an intensification of river bed erosion and an aggradation of the valley bottom and levelling of its relief differences.

4.1.2 Changes in river courses and channels

In general, rivers can change their *course* by leaving their old valley or by formation of a new channel within their hitherto existing valley. Rivers can be forced to leave old valleys through tectonics, retrograde erosion or glacier damming. The accordant timescale mostly is a few to hundreds thousands of years (in phase with climatic *evolution*). Smaller changes in the *channel* pattern ('fluvial style') lead to new river beds within existing valleys, which are predominantly initiated by climate-driven changes of drainage (frequency, magnitude), erosion and bedload. This spans a timescale of tens to hundreds of years (in phase with climatic *changes*; Vandenberghe 1995b).

Of the regional rivers, only the Elbe has been investigated for changes in its course. In the Tertiary to mid-Pleistocene, *large-scale* river course changes (lateral river bed deviation of max. c. 150 km) occurred due to tectonic processes and to river damming triggered by glaciations. It was not until the end of the Saalian that its present course was substantially formed (e.g. Thieke 2002). *Small-scale* river course changes (max. c. 25 km) occurred in the Elbe-Havel River region ('Elb-Havel-Winkel' in German) still in historic times (early 18th century AD), when the river, caused by strong floods, was following older courses in the deeper lying Havel River valley (Schmidt 2000). Finally, evidence for river channel changes (max. c. 5 km) is available for the river section between Magdeburg and Wittenberge, showing that the present-day single channel river was a Holocene anastomosing system in this section up to the mid-18th century AD (Rommel 1998, Caspers 2000).

A few records are available on channel pattern changes in the region (Fig. 5). The mean *present-day* annual discharge of accordant rivers, however, varies extremely (0.3 to 550 m^3 s^{-1}). Six types of channel patterns were identified (braiding, meandering with large and small meanders, anastomosing, V-shape valleys/straight course, and inundation/valley mire formation). The type formed depends on several hydraulic parameters (bed gradient, load, flow velocity, discharge volume and temporal distribution; Miall 1996). In the late Pleniglacial and early Lateglacial all rivers investigated were braided systems caused by high load and strongly episodic discharge after heavy snow melting under periglacial conditions (e.g. Mol 1997). An incision phase took place in the early Lateglacial, when the regional erosion base in the Baltic Sea and North Sea basins was low or when the local erosion base was lowered by dead-ice melting. The (early) Lateglacial is characterised by the formation of so-called large meanders, which are attributed to short-term high discharges following extreme snow melting (Vandenberghe 1995a). For the Spree River, a distinct radius downsizing of sequenced meander generations was postulated (large meanders: 900–1000 m, small meanders: 600–900 m, recent me-

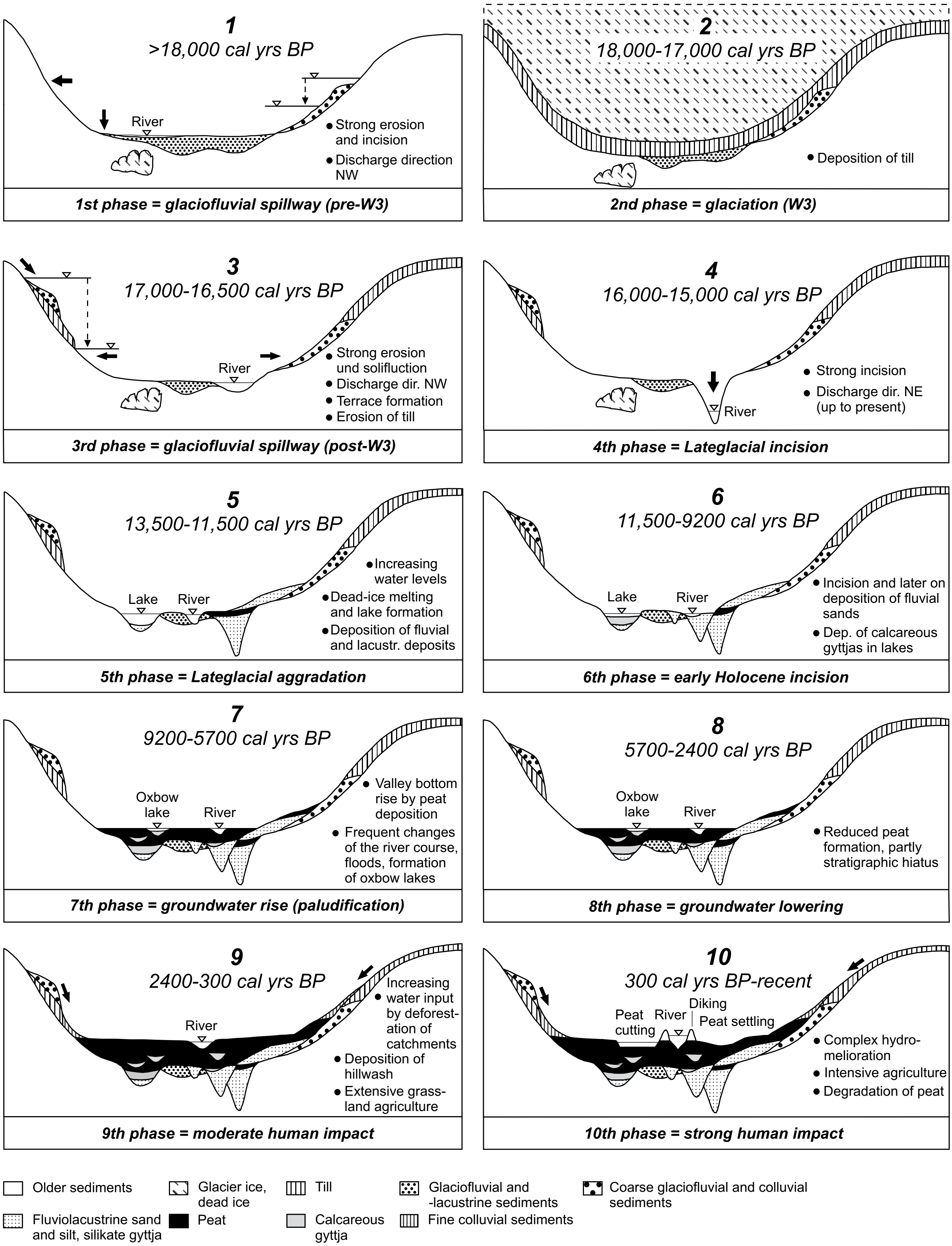

Fig. 4: Model of the geomorphic development of low-lying river valleys in Vorpommern (after Kaiser 2001, Janke 2002, adapted); a schematic geologic cross-section through a river valley is depicted. The term 'W3' used for phases 1–3 refers to the late Pleniglacial inland-ice advance of the Mecklenburgian Phase (Weichselian3/W3)', which is approximately dated by radiocarbon data from the Pomeranian Bay, southern Baltic Sea (Görsdorf & Kaiser 2001).

Abb. 4: Modell der geologisch-geomorphologischen Entwicklung tiefliegender Flusstäler in Vorpommern (nach Kaiser 2001, Janke 2002, verändert). Dargestellt ist ein schematischer geologischer Schnitt durch ein Flusstal. Der Begriff „W3", genutzt für die Talentwicklungsphasen 1–3, bezieht sich auf den spätpleniglazialen Inlandeisvorstoß der Mecklenburger Phase (Weichsel3/W3). Dieser Eisvorstoß ist näherungsweise durch Radiokohlenstoffdaten aus der Pommerschen Bucht/südliche Ostsee datiert (Görsdorf & Kaiser 2001).

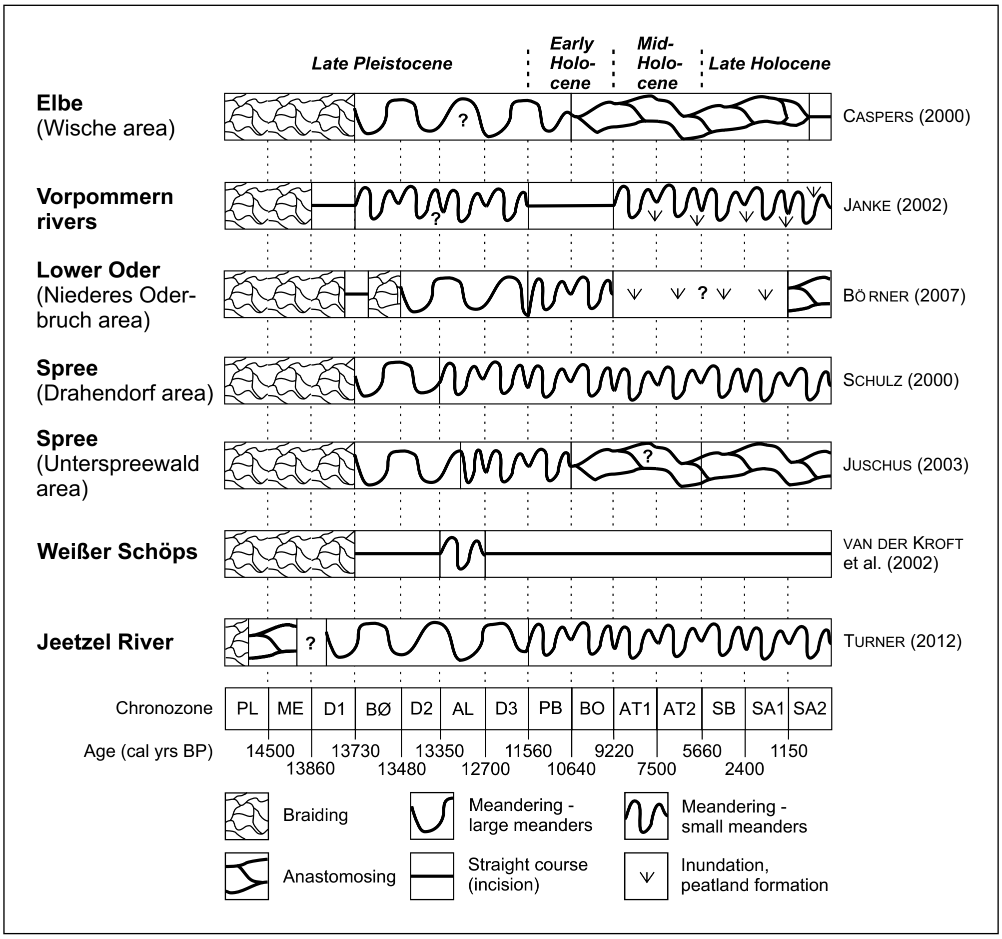

Fig. 5: Late Pleistocene and Holocene channel pattern changes in river valleys in northeast Germany (after various authors, adapted). Note missing data or questionable records are indicated by question marks.

Abb. 5: Spätpleistozäne und holozäne Veränderungen der Gerinnebettmuster in Flusstälern Nordostdeutschlands (nach verschiedenen Autoren, verändert). Fehlende Daten oder fragliche Befunde sind mit Fragezeichen gekennzeichnet.

anders: 150–300 m; SCHULZ 2000), which generally indicates decreasing (seasonal) discharge volumes. Beginning in the late mid-Holocene but strengthened in the late Holocene, some low-lying river sections were temporarily inundated and were generally transformed into peatlands (e.g. lower Oder River and some Vorpommern rivers).

In the last c. 800 years, human impact has considerably changed both the floodplain structures and courses of regional rivers by deforestation, artificial river-bed removing and strengthening as well as dyking, settlement and infrastructure construction (e.g. SCHICH 1994, SCHMIDT 2000, DRIESCHER 2003). For example, a dense network of canals for inland navigation has been built, beginning in the 16th century AD and culminating in the late 19th to early 20th century AD (UHLEMANN 1994, ECKOLDT 1998), in addition to the construction of innumerable drainage ditches.

4.1.3 Palaeodischarge and palaeoflood characteristics

Quantitative estimations of palaeohydrologic parameters for rivers usually aim at describing palaeodischarge (mean annual discharge, bankfull discharge) and palaeoflood characteristics (magnitude, frequency, risk; e.g. GREGORY & BENITO 2003, BENITO & THORNDYCRAFT 2005). Whereas in the adjacent Polish territory, palaeodischarge and palaeoflood studies were performed quite early (e.g. ROTNICKI 1991, STARKEL 2003), corresponding studies for northeast Germany are generally rare and of more recent status.

One recent study of the Elbe River mouth (German Bight, North Sea) produced a high resolution 800-year-long proxy record of palaeodischarge, based on a $\delta^{18}O$-salinity-discharge relationship (SCHEURLE et al. 2005; Tab. 4). The reconstructed variance of mean annual discharge (MAD), revealing a minimum-maximum span of 100–1375 $m^3\ s^{-1}$, is linked to long-term changes in precipitation. Four main periods of palaeodischarge/palaeoprecipitation become apparent, with higher and lower values than at present.

For the lower Oder River, a coupled climatic-hydrologic model estimated MADs for the early and mid-Holocene similar to those of today (WARD et al. 2007; Tab. 4). These modelling results coincide with local palaeohydrologic data from the Prosna River (a tributary of the Oder via the Warta in Poland; ROTNICKI 1991), which show that discharges there in the early and mid-Holocene were broadly similar to those in the period 1750–2000 AD.

For the Spree River, late Holocene palaeomeanders were investigated (HILT et al. 2008). Reconstructions show narrower and shallower channels for the undisturbed lower Spree as compared to recent conditions, which are strongly influenced by mining drainage water input (GRÜNEWALD 2008). Flow velocities and discharge at bankfull stage (Tab. 4) were smaller in palaeochannels and flow variability was higher. Furthermore, the increase in bankfull discharge was attributed to deforestation and drainage of the catchment as well as channelisation, bank protection and river regulation measures.

For the joint area of Vorpommern and northeast Brandenburg, BORK et al. (1998) estimated a regional water balance

Tab. 4: Holocene palaeodischarge estimations for Elbe, Oder and Spree Rivers after Scheurle et al. (2005), Ward et al. (2007) and Hilt et al. (2008), respectively.

Tab. 4: Abschätzungen der holozänen Paläoabflüsse für die Elbe (Scheurle et al. 2005), die Oder (Ward et al. 2007) und die Spree (Hilt et al. 2008).

River	**Elbe**	**Oder**	**Spree**
Gauging site	Neu Darchau (upstream of Hamburg)	Gozdowice (downstream of Frankfurt/Oder)	Neubrück (downstream of Cottbus)
Recent discharge ($m^3 s^{-1}$)	720 (100 %)[1]	527 (100 %)[1]	52 (100 %)[2]
Gauging period	1900-1995	1901-1986	present
Approach used	proxy record of palaeodischarge using a $\delta^{18}O$-salinity-discharge relationship	coupled climatic-hydrologic model	proxy record of bankfull palaeo-discharge using hydraulic properties of palaeomeanders
Palaeodischarge ($m^3 s^{-1}$)	1300 AD: 800 (111 %)[1] 1400 AD: 900 (125 %)[1] 1500 AD: 700 (97 %)[1] 1600 AD: 500 (69 %)[1] 1700 AD: 1000 (139 %)[1] 1800 AD: 900 (125 %)[1] 1900 AD: 500 (69 %)[1] Max. c. 1580 AD: 1375 (191 %)[1] Min. c. 1260 AD: 100 (14 %)[1]	early Holocene (9000-8650 cal yrs BP): 522 (99 %)[1] mid-Holocene (6200-5850 cal yrs BP): 538 (102 %)[1]	late Subboreal-early Subatlantic (3200-2500 cal yrs BP): 8 (15 %)[2]
Reference	Scheurle et al. (2005)	Ward et al. (2007)	Hilt et al. (2008)

[1]mean annual discharge
[2]bankfull discharge

for the time steps 650 AD, 1310 AD and today, which shows a maximum discharge value for the late Medieval period. This was caused by the lowest amount of forested areas (thus relatively low amounts of evapotranspiration and interception) during the late Holocene (Tab. 5).

Data on palaeoflood characteristics in the region are primarily available for the Elbe (Brázdil et al. 1999, Glaser 2001, Mudelsee et al. 2003), Oder (Glaser 2001, Mudelsee et al. 2003) and Spree Rivers (Rolland & Arnold 2002). Sporadic historical records start in the 11th century AD, while more continuous records are not available until the 16th century AD. As an example, for the Elbe River Mudelsee et al. (2003) detected significant long-term changes in flood occurrence rates from the 16th to the 19th century AD. A first maximum in the flooding rate was reached in the mid-16th century AD. At this time, rivers in central and southwest Europe experienced a similar increase in floods, which has been attributed to higher precipitation (Brázdil et al. 1999). Later on winter floods reached an absolute maximum (around 1850 AD) and then finally decreased. Mudelsee et al. (2003) concluded by means of statistical correlations for the Elbe and Oder Rivers that reductions in river length, construction of reservoirs and deforestation have had only minor effects on flood frequency. Furthermore, they arrived at the conclusion that there is no evidence from both historic data and modern gauging for a recent upward trend in the flood occurrence rate (in this context see Petrow & Merz 2009). This represents an important *regional* finding with respect to the current debate on regional hydrologic changes initiated by global climate change, emphasising the importance of temporally long hydrologic data series.

4.2 Lakes

In general, lake basins ubiquitously provide sedimentary archives from which both the local and to a certain extent even the regional landscape development can be reconstructed.

The lake basins in the northern part of the region (Mecklenburg-Vorpommern) were formerly classified by size as 'large glaciolacustrine basins' (former proglacial lakes, >100 km^2), 'medium-sized lakes' (0.03–100 km^2), and 'kettle holes' (<0.03 km^2; Kaiser 2001, Terberger et al. 2004). Although designed for a specific area, this classification by size can also be applied for the whole morainic area, additionally taking into account some local characteristics. Regional research on lake genesis performed so far mainly concentrated on (1) lake basin development (e.g. dead-ice dynamics and depositional changes) and on (2) palaeohydrology (lake-level and lake-area changes). Both aspects will be presented in the following.

4.2.1 Lake basin development
4.2.1.1 Dead-ice dynamics

Most of the medium- and small-sized lake basins in the Weichselian glacial belt originated from melting of buried stagnant ice, usually called 'dead ice' (e.g. Nitz et al. 1995, Böse 1995, Juschus 2003, Niewiarowski 2003, Kaiser 2004a, Lorenz 2007, Błaszkiewicz 2010, 2011). This term refers to the temporary *local* conservation/incorporation of ice in depressions and/or in sedimentary sequences; either coming from the freezing of pre-existing water bodies (e.g. shallow lakes) before being overridden by glacier ice or as a direct remnant from the glacier. Glacially- and melt water-driven

Tab. 5: *Estimation of the water balance for the northern part of northeast Germany considering the Vorpommern and Uckermark areas (after* Bork *et al. 1998, adapted).*

Tab. 5: *Abschätzung der Wasserbilanz für den nördlichen Teil von Nordostdeutschland (Vorpommern und Uckermark; nach* Bork *et al. 1998, verändert).*

Time step	**650 AD**		**1310 AD**		**Present**	
Land cover parameter	km^2	%	km^2	%	km^2	%
Total area	10000	100	10000	100	10000	100
Arable land and grassland	100	1	7900	79	6800	68
Forest (including uncultivated land)	9400	94	1500	15	2400	24
Surface waters	500	5	500	5	500	5
Other areas	<100	<1	100	1	300	3
Hydrological parameter	mm a^{-1}	%	mm a^{-1}	%	mm a^{-1}	%
Mean annual precipitation	595[1]	100	595[1]	100	595	100
Total runoff	40	7	140	24	120	20
Surface runoff	<1	0	10	2	3	<1
Subterraneous runoff	2	<1	5	1	4	<1
Mean evapotranspiration and interception	555	93	455	76	475	80

[1]assumed as today

erosive processes produced variously formed depressions (wide basins, channels, kettle holes), which were filled by dead ice during the glacier's decay. After the melting of these ice 'plombs', water-filled basins of varying size could appear, depending on the local hydrologic situation. Between dead-ice formation/burial and dead-ice melting, thousands of years, occasionally tens of thousands of years passed by. In contrast, the rare *present-day* natural lakes in the Saalian belt owe their existence mainly to local endogenic processes triggered by the dynamics of Zechstein salt deposits in the deep underground.

Dead-ice dynamics can be sedimentologically detected either by dislocation of sediment layers or by unusual succession of certain sediments. In the region, the first was repeatedly demonstrated by the record of heavily tilted peats and gyttjas (e.g. Kopczynska-Lamparska et al. 1984, Nitz et al. 1995, Strahl & Keding 1996, Kaiser 2001). The latter is normally attributed to the occurrence of basal peats below gyttjas, partly below a present-day water body of several decametres thickness (e.g. Kaiser 2001, Błaszkiewicz 2010, 2011).

Subsequent to the melting of dead ice in the basins and valleys, swamps/mires and lakes began to occupy the depressions. For parts of the study area, overviews on this onset of lacustrine sedimentation in medium-sized lakes and kettle holes are available (Kaiser 2001, 2004b, Brande 2003, de Klerk 2008). According to Kaiser (2001), in about 90 % of the lake basins compiled for Mecklenburg-Vorpommern and northern Brandenburg (total profile number analysed = 99) the process of sedimentation began in the Lateglacial, 38 % alone in the Allerød (Fig. 6).

In general, basin-forming dead-ice melting processes occurred from the Pleniglacial up to the early Holocene, with a concentration in the Allerød. Final dead-ice melting was assumed or reported for the Preboreal (e.g. Böse 1995, Niewiarowski 2003, Błaszkiewicz 2010, 2011). Over a third of the profiles analysed for Figure 6 include basal peats mainly from the Allerød, which ended regularly in a secondary position due to settling as the result of dead-ice melting.

4.2.1.2 Depositional changes

The deposition of fine silicate clastic gyttjas is characteristic for the cold Lateglacial stages. Peats and gyttja deposits rich in carbonates and organic matter mainly originate from the relatively warm Allerød. The dominant minerogenous input during the Lateglacial is caused by a very thin vegetation cover and an unstable overall relief (ablation, deflation, gully erosion, dead-ice melting, braiding). Besides basal peats from the Allerød, higher-lying peats of the same age buried by lacustrine and fluvial sands occur. They indicate a significant intensification of lacustrine and fluvial deposition during the subsequent Younger Dryas, which has been recognised throughout northeast Germany, triggered by renewed cold-climate conditions (e.g. Helbig & de Klerk 2002, Kaiser 2004b, de Klerk 2008). Although the increase in fluvial and erosional dynamics during the Pleistocene-Holocene transition constitutes a more general trend throughout the region, on an individual basis, some sedimentary records show that changes occurred rapidly and were often triggered by local relief instabilities and small scale catastrophic drainage events (e.g. Kaiser 2004a).

Sedimentation of organic and calcareous gyttja as well as peat generally characterises the Holocene. This is mainly due to a reduction in clastic input following a dense veg-

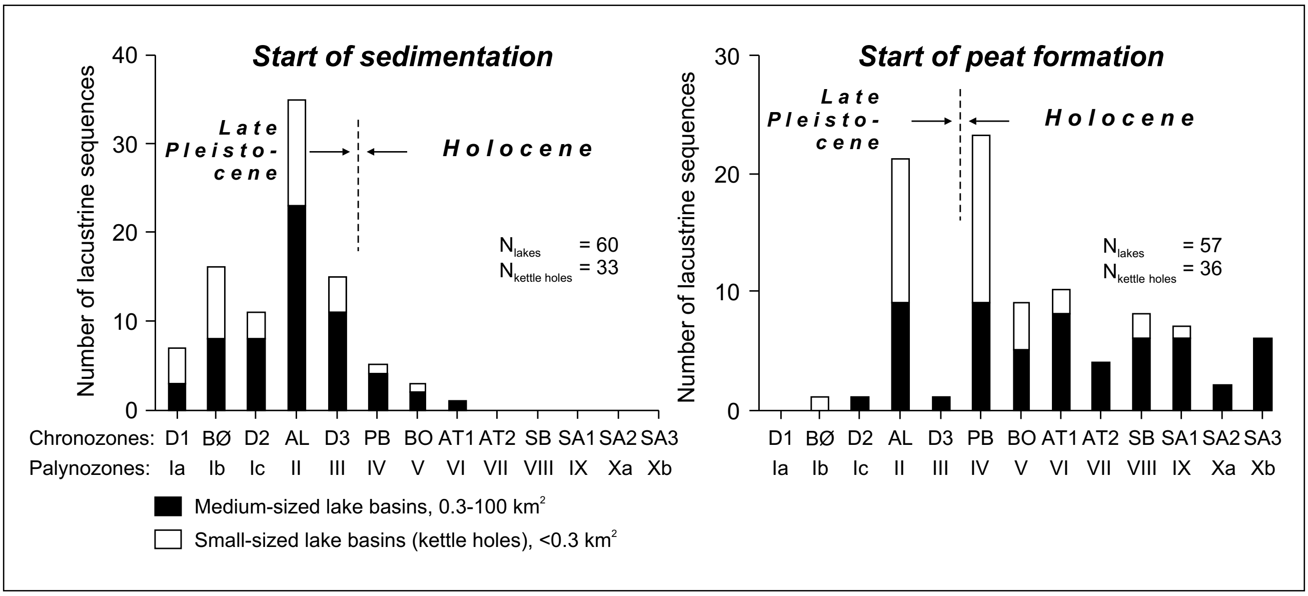

Fig. 6: Onset of lacustrine sedimentation (left) and peat formation (right) in lake basins of northeast Germany (areas of Mecklenburg-Vorpommern and northern Brandenburg; after Kaiser *2001, adapted).*

Abb. 6: Beginn der limnischen Sedimentation (links) und der Torfbildung (rechts) in Seebecken in Nordostdeutschland (Mecklenburg-Vorpommern und nördliches Brandenburg; nach Kaiser *2001, verändert).*

etation cover and a reduced geomorphic activity. In parallel the lake bioproduction increased. Deposition of gyttjas and, to a lesser degree, of fluvio-deltaic sequences filled shallow lacustrine basins. The common occurrence of fluvio-deltaic sequences, called (palaeo-) fan-deltas or Gilbert-type deltas (Postma 1990), in dead-ice depressions represents a previously undescribed geomorphic feature in the Weichselian glacial belt of northeast Germany (Kaiser et al. 2007), which corresponds to fan-deltas described from northwest Poland (Błaszkiewicz 2010).

Peat accumulation causing (natural) aggradation of lakes became a widespread regional phenomenon during the mid- to late Holocene. Commencing in the Subboreal and increasingly during the Subatlantic, human impact led to noticeable effects on the lake development. Increases in lacustrine sedimentation rates and clastic matter influxes since c. 1250 AD are evidence of erosion following forest clearing and systematic land use including anthropogenic lake-level changes and lake drainages (e.g. Brande 2003, Lorenz 2007, Selig et al. 2007, Enters et al. 2010). In the late 19th century AD, but enormously strengthened in the mid-20th century, human induced eutrophication by nutrient loading through agriculture, industry, sewage release, and soil erosion became a major threat to regional lakes (e.g. Scharf 1998, Mathes et al. 2003, Lüder et al. 2006). This eutrophication, partly in conjunction with human- and climate-driven hydrologic processes (e.g. Germer et al. 2011, Kaiser et al. 2012b), caused both depositional and hydrographic changes (increasing deposition rates, formation of anoxic sediments, partly shrinkage of lakes by aggradation).

The former vast ice-dammed (proglacial) lakes at the Baltic Sea coast underwent, in comparison to the medium- and small-sized inland lakes described above, a different development during the late Pleistocene and Holocene (Fig. 1). These late Pleniglacial lakes received water both from the melting inland-ice in the north and the stagnant (non-buried) ice in the immediate lake surroundings as well as from the ice-free area in the south. The largest lakes reconstructed are the 'Haffstausee' (c. 1200 km²; Janke 2002, Borówka et al. 2005) in the vicinity of Szczecin and the 'Rostocker Heide-Altdarss-Barther Heide-Becken' (>700 km²; Kaiser 2001) in the vicinity of Rostock. During deglaciation around 17,000 cal yrs BP, up to 25 m-thick glaciolacustrine sediments (clays, silts, sands) were accumulated. Local littoral gyttjas and aeolian sands dated to the Lateglacial have been found, indicating the end of the *large*-lake phase still within the Pleniglacial due to the decay of the basin margins consisting of ice (Kaiser 2001). For the Allerød and the early Younger Dryas, soils, peats, littoral gyttjas and Final Palaeolithic archaeological sites indicate widely dry conditions in these basins, in which only local lakes and ponds existed. In the late Younger Dryas, over large areas the basin sands were re-deposited by wind. The Holocene, on the one hand, is terrestrial, or locally also lacustrine, fluvial and boggy in form (e.g. Bogen et al. 2003, Terberger et al. 2004, Borówka et al. 2005, Kaiser et al. 2006, Börner et al. 2011). On the other hand, the lower parts of the glaciolacustrine basins came under marine influence, thereby becoming integrated into the Baltic Sea or the coastal lagoons (Lampe 2005, Borówka et al. 2005, Lampe et al. 2010)

4.2.2 Palaeohydrology

4.2.2.1 Lake-level changes

In general, lake-level records offer an important palaeohydrologic proxy as they can document past changes in the local to regional water budget in relation to climatic oscillations. Lake levels are influenced by climatic parameters affecting both evaporation and precipitation. But they can also be influenced by a variety of local, non-climatic factors such as local damming of the outflow by geomorphic processes and vegetation, animals (beaver) and man, or by land-

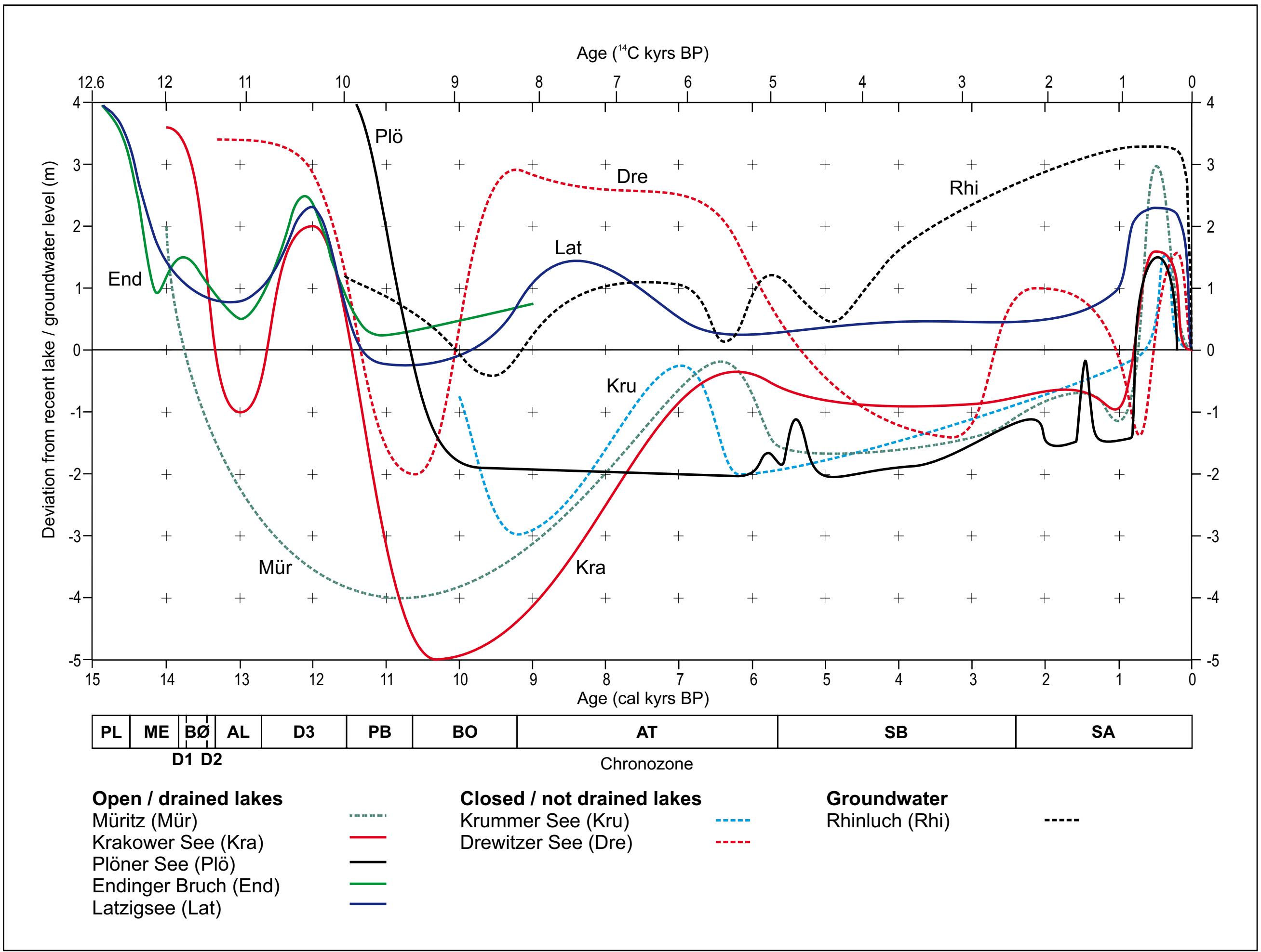

Fig. 7: Reconstruction of late Quaternary lake levels from northeast Germany (Lake Müritz: Kaiser et al. 2002, Lampe et al. 2009; Lake Endinger Bruch: Kaiser 2004a; Lake Latzigsee: Kaiser et al. 2003, Kaiser 2004b; Lake Krakower See: Lorenz 2007; Lake Großer Plöner See: Dörfler 2009; Lake Krummer See: Küster 2009). Additionally the reconstruction of the groundwater level in the Rhinluch peatland is shown (Gramsch 2002). All curves are adapted.

Abb. 7: Rekonstruktion spätquartärer Seespiegel in Nordostdeutschland (Müritz: Kaiser et al. 2002, Lampe et al. 2009; Endinger Bruch: Kaiser 2004a; Latzigsee: Kaiser et al. 2003, Kaiser 2004b; Krakower See: Lorenz 2007; Großer Plöner See: Dörfler 2009; Krummer See: Küster 2009). Ergänzend wird die Rekonstruktion des Grundwasserspiegels für das Rhinluch abgebildet (Gramsch 2002). Alle Kurven sind verändert.

cover changes in the catchment area influencing runoff and groundwater recharge (e.g. Gaillard & Digerfeldt 1990, Digerfeldt 1998, Duck et al. 1998, Harrison et al. 1998, Magny 2004).

Long-term ('continuous') records on the regional lake-level dynamics are available almost exclusively for the young morainic area north of Berlin. These records have been synthesised and are shown in Figure 7. Some further lake-level records that exist for the region have several constraints (e.g. coarse resolution, comparative only, temporally very fragmented, very synthetic/tentative; e.g. Brande 1996, Böttger et al. 1998, van der Kroft et al. 2002, Wennrich et al. 2005).

The records shown in Figure 7 span different time segments (i.e. chronozones) over the last 15,000 years. The manner of reconstructing past lake levels varied in the investigations (e.g. using subaquatic peats, lacustrine terraces and beach ridges, subaquatic wood remains and archaeological sites, historic documents), so the levels are based on data with different precision. The original records are referenced to absolute topographic levels (m a.s.l.), whereas the synoptic presentation in Figure 7 uses the (relative) deviation from the recent lake level for better comparison. Generally, the records available have a relatively low resolution, comprising often only one data point per chronozone. Thus the lake-level *curves* actually represent links of discrete data *points*, not continuous records. Consequently, far more (short-term) lake-level fluctuations can be expected than suggested by these curves. Despite these constraints, however, some *general trends* can be derived:

In the Pleniglacial and in parts of the Lateglacial, all lakes investigated had distinctly higher levels than at present. This was initially caused by deglaciation processes occurring at higher terrain levels, and later on caused by several geomorphic processes specific to the Pleistocene-Holocene transition, such as dead-ice melting, phased initiation of fluvial runoff and permafrost dynamics. After a distinct lowering in the early Holocene, lake-levels in one portion of lakes remained below present levels until the late Holocene, accompanied by fluctuations. Another portion of lakes shows temporally higher Holocene lake levels than at present. Common to all lakes, however, are the sudden and large changes in levels, initially positive, later on negative, that occurred in the late Holocene, after c. 1250 AD.

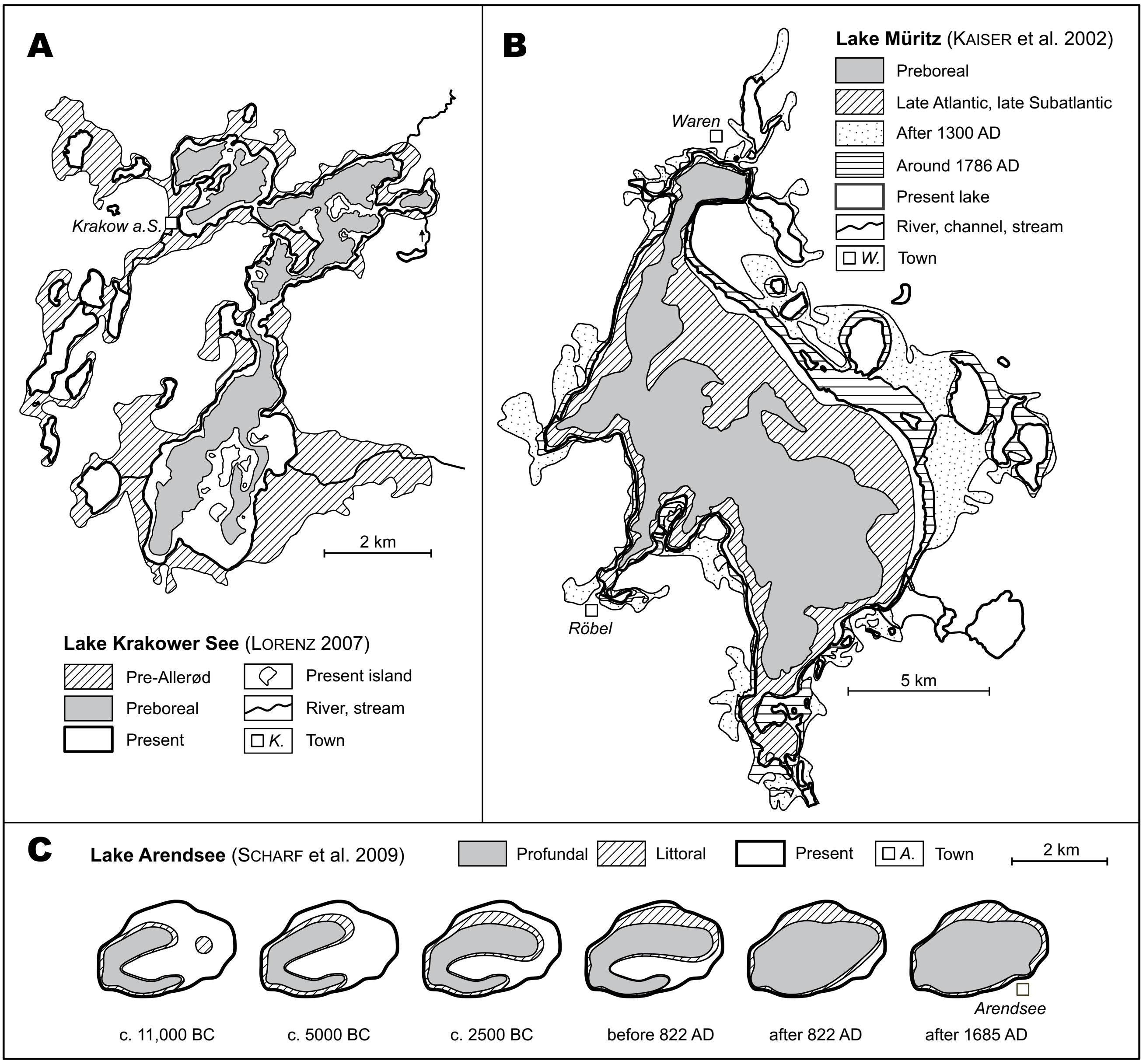

Fig. 8: Reconstruction of late Pleistocene and Holocene lake contours from northeast Germany. A: Lake Krakower See (Lorenz 2007). B: Lake Müritz (Kaiser et al. 2002). C: Lake Arendsee (Scharf et al. 2009). All subfigures are adapted.

Abb. 8: Rekonstruktion spätpleistozäner und holozäner Seeflächen in Nordostdeutschland. A: Krakower See (Lorenz 2007). B: Müritz (Kaiser et al. 2002). C: Arendsee (Scharf et al. 2009).

More specific, distinct phases of relatively low and relatively high lake levels can be deduced for the young morainic area (Fig. 7). Low lake levels in the Allerød and high lake levels in parts of the Younger Dryas were repeatedly detected (e.g. Helbig & de Klerk 2002, Kaiser 2004a, Lorenz 2007), which can be explained by climatic and geomorphic changes in that time. During the Allerød a moderate warm climate, forest vegetation and dominant dead-ice melting prevailed. The Younger Dryas, in contrast, was characterised by a cold climate with regional reestablishment of permafrost conditions, tundra vegetation and enhancement of surficial drainage. Similar observations have been made for the Baltic Sea near-coastal regions of Poland and Sweden (Berglund et al. 1996b, Ralska-Jasiewiczowa & Latałowa 1996). The early Holocene (Preboreal, Boreal) is widely characterised by low lake levels that can be ascribed to climatic warming and a fully forested landscape, as well as final dead-ice melting and intensification of erosive fluvial processes. In that time all lakes presented reached their Holocene minimum, partly lying 5–7 m below the present lake level (e.g. Lorenz 2007). In the mid-Holocene warm-wet Atlantic period the lakes initially rose, despite the fact that forests had their Holocene maximum extent and vigour (Lang 1994), potentially leading to high evapotranspiration rates in the lake catchments. This is in contrast to north Polish findings where predominantly low lake levels during the Atlantic have been detected (Starkel 2003). After the decreases in levels during the late Atlantic and Subboreal, some, partly strong, undulations took place in the Subatlantic (e.g. Kaiser et al. 2002, Lampe et al. 2009). The last c. 800 years saw almost identical dynamics, with increases in lake levels in the 13^{th}–14^{th} century AD (partly up to the 17^{th}–18^{th} century AD) and decreases in the

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18th–19th century AD. These changes are primarily caused by man, who became a major factor in lake hydrology due to the construction of mill and fish weirs, drainage improvement, canal construction and forest clearing (e.g. Jeschke 1990, Schich 1994, Kaiser 1996, Bork et al. 1998, Driescher 2003, Lorenz 2007, Küster & Kaiser 2010).

4.2.2.2 Lake-area and lake-contour changes

The late Quaternary lake-level changes caused to some extent drastic changes in the lake topography (volume, area, contour). However, only a few areal calculations and topographic (map) reconstructions exists for the region so far, showing that the lake areas and contours varied substantially (Fig. 8). For example, Lake Müritz, with a current area of 117 km^2 (100 %), varied from a minimum area of c. 74 km^2 (63 %) in the Preboreal to a maximum area of c. 188 km^2 (161 %) in the beginning of the 14th century AD (Kaiser et al. 2002).

4.3 Peatlands

Analogously to rivers and lakes, peatlands can also act as late Quaternary palaeohydrologic archives primarily indicating groundwater dynamics (e.g. Chambers 1996). Knowledge on their contribution, function, stratigraphy and development in northeast Germany is well-developed with an increasing number of studies and publications in the last c. 20 years (Succow & Joosten 2001). The following overview offers (1) a presentation of generalised phases of regional peatland formation and information on long-term groundwater dynamics, (2) the identification of genetic relationships between rivers, lakes and peatlands, and (3) an outline of the impact of historic mill stowage on peatlands and lakes.

4.3.1 Peatland formation and groundwater-level changes

4.3.1.1 General development

In central Europe, eight hydrogenetic mire types – mires are undrained virgin peatlands (Koster & Favier 2005, Joosten 2008) – can be distinguished (Succow & Joosten 2001). They are defined by the topographic situation, the hydrologic conditions (water input) and the processes by which the peat is formed. This hydrogenetic setting is of great importance in deciphering (palaeo-) hydrologic information.

A statistical analysis of 168 palynostratigraphically investigated profiles from peatlands in northeast Germany reveals distinct periods of specific hydrogenetic mire formation (Couwenberg et al. 2001; Fig. 9). With a maximum age of c. 12,400 ^{14}C yrs BP (c. 14,600 cal yrs BP), swamp mires are the oldest peat-forming systems in the region. The first lake mires developed still in the Lateglacial at c. 11,500 ^{14}C yrs BP (c. 13,400 cal yrs BP), whereas first kettle-hole and percolation mires did not develop until the early Holocene. The first rain-fed mire development started as recently as in the mid-Holocene at c. 7500 ^{14}C yrs BP (c. 8300 cal yrs BP). Partly, this temporal sequence reflects a stratigraphic succession of different mire types at the same location. The comparatively late increase and onset of percolation mire and rain-fed mire formation could reflect the mid- to late Holocene increase of regional humidity. Furthermore, there is a conspicuous peaking for the formation of some mire types in Figure 9, partly followed by a rapid decline. Between c. 1000–500 yrs BP, swamp mires show a maximum formation period, which was attributed to strong anthropogenic deforestation (e.g. Brande 1986, Jeschke 1990, Bork et al. 1998, Wolters 2005). The declining number of kettle-hole, percolation and rain-fed mires in the last 1000 to 2000 yrs, on the other hand, reflects direct human impact in the form of hydro-melioration measures and peat cutting. This caused the cessation of peat formation and the disappearance of older peat layers.

In contrast to the numerous pollen diagrams from peatlands and accordant estimations of the local *relative* groundwater dynamics, only two curves of *absolute* groundwater levels exist so far for northeast Germany. For the Reichwalde lignite open cast mine (Niederlausitz area), a short-term curve covers the Lateglacial Bølling to Allerød chronozones, i.e. a total of c. 1400 years, showing the development from a relatively stable low to an instable high groundwater level (van der Kroft et al. 2002). The Holocene groundwater dynamics derived from the c. 11,500 years-long synthetic Rhinluch peatland record (west of Berlin) reveal a low level at the end of the early Holocene, an increasing level accompanied by fluctuations during the mid-Holocene and a maximum level in the late Subatlantic (Gramsch 2002; Fig. 7). A marked decrease of the groundwater level of c. 3 m occurred in the very late Subatlantic (18th-19th century AD), which was caused by local hydro-melioration measures (e.g. Zeitz 2001).

4.3.1.2 Peatlands in large river valleys

Close relationships between the development of rivers, lakes and peatlands existed particularly during the late Holocene complex paludification processes in large river valleys of the region. They are caused, on the one hand, by natural climatic and hydraulic changes and, on the other hand, by direct anthropogenic impact in the form of mill stowage (for the second see chapter 4.3.2).

The largest peatlands in the region are located in former ice marginal spillways of Brandenburg and Mecklenburg-Vorpommern. Beside local lake mires and widely-stretched (but typically small) floodplain mires accompanying the abundant rivers, vast swamp (paludification) mires occur.

The Havelländisches Luch (c. 300 km^2) and Rhinluch (c. 230 km^2) peatlands, for instance, form wide elongated depressions which were formed by glaciofluvial and glacial erosional processes during the Weichselian glaciation and by (glacio-) fluvial processes during deglaciation and afterwards (Weisse 2003). After a Pleniglacial fluvio-lacustrine phase leading to the deposition of vast amounts of sands ('Beckensand' in German), a number of small shallow lakes developed following dead-ice melting in the Lateglacial. During the early Holocene most lakes aggraded by both sedimentary infill and groundwater lowering (Fig. 7), forming local lake mires (Succow 2001a). Dated palaeosols in peat, fluvial and lacustrine sequences (8770 ± 160 to 4170 ± 150 cal yrs BP; Mundel 1996, Kaffke 2002) form a stratigraphic hiatus, which indicates regional groundwater lowering and reduced fluvial activity in the mid-Holocene to the early phase of the late Holocene. The former vegetation of the Havelländisches Luch peatland with dominating sedges and reed was largely

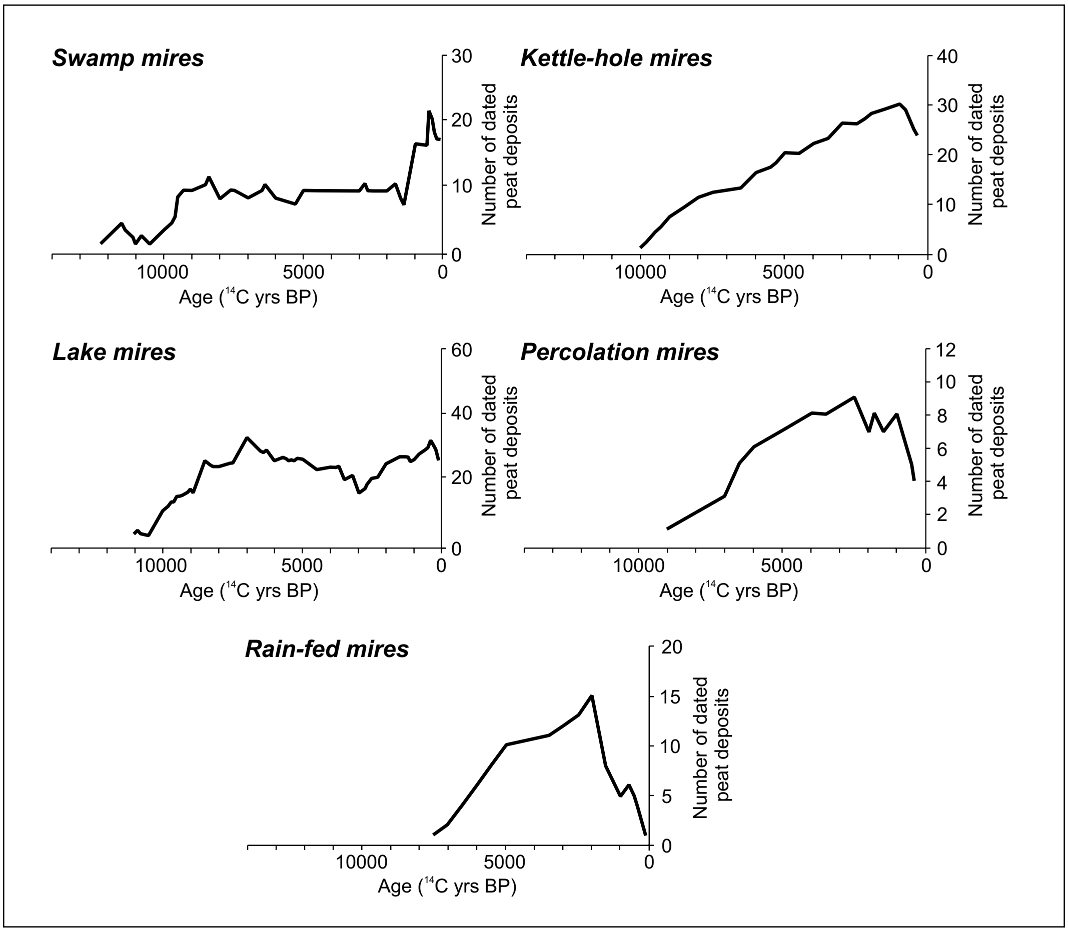

Fig. 9: Temporal distribution of palynologically dated peat and gyttja deposits of selected hydrogenetic mire types in northeast Germany (Couwenberg et al. 2001, adapted).

Abb. 9: Zeitliche Verteilung palynologisch datierter Torf- und Seeablagerungen ausgewählter hydrogenetischer Moortypen in Nordostdeutschland (Couwenberg et al. 2001, verändert).

replaced by wet forests consisting of oaks and alders (Kloss 1987a). In general, although local mire development in northeast Germany varies considerably, many peat sequences are characterised by this mid- to early late Holocene stratigraphic hiatus (e.g. Brande 1996, Succow 2001a, Wolters 2002, Janke 2004, Brande et al. 2007) reflecting the regional dry climatic conditions in that time. Looking at this from a wider perspective, this northeast German peatland-palaeosol (and hiatus) is apparently comparable with the so-called 'Black Floodplain Soil', a polygenetic buried humic soil horizon (Boreal-Atlantic) found in river valleys and basins of central and southern Germany (Rittweger 2000). Between c. 3800 cal yrs BP (Mundel 1996) and c. 2600 cal yrs BP (Kaffke 2002) an increase in groundwater occurred, causing regional paludification and local lake levels to rise. The vegetation shifted back from wet forests to reeds. Basically in that time vast swamp mires were formed transgrading onto former areas without peats. Two possibly superimposing reasons have been identified for this, namely a supra-regional late Holocene climatic shift to relatively wet-cool conditions (Mundel 1996; see more general: e.g. Zolitschka et al. 2003) and a regional damming-effect of the rising Elbe River bed, which was driven by the eustatic rise of the North Sea (Mundel 1996, Küster & Pötsch 1998; see more general: e.g. Behre 2007). This damming effect was linked to relatively high aggradation rates in the Elbe valley versus low rates in the Havel valley. The abundant lake basins in the Havel course serve even now as effective traps for river load (Weisse 2003). Thus the drainage of the Havel and its tributaries was impeded, causing a rise in the regional groundwater. No later than the mid-18th to early 19th century AD, regional peat growth stopped again, this time caused by hydro-melioration measures for agricultural use and peat cutting.

Close relationships between fluvial-lacustrine processes and mire development are also a characteristic of several low-lying river valleys of Vorpommern close to the Baltic Sea coast, which were strongly forced by marine influence (Janke 2002, Michaelis & Joosten 2010; see chapter 4.1.1).

4.3.2 Human impact on peatlands and lakes by mill stowage

In general, until the late 12th to early 13th century AD landscape hydrology in northeast Germany was dominantly driven by climatic (e.g. wet and dry phases), geomorphic (e.g. fluvial aggradation and incision) and non-anthropogenic biotic (e.g. beaver activity) factors. However, since the Neolithic, localised and phased hydrologic changes in catchments due to land-cover changes can be assumed.

During the German Medieval colonisation, the water mill technology was introduced by the west German and Fleming/Dutch settlers in eastern central Europe. For water mill

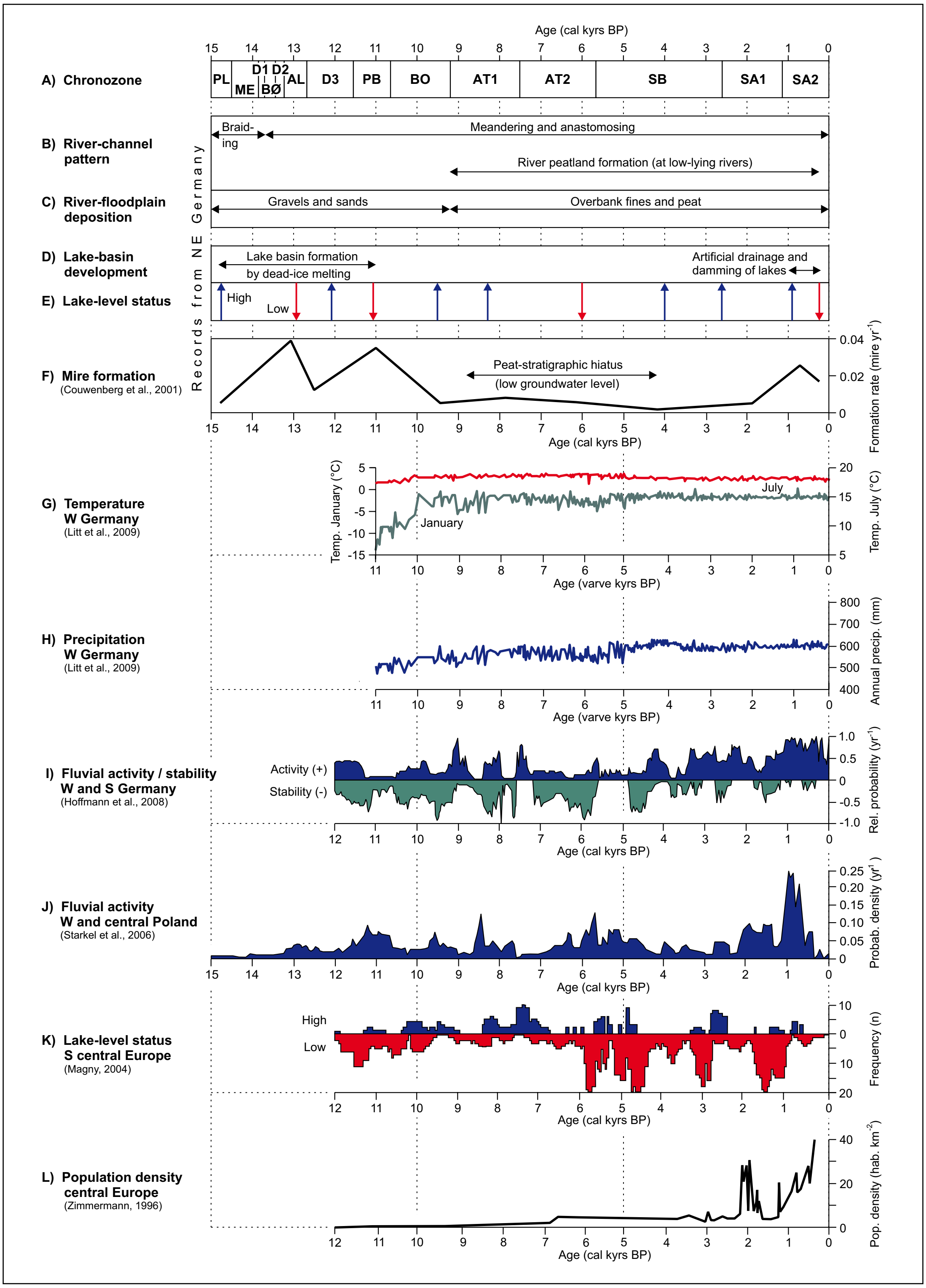

Fig. 10: Late Quaternary hydrologic changes in northeast Germany (B-F) plotted alongside further palaeoclimatic and palaeohydrologic proxy records (G-K) as well as population data (L) from central Europe. G-H: January and July temperatures and annual precipitation reconstructed from pollen data from annually laminated (varved) sediments of Lake Meerfelder Maar (Eifel region, west Germany), using pollen-transfer functions (Litt et al., 2009; adapted). I: Geomorphic activity (positive probability values) and stability (negative probability vales) based on CDPF analysis of west and south German fluvial deposits (Hoffmann et al. 2008, adapted). J: Geomorphic activity based on CDPF analysis of west and central Polish fluvial deposits (Starkel et al. 2006, adapted). K: Lake-level status reconstructed from lakes of southern central Europe (Jura, French Pre-Alps and Swiss Plateau; Magny 2004, adapted). L: population density of central Europe reconstructed from archaeological evidence (Zimmermann 1996, adapted).

Abb. 10: Spätquartäre hydrologische Veränderungen in Nordostdeutschland (B-F) dargestellt mit weiteren paläoklimatischen und paläohydrologischen Proxydaten (G-K) sowie paläodemografischen Daten (L) aus Mitteleuropa. G-H: Januar- und Juli-Temperaturen sowie Jahresniederschlag rekonstruiert anhand von Pollendaten (mittels Pollen-Transferfunktionen) aus den jahreszeitlich geschichteten (warvierten) Sedimenten des Meerfelder Maars (Eifel, Westdeutschland; Litt et al., 2009, verändert). I: Geomorphodynamische Aktivität (positive Wahrscheinlichkeitswerte) und Stabilität (negative Wahrscheinlichkeitswerte) basierend auf der CDPF-Analyse west- und süddeutscher fluvialer Ablagerungen (Hoffmann et al. 2008, verändert). J: Geomorphodynamische Aktivität basierend auf der CDPF-Analyse west- und mittelpolnischer fluvialer Ablagerungen (Starkel et al. 2006, verändert). K: Seespiegelstatus rekonstruiert für das südliche Mitteleuropa (Jura, französische Voralpen, Schweizer Mittelland; Magny 2004, verändert). L: Bevölkerungsdichte in Mitteleuropa rekonstruiert anhand archäologischer Befunde (Zimmermann 1996, verändert).

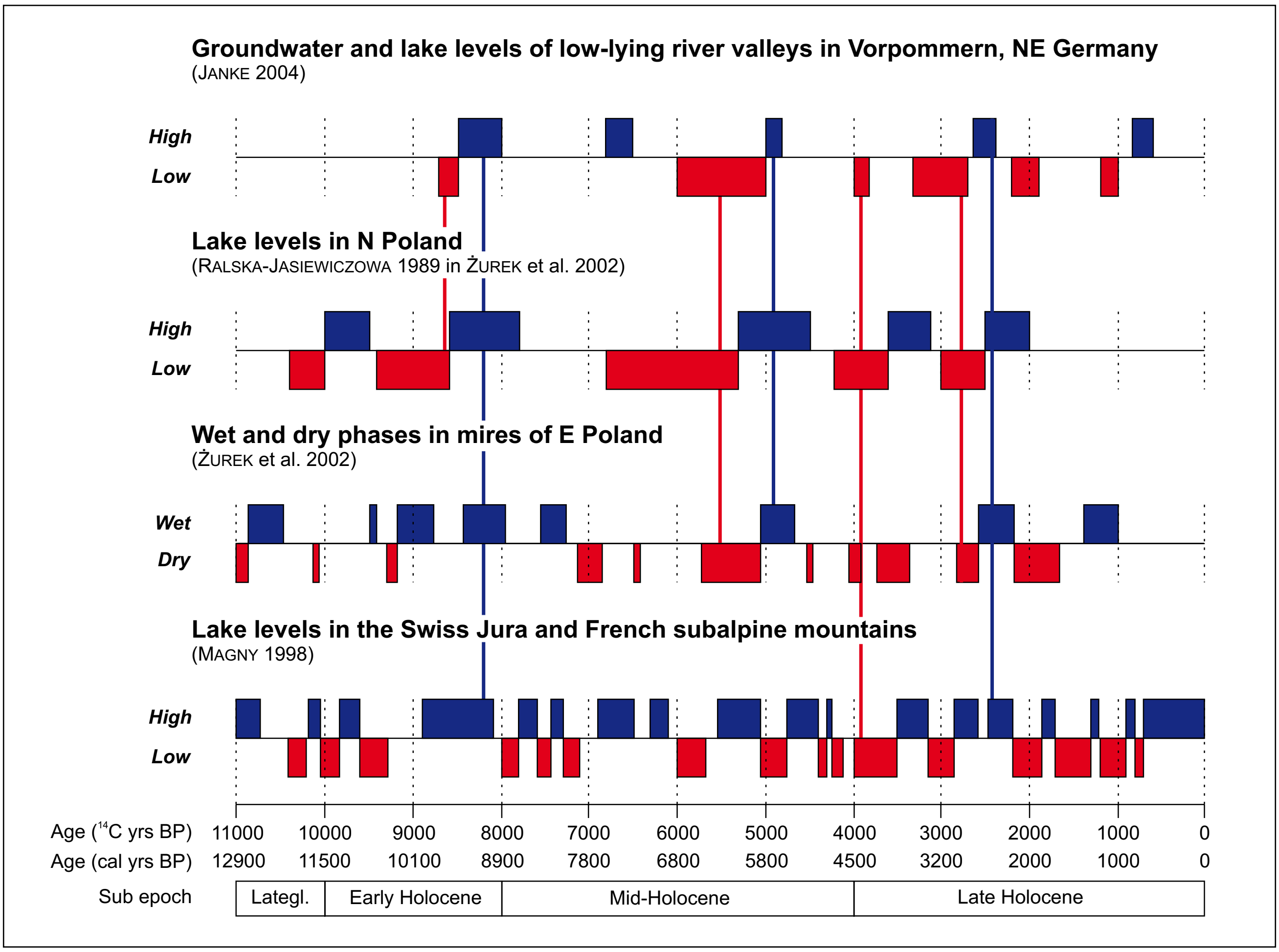

Fig. 11: Lateglacial and Holocene lake- and groundwater-level data from northern and southern central Europe (after various authors, adapted). Vertical bars mark synchronicity of wet (in blue) or dry (in red) phases.

Abb. 11: Spätglaziale und holozäne Seespiegel- und Grundwasserspiegeldynamik im nördlichen und südlichen Mitteleuropa (nach verschiedenen Autoren; verändert). Die vertikalen Linien markieren Synchronität feuchter (in blau) oder trockener (in rot) Phasen.

operation, a local water level difference of c. 1 m at minimum is required. This led to the construction of a multitude of mill dams and, accordingly, of dammed (mostly originally natural) lakes and of rising groundwater levels upstream (e.g. SCHICH 1994, KAISER 1996, DRIESCHER 2003, BLEILE 2004, NÜTZMANN et al. 2011). The operation of hundreds of water mills (together with fish weirs) drastically changed the Medieval hydrology in the region.

The phenomenon of 'mill stowage' ('Mühlenstau' in German) and its implications for settlement, economy and landscape was first systematically investigated by BESCHOREN (1935) and HERRMANN (1959) particularly for the Spree and Havel Rivers, and later on extended by DRIESCHER (2003) in the form of a multitude of local case studies in the wider region. The impacts of mill stowage on groundwater and lake levels, mire development and sedimentary processes are particularly well-investigated in the Berlin region. For the time-span 12th–14th century AD dated sequences from peatlands typically show a sudden change from highly to weakly decomposed peats or an inversion of the aggradation sequence (lake deposits overlying peats). These records were interpreted in terms of an intensification of mire formation and rising lake levels, respectively, by mill stowage (BRANDE 1986, 1996, BÖSE & BRANDE 1986, 2009, KÜSTER & KAISER 2010). Medium-scale rivers and their riparian zones, such as the Havel and Spree, were in part drastically influenced by these processes, whereas the large-scale Elbe River had no damming constructions but boat mills (GRÄF 2006). Along the low-gradient middle Havel course, mill weirs in the cities of (Berlin-) Spandau and Brandenburg/Havel, which were constructed in the late 12th/early 13th century AD (SCHICH 1994), caused large-scale lake enlargements and paludifications (KAISER et al. 2012b). Some smaller rivers and streams had a multitude of water mills ('mill staircases'). For instance, along a 20 km section of the upper Dahme River (Brandenburg) 14 mills were operated, some since the 13th/14th century AD (JUSCHUS 2002), strongly changing the river gradient, the discharge process and the local groundwater level.

5 Synopsis

The temporal focus of this overview is the Late Quaternary comprising here the last c. 20,000 years and using a millennial scale. Accordingly, this synopsis will concentrate on this time span, comparing the regional results with those from other parts in central Europe and adding information (e.g. on climatic evolution), which is important for the understanding of the results presented. However, as the regional drainage system is influenced, on the one hand, by very long-term endogenic processes, and, on the other hand, has experienced a historically unprecedented strong change during the last c. 300 years (e.g. Blackbourn 2006, Kaiser et al. 2012b), these both time perspectives shall be touched at least by a glimpse.

5.1 Impact of neotectonic processes

As outlined in chapter 2, successive Pleistocene glaciations have formed the main relief and sedimentary settings in the North European Plain. However, the river system shows several conspicuous patterns – e.g. the asymmetric (right-skewed) catchments of the Elbe and Oder Rivers, a west oriented turn of the Havel and Spree Rivers, the nearly orthogonal valley grid of Vorpommern (Fig. 1) – which suggests the impact of neotectonic processes. Accordingly, several authors (e.g. Schirrmeister 1998, Reicherter et al. 2005, Sirocko et al. 2008) have asked to what extent does the present-day topography of the so-called 'North German Basin' mirror the heterogeneous structure of the basement? They stated that several river valleys (including spillways) and terminal moraines in northern Germany apparently run parallel to the major tectonic lineaments and block boundaries. Moreover, the drainage pattern and the distribution of lakes in north Germany exactly follow block boundaries and, hence, mark zones of present-day subsidence. The Tertiary morphology in that area was apparently draped by Quaternary glacial deposits, but rivers and lakes that dominate the topography of the modern landscape still reflect the geodynamic centres of Tertiary tectonism and halokinesis (Sirocko et al. 2002).

5.2 Climate impact

The synoptic Figure 10A-F shows a selection of results on Late Quaternary river, lake and peatland formation, which represent the regionally typical processes discussed in the previous sections. The temporal resolution of those results is quite coarse mostly covering a chronozone or representing, in general, a millennial scale. By contrast, comparing climatic, hydrologic, geomorphic and historic data from central Europe partly represent centennial- up to decadal-scale records (Fig. 10G-L). It should be borne in mind that the statistical basis for certain evidence in northeast Germany partly is still small (e.g. on the lake-level status).

In this region, climate was the dominant driver for geomorphic and hydrologic changes up to the late Holocene intensification of land use by man. With the exception of neotectonic processes of yet inadequately known impact even the partly considerably effective sea level rise of the North and Baltic Seas is ultimately climate-driven. Within this climate-controlled setting local geomorphic and biotic processes operated (e.g. dead-ice melting, fluvial aggradation/incision, mire formation).

However, there is no specific (high-resolution) Late Quaternary *climate record* from northeast Germany available so far except those that cover relatively short periods (e.g. Büntgen et al. 2011). Hence the pollen-based high-resolution record from Lake Maarfelder Maar (Eifel region, west Germany; Litt et al. 2009) can be used to characterise some climatic trends at least for the Holocene, which, in general, can be assumed even for northeast Germany. In particular for the relatively dry-warm early Holocene and the wet-warm mid-Holocene ('Holocene optimum' between c. 8000–5000 varve yrs BP; Wanner et al. 2009) some simultaneous hydrologic phenomena of northeast Germany (e.g. early Holocene lake-level lowstands, mid-Holocene groundwater lowering in peatlands) can be presumably ascribed to direct climatic impact. Geochronological data from river catchments of west and south Germany as well as west and central Poland allow some general assumptions on *palaeodischarge and palaeoflood dynamics*, which can be hypothesised even for the region under study. Large datasets of ^{14}C ages obtained from late Quaternary fluvial units were analysed using cumulative probability density functions (CPDFs) in order to identify phases of fluvial activity (floods) and stability (Starkel et al. 2006, Koślacz et al. 2007, Hoffmann et al. 2008; Fig. 10I, J). In the west and south German record (Fig. 10I), several periods of fluvial activity were identified and compared to climatic, palaeohydrologic and human impact proxy data. Until c. 4250 cal yrs BP, events of fluvial activity are mainly coupled to wetter and/or cooler climatic phases. Due to growing population and intensive agricultural activities during the Bronze Age the increased fluvial activity between c. 3300 and 2820 cal yrs BP cannot unequivocally be related to climate. Since 875 AD the growing population density (Fig. 10L) is via landcover changes in the catchments (increasing arable land and pastures, decreasing forests) considered as the major external forcing (Hoffmann et al. 2008). Similar curve characteristics of CPDFs from ^{14}C data on fluvial units show records from west and central Poland (Starkel et al. 2006; Fig. 10J), allowing corresponding conclusions.

From southern central Europe a data set of 180 radiocarbon, tree-ring and archaeological dates obtained from sediment sequences of 26 lakes was used by Magny (2004) to construct a regional Holocene lake-level record (Fig. 10K). The dates form clusters suggesting an alternation of lower and higher, climatically driven lake-level phases. The comparison of *relative* Holocene lake- and groundwater-level data from north and south central Europe reveals some distinct synchronicities of wet and dry phases, but also some distinct disparities (Ralska-Jasiewiczowa 1989, Kaiser 1996, Magny 1998, Wojciechowski 1999, Kleinmann et al. 2000, Żurek et al. 2002, Janke 2004; Fig. 11). In general, synchronic correlation of identical phases works far better within nearby German and Polish sites of northern central Europe. In comparison to the southern central European record, however, these records appear to be somewhat 'monolithic', which probably is caused by a low temporal resolution. For the early Holocene, the lake-level record in northeast Germany and north Poland shows a clear tendency towards low levels. This is not reflected in the mire record of east Poland that widely indicates a wet phase. The late Boreal and partly the early Atlantic is characterised by increasing lake and groundwater levels followed by a decrease in the late Atlantic. The beginning and mid-late Holocene (c. 2500 cal yrs BP) reveals wet phases, whereas a dry phase lies in between. The general wet-dry pattern inferred correlates well with major Holocene climatic episodes (e.g. Harrison et al. 1993, Magny 2004, Litt et al. 2009, Wanner et al. 2009).

A synoptic view on the northeast German results (Fig.

Tab. 6: Examples of new and promising palaeohydrologic research topics for northeast Germany.

Tab. 6: Beispiele für neue, vielversprechende Forschungsthemen zur Paläohydrologie/Historischen Hydrologie in Nordostdeutschland.

Research field	Remarks
Exploration and combination of proxies	Using new proxies and new combinations of proxies for deciphering and validating of palaeohydrologic information (e.g. tree ring data, near-shore and shoreline sediments of lakes, palaeosols of wetlands)
Human induced lake drainage	Exploring the occurrence and the rewetting potential of lake basins drained by historic anthropogenic hydromelioration
Human induced lake formation	Exploring the properties and genesis of lakes and ponds formed in Medieval times and afterwards; deciphering historic hydrologic information from young deposits / geoarchives
Long hydrologic time series	Linking instrumental records of specific hydrologic parameters (observations e.g. by gauging) with proxy records from geoarchives
Quantitative palaeohydrology	Combining palaeohydrologic field records with hydrologic modelling at different areal and temporal scales
Reference status of wetlands	Reconstructing the (near-) natural status of wetlands; i.e. before human impact has sustainably changed the aquatic environments

10B-F) and on evidence from other central European regions (Fig. 10G-L) reveals some concordances. But even discrepancies become apparent, partly within a type of proxy. Reasons for this might be, on the one hand, real differences in the regional hydrologic evolution, which are partly caused by different (pre-)historic human impact. On the other hand, a partly drastically different statistical base for the parameters presented is to consider. For example, a few data on the lake-level status in northeast Germany contrast a large database in southern central Europe. Thus future research possibly will modify the regional information available.

5.3 Pre-modern and modern human impact

First *intended* changes of the regional hydrography date from late Medieval times (since the late 12th century AD). In parallel the regional forests were widely cleared (Bork et al. 1998; Tab. 5), causing several *unintended* hydrologic changes such as rising groundwater and lakes followed by increasing fluvial discharge. In the period 18th to first half of the 20th century AD most of the peatlands were transformed by hydromelioration into *extensive* grasslands (Schultz-Sternberg et al. 2000). In parallel a dense network of channels for inland navigation was formed and most channels of large rivers were modified by hydraulic engineering.

The most intensive changes, however, did not occur until the last c. 50–60 years. The peatlands were nearly totally transformed into *intensive* grassland and arable land by complex melioration measures (e.g. Succow 2001b); only 2 % of the original mires remained in a near-natural status (Couwenberg & Joosten 2001). In Mecklenburg-Vorpommern, for instance, for the period 1965–1995 a total loss of c. 290 km^2 peatlands by peat decomposition was calculated, which accounts for c. 13 % of the state's pre-modern peatland area (Lenschow 2001). Even as a consequence of hydro-melioration measures in parallel with climatic and land-cover changes, the groundwater level of the first aquifer significantly dropped (1–2 m) at a regional scale, particular in Brandenburg, causing in many cases lake-level lowerings (e.g. Germer et al. 2011, Kaiser et al. 2012b).

Furthermore, lignite open cast mining has drastically changed vast areas in some regions. In the Niederlausitz region a total of c. 800 km^2 was required for mining activity so far. A large number of artificial lakes and connective canals were formed by flooding of disused open-cast mines, forming the 'Lausitzer Seenland' (Grünewald 2008). In the near future the lake area here amounts to a total of c. 250 km^2. In the Leipzig-Halle-Bitterfeld area the total area of anthropogenic lakes in the mid-21st century AD will result in c. 70 km^2, forming the 'Neue Mitteldeutsche Seenlandschaft' (Czegka et al. 2008). If the total area of natural lakes in northeast Germany is considered (c. 1300 km^2; Korczynski et al. 2005), artificially dug 'new lakes' (c. 320 km^2) will form a portion of c. 20 % of the total lake area soon of c. 1620 km^2.

Thus in modern times, man became by several impacts a very important geomorphic and hydrologic factor in the region. With respect on the pace and magnitude his influence exceeds natural changes by climate and natural geomorphic processes.

5.4 Final remarks and research perspectives

The results presented here on the partly interdependent development of the main aquatic (inland) environments in northeast Germany hold treasures for those seeking to understand the long-term hydrologic dynamics of these ecosystems. Many modern day issues, such as understanding the causes of present hydrologic changes, re-evaluation of land use strategies and implementation of restoration measures, can profit from being looked at from a longer temporal perspective. Periodic hydrologic change is the 'normal status' of the environments discussed. But even if the current regional hydrologic change, probably strongly triggered by man-made global climate change, should be exceptional with respect to its pace and magnitude, historic analogies may help to understand or even foresee complex future landscape dynamics.

As shown above, a number of principle questions on regional palaeohydrology have been posed periodically – gaining in significance each time they resurface. Research over the last c. 20 years has generally made progress in terms of expanding the regional thematic knowledge base. What is new for the region is the growth in well-documented *local* field findings with a broad range of accompanying lab analyses, particularly of geochronological and palaeoecological data. These have been complemented with (semi-) quantitative data on the development of specific hydrologic parameters (e.g. on river-channel patterns or lake-level status) as well as summaries of certain processes (e.g. on peatland formation) in several studies. Even some specific geomorphic and sedimentary-pedologic features were newly discovered for the region (e.g. fluvio-deltaic sequences / fan-deltas in lake basins, palaeosols / hiatuses in peatlands).

New research is needed to refine knowledge on the long-term development of the regional drainage system and its specific aquatic environments. This includes the establishment of hydrologic records with high temporal resolution, which are widely missing in the region so far. In addition, new and particularly promising regional research aspects still abound; some examples are listed in Table 6.

6 Conclusions

(1) Regional research performed on late Quaternary palaeohydrology has largely concentrated on single aquatic environments and single hydrologic parameters so far. But the drainage pattern evolution as a system was rarely in focus. This first comprehensive overview on drainage system evolution in northeast Germany has shown in detail how rivers, lakes and peatlands developed partly interdependently during the last c. 20,000 years.

(2) Until the late Holocene (c. 12th/13th century AD), landscape hydrology in northeast Germany was predominantly driven by climate, including geomorphic and non-anthropogenic biotic factors. Furthermore, initial structural geologic findings suggest that tectonic and halokinetic influence played a more pronounced role on late Quaternary hydrographic evolution than previously assumed. The first *indented* anthropogenic changes of the regional hydrography date from the late Medieval. Strong human impacts on a regional scale occurred from the 18th century AD onwards. In modern times, man's impact exceeds the natural changes caused by natural climatic and geomorphic processes.

(3) Although certain aspects of regional drainage network evolution have attracted considerable interest, the general state of thematic knowledge can be characterised as 'moderate' at best. For example, (a) the late Quaternary development of the large rivers (Elbe, Oder) and most of the medium-scale rivers (e.g. Havel, Spree) is only initially known; (b) high-resolution lake-level records are not yet available; (c) estimations of palaeodischarge and palaeoflood characteristics are widely lacking; (d) the aspect of climatic versus human forcing of past hydrologic processes has rarely been pursued so far; and (e) the role of beavers as effective 'engineers' forming Holocene aquatic landscapes has not yet been approached in the region.

(4) To overcome these deficiencies new research is necessary. Several current and planned projects on river valley and peatland restoration in the region open promising opportunities for the regular integration of palaeohydrologic work into present issues. Future research, more than previously, should aim at developing and integrating multiproxy records from a variety of scientific perspectives. Close links to high resolution records of climate and human impact, which regionally are still to be established, must be encouraged and fostered.

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Younger Middle Terrace – Saalian pre-Drenthe deposits overlying MIS 7 Nachtigall interglacial strata near Höxter/Weser, NW-Germany

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Abstract: Subrosion of Lower Triassic evaporites by subsidence has preserved peat coal and clay as Pleistocene warm and cool stage deposits near Höxter in the northwest-German uplands. There, in the upper reaches of the River Weser, peat coal and clay are sandwiched by gravel and sand of two river terraces; they were exploited in a small pit called Grube Nachtigall. Exploration drillings from 1997/98 enabled investigating a 13.5 m core with the warm and cool stage deposits in respect of sedimentology, palynostratigraphy and radiometric dating. In 2011 the results concerning this "Nachtigall-Complex" have been published separately by two groups of authors also concerned with the project. Since 1994 also analyzing lithostratigraphy and structure of the Pleistocene framework, that is to say about 25 square kilometers of valley floor and hillside landscape, has been the aim of the methodologically independent study presented here. The subrosion structure turned out to be partly limited by faults. The most important Pleistocene mapping units are the deposits of 4 river terraces; in descending chronological order these are: – youngest Upper Terrace, not dislocated – Older Middle Terrace, subsided – Younger Middle Terrace, not subsided – Lower Terrace, not subsided, its sediments partially lying upon subsided Older Middle Terrace deposits. The layers of Nachtigall-Complex likewise directly overlie subsided Older Middle Terrace deposits and with an angular unconformity are overlain by Younger Middle Terrace deposits; they are subsided and deformed. According to generally accepted traditional mapping outside the study site, the Older Middle Terrace has to be assigned to the MIS 8 equivalent in the lower part of the Saalian Complex, and the Younger Middle Terrace to the deep MIS 6 equivalent in the deepest portion of the upper Saalian Complex, i.e. pre-Drenthe. Hence the palynological as well as the radiometric MIS 7 age of the recently defined Nachtigall 1 Interglacial correspond to the lithostratigraphic model inferred from structural analysis.

Die vor-drenthe-zeitliche Jüngere Mittelterrasse über dem MIS 7 Nachtigall-Interglazial – Elemente der Subrosionsstruktur Albaxen bei Höxter/Weser, NW-Deutschland

Kurzfassung: In einer Subrosionsstruktur über Evaporiten des Oberen Buntsandstein sind im Bergland am Oberlauf der Weser zwischen Höxter und Holzminden pleistozäne warm- und kühlzeitliche „Tone"- und „Torfe" erhalten geblieben. Sie trennen hier kaltzeitliche Terrassen-Kiese und -Sande der Weser. Als Rohstoff wurden sie in dem kleinen Tagebau Nachtigall abgebaut. Aus Bohrungen, die 1997/98 zur Erkundung der Lagerstätte u.a. als Rammkernbohrungen niedergebracht wurden, standen für geowissenschaftliche Untersuchungen 33 m nahezu durchgehende Kernstrecke zur Verfügung. Daraus konnten 13,5 m für sedimentologische, palynostratigraphische und radiochronologische Untersuchungen ausgewählt werden. Diese warm- und kühlklimatischen Ablagerungen wurden als Nachtigall-Complex zusammengefasst und 2011 von zwei Autorengruppen des Projekts in zwei methodisch unterschiedlichen Artikeln veröffentlicht. Als geologischer Rahmen wurden 25 km² Flussniederungs- und Hanglandschaft seit 1994 lithostratigraphisch und strukturgeologisch mit dem hier vorgelegten Ergebnis analysiert. Die zuvor schon bekannte Subrosionsstruktur erwies sich als z.T. von Störungen begrenzt. Die wichtigsten pleistozänen Kartiereinheiten sind die Sedimentkörper von 4 Flussterrassen; nach abnehmendem Alter sind dies: – jüngste der Oberterrassen, nicht abgesunken – Ältere Mittelterrasse (ÄMT), abgesunken – Jüngere Mittelterrasse, nicht abgesunken – Niederterrasse, z.T. auf Schichten der abgesunkenen ÄMT. Die Schichten des Nachtigall-Complex liegen direkt über Schichten der ÄMT und werden von Schichten der Jüngeren Mittelterrasse diskordant überlagert; sie sind abgesunken und verformt. Außerhalb des Untersuchungsgebietes wird die ÄMT – ihrer Lage in der Terrassentreppe gemäß – dem Marinen Isotopen-Stadium MIS 8 im unteren Teil des Saale-Komplex zugeordnet, die Jüngere Mittelterrasse dem älteren Abschnitt von MIS 6 im tiefen oberen Teil des Saale-Komplex vor dem Drenthe-Stadium. Damit wird das sowohl palynostratigraphisch als auch radiometrisch ermittelte MIS 7-Alter des zwischen beiden liegenden jüngst definierten Nachtigall 1 Interglazial lithostratigraphisch gestützt.

Keywords: *Younger Middle Terrace, Wehrden-Niveau, Nachtigall 1 Interglacial, Older Middle Terrace, Reiherbach-Niveau, Saalian Complex, pre-Drenthe period, MIS 6, MIS 7, MIS 8, mapping, lithostratigraphy, structural analysis, river terrace, staircase-position, stack-position, Weser upper reaches, Albaxen subrosion structure, NW-Germany*

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1 Introduction

If we look back to the starting point of the present study in 1998, we find:

- a pit for exploiting peat coal and clay for manifestly more than a century —
- ice age deposits of interglacial character taken note of for at least a hundred years —
- questionable assignment of the warm period strata to the Pleistocene stratigraphy —.

We also find a drilling project for expanding the area of open cast working to the north. Jochen Lepper located and supervised the drillings, evaluated the drilling project (Lepper 1998, 2011) and made available core KB 1 for scientific investigation. He also coordinated the studies that are based on structural analysis plus lithostratigraphy, on sedimentology, on palynology plus palynostratigraphy, and on radiometric dating. An initially envisaged common publication regrettably has not been realized as it turned out to be too complex and too long. Special mention should be made of the comprehensive palaeobotanic analyses of Helmut Müller, by which he unravelled intricate depositional environments of peat coal and clay. Thus he determined the stratigraphic position of Nachtigall-Complex, got aware of its contemporaneity with the succession of Göttingen / Ottostraße (Grüger 1996), and found out the analogy with the Velay profile, Massif Central, south-central France (Reille et al. 2000, Beaulieu et al. 2001). His death called for finally supporting his palynostratigraphic conclusions. The way in which the authors accounted for the bio- and chronostratigraphic results has been outlined by Kleinmann, Müller, Lepper & Waas (2011).

Peter Rohde could discuss thoroughly his results of terrace stratigraphy and his ideas concerning the source of certain fine-grained clastic matter with Helmut Müller. His terrace model is supported by results of the deceased field geoscientist and highly appreciated colleague Wolfgang Thiem, Hannover University, Geographic Institute (Thiem 1988; Rohde & Thiem 1998). Wolfgang Thiem has taken active part in the field work in Nachtigall pit. The present paper has been initiated by Jochen Lepper, who above all contributed the geological setting.

We would like to point to the end of text where abbreviations are explained.

2 Study site

The study area in the Weser Valley in NW-Germany is situated in the Mesozoic uplands, about 60 km south of their northern edge. Here, downstream of the medieval town of Höxter, about 25 square kilometers of valley floor and hillside landscape are included (Fig. 1). For orienting oneself, a more prominent place may be Corvey Abbey near Höxter, a world heritage name because of the 1130 years old, well preserved Carolingian westwork.

The so called Zeche NACHTIGALL (Nightingale pit) is at a place that since the middle of the 16th century is called „Bergstette“ in terms of a prospective site (Michael Koch, Höxter, personal communication 2010). It has been exploited maybe since 1795 (⟹ Tonenburg – in Wikipedia [27 Febr.2012]), maybe from the 1840s (Schlegel 1997). Documentarily mentioned, subsurface peat-coal mining was practiced at least since 1857 (Schlegel 1997: map of mine 1866) up to a date well before 1884/86 (Carthaus 1886), and about 1920 until 1923. Besides, “clay” has been extracted temporarily by open cast work having started some time between 1866 and 1884. In 1895 the Feldbrandziegelei Johann Buch was founded. Recently the brickworks Ziegelwerk Buch at Höxter-Albaxen, in its small Nachtigall clay pit (Fig. 1, 2, 3) extracted minor quantities of peat coal, but primarily carbonaceous pelites as before.

It is worth noting that at a distance of about 150 km, at Witten-Bommern, Ruhr district, there is another ancient pit with the same name Zeche Nachtigall. 1743–1892, this former colliery produced carboniferous hard coal, and 2003 has been transformed into the identically named Industrial Museum.

3 Previous research

The peat and clay deposits of Nachtigall pit have attracted interest since the 19th century (Dechen 1884; Carthaus 1886; Koken 1901; Siegert 1913, 1921; Soergel 1927, 1939; Stoller 1928). In 1908/09 and 1927 first thorough studies were conducted to present the geological map 1 : 25 000, sheet Holzminden, today‘s number 4122 (Grupe 1912, 1929). Much later the site-specific extraordinary Pleistocene succession in the south of the mapped area was dealt with again (Brelie et al. 1971, Müller 1986, Rohde 1989, Rohde & Thiem 1998). Mangelsdorf (1981) studied geologically the Nachtigall site and investigated palynologically the succession of its Middle-Pleistocene organic deposits. The hydrogeological aspect of the area has been investigated by Straaten (1982) and Fischer et al. (1990). The latest information was given by an unpublished report by Lepper (1998) in the context of the company’s request to the authorities for extending the hitherto granted exploitation of the carbonaceous brick clay. For archived supporting documents see Lepper (2011).

4 Borehole evidence

The present structural study and map are based on boreholes of very different informative value (Fig. 2, 3) that may be divided into 5 groups.

1) Three percussive core / wash boreholes as exploration boreholes for exploring the extension of the clay and peat deposit, additionally used as groundwater observation-wells / Grundwassermeßstellen (Lepper 1998, 2011).

   KB 1 (1998), cored 10–43 m; the core was investigated sedimentologically and palynologically in great detail (Kleinmann et al. 2011), partially also by radiometric dating (Waas, Kleinmann & Lepper 2011); Gauss-Krueger grid reference R 35 27 755 / H 57 41 810, elevation +108.55 m NN; geographic coordinates 51°48′34.8″ N / 9°24′04.7″ E.

   KB 2 (1998), cored 10–13 m / 15–16 m / 18.2–32 m, washed for the rest; Gauss-Krueger grid reference R 35 27 855 / H 57 41 805, elevation +105.27 m NN.

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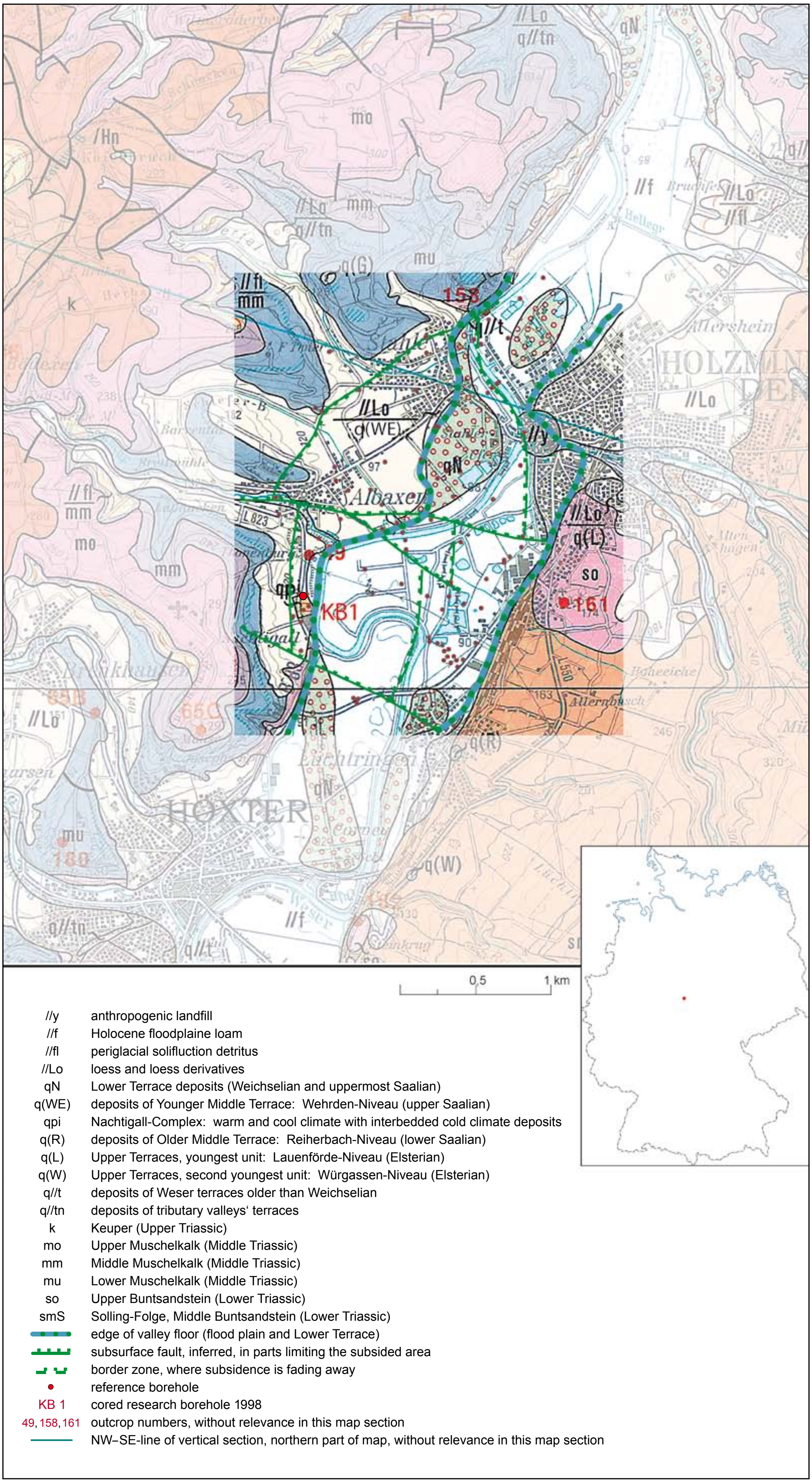

Fig. 1: Geological Setting. Taken from Lepper & Mengeling *(1990), with modifications. As to further details see Fig. 2.*

Abb. 1: Geologischer Rahmen. Nach Lepper & Mengeling *(1990), mit Änderungen. Weitere Signaturen siehe Abb. 2.*

KB 3 (1998), partly cored 15–17 m, cored 17–28 m depth, washed for the rest; Gauss-Krueger grid reference R 35 27 888 / H 57 41 665, elevation +102.22 m NN.

2) Three additional wash boreholes as exploratory and groundwater observation wells; these are B 1, B 2, B 4 (1998) of 36 m / 27.5 m / 31 m, having penetrated the clay and peat deposit (Lepper 1998).

3) Eight percussion boreholes S 1 – S 8 (1997), up to 13 m deep, for groundwater observation and for investigating the strata that overlie the clay and peat deposit (Lepper 1998).

4) Approximately 50 boreholes for gravel-exploration and groundwater survey in the surroundings of the clay and peat deposit (1952 and 1972-1998, data bases of the state geological surveys Landesamt für Bergbau, Energie und Geologie, Hannover, as well as Geologischer Dienst Nordrhein-Westfalen, Krefeld).

5) According to Grupe (1929) five historical exploration boreholes in the context of coal mining activities, 54 m (about 1905) and 18 m / 20 m / 28 m / 29 m (about 1919).

5 Methods

The geological study is principally concerned with the structural analysis of a 25 km^2 area around Nachtigall pit with its deposits of interglacial character (Nachtigall-Succession, Kleinmann et al. 2011). It distinguishes lithostratigraphic units and decodes the intricate structure of their horizontal and vertical relative positions. Finally it aims at assigning the elements of the structural geological model to Pleistocene stratigraphy (Litt et al. 2007) and to the chronostratigraphically calibrated MIS system of Marine ^{16}O-Isotope Stages (e.g. Petit et al. 2000).

The lithostratigraphic and the structural investigations comprise geological mapping in the pit and applying the detailed geological map 1 : 25 000 by Grupe (1929) and also the diploma thesis by Straaten (1982). Moreover it comprises evaluating historical mine plans as well as evaluating, interpreting and correlating the petrographic borehole data of all sorts of borehole records available (see above). Ground level elevation refers to German Ordnance datum NN (Normal Null) mainly according to Topographic Map 1 : 5 000, partly to Topographic Map 1 : 25 000.

Assigning the sedimentary bodies to specific units of the Pleistocene stratigraphy has been carried out on the base of

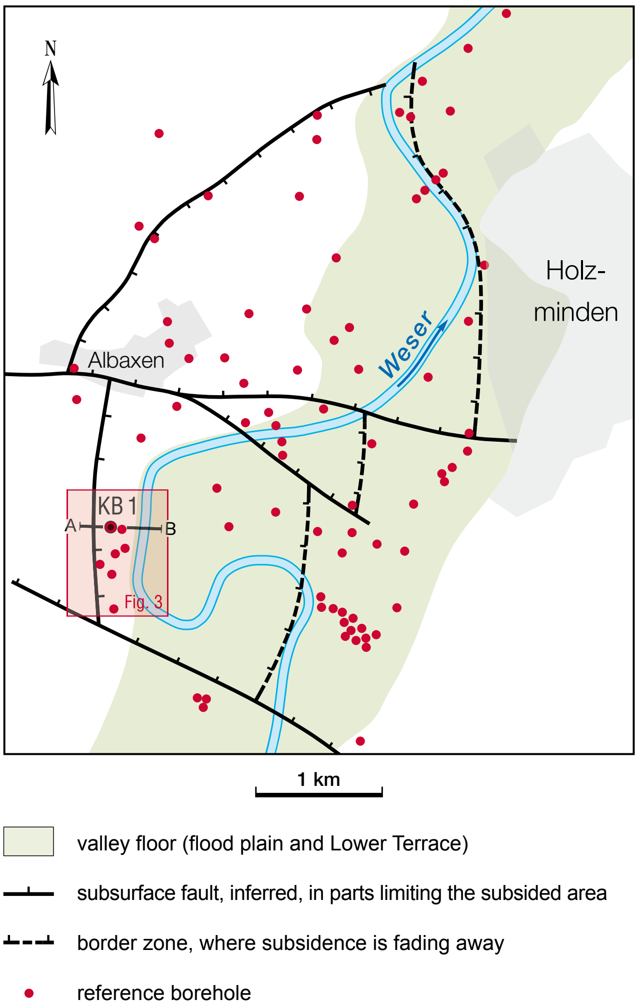

Fig. 2: Albaxen subrosion structure. Central part of Fig. 1, augmented. Mapping 2008 by Peter Rohde.

Abb. 2: Subrosionsstruktur von Albaxen. Innerer Teil von Abb. 1, vergrößert. Kartierung 2008, Peter Rohde.

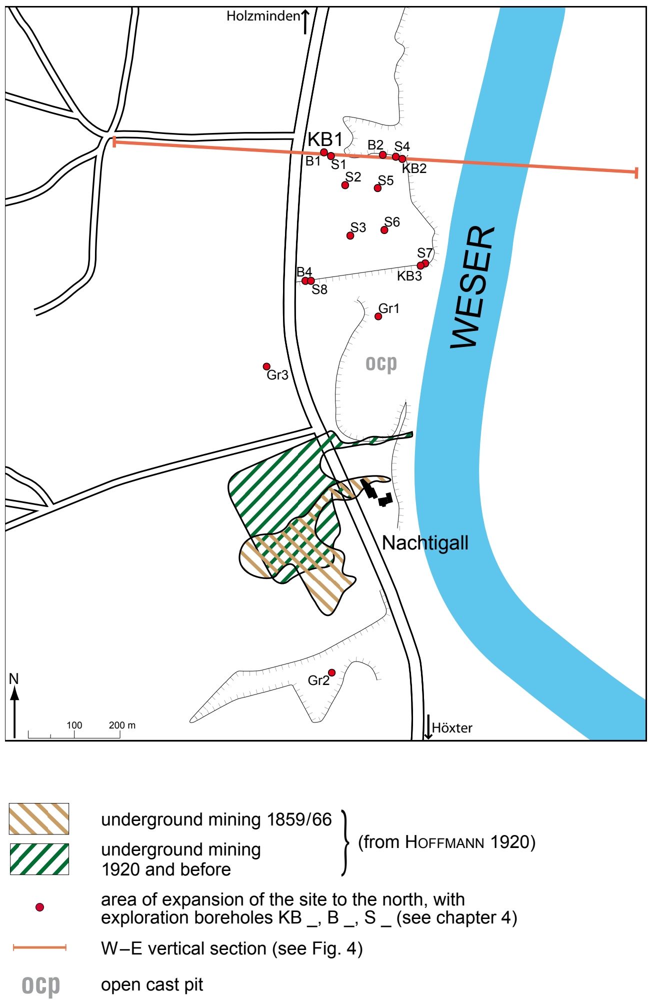

Fig. 3: Plan of Nachtigall pit. Topography from Deutsche Grundkarte 1 : 5 000, Nachtigall, 3526 Rechts, 5740 Hoch.

Abb. 3: Geländegrundriss der Grube Nachtigall. Topographie nach Deutsche Grundkarte 1 : 5 000, Nachtigall, 3526 Rechts, 5740 Hoch.

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Tab. 1: Quaternary lithostratigraphic units in Albaxen site near Höxter, NW-Germany. The site includes the Albaxen subrosion structure with the warm, cool, and cold stage sedimentsts of Nachtigall-Deposit. Data concerne base level, surface level and thickness of lithostratigraphic units (Rohde 2008). Maximum subsidence suggested 62–65 m (? 78 m).

Tab. 1: Quartärzeitliche lithostratigraphische Einheiten im Gebiet Albaxen bei Höxter in NW-Deutschland. Das Gebiet umfasst die Subrosionsstruktur von Albaxen mit den warm-, kühl- und kaltzeitlichen Nachtigall-Lagerstättenschichten. Erfasst sind die Einheiten nach Basis- und Oberflächen-Höhenlage und Mächtigkeit (Rohde 2008). Die Absenkung durch Subrosion beträgt maximal vermutlich 62–65 m (? 78 m).

deposits	**base:** meter relative to		**recent surface**	**recent thickness:** meter		
	NN	WOF [1)] {...} km 27 → 117	meter NN	mean	minimum	maximum
Upper Terraces, Lauenförde Niveau	116 → 126	{28 → 37}	132 (?)			
Older Middle Terrace (Reiherbach Niveau) ditto: subsided	103 → 110	{15 → 22}				
- borehole KB 3 / 1998 - borehole KB 2 / 1998 - borehole GRUPE 1 (1919?)	< 74.2 < 73.3 < 69.9	< -13.6 < -14.4 < -18			≥ 4 ≥ 5 ≥7.3	
OMT, subsided and undistinguishably covered by Lower Terrace OMT, subsided, base extremely low-lying	71.0 → 75.5 40.8/38.0/?25.1			11.4 (n=8)	8.0	13.6 46.2
Nachtigall-Deposit, subsided ditto: top low-lying due to erosion	72.5 → 82.4		94.2 → 99.0 87.1 → 89.1	19.2 (n=3)	13.3 6.7	25.0 8.7
Younger Middle Terrace (Wehrden Niveau)		{-4 → -1}				
- borehole KB 2 / 1998 - borehole B 2 / 1998 - borehole KB 3 / 1998	87.07 87.56 89.12	-0.7 -0.1 +1.4	94.7 95.3 92.2	6.0(n=3)	7.4 7.7 2.8	
solifluction matter incl. "loess" "loess": (sandy) loess, partly reworked	92.2 → 99.0		ground level exceeding that of Lower Terrace	10.4 (n=10) 2.5 (n=12)	9.0 2.0	11.0 3.1
Lower Terrace ditto: top low-lying due to erosion and covered with floodplain loam	 76.2 → 80.3	{-11 → -6} ca -9	86.2 → 88.5 84.5 → 87.0	ca 9 7.6 (n=33)	 5.0	 10.2
floodplain loam (Holocene)	84.5 → 87.0		(88.5?→)86.2(84.8)	2.6 (n=40)	0.6	3.5
[1)] WOF: Water-level of over-bank flooding; corresponding to Mittleres Hochwasser MHW (Mean annual highest water-level) 1941–1980, Weser-km 73–83			88.5 → 84.8			

mapping experience both in the uplands of Lower Saxony with their fluvial terraces (e.g. Rohde 1989) and in the lowlands with their drift sediments and, moreover, in the borderland in between. For broad understanding the present article refers to the MIS-system (Marine isotope stages) as has been taken into account by Litt et al. (2007).

6 Geological setting: the solid rock

The Nachtigall clay pit is situated at the western flank of the anticlinal structure "Solling-Gewölbe" built up by red beds of the Buntsandstein group (Germanic Lower Triassic) that comprises a clastic, pelitic sequence with evaporitic interbeds in its upper portion (Fig. 1). Only 0.7 km west of Nachtigall pit shallow marine sediments of the Muschelkalk group (Germanic Middle Triassic) that overlays the Buntsandstein group are outcropping. The bottom of the Buntsandstein is built up by an Upper Permian (Zechstein) succession of marine evaporites and carbonates underlain by Lower Permian (Rotliegend) red beds which rest unconformably on the folded Variscan basement.

After updoming of the Mesozoic sediment pile in the course of late Cretaceous to Tertiary times, the antecedent River Weser subsequently incised an isoclinal valley from late Tertiary to late Pleistocene in the zone between the western flank of the Buntsandstein-anticline and the outcropping Lower Muschelkalk to the west of it. In the study area the Weser valley generally follows the strike of the Upper Buntsandstein beds with pelites and evaporites easily to be eroded.

Within the Upper Buntsandstein mudstone and evaporites, leaching processes have affected chloride and sulfate layers. The process started in the evaporite outcrop area at the western flank of the "Solling-Gewölbe" and successively prograded from east to west and from surface to subsurface according to the regional dip. The original evaporite thickness might have been up to approximately 100 m.

Therefore in the western part of the present valley and on the adjacent hillside this general geological setting is complicated by a subrosion structure as a sediment-„trap" of Pleistocene age which cannot be recognized morphologically at the ground surface but which is reflected by the „trapped"

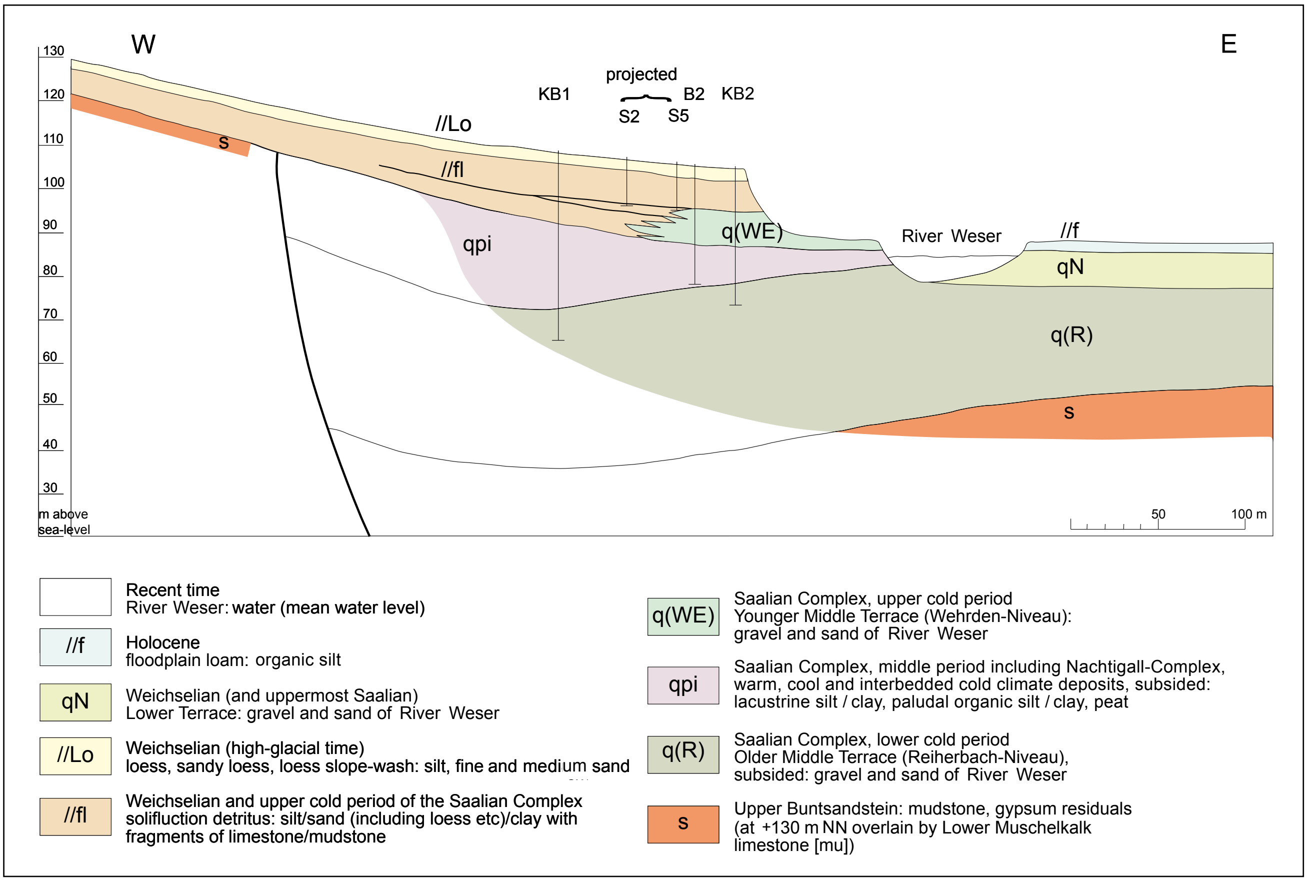

Fig. 4: MIS 7 Nachtigall interglacial deposits between Saalian Older Middle Terrace and Younger Middle Terrace of the River Weser in stack-position (!not: staircase-position). Cross section; main focus upon Nachtigall concession near Albaxen, N of Höxter, NW-Germany. Draft: Peter Rohde in 2008. As to the up-to-date stratification of KB 1 borehole profile see Kleinmann *et al. (2011) and present paper, section 7.2.*

Abb. 4: Ablagerungen des MI-Stadium 7 mit Nachtigall-Interglazial. Vorkommen zwischen der Älteren und der Jüngeren Weser-Mittelterrasse aus der Saale-Kaltzeit in Stapel-Lagerung (und nicht Treppen-Lagerung). Vertikaler Querschnitt, im Wesentlichen durch das Gelände der Konzession Nachtigall bei Albaxen nördlich von Höxter in NW-Deutschland. Entwurf: Peter Rohde i.J. 2008. Zur aktuellen Gliederung der Schichtenfolge in Bohrung KB 1 siehe Kleinmann *et al. (2011) sowie Abschnitt 7.2 der vorliegenden Veröffentlichung.*

sediments („Albaxen subrosion structure“, Fig. 1 and 2) and is dealt with in chapter 7.

7 The Quaternary deposits within and adjacent to the subsided area

The fossil subrosion structure is 5 km by 2 or 3 respectively, the long axis following the SSW–NNE trending Weser valley (Fig. 1 and Fig. 2; Rohde 2008). The Pleistocene depositional environment comprises two different sedimentary complexes belonging to different morphological units.

7.1 Today's valley floor

This flat ground extends as far as the mainly Weichselian Lower Terrace and the Holocene flood plain (Fig. 1). The surface level of the Lower Terrace is at +86.2 to +88.5 m NN, whereas that of the flood plain only is at 86.2 m on average (84.8–88.5 m; Table 1). The top layers of the **Lower Terrace** comprise 2 m, in some parts 3 m of sandy loess-like loam, deposited by a periglacial river in the braided river plain during a period of very low river activity. Yet in the flood plain there have been accumulated 2.5 m (min: 0.5 m, max: 3.5 m) of organic sandy or argillaceous silt, deposited as over-bank sediments during floods of the meandering Weser. Beneath both of them there are periglacial braided river sand and gravel of the Lower Terrace with a supposed base level of +76.2 to +80.3 m NN, that is 9 m below recent Water-level of over-bank flooding (WOF), on average.

In a certain area similar sand and gravel are also found below this level (Fig. 2; Fig. 4: q(R)), in eight boreholes down to +71 m NN, and in a few boreholes still deeper, with a minimum value of +40.8 and possibly even +25.1 m NN. The resulting overall thickness of sand and gravel is 11.5 m on average and 46 m as maximum in comparison with 7.5 m concerning the Lower Terrace separately. This difference of thickness is explained by subsidence due to subrosion, the lower part of the sediment stack interpreted as Older Middle Terrace sediments (see below); the actual situation suggests deposition of only that series during subsidence of the bedrock beneath it.

7.2 Today's hillside

The subrosion structure is not confined to the valley floor, but westward also comprises the lower part of the hillside downhill the Muschelkalk escarpment and also includes the Nachtigall open cast pit. The surface of the slope is built up by **loess** and **solifluction detritus** (gelifluction detritus) of limestone and mudstone with thin slopewash intercalations

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Fig. 5: Temporary outcrop of Nachtigall 1 Interglacial with peats A, B, C and D. As to lithostratigraphy see chapter 9. Folding ruler 1 m. Photo Jochen Lepper, 18 August 1998. Approximate thicknesses are: - peat (D) 1.40 m, upper part not visible, - clay and silt 1.90 m, - peat (C) 0.05 m, - clay and silt 1.00 m, - peat (B) 0.40 m, - clay and silt (lower part dark) 0.85 m, - peat (A) 1.30 m, lower part not visible.

Abb. 5: Kurzzeitiger Aufschluss des Nachtigall 1 Interglazial mit den Torfen A, B, C und D. Zur Lithostratigraphie siehe Kapitel 9. Messstab 1 m. Foto Jochen Lepper, 18.8.1998. Die Mächtigkeit der Schichten beträgt etwa: - Torf (D) 1,40 m, oberer Teil nicht sichtbar, - Ton und Schluff 1,90 m, - Torf (C) 0,05 m, - Ton und Schluff 1,00 m, - Torf (B) 0,40 m, - Ton und Schluff (unterer Teil dunkel) 0,85 m, - Torf (A) 1,30 m, unterer Teil nicht sichtbar.

of red, originally eolian sand. These periglacial sediments in the northern pit area are 9–11 m thick (Fig. 4, Tab. 1) and include an Eemian palaeosol of decalcification. They are down-dipping towards the adjacent valley floor in the east. In the open cast pit they overlie their substratum at an angular unconformity, as the substratum got trough-shaped by subrosion, and in its eastern part was tilted to the west and eroded at its former surface.

Concerning the substratum up to 25 m of lacustrine silt, clay and mud are interbedded with several seams of slightly calciferous peat. In the recent quarry all these sediments (Fig. 5) are dipping in westerly directions towards the slope; the bedding would assign the investigated section to the eastern flank of a very shallow syncline if compared with the former subsurface mining area farther in the south. The interglacial, interstadial and included stadial layers, in borehole KB 1 core comprising 10 m of sediments, by A. Kleinmann have been termed Nachtigall-Succession (Kleinmann et al. 2011; see Chapter 9). The whole 13.5 m section studied palynologically in KB 1 core by A. Kleinmann has been termed Nachtigall-Complex; it encloses 3.5 m of cold and cool climate lacustrine mud and fen peat in its upper part. – With additional cold climate lacustrine (7.22 m) and ambiguous lacustrine or slope sediments (4.28 m) the KB 1 core section is called **Nachtigall-Deposit** in the present paper (Tab. 1).

In summary the terms used for sections of KB 1 core and in the present paper are

Nachtigall-Succession: contains the deposits of interglacial / interstadial character (36.00–26.02 m); Nachtigall-Complex: strata with arboreal pollen >20%, i.e. at least „tree tundra“ (36.00–22.50 m); Nachtigall-Deposit: contains peat and clay in terms of natural mineral resource (36.00–11.00 m).

Within the Nachtigall-Succession and its lowest section, i.e. Nachtigall 1 Interglacial, the base of Allochthonous Unit I (cf. Chapter 9) includes dropstones of Weser characteristics without nordic material (pebbles 2–20 cm, sample K.D. Meyer, August 1998, analysis P. Rohde & J. Lepper, December 1998). Similarly the base of Allochthonous Unit II includes an erratic boulder of Swedish Dala quartzite (Grupe 1929: “Bohrung 4”, meaning a shaft, but localisation regrettably not to be verified). It must have been transported primarily by the Elsterian inland ice, then might have been deposited in Thuringia and ice drifted from there on a lake or by the river. Alternatively it might have been drifted directly from the north to Nachtigall site on an Elsterian ice dammed lake (e.g. Winsemann et al. 2011; see also Chapter 10) and perhaps furthermore slid down at the slope above the pit area from an unknown Elsterian secondary deposit. The extraordinary lateral supply of mineral matter in warm climate periods still is worth discussing (see Chapter 10).

Adjacent to the present valley floor, the Nachtigall-Depos-

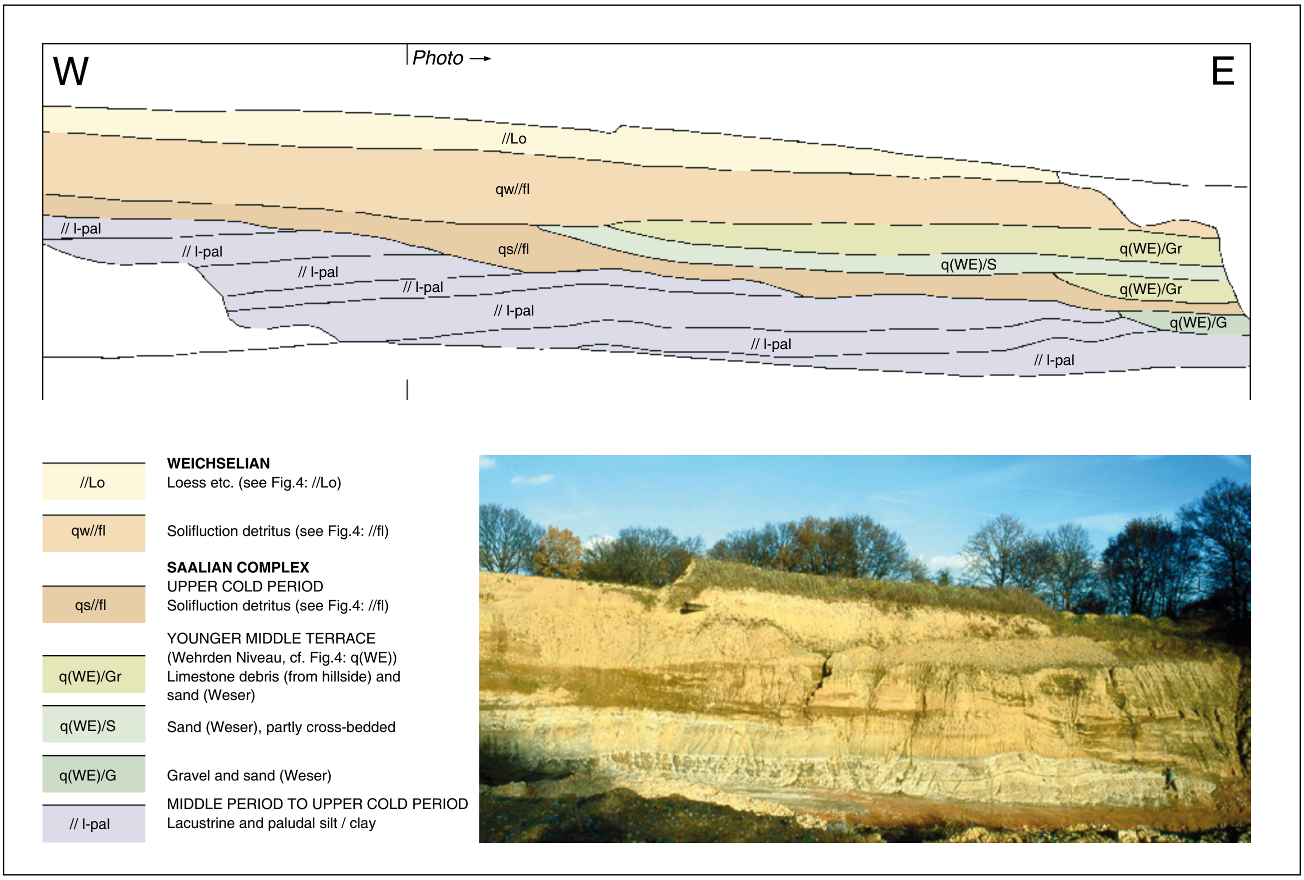

Fig. 6: Younger Middle Terrace deposits interfingering with solifluction detritus. Quarry face, autumn 2011. Drawing: – left margin R 35 27 760, H 57 41 690, elevation +108.7 m NN, height above bottom 16.4 m, – right margin R 35 27 875, H 57 41 685, elevation +103.5 m NN, – width 115 m. Photo: – left margin R 35 27 790, H 57 41 700, elevation +107.3m NN, – width 85 m. Above numerical data approximate only. Folding rule in middle of photo: 1 m; at right margin: 2 m. Photos Jochen Lepper, 7 Oct. and 30 Nov. 2011. Photo interpretation Peter Rohde. For the first time the quarry face exposes the complete overburden stack of the Nachtigall-Deposit in a W–E section, appr. hundred meters south of section AB (Fig. 3, Fig. 4).

Abb. 6: Ablagerungen der Jüngeren Mittelterrasse, verzahnt mit Solifluktionsschutt. Grubenwand, Herbst 2011. Zeichnung: – linker Rand R 35 27 760, H 57 41 690, Gelände +108,7 m NN, Wandhöhe 16,4 m, – rechter Rand R 35 27 875, H 57 41 685, Gelände +103,5 m NN, – Wandlänge 115 m. Foto: – linker Rand R 35 27 790, H 57 41 700, Gelände +107,3 m NN, – Wandlänge 85 m. Alle Lage-Angaben als Circa-Werte. Messstab im Foto, Mitte: 1 m; rechter Rand: 2 m. Fotos Jochen Lepper, 7.10. und 30.11.2011. Foto-Auswertung Peter Rohde. Erstmals erschließt die Grubenwand die vollständige Schichtenfolge über der Nachtigall-Lagerstätte in einem W–E-Schnitt etwa 100 m südlich des Schnitts AB (Abb. 3, Abb. 4).

it partially was eroded at the top. Forming an angular unconformity, the erosion surface partially has been covered by up to 8 m – or perhaps more – of fluvial gravel and sand attributed to the younger Saalian Weser terrace (**Younger Middle Terrace** or Wehrden-Niveau, Table 1, Fig. 4, Fig. 5; Rohde 1989). The strata have not been dislocated, their base level is at +87 to +89 m NN (Grupe 1929: geol. Map), i.e. minus 0.7 m to plus 1.4 m relative to WOF. These dates are in accordance with results of a reference survey (Rohde 1989). Partially the sediments interfinger with the solifluction detritus mentioned above. Later, the fluvial gravel and sand were covered with different periglacial sediments, which also capped the western, less or not eroded part of the Nachtigall-Deposit.

In KB 1 borehole above the Nachtigall-Deposit no fluvial, but only sediments of the slope were encountered (Fig. 4). In contrast KB 2 core drilling revealed Younger Middle Terrace sediments top-down consisting of:

1.9 m medium-grained sand with silt, fine sand, coarse sand, little fine grit, red brown, fluvial (Weser)

5.7 m gravel, with sand and coarse pebbles, red to grey, fluvial (Weser).

This succession is typical for a fluvial sediment series formed during periglacial conditions.

The Nachtigall-Deposit is underlain by fluvial gravel and sand assigned to the older Saalian Weser terrace (**Older Middle Terrace** or Reiherbach-Niveau, Table 1; Rohde 1989). By the pit-extension drillings its base was not reached, lying deeper than +74 m to +70 m NN. In the valley-floor where this gravel and sand are also found their base might be at about +41 m NN minimum (or even +25 m NN, see above).

7.3 Basin structure and development

The two morphological elements, that means valley floor and hillside (section 7.1, section 7.2), are divided by a young scarp at the foot of the hillside (Fig. 4: 10 m east of KB 2 borehole) and close to it by an escarpment visible as steep western river bank and extending down below the river surface (Fig. 4: 80 m east of KB 2). Both faces are due to fluvial downcutting. The latter, being the older one, originated when the Lower Terrace river formed the deep incision and the wide periglacial braided river plain at the end of the

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Saalian period. The former resulted from partial erosion of the Lower Terrace sediment stack near to its former surface at late-glacial time.

Yet the overall structure is complex. Information from outcrops, boreholes and from the geological map especially of the pit area suggest that subsidence by subrosion affected only the Older Middle Terrace and the Nachtigall-Deposit. Faults (Fig. 1, Fig. 2) are supposed to have been effective as slides during subsidence. At least partially they may have been caused by subrosion. Only the aforementioned sediments were dislocated. There are neither indications of subsidence concerning Elsterian deposits (Tab. 1; Rohde 1989: Lauenförde Niveau; Grupe 1929: geol. Map: Obere Terrasse between +120 m and +132 m NN) nor are there indications concerning the sediments of Younger Middle Terrace and of Lower Terrace and of the periglacial slope deposits. In the valley floor, larger differences of thickness of Pleistocene deposits beneath the Holocene flood plain also must be explained by subsidence. There, this process exclusively has affected the Older Middle Terrace sediments underlying the Lower Terrace sediments in great parts. Towards the east, beneath floodplain loam and Lower Terrace gravel and sand, the preservation of Older Middle Terrace deposits by subrosion is fading away. It is important to note that Older Middle Terrace sediments subsided by subrosion have been identified at several sites of the Weser valley, so near Polle, Bodenwerder and maybe also near Hameln (Thiem, Fleig, Kliem partly joint in Rohde & Thiem 1998: 116, 120, 141; besides orally by W. Thiem).

As to maximum subsidence there is no validated evidence. From three dates of Older Middle Terrace sediments a subsidence of ca 65 m, perhaps even of ca 78 m, relating to their undisturbed base level in an upper Weser reference survey by Rohde (1989) are worth being considered.

Generally the base of the Quaternary deposits seems to be irregular as a result of different degrees of subrosion. There might exist various sinking centers within the basin. Tracing a contour map of the base of the Quaternary sediments in a coherent manner regrettably remains left till additional boreholes provide less scattered and less discontinuous data distribution.

7.4 Postscript findings in autumn 2011

The advance of the open cast work to the north of the pit area made possible studying the complete overburden stack of the Nachtigall-Deposit in a vertical and straight quarry face for the first time. The study corroborates principally the borehole data evaluation as presented in the cross section of Fig. 4. Beyond that, explicitly examined it reveals the onlapping superposition of younger units on deformed and hill-ward dipping clay/silt layers of Nachtigall-Deposit. On an erosive basal face the onlapping units comprise horizontal Younger Middle Terrace gravel and sand in the east as well as the lowest solifluction unit in the west. Most instructive to see how the younger fluvial sediments are directly supplied by limestone debris, that have been transported downslope by periglacial solifluction. The overlying fluvial red sand displays evidently that the periglacial fluvial activity weakened in the final stages as generally is known from the latest phase of terrace building accumulation, e.g. concerning Lower Terrace top layers (Rohde & Becker-Platen 1998: 42; 2002: 106). Fig. 6 summarizes our supplementary findings. It displays the quarry face in direction of the hill slope and across the strike of the clay deposits, as exposed in autumn 2011. It refers to section 2 of chapter 7 and supports evidently the suggested synthesis. The drawing is based on reconnaissance examination in the pit, supplemented by a relevant photo evaluation.

At the time mentioned the brickworks being in liquidation, the outcrop is in risk to get lost, unfortunately.

8 Discussion: Younger Middle Terrace

The Younger Middle Terrace seems to be a weakly developed geologic and hardly to distinguish morphologic element. It might be overlooked therefore in most cases. Its **base** in staircase position is minus 4 m (to minus 1 m) relative to WOF mapped by a reference survey (Rohde 1989). That implies only ca 5 m of difference to the base of the Lower Terrace and a position lower than the Lower Terrace surface. Like the sediment package of the Older Middle Terrace it is one of the geological units that build up the slope along the side of a recent valley. Consequently its **surface** lying at the bottom of a slope is hidden by slope deposits and / or loess. Possibly in upland positions Younger Middle Terrace deposits with a rather low surface level escape the geologist‘s notice at the edge of the Lower Terrace or are visible only in outcrops.

Thus field mapping is in risk not to discern the deposits where both terraces really are existing and then simply defines a mapping unit ‚Middle Terrace‘ in the sense of the common Older Middle Terrace, the more so as this is supposed to be the main Saalian fluvial formation in Lower Saxony (yet Wansa in Litt et al. 2007: 38). If the sediments of different terraces occur as stack, as is the case due to subrosion (W. Thiem concerning Weser valley: Thiem 1988; Rohde & Thiem 1998) and generally is the case in lowland regions, they might be complete, but inevitably are difficult to identify in case they lack interbedded interglacial deposits.

9 Conclusion: MIS 8 -, MIS 7 - and prior to Saalian inland glaciation MIS 6 -deposits

The fluvial deposits of the two Middle Terraces have to be assigned stratigraphically to pre-Drenthe-stadial periods of the Saalian Complex (Litt & Turner 1993, Litt et al. 2007). In and around Nachtigall pit the deposits match the stratigraphic units as tabulated below (Tab. 2).

The direct contact of the sediments which in Tab. 2 are accentuated by heavy print, contrasts with common terrace staircases in uplands; this fact must be emphasized sufficiently. In their special position demonstrated in Fig. 4, the two terraces corroborate that the Nachtigall-Succession with its interglacial and stadial layers has been deposited during Marine Isotope Stage 7 (MIS 7) and neither during the Holsteinian (MIS 9e, Geyh & Müller 2005) nor during the Eemian (MIS 5e) interglacial periods. The correlation of the Holsteinian interglacial alternatively to MIS 11 cannot be discussed here.

As has been mentioned the fluvial activity leading to the

Tab. 2: Deposits of Nachtigall site matched to the stratigraphic scheme for North Germany. Loess and solifluction deposits (Weichselian, Warthe and period of Younger Middle Terrace) omitted.
Wehrden Niveau, Reiherbach Niveau: according to Rohde 1989.
Bouchet 2, Bouchet 3: according to Beaulieu et al. 2001 and others (see chapter 9).
N. Nachtigall
⇓ deposits subsided by subrosion.

Tab. 2: Schichteinheiten im Gebiet der Grube Nachtigall im stratigraphisches Schema für Nord-Deutschland.
Löss und Solifluktionsablagerungen (Weichsel-Kaltzeit und Warthe-Stadium und Zeit der Jüngeren Mittelterrasse) nicht einbezogen.
Wehrden-Niveau, Reiherbach-Niveau: nach Rohde (1989).
Bouchet 2, Bouchet 3: nach Beaulieu et al. 2001 und weiteren (s. Kap. 9).
N. Nachtigall
⇓ Ablagerung durch Subrosion abgesenkt.

<table>
<tr><th colspan="4">NORTH GERMANY
Litt et al. 2007, with regard to ongoing discussion
modified according to proposal of Kleinmann et al. 2011</th><th colspan="2">NACHTIGALL SITE</th></tr>
<tr><td>Weichselian</td><td>MIS 2-5d</td><td colspan="2">cold period with interstadials</td><td colspan="2">Lower Terrace</td></tr>
<tr><td>Eemian</td><td>MIS 5e</td><td colspan="2">Interglacial</td><td colspan="2">[decalcification on slope]</td></tr>
<tr><td rowspan="10">Saalian Complex</td><td rowspan="4">MIS 6</td><td rowspan="2">Warthe- + Drenthe- Stadium</td><td></td><td colspan="2">[? Lower Terrace, older part]</td></tr>
<tr><td>[glaciation]</td><td colspan="2"></td></tr>
<tr><td colspan="2" rowspan="2">cold stage
[?Delitzsch-Phase]
with interstadials</td><td colspan="2">Younger Middle Terrace
[Wehrden Niveau]</td></tr>
<tr><td colspan="2">Stadial of N.-Deposit ⇓
Stadials + 5 Interstadials of
N.-Complex / -Deposit ⇓</td></tr>
<tr><td>MIS 7a</td><td colspan="2">warm stage [cf. Velay, France: Bouchet 3]</td><td rowspan="3">Nachtigall-Succession ⇓
[i.e. lower part of N.-Complex / lowest part of N.-Deposit]</td><td>N.2 Interstadial</td></tr>
<tr><td>MIS 7b</td><td colspan="2">cool stage</td><td>Albaxen Stadial</td></tr>
<tr><td>MIS 7c</td><td colspan="2">Interglacial [cf. Velay, France: Bouchet 2]</td><td><u>N.1 Interglacial</u></td></tr>
<tr><td>MIS 7d, e</td><td colspan="2">[cool stage, warm stage]</td><td colspan="2"></td></tr>
<tr><td>MIS 8</td><td colspan="2">cold stage</td><td colspan="2">Older Middle Terrace
[Reiherbach Niveau] ⇓</td></tr>
<tr><td>MIS 9a-d</td><td colspan="2">[2 warm, 2 cool stages]</td><td colspan="2"></td></tr>
<tr><td>Holsteinian</td><td>MIS 9e</td><td colspan="2">[Interglacial]</td><td colspan="2"></td></tr>
<tr><td rowspan="3">Elsterian</td><td rowspan="3">MIS 10</td><td rowspan="3">cold period</td><td></td><td colspan="2"></td></tr>
<tr><td>[glaciation]</td><td colspan="2"></td></tr>
<tr><td></td><td colspan="2">youngest Upper Terrace</td></tr>
</table>

accumulation of the Younger Middle Terrace occurred before the Drenthe-stadial ice had covered northern Germany. For the time in question temperature and dust records from Antarctica suggest a span between ca 190 (200) and 160 ka b.p. (temperature-against-time-graph of Vostok ice core, see Petit et al. 1999, Petit et al. 2000).

On the basis of palynological and sedimentological studies in KB 1 core, A. Kleinmann distinguished the following units (Kleinmann et al. 2011):

Nachtigall 1 Interglacial, 36.0–28.60 m, corresponding to Bouchet 2 in the Velay sequence (Beaulieu et al. 2001, Reille et al. 2000, Tzedakis et al. 1997) and to MIS 7c. The warm period sediments consist of peat, clay and mud with intercalated layers of allochthonous material:

peat A 36.00–34.70 (Grupe 1929: Unterflöz des Haupttorflagers)
Allochthonous Unit (I) 34.70–33.85
peat B and lacustrine clay / mud 33.85–33.25 (Grupe 1929: Oberflöz des Haupttorflagers)
Allochthonous Unit (II) 33.25–30.00, with peat C at 32.00–31.97 m
peat D 30.00–28.60.

Albaxen Stadial, 28.60–27.05 m, corresponding to Bonnefond in the Velay sequence and to MIS 7b.

Nachtigall 2 Interstadial, 27.05–26.02 m, corresponding to Bouchet 3 in the Velay sequence and to MIS 7a. At 27.00–26.54 it contains dark clay (organogenic layer E).

Stadials 1–5, 26.02–22.50 m, separated by the minor Interstadials 1–4, corresponding to Costaros in the Velay sequence and to the beginning of MIS 6.

Dating of core samples from KB 1 borehole by TIMS ^{230}Th/U method yielded the following isochron corrected ages (Waas, Kleinmann & Lepper 2010):

227 +9 / -8 ka for birch fen peat sample 34.87–34.83 m (closed system within peat A)
201 +15 / -13 ka for brown moss peat sample 33.63–33.47 m (closed system within peat B)
206 ±6 ka for fen peat sample 28.90–28.82 m (closed system within peat D).

An attempt at dating the clay sample 26.8–26.7 m with ca 15 % organic matter (organogenic layer E) yielded an age 177 ±8 ka which seems to be not consistent with the palynological results, a higher age being expected.

As a result of these contemporary research activities by three-way approach, the warm climate deposits of Nachtigall pit, known since a long time, actually have to be assigned to MIS 7 warm stage periods. The present study contributes lithostratigraphic aspects: directly subjacent as well as directly overlying gravel deposits represent Older Middle Terrace and Younger Middle Terrace respectively and thus also establish the age of the Nachtigall warm climate deposits as mid-Saalian MIS 7.

10 Further tasks

Allochthonous Unit I as well as Allochthonous Unit II of Nachtigall-Complex (Chapter 7.2, Chapter 9) have equivalents in the succession of Göttingen-Geismar, Ottostraße, 50 km away from Nachtigall pit (Grüger et al. 1994; Kleinmann et al. 2011, mentioning this correlation as orally communicated by Helmut Müller in or just before 2007). The respective sections with their striking amounts of mineral matter from the slope occur at the same palynostratigraphic positions during warm climate. Catastrophic rainfalls during the periods in question might have torn open the plant cover at slopes and thus made possible the anomalous mass transport (Kleinmann et al. 2011). Apparently the events on which the contemporaneity is based were not triggered by seismic activities: there is no evidence of earthquakes in sedimentological record, and in fact the palynological record even revealed certain vegetational developments in Allochthonous Unit I and Allochthonous Unit II (A. Kleinmann, orally). A conclusive explanation of the circumstances that determined the contemporaneity of the peculiar sedimentation at both sites seems to be worth further study.

By their climatic conditions of accumulation the Nachtigall warm and cool climate deposits are expected to yield archaeological finds that have not been taken into account so far.

As to lithostratigraphy it is an open question, how the inland ice influenced sedimentation at the study site from its margin about 20 km in the north. We have to assume that ice dammed lakes covered the upper valley of the River Weser (glacial Lake Weserbergland, Wortmann 1998; glacial Lake Weser, Winsemann et al. 2009, 2011). The surface of the Drenthe valley floor might have been at a level of about +95 m NN at the least, being formed by the Younger Middle Terrace. Up to now glaciolacustrine finegrained sediments or delta or subaqueous fan deposits have not been identified in or around Nachtigall pit. – According to Grupe (1929) an erratic boulder of Swedish Dala quartzite was found in the pit area, beneath 0.5 m of Weser gravel and 17.4 m of "clay" with some peat layers at a depth of 23.9 m. Yet it was interbedded in allochthonous deposits of Nachtigall 1 Interglacial and not in glacial sediments (cf. Chapter 7.2).

A fundamental stratigraphic detail is concerning the base of the Lower Terrace deposits. This means the question whether the Weser as a braided river formed the deep incision and the wide periglacial valley plain by downcutting really during the deglaciation at the end of the Saalian Complex. Subsequently still prior to the Eemian interglacial the Lower Terrace aggradation started during cold and wet climate and continued after the interglacial. This model applies accordingly to older terraces (Wortmann 1968, Wortmann & Wortmann 1987, Bridgland 1994, Schreve 2004, McNabb 2007). On closer inspection the very important process of downcutting already may have been initiated as fluvial response to sea level lowering prior to the advance of the inland ice. This model may be inferred from Elsterian fluvial and fluvioglacial deposits („Flußrinnen-Schichten") in the former Weser valley in and near Hannover (Rohde 1983, 1994, Rohde & Becker-Platen 1998: 38f., 138f., 34f.).

In Memoriam

This work has been accomplished in reverence for Helmut Müller (20 July 1924 – 18 June 2008) who opened the palaeobotanical secrets of the Nachtigall deposits and stratigraphically assigned them to a still incompletely known mid-Saalian warm stage.

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Abbreviations

ka b.p. Kilo annos (thousand years) before present (1950)

MIS Marine isotope stage (warm or cold stage of Quaternary period, established by temperature-dependent $^{18}O/^{16}O$ oxygen isotopes ratio of marine calcareous microfossil shells [foraminifers] or glacier ice). Alternative notation: MI stage. In general odd numbers are applied for warmer stages, even numbers for colder stages

NN German Ordnance datum (Landkarten-Bezugshöhe Normal Null)

OMT Older Middle Terrace (older Saalian fluvial terrace)

TIMS Thermal ionisation mass spectrometry

WOF Water-level of over-bank flooding. Correspondingly: level of younger alluvial clay (Jüngerer Auelehm). Also correspondingly: 30/40 years mean annual highest water-level (Mittleres Hochwasser MHW 1941–70/80)

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Interrelation of geomorphology and fauna of Lavrado region in Roraima, Brazil – suggestions for future studies

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Abstract: The authors discuss the relevance of geomorphology and biology interaction under the concepts of the Brazilian morphoclimatic domains. The discussion is focused on biogeographical and ecological aspects. The open areas of Roraima – the lavrado – localized between Brazil, Venezuela and Guyana, in the Northern portion of the Amazon morphoclimatic domain, is the region where the present case study was carried out. Remote sensing techniques were applied to determine the relief and field biology characterization. The generated products were useful for describing the habitats and local distribution of the lavrado's fauna.

Die Wechselbeziehung von Geomorphologie und Fauna in der Lavrado Region in Roraima, Brasilien: Vorschläge für zukünftige Studien

Kurzfassung: In der vorgelegten Arbeit wird die Abhängigkeit von Geomorphologie und biologischen Interaktion unter Verwendung des Konzeptes morphoklimatischer Regionen Brasiliens vorgestellt. Die Diskussion fokussiert hierbei auf biogeographische und ökologische Aspekte. Die vorgelegte Studie wurde in den offenen Bereichen von Roraima – Lavrado – zwischen Brasilien, Venezuela und Guyana durchgeführt. Dieses Gebiet liegt im nördlichen Teil der morphoklimatischen Region Amazoniens. Zur Anwendung kamen Techniken der Fernerkundung, um das Relief der Region zu ermitteln und biologische Charakterisierungen durchzuführen. Die hierdurch erzielten Ergebnisse wurden genutzt, um Lebensräume der Region und die Verteilung der Lavrado Fauna zu beschreiben.

Keywords: *Biogeomorphology, Amazon morphoclimatic domain, Roraima, lavrado*

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1 Introduction

On a geomorphological viewpoint, the Amazon rainforest can be characterized by its lowland relief and extensive forested areas, by the dichotomy between main allochthonous and small autochthonous rivers, as well as combining latosol and podzol low fertility soils. The annual thermal range is relatively homogeneous, 24°C to 26°C; the rainfall is heterogeneous, 1750 to 2300 mm/year, outside the Andes, which is around 7000 mm/year; and the vegetation is a complex net distributed over periodically flooded forest and upland, *várzea* and *terra firme* respectively (Victoria et al. 2000; Ab'Saber 2003). These combined geomorphological features form an area of approximately 7 million km², called the Amazon morphoclimatic domain. What in Brazil we call a morphoclimatic domain is an area of sub-continental dimensions, with characteristic patterns of relief, drainage, climate, soils and vegetation (Ab'Saber 1967).

One important feature of the Amazon morphoclimatic domain is the physiognomy of its vegetation, which can be open (scrubs, herbs and small trees) or closed (tall trees, with some emerging). Taking only this aspect into account, the open areas that occur in the Amazon region can be quite similar to those occurring in other domains, for example, the open vegetation of *cerrado* in the Central Brazilian ecosystem, the Bolivian *Chaco* or the *lhanos* in Venezuela. However, there are many ecological and physiological differences between these open formations, such as floristic composition, soil formation, geomorphological genesis, drainage and climate (Vanzolini & Carvalho 1991; Eiten 1992, 1994).

We can focus on this physiognomic dichotomous property of the Amazon vegetation with different lenses, depending of the goal. From the biogeographical viewpoint, for instance, these two morphological aspects of vegetation, open and closed areas, are important for understanding the distribution of organisms, principally when we consider the pulsation of the forest over the last 20.000 years – the open areas entering the forest during the Pleistocene glacial dry periods and the expansion of the forest during the interglacial wet periods throughout South American ecosystems (Vanzolini 1988; Ab'Saber 1977; Pessenda et al. 2009).

In the Brazilian Amazon there are expressive open veg-

Fig. 1: Examples of open areas within the Amazon morphoclimatic domain. 1 – Roraima, Venezuela and Guyana. 2 – Amapá state, Brazil. A – Monte Roraima; B – Branco River; C – Lacustrine Systems; D – Serra da Lua; E – Serra Marari; F – Maracá Island; G – Serra do Tepequém.

Abb. 1: Beispiele von Freiflächen innerhalb der morphoklimatischen Region Amazonas. 1 – Roraima, Venezuela und Guyana. 2 – Bundesstaat Amapá, Brasilien. A – Monte Roraima; B – Branco River; C – Lacustrine Systems; D – Serra da Lua; E – Serra Marari; F – Maracá Island; G – Serra do Tepequém.

etation areas in the states of Pará, Amapá and Roraima, occurring as enclaves inside extensive forested areas (Murça-Pires 1974; Carvalho 2009; Vanzolini 1992) and along the major rivers, such as the Trombetas (Egler 1960), Negro (Ducke & Blake 1953), in the mouth of the Tapajós (Radambrasil 1975) and in the Madeira (Murça-Pires 1974). These open areas comprise several landscapes, such as plains, plateaus, hills and mountains. Associated with these geomorphological features there occur the scrubs, herbs, grasses and cactacean adapted to these physical formations, constituting very particular habitats where can live and reproduce different species of animals.

The most relevant fact concerning the distribution of animals and plants is that they are not randomly distributed along their areas of occurrence; on the contrary, there are specific habitats where they can live. In this way, it is our thought that: i) biological aspect concerning the distribution of organisms among the various habitats that form ecosystems cannot be understood without the understanding of the physical structure of these habitats, ii) this comprehension can be given by a geomorphological approach.

The rational of our thought is tied to the concepts establishing that the distribution of organisms reflects their sets of adaptations to the immediate environment, a concept known as ecological niche (Vanzolini 1970; Pianka 1994). This idea is the soul of classical studies approaching biology (zoogeography) and geomorphology, which were carried out by Vanzolini & Williams (1970), Vanzolini (1970, 1981) and Ab'Saber (1967), currently incremented by news geoprocessing techniques (Carvalho & Ramirez 2008; Carvalho 2009a; Metzger 1997).

In this context, the aim of the present study is focused on the landscape and habitats that occur in open areas inside the Amazon region. The scenario of this discussion encompassing the field of biogeomorphology comprises three ways: i) concepts of morphoclimatic domains and biogeography, ii) the case study of a very interesting open area known as *lavrado*, situated in the Northern Amazon region – the Brazilian state of Roraima, iii) use of geoprocessing techniques for identifying and describing habitats.

2 The case study area

The general region described in this report (Fig. 1), comprised in the Guyana Shield (Hammond 2005), is a very peculiar open area of some 69.000 km², mostly situated in the northern portion of the Amazon morphoclimatic domain, overlying three countries. We estimate, by remote sensing, that this area covers some 45.000 km² in the Brazilian state of Roraima, 10.000 km² in Venezuela and 14.000 km² in Guyana.

In Venezuela this portion of open areas is about 1200–1600 meters above sea level. It is characterized by the presence of ruiniform tabular mountains, individually called *tepuy*. The *tepuyes* are part of a geomorphological formation known in Venezuela as *Gran Sabana*. In the Brazilian territory the best known *tepuy* is the Roraima Mount (05°11'S, 60°49'W), around 2800 meters high, situated on the triple border of Brazil, Venezuela and Guyana. We do not consider this Venezuelan region to be part of the Amazon morphoclimatic domain (see Ab'Saber 2003).

In the Guyana region this Northern Amazon open area is mostly situated on the basin of the Rupununi River, an affluent of the main Guyanese river, the Essequibo. This open area, locally known as *Rupununi Savanna*, is separated by the Tacutu River. This river, that forms the border of Brazil and Guyana, runs in a geological fissure from South to North, where it turns westward to flow into the Uraricoera River in Brazil (approximately at 03°01'N, 60°28'W), both rivers forming the Branco, which flows southward into the Negro River in the Brazilian state of Amazonas.

In the Brazilian portion, the state of Roraima, this area is known as *lavrado*, an old Portuguese term for open vegetation (Vanzolini & Carvalho 1991; Carvalho 2009). The *lavrado* has its own socio-cultural and ecological identity, integrated by complex networks of interactions among the local people with the landscape, and by a characteristic local fauna and flora adapted to the *lavrado* ecosystem (Nascimento 1998).

This open area is formed by peculiar geomorphological features, such as boulders, alluvial plains, lakes and gallery forests along the rivers. Isolated patches of forest, scrubs and herbs, are present throughout the area. Gallery forests occur in the banks of the rivers. These features form the *lavrado* habitats, harboring many species of plants and animals, whose biological aspects of their distribution along these regional habitats are also focused in this study.

3 Material and Methods
3.1 Geomorphology

To describe the morphology of the case study relief we used remote sensing techniques (hypsometry, shaded relief, topographic profiles and RGB composition) from elevations model of SRTM (Shuttle Radar Topography Mission) and Landsat 5 images. The elevation model from SRTM is a radar image, acquired by interferometry method in 2001 for entire globe, used for geomorphometrics analysis of the terrain.

The software ENVI 4.3 was used to resize the SRTM data to 30 meters, by interpolation, from original spatial resolution of 90 meters. This digital elevation model was important to identify the different altimetry values, and the morphology of denudation forms (ranges and hills) using shaded re-

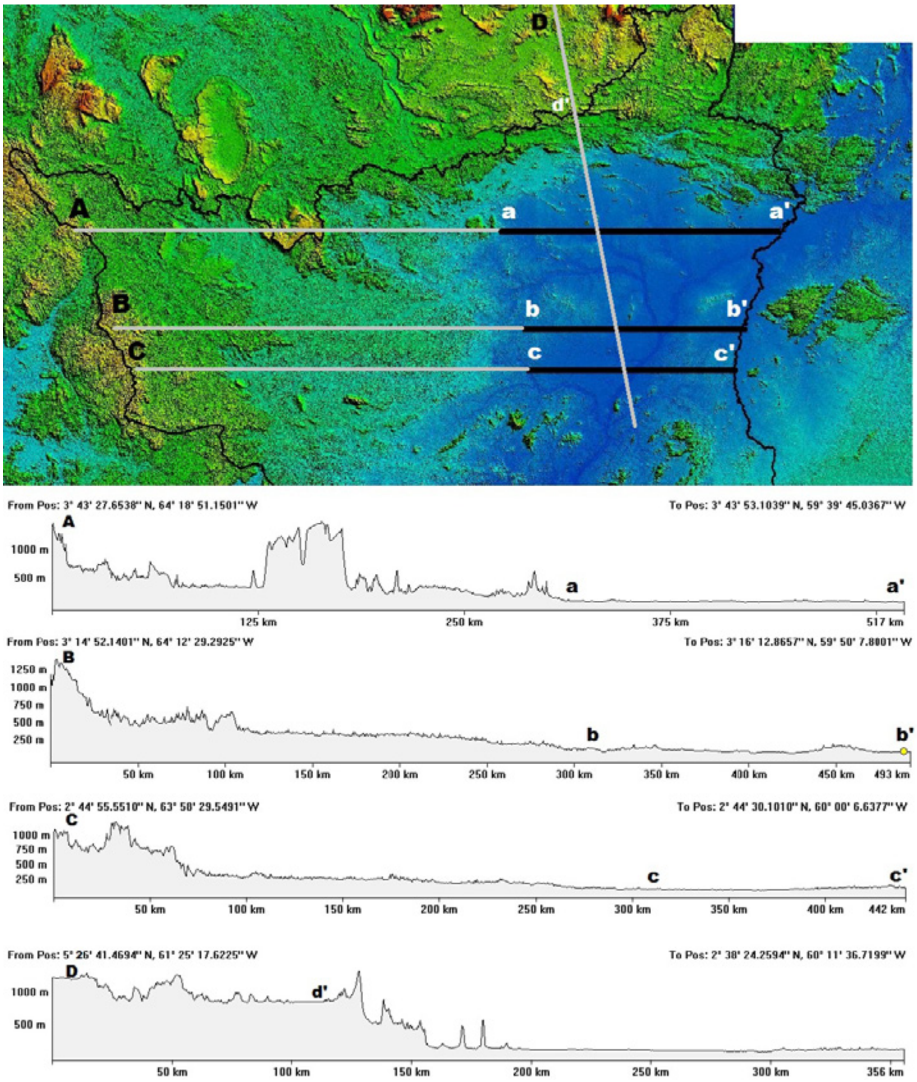

Fig. 2: Roraima, topographic transects profiles a-a' to c-c'; Venezuela – Roraima, transect D.

Abb. 2: Roraima, topographische Transekte a-a' to c-c'; Venezuela – Roraima, Transekt D.

lief and topographic profiles techniques. The optical images of Landsat 5 are mostly used for environmental studies, with 30 meters spatial resolution, and were used for identifying agradational morphologies, like fluvial plains, lakes and vegetation aspects by visual interpretation.

The Landsat 5 images RGB composition was applied in ENVI 4.3, using bands 5,4 and 3. The Landsat 5 images were achieved in 2005, from December to April, which corresponds to dry season (without clouds), patch-rows were 232(56,57,58); 231(57,58). These images were acquired at the National Institute of Spatial Research (INPE) – *www.dgi.inpe.br/CDSR/ – and Embrapa Relevo – www.relevobr.cnpm.embrapa.br/.*

3.2 Fauna examples

Case study of faunal elements, in the present context, were determined through field work conducted by the Instituto Nacional de Pesquisas da Amazônia (National Institute of Amazonian Research – INPA) in Roraima throughout the past two decades, mainly on the lavrado area (see Carvalho 2009). We take as examples the vertebrate fauna of the area, mainly aspects of its distribution along the habitats comprised by geomorphological features determined through geoprocessing techniques (Figures 8, 9, 10).

4 Results and Discussion

4.1 Geomorphological features of the *lavrado*: remote sensing

One can see the position and topographic profile of the *lavrado* and adjacent forested areas just looking at the region through transects, for example covering the forests of the West portion of Roraima up to the open areas in the East, or covering part of the Venezuelan *Gran Sabana*, until the *lavrado* areas (Fig. 2). At the same way, through transects (Fig. 3) we can see the main features of the relief, like high altitudes (more than 1500 meters high), with tabular relief (*tepuys*), agradational and denudational processes, moderated dissection and low structural control (Fig. 3 "1"); intermediary altitude, somewhat of 500–1500 m, with denudational processes, high dissection and strong structural con-

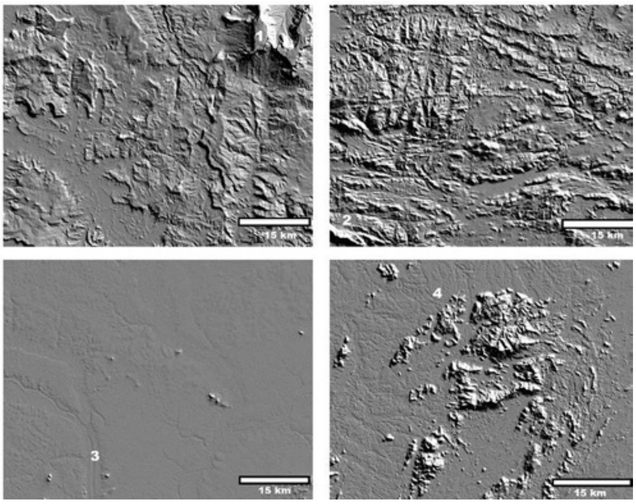

Fig. 3: 1 – Mount Roraima, moderate relief dissection, border of Venezuela and Brazil (05°11'N, 60°49'W); 2 – Serra Marari, moderate to strong dissected relief (04°16'N, 60°46'W); 3 – Uraricoera River, low relief dissection (03°19'N, 60°25'W); 4 – Serra da Lua, low and strong relief dissection (02°27'N, 60°28'W).

Abb. 3: 1 – Mount Roraima, moderate Reliefzergliederung an der Grenze zu Venezuela und Brasilien (05°11'N, 60°49'W); 2 – Serra Marari, mäßige bis starke Reliefzergliederung (04°16'N, 60°46 ,W); 3 – Uraricoera River, niedrige Reliefzergliederung (03°19'N, 60°25'W); 4 – Serra da Lua, niedrige und starke Reliefzergleiderung (02°27'N, 60°28'W).

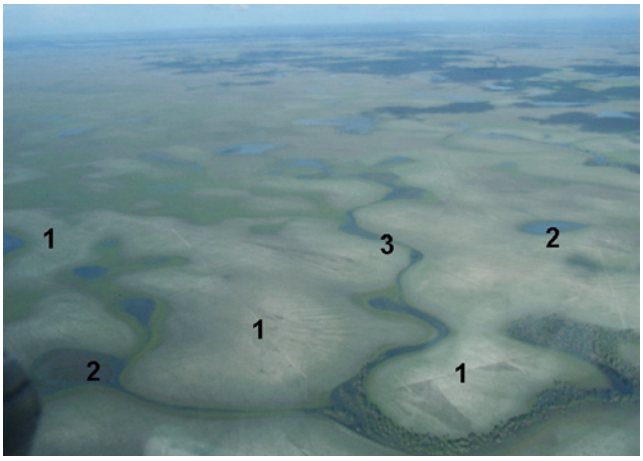

Fig. 4: 1 – Tesos (low hills) convex morphologies; 2 – lakes; 3 – small elevations between the streams.

Abb. 4: 1 – Tesos (flaches Hügelland) konvexe Morphologie; 2 – Seen; 3 – kleine Erhebungen zwischen den Strömen.

trol (Fig. 3 "2"); low sedimentary plains, around 70–100 m, with agradational processes like fluvial plains and lacustrine systems (Fig. 3 "3"); and isolated hills, inselbergs, with structural control (Fig. 3 "4").

In contrast with the high elevation of the Venezuelan *Gran Sabana*, the elevation of the *lavrado* area is relatively low, around 70–200 m a.s.l. This area is drained by the Branco River, which is composed by a system of low hills, with low dissected relief, isolated residual peaks (*inselbergs*), surrounded by lakes in the headwaters, the flowing of which creates a interconnected streams (*igarapés*) separated by small elevations, known as *tesos,* forms drainage dissection around the lakes and streams (Fig. 4). Also we can see in the area the sugar-loaf formations (*pão-de-açucar*) and laterite layers exposed on the soil (*lajeiro*).

In all *lavrado* areas narrow lines of palm trees remind one of the landscapes of the morphoclimatic domain of the Central Brazil *cerrados.* Of course this resemblance is only apparent, since the *cerrado* is a very distinct ecosystem, situated a few thousands kilometers from the *lavrado.* The reconnaissance of the distinctiveness between both ecosystems – *lavrado* and *cerrado* – has a very important ecological and biogeographical significance (Eiten 1963; Coutinho 1978; Vanzolini & Carvalho 1991; Carvalho 2009).

The predominant declivity of the *lavrado* is between 5°–8°, with low energy, forming a region that receives sedimentary material, mainly sand coming from the surrounding crystalline uplands (Guyana Shield). The *lavrado* central portion's relief low energy favors the formation of a complex lacustrine system, composed by more or less circular up to 300 meters long lakes, most of which are temporary (Fig.5). These lakes are independent, interconnected by narrow streams, forming dendritic, rectangular and subdendritic patterns. These lakes are fed by ground water and resemble the lakes of the morphoclimatic domain of the *cerrado* (Carvalho & Zucchi 2009). Figures 6–7 show some aspects of the fluvial plain and vegetation of the *lavrado.*

The rivers that cross the *lavrado* are autochthonous, with its headwaters in the elevated *serras* that make the border of Brazil and Venezuela, the Parima-Pacaraima system. The *lavrado* drainage, formed by well developed fluvial plains, is directed to the Negro River, which runs from the Andes until its confluence to the Solimões River, in the Central Amazon Basin. The main rivers that run in the *lavrado* have banks (*dique marginal*) and beyond these a formation called *várzea*, a floodplain area formed along the main rivers during the rainy season.

The vegetation of the *lavrado* is composed by interesting formations (Beigbeder 1959; Ab'Saber 1997). Throughout this open area one can see near 8–10 meters height and less than 0.5 ha, wood patches, surrounded by grouped or more disperse scrubs and small trees. The ground is covered by grasses and grass-like plants (family Cyperaceae). Lines of palm trees (*Mauritia flexuosa*), known as *buritizais*, due to the popular name *buriti* (family Palmae) for the palm tree, is an important element of the *lavrado* landscape, starting in small lakes and running toward the main rivers, for a distance of around 300–800 meters. The *lavrado* is surrounded by 15–20 meters high forest, soil with shallow litter, some emerging trees and somewhat unstructured understory.

4.2 Geomorphology and fauna
4.2.1 The approach

We can look at this interaction between geomorphology and biology from the point of view of different related areas of knowledge. Whatever the area, the main idea of this interaction is focused on species and populations distribution, local or along large areas. On the regional distribution, one may be interested in describing the species richness between habitats within an ecosystem, to understand aspects of the local biodiversity. On the other hand, we can focus on the distri-

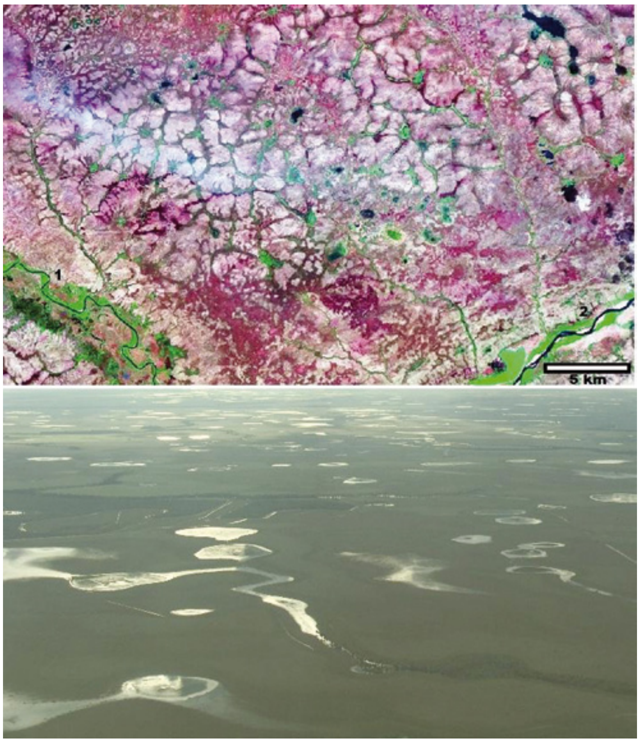

Fig. 5: Lacustrine system (03°37'N, 60°15'W); 1 – Surumu River; 2 – Tacutu River.

Abb. 5: Lakustrines System (03°37'N, 60°15'W); 1 – Surumu River; 2 – Tacutu River.

bution of species and populations among ecosystem, in order to understand extensive distribution patterns, for example, populations in contact or separated by geomorphic barriers that have occurred in the present or past events.

Two emeritus herpetologists together with one geologist were the first to approach geomorphology to biology in 1969–1970. The zoologists are the Brazilian scientist Paulo Emílio Vanzolini and his North American colleague Ernst Williams. They have formulated a very elegant South American lizard (genus *Anolis*, family Polychrotidae) speciation model based on forest expansion and retraction, paleoclimatic events that occurred under the influence of the Pleistocene dry and wet periods over the past 20.000–10.000 years.

This model of speciation formulated by Vanzolini & Williams (1970) establishes that because of the forest fragmentation occurring during paleoclimatic dry periods (glaciation) animals became isolated in forests patches, which resulted in ecological barriers – forest species do not live in open areas. These barriers, in turn, determined the interruption of gene flow between populations. Dry events of the past can be inferred at present by geomorphologic features, such as the stone-lines (paleosols formed in dry paleoclimatic periods and buried in sedimentary deposits), indicating that a forested area today was open in the past (Ab'Saber 2003; Hiruma 2007). Another way to infer past dry events is through palynological records and ^{14}C dating of sediments (Absy 2000; Salgado-Laboriau 1982).

During the humid phase (interglacial) the forest coalesced and what was fragmented forest became continuous forested area; however, many animal species did not change gene again, because their populations were isolated for a period in which several biological and physiological changes occurred in each one. The result of these processes was the formation of distinct species. The model focused mainly on the pulsation of the forest in the Amazon region; however, the idea was applied for other regions and species (Vanzolini 1988, 2002; Wüster et al. 2005).

Following another way of the same theme, the German geologist Jürgen Haffer studying Amazonian birds, in 1969 came to the same conclusion and model of speciation as did Vanzolini and Williams for lizards in early 1970. This model of speciation, taking geomorphological evidences of expansion and retraction of the forest, became classic in biogeography and is well known as Pleistocene Refugia Model and Refugia Theory (Vanzolini 1970; Absy et al. 1991; Haffer & Prance 2001; Haffer 1969; Ab'Saber 1982).

The morphoclimatic domains concepts, adopted by Vanzolini and Williams as vegetation criteria for their study of species distribution and refuges, were first formulated by Aziz Nacib Ab'Saber in 1967. Prior to this Brazilian geographer and geomorphologist, the vegetation of the regions in Brazil was identified through fragmented floristic features. The model of Ab'Saber gave the necessary strength in identifying large vegetal formations, instead of patches inside the same ecological and geomorphologic formation. Ab'Saber used the climate, vegetation, soil, Hydrography and relief as features to recognize what he called the area "core" in a domain, in a sub-continental scale. All kinds of regional geomorphological facies could, then, be included in one domain or another – *cerrado, caatinga, mata atlântica and hiléia* (the Amazon) – recognizing the transitional zones.

The model formulated by Ab'Saber was a great advance to the fields of geography and geomorphology, since regions could then be identified as a continuous unit with related geomorphologic features. To biologists interested in ecology and biogeography this geomorphologic model integrated formerly scattered data, enabling one to come to a better understanding of species distribution.

4.3 Habitats and faunal distribution: the *lavrado*

All those geomorphological formations comprised in the *lavrado*, such as hills, rock outcrops, lakes, small patches of forest, scrubs and the gallery forests along the rivers, with the back-swamps (*várzea*) of the major ones, form the habitats inhabited by many organisms. Identifying these habitats is the first step to comprehend the biology of any species that live in the *lavrado*, in terms of adaptations and gene flow among individuals and populations. Some geomorphological features can illustrate this point of view, such as the granite and laterite extrusions, hogbacks, inselbergs and sparse or grouped boulders at various sizes (*matacões*) in the plain and low hills present in the *lavrado* (Ruellan 1957). In addition to the geomorphological interpretation, these formations also have their ecological identity, forming complex microhabitats inhabited by birds, bats, rats, snakes, frogs, lizards and many species of invertebrates (Vanzolini & Carvalho 1991; Carvalho 2009; Nunes & Bobadilha 1997; Rafael et al. 1997).

The rocks are distributed throughout the area and are directly exposed to the sunlight. These features led to many relevant biological questions, such as: How many species of vertebrates and invertebrates are associated with these

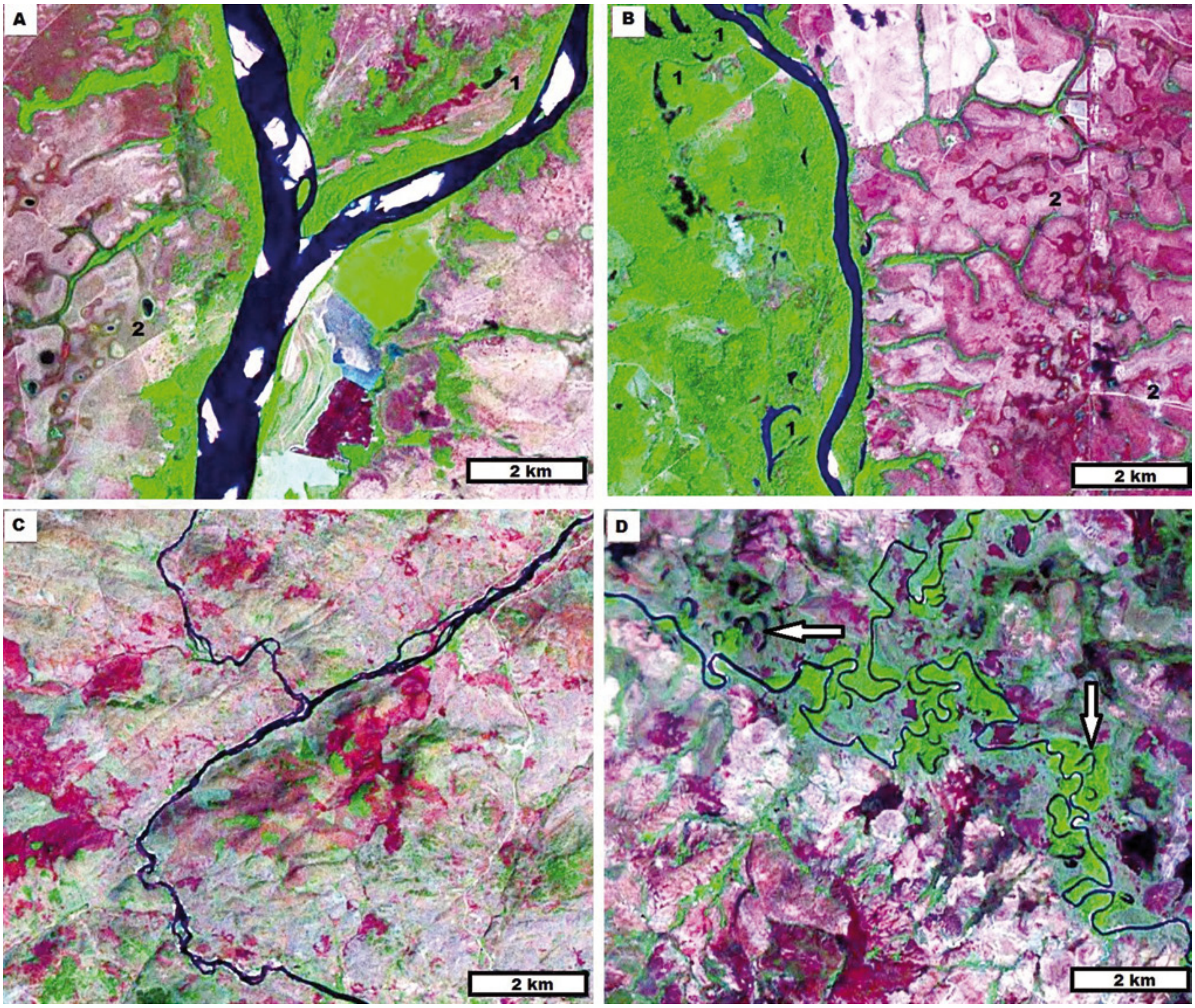

Fig. 6: "A-B" – fluvial plains, lowlands (03°01'N, 60°29'W and 02°36'N, 60°54'W); "C-D" – fluvial plains, highlands (04°17'N, 60°32'W and 04°56'N, 61°14'W). 1 – abandoned chanel lakes of flood plain; 2 – Lakes of flat plain. A – Mouth of the Tacutu River in the Branco River; B – Mucajaí River; C – Cotingo River with structural control, non flood plain; D – meandriform river developed at structural control, small flood plain with lacustrine systems (oxbows lakes).

Abb. 6: "A-B" – Flussniederungen, Tiefland (03°60°N, 29º01'W und 02°36'N, 60°54'W); "C-D" – Flussniederungen, Hochland (04°17'N, 60°32'W und 04°56'N, 61°14'W). 1 – aufgegebene Kanalseen der Flussaue; 2 – Seen der Tiefebene. A – Zusammenfluss von Tacutu in den Brancoss; B – Mucajaí; C – Cotingo, nicht zur Flussaue entwickelt; D – mäandernder Fluss, kleinflächige Flussaue mit lakustrinen Systemen (Altwasserseen).

habitats? How many diet and reproductive adaptations have these species undergone so as to be able to survive in these geomorphological units? How can be genetically characterized the populations of the same species inhabiting the *lavrado*? There are any preferences of some particular species in residing certain geomorphological features, such as granite and laterite? How these rock outcrops are distributed (grouped or dispersed) and how to interpret the distribution pattern?

An interesting case of animal distribution in these granite habitats come from the frog *Leptodactylus myersi*, a species that seems to be endemic to the *lavrado*, living on the rocks, at least the main populations (Heyer 1995). Each population of this frog seems to be separated by several kilometers, which is the distance between the boulders. Questions based on this example may include: How are these frog populations distributed, taking into account they are directly associated with the boulders distribution? How to characterize the adaptations of this frog, in terms of reproduction and diet? Where they lay their eggs, considering the extreme exposure to this habitat to sunlight and dry environments? These are questions being currently studied.

Another species very common in the rock formations of the *lavrado* is the lizard *Tropidurus hispidus* (family Tropiduridae). The biological questions that can be applied to the populations of this lizard are associated with the habitats where they live, such as the boulders, small trees, border of the forest and in the small patches of forest. For example: Do all these lizards have the same set of adaptations? Is it possible to determine the populations of this lizard precisely by identifying the habitats through geoprocessing techniques?

Among mammals there are some interesting distribution in habitats composed by lacustrine system in general, low hills and dissected relief, vegetation of the margins of rivers (*mata ciliar*) and lines of palm trees (*buritizais*), laterite layers (*lajeiros*) and boulders. All these geomorphological

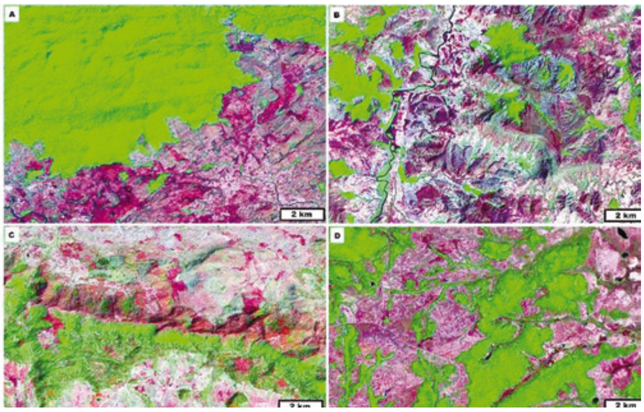

Fig. 7: A – Venezuela-Roraima border, transition of the forest to lavrado – grassland with tall shrubs and small trees (04°02'N, 61°03'W); B – Venezuelan open areas, patches of forest with well developed rills and structural control (4°50'N, 60°57'W); C – Serra da Memória, shrubs and small trees, vegetation slope with tors and blocks (04°10'N, 60°57'W); D – island vegetation with small lakes at flat plain (3°12'N, 60°57'W).

Abb. 7: A – Grenze Venezuela – Roraima, Übergang von Wald zu lavrado – Grasland mit hohen Büschen und niedrigen Bäumen (4°2'N, 61°3'W); B – Venezuelanische Offenlandschaft mit gut entwickelten Bächen (4°50'N, 60°57'W); C – Serra da Memória, Sträucher und niedrige Bäume, bewachsene Hänge mit Blöcken (4°10'N, 60°57'W); D – inselartige Vegetation mit kleinen Seen in einer flachen Ebene (03°12'N, 60°57'W).

units comprise the environment where many mammal species can live, such as *Myrmecophaga tridactyla* (*tamanduá*) and *Tamandua tetradactyla* (*mambira*), two related species of the family Myrmecophagidae (Order Pilosa) that feed on termites and ants. These two mammal species also have the patches of forest as refugia during the night.

Another species that have its habitat associated with the geomorphological units of the open areas is the little mammal *Nasua nasua* (*quati*) of the family Procyonidae (Order Carnivora), an inhabitant of the boulders of the plains and hills. During the day it is common to see this animal in that habitat, looking for food, mainly earthworms, insects and some fruits. The vegetation of this habitat is composed by herbs, grasses, scrubs and isolated trees, where *N. nasua* can be found climbed at night. Again, geomorphology gives the direction for describing these habitats.

Among birds we can also have some representative species currently endemic to Roraima, such as *Aratinga solstitialis* (*jandaia-sol*) of the family Psittacidae, that live in habitats comprised by the gallery forests or on the forest edge (approximately 03°52'N, 59°37'W). The precise localization of these endangered species habitats can be obtained by geoprocessing techniques, like the other endangered species *Synallaxis kollari* (*joão-de-barba-grisalha*) of the family Furnariidae. This small bird can live in habitats formed by low hills and dissected relief, scrubs and small trees, up to the right bank of the Tacutu River, in Guyana territory. Some populations of this bird can also be found in Roraima, in gallery forest.

The same rational can be applied to the botanical species present in the *lavrado*. For example, there is a small and interesting cactus genus *Melocactus* that occurs on the rocks forming clusters. The distribution of this cactacean can be easily established through the identification of the rock extrusions. Another cactacean present in the *lavrado*, the distribution of which can be ascertained through geoprocessing techniques, is the Brazilian popular *mandacaru* genus *Cereus*, whose main distribution may be associated with the soil, as well with clusters of termite nests genus *Cornitermes* (approximately 03°52'N, 59°37'W).

It is also very useful and informative to apply the geoprocessing techniques for understanding the *lavrado* vegetation. These features of the landscape in this area are made up by a complex net of small more or less rounded forest patches (island forests) some 0.5 ha or less, palms trees (linear or almost rounded), described having the focus on the habitat of animals. But we can also focus the question with another lens. How the forest patches of the *lavrado* are distributed? Is there any pattern accounting for forest patches distribution and soil? The relevance of these questions is not restricted to the present, but imply in considerations such as how the landscape change and what would be the implications for the fauna and flora.

These questions lead us to look at the *lavrado* vegetation under another focus, which is the pulsation of open and closed vegetal formations under climate changes. It is quite possible that the expansion and retraction of the forest during the Pleistocene have influenced the gene flow of many species living today in these kinds of vegetation, connecting or interrupting definitely or temporarily the patches of forest. How the various species of the *lavrado* terrestrial vertebrates, for example, were locally affected by the events during the dry and wet paleoclimate periods? What to say about the forest pulsation and climate change that might be undergoing at present?

Recognizing evidence of pulses in the *lavrado* vegetation, through geoprocessing data associated with the local distribution of species, might certainly elucidate several of these questions. This is the case, for instance, of three sympatric species of lizards of the genus *Gymnophthalmus* (family Gymnophthalmidae) that occur in the open areas of Roraima and in the forest edge, in contact with the *lavrado*. The species are *G. leucomystax* associated with termite nests, *G. vanzoi* in the contact forest open areas, and *G. underwodii* in the continuous forest (Vanzolini & Carvalho 1991; Carvalho 1999). In a 1.5 kilometer transect, we can find these three lizard species, each one in its specific habitat. These three species are so tightly taxonomically related, that it is difficult to recognize them at a first look, and we can imagine how many geomorphological events might have occurred for the speciation of these three lizards species. We can map the distribution of these lizards through geoprocessing techniques.

Looking again to the landscape of the *lavrado* and its associated fauna, another example of biogeomorphology applied to the biological distribution of populations comes from the termites. At least two species of these social insects of the family Termitidae build their nests on the ground (epigeous nests): *Nasutitermes minimus* and the *Cornitermes ovatus* (Bandeira 1988). Both species of termites construct nests in different parts of the *lavrado*, maybe due to soil factors, vegetation cover or both features together. The nest of *N.minus* is rounded on the top, around 30–40 centimeters high, and the base is 20–30 centimeters in diameter, are constructed mainly over the hills (approximately 03°20'N, 61°24W) . The

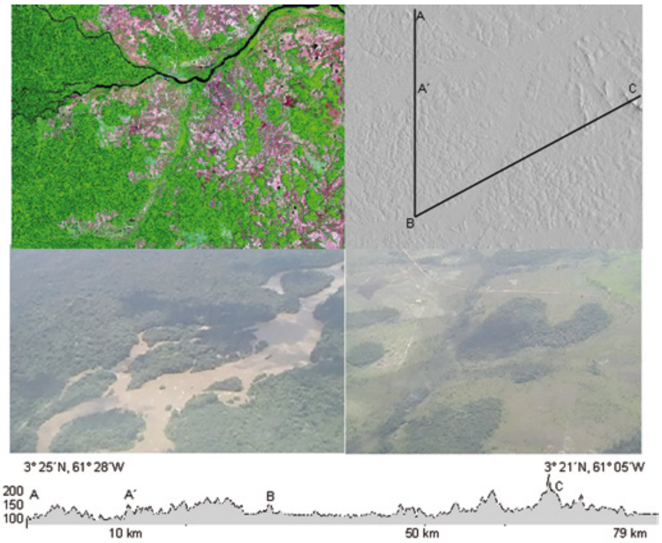

Fig. 8: "A-A´" – Uraricoera River (left photo); B – Grande River; C – Serra do Tabaio. Region of the endemic lizards Gymnophthalmus vanzoi and G. leucomystax (family Gymnophthalmidae). Forest and lavrado contact, with small patches of forest (right photo). Transition of denudational and aggradational relief, with isolated hills, drained by a not well development fluvial plain.

Abb. 8: "A-A´" – Uraricoera (Foto links); B – Grande River; C – Serra do Tabaio. Region mit den endemischen Eidechsen Gymnophthalmus vanzoi und G. leucomystax (Familie Gymnophthalmidae). Wald und lavrado-Kontakt mit kleinen Waldinseln (Foto rechts). Übergang von Abtragungs- zu Aufschüttungsrelief mit isolierten Hügeln, die Entwässerung erfolgt über eine schlecht entwickelte Flussniederung.

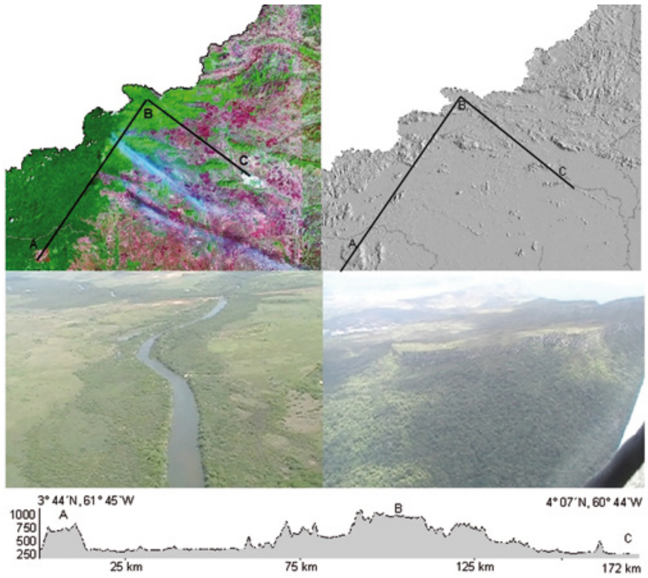

Fig. 9: Brazil-Venezuela border. A – Tepequém tepuy (right photo); B – Parima River, hills; C – Surumu River (left photo), type locality of the amphibian Elachistocleis surumu. Not well developed fluvial plains (Surumu River), with temporary lakes. Litholic soils with boulders, tors and scrub-herbs vegetation.

Abb. 9: Grenze Brasilien-Venezuela. A – Tepequém tepuy (Foto rechts); B – Parima, Hügel; C – Surumu (Foto links), Typuslokalität der Amphibie Elachistocleis surumu. Schlecht entwickelte Flussniederungen (Surumu) mit temporären Seen. Litholic-Böden mit Felsblöcken, strauch- und krautreiche Vegetation.

nest of *C.ovatus* is pointed on the top; the construction is very hard, around 2.0 meters high, and the base 1.0–1.5 meters in diameter, mainly constructed on the plains (approximately 03°52'N, 59°37'W), at the same region of the cactacean *Cereus*.

There are many animals associated to the nests of both of these termite species. The rattlesnake *Crotalus ruruima*, the lizards *Tropidurus hispidus* (Family Tropiduridae), *Cnemidophorus lemniscatus* (Family Teiidae) and the gekko *Hemidactyulus mabouia* (Family Gekkonidae) are tenants of these nests. Also some species of rats and opossuns, spiders and many invertebrate species live in those nests. There are interesting biological questions associated with the distribution of those termite species. With the help of geoprocessing techniques so as to identify the areas of occurrence of both, nests and soil, any approach related to these termites becomes more practical.

The distribution of rare or endemic species that occur in Roraima can be illustrated on maps using geoprocessing techniques, exemplifying species distributed in the *lavrado* and surrounding areas of this open vegetation ecosystem (Fig. 8, 9, 10).

5 Conclusions

Examples exposed in the present discussion can guide the focus of the biogeomorphology approach in two directions: i) at a regional scale or ii) at a sub-continental level, within or among large vegetal formations. Either way, the questions regarding species and habitats distribution should be made involving geomorphology as a backdrop of the whole scenery.

At a sub-continental level, considering large vegetal formations, the questions leads to problems related to speciation and its process. The recognition of the geographic units of the species been studied – the morphoclimatic domains – is fundamental for that approach, because the whole distribution area of a single species or groups of species will be compared through biological aspects, which may vary significantly or not. The main questions that arise at this level may include: How many vegetal formation can be recognized inside the domain (or domains) been studied? How are the soil, topography and hydrography characteristics in each studied region? Are these geomorphic features acting as barriers for gene flow among populations?

At a regional scale, such as that of the *lavrado* area, before the formulation of specific biological questions it is also imperative to locate the geographic insertion of the region within the main ecosystem. Once recognized the geographic context of the study site, we turn the eyes to the diversity and composition of the regional geomorphological units, such as the boulders, plains, hills, montains, lacustrine system, drainage and regional vegetal formations, which can be done applying remote sensing and geoprocessing techniques.

The geomorphological features will then characterize the habitats. Taking these features as criteria for categorize the compartments of the region, we can focus on the questions to be worked, which can be directed to analyze species richness, regional distribution of a group of species or distribution of a single species, habitat change and modification of

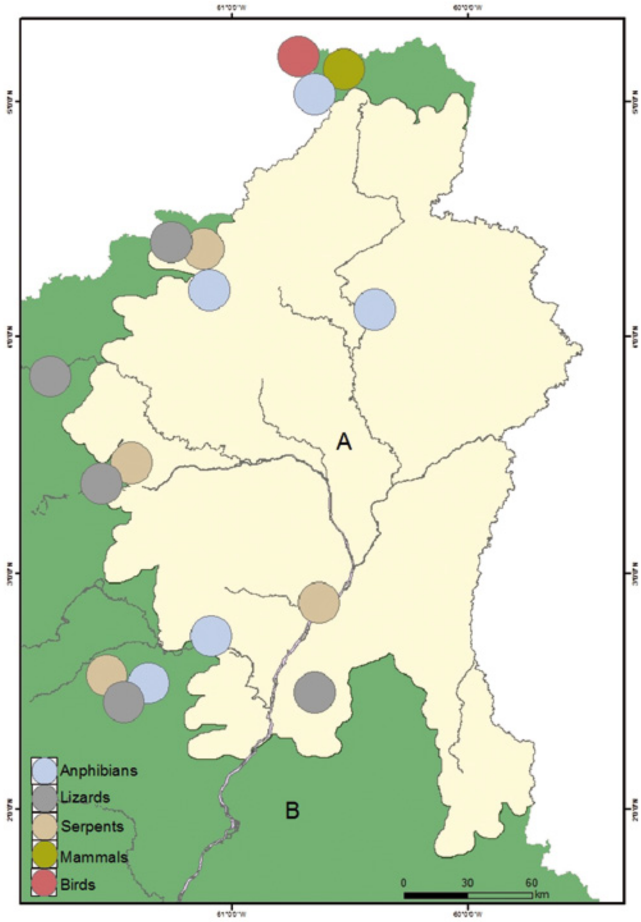

Fig. 10: Distribution of endemic species in northern of Roraima, lavrado. A – Lavrado area; B – Forest area. Some species: Lizards: Mabuya carvalhoi (Sauria: Scincidae), Gymnophthalmus leucomystax (Sauria: Gymnophthalmidae), Gymnophthalmus vanzoi (Sauria: Gymnophthalmidae); Amphibians: Dendropsophus benitezi; Elachistocleis surumu; Serpents: Micrurus pacaraimae; Birds: Schistocichla saturata ; Herpsilochmus roraimae; Syndactyla roraimae; Myiophobus roraimae; Thamnophilus insignis; Megascops guatemalae; Mammals: Nasua nasua vittata.

Abb. 10: Verteilung endemischer Arten im Norden von Roraima, lavrado. A – Lavrado Bereich; B – Waldbereich. Einige Arten: Echsen: Mabuya carvalhoi (Sauria: Scincidae), Gymnophthalmus leucomystax (Sauria: Gymnophthalmidae), Gymnophthalmus vanzoi (Sauria: Gymnophthalmidae); Amphibien: Dendropsophus benitezi; Elachistocleis surumu; Schlangen: Micrurus pacaraimae; Vögel: Schistocichla saturata ; Herpsilochmus roraimae; Syndactyla roraimae; Myiophobus roraimae; Thamnophilus insignis; Megascops guatemalae; Säugetiere: Nasua nasua vittata.

the landscape, population ecology and conservation. We get these data mainly through inventories, which should start someplace.

If an exhaustive faunal survey is a hard task, because of its high costs and need of experienced personal involvement, a reliable alternative is to select habitat samples by mapping the regional morphoclimatic units. This can be done applying remote sensing and geoprocessing techniques. Where to start will depend on the question. A good set of suggestions can be found in Heyer at al. (1994) and Carvalho (2009) for the *lavrado* area.

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Abstract: This paper presents the morphostratigraphy, lithostratigraphy and sedimentary structures of Terna River basin in the Deccan Basaltic Province (DBP) of West Central India. These Quaternary deposits have been divided into three informal formations (i) dark grey silt formation – Late Holocene, (ii) Light grey silt formations – Early Holocene, (iii) Dark grayish brown silt formation – Late Pleistocene with the older Quaternary Alluvial deposits of Upper Pleistocene age. The fine clay and silt formations in the lower reaches reflect that the streams are of low gradient and more sinuous. The river shows evidences of channel movement by avulsion, largely controlled by lineaments. Palaeo-levees, in the form 4–5 m high ridges exist along the Terna River floodplain, specifically in the Ter, Killari, Sastur, Dhuta and Makni villages. Several lineaments occur along NE-SW, NW-SE, E-W and WNW-ESE directions, which control the basement structure in the study area. The values of the Topographic Sinuosity Index (TSI) indicate rejuvenation of the area leading to the dominance of topography on the sinuosity of the river channels. The break in slope in the long profile is also indication of the Quaternary tectonic uplift of the area. Radiocarbon dating of some charcoal fragments collected from folded beddings indicates that paleoseismic activity might have taken place along the basin between AD 120 and AD 1671.

Quartärgeologie und Geomorphologie des Terna Beckens im westlichen Zentralindien

Kurzfassung: Im vorgelegten Artikel werden die Morphostratigraphie, Lithostratigraphie sowie die Sedimentstrukturen des Terna Beckens in der Deccan Basaltic Province (DBP) im westlichen Zentralindien vorgestellt. Die Quartärablagerungen können in drei große Einheiten unterteilt werden (i) dunkelgraue Schluffablagerungen – Spätes Holozän, (ii) hellgraue Schluffablagerungen – Frühes Holozän, (iii) dunkelgrau-braune Schluffablagerungen – Spätpleistozän mit altquartären alluvialen Absätzen mit oberpleistozänen Altern. Die feinen tonig-schluffigen Ablagerungen im Unterlauf des Flusses deuten auf ruhige Ablagerungsbedingungen und einen sinusartigen Abfluss hin. Der Fluss zeigt Tendenzen zu abschwemmungsbedingten Gerinneverlagerungen, die wiederum durch vorhandene Bruchlinien gesteuert wurden. Entlang des Terna-Flusses konnten weiterhin Paläouferrücken in Form von 4–5 m hohen Rücken nachgewiesen werden, hier vor allem im Bereich der OrtschaftenTer, Killari, Sastur, Dhuta und Makni. Einige nachgewiesene Bruchlinien treten vor allem in NE-SE, NW-SE, E-W und WNW-ESE-Richtung auf und bestimmen die Struktur des Grundgebirges im Untersuchungsgebiet. Die TSI-Werte (Topographic Sinousity Index) verdeutlichen einen Erosionswechsel im Untersuchungsgebiet mit einer Verstärkung des topographischen Einflusses auf die Ausformung der Abflussbahnen. Die im Profil sichtbare Geländekante zeugt weiterhin von einer tektonischen Hebung des Gebietes im Quartär. Radiokohlenstoffdatierungen, die an einigen Holzkohlefragmenten durchgeführt wurden, die aus gefalteten Ablagerungen entnommen wurden, deuten darauf hin, dass eine seismische Aktivität in der Zeitspanne zwischen 120–1671 n. Chr. stattgefunden haben kann.

Keywords: *Quaternary Geology, Lithologs of Quaternary sediments, morphostratigraphy, Geomorphology, Terna River*

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1 Introduction

The Peninsular India was considered to be tectonically stable, until the Killari-Latur earthquake in 1993 (Fig. 1), which was followed by another event at Jabalpur in 1997 (Fig. 1) and continued episodes of reservoir induced earthquakes at Koyna (Maharashtra India) since 1967 (Fig. 1). The epicenter of the devastating Killari-Latur earthquake (mb=6.3) of September 30, 1993 is located in the Terna drainage basin (Fig. 2). This event is one of the rare occurrences of an earthquake in shield area and brought into focus several unresolved questions regarding the intracratonic earthquakes. The seismicity recorded in this region in the last few decades apparently contradicted the traditional notion of the tectonic stability of the Deccan Volcanic Province (DVP). These earthquakes also demonstrated the catastrophic effects and the risk of annihilating earthquakes occurring in the DVP in Peninsular India in response to ongoing neotectonic activity in the region. The observed seismicity has so far remained unexplained within a neotectonic framework in the absence of such studies in the region. No study has been attempted to so far on documentation of neotectonic evidences and its influence on

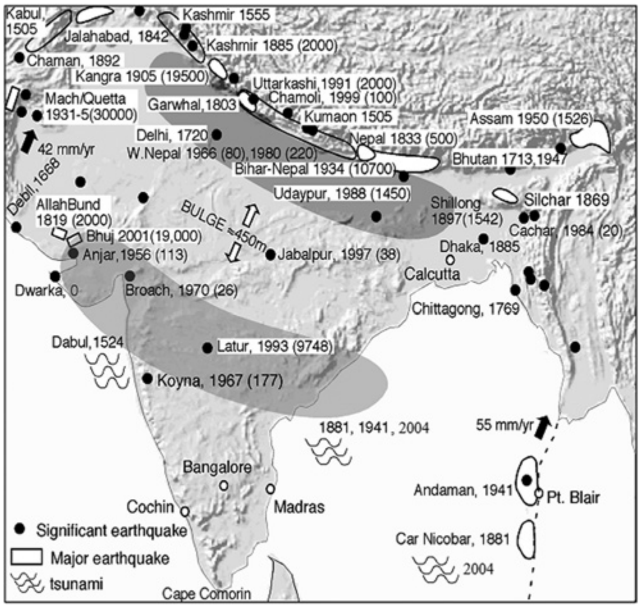

Fig 1. Schematic views of Indian Tectonics and historic Earthquakes map of India (Modified after Bilham 2004). The values in bracket indicate the number of casualties because of the earthquake. Shading indicates flexure of India: a 4 km deep trough near the Himalaya and an inferred minor (40 m) trough in south central India are separated by a bulge that rises approximately 450 m.

the shaping of the landscape in the recent geologic past. The Quaternary deposits occurring in the Terna basin have also remained uninvestigated so far, which are potential archives for delineating past neotectonic and seismic events and Late Quaternary evolutionary history.

The earthquake near Killari-Latur along Terna River demonstrated the need for detailed neotectonic reappraisal of the DVP, which consists primarily of a thick pile of trappean lava flows and narrow fringe of Quaternary sediments. Although, the lava flows have been studied extensively in terms of their petrological characteristics and geochemistry but studies on their structural aspects, geomorphological and neotectonic evolution are virtually sporadic. The alluvial deposits in western uplands of Maharashtra have been studied with respect to Neogene uplift of Peninsular India and Quaternary paleoclimatic changes (Radhakrishna 1993; Rajaguru et al. 1993; Rajaguru & Kale 1985). The studies carried out so far (Babar et al. 2000) indicate a control of structure and neotectonics on the geomorphic set up and drainage configuration of the Terna basin. Lineament and fault controlled drainage pattern, entrenched meanders, incised cliffs of Quaternary sediments and bedrock and a rejuvenated topography points to a dominant control of neotectonic activity on the landscape evolution of the area.

The paper represents the Quaternary Geology and geomorphology of Terna River basin in the DVP of West central India. The location of sites is given in Fig. 2. The Quaternary geological mapping was carried out in the area in order to generate the data on morphostratigraphy and lithostratigraphy. The lineaments occurs along NE-SW, NW-SE, E-W and WNW-ESE directions (Fig. 3), which has influenced the drainage network of the area and the tributaries of the Terna River.

2 Geology of the area

Geologically, the entire study area belongs to DVP of Peninsular India (Fig. 4). Deccan volcanism is considered to be a manifestation of original tectonic regime developed within the continental lithospheric plate (Chandrasekharam 1985; Cox 1989; Cox 1991; Bose 1996). The stress conditions in the Indian peninsula initiated formation of fissure swarms and with increasing intensity and developed miniature Continental rifting. The Killari-Latur 1993 earthquake rejuvenated the debate over the existence of rift valleys underneath the DVP (Valdiya 1993; Kailasam 1993).

The Deccan Traps, which cover an area of more than 600,000 sq km of this region, consist of a number of flows ranging in thickness from a few meters up to about 100 m with the successive flows being separated by red bole or Inter-trappean beds and are characterized by compact basalt at the bottom part succeeded by a vesicular zone (Gupta & Dwivedy 1996). The Deccan trap sequence, in general, is classified into stratigraphic units on the basis of chemical composition of various flows (e.g. Mitchell & Widdowson 1991). The southern part of Deccan volcanic province in the eastern Maharashtra is composed of Poladpur and Ambenali Formations of the Wai sub group (Mitchell & Widdowson 1991; Bilgrami 1999).

The Deccan basalt flows, in general, are broadly horizontal in disposition and exhibits gentle gradients. The gradient is towards ENE and SE. Drilling at Killari (Gupta & Dwivedy 1996; Gupta et al. 1998) indicates that the total thickness of basaltic layers is about 338 m with about 12–15 flows. The lava flows are underlained by 8 m thick infratrappean sequence comprising 1–2 m oxidized shale followed by a conglomeratic grit-sandstone. This layer overlies the Precambrian granitic basement (biotite-granitic gneisses to pink granite). In the present study area there are nine basaltic lava flows as given in Table 1.

Closely spaced gravity survey and modeling along the two profiles (Mishra et al. 1998) across the epicentral area of 1993 Killari-Latur earthquake suggest some high and low density bodies of shallow origin indicating highly heterogeneous basement. Under these circumstances the most convincing evidence of paleoseismicity as well as tectonic activity, which may have occurred in this region, is most likely to come from the sediments, which have been preserved along the rivers.

3 Methodology

3.1 Satellite Data

For the present study the IRS P6 LISS III 2010 satellite data was used to delineate Quaternary litho units of the Terna River. Active channels and floodplain features were mapped. The digital data format from Indian remote sensing satellite (IRS P6) of LISS-III 2010 with 24 m spatial resolution with four spectral bands was used to meet the requirement of area under study. The image taken was false colour composite (FCC) on 1:50,000 scale, having band combination of 4:3:2:1 (NIR: red: green). The SOI toposheets and digital satellite data were geometrically rectified and geo-referenced and merged using Arc GIS 9.3.

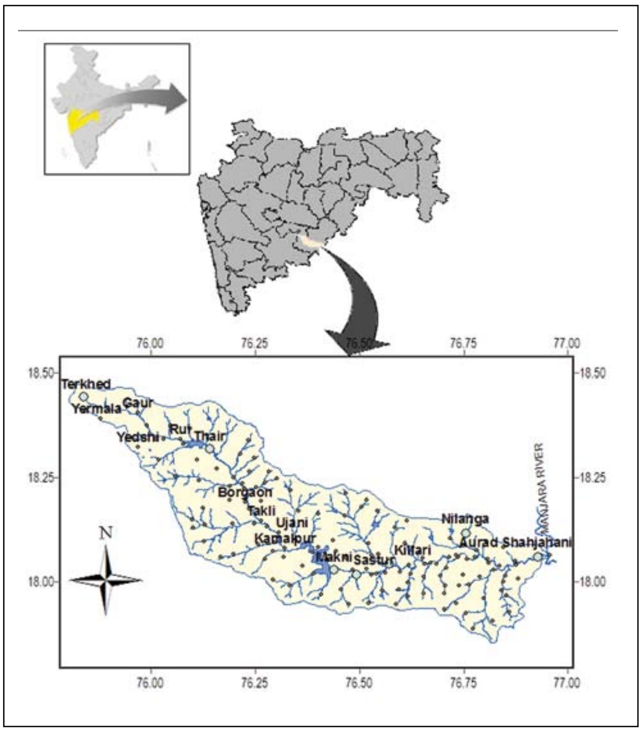

Fig. 2: Location map of study area for sites of lithologs along the Terna River Basin.

3.2 Field Work

The geomorphic study carried out was based on the satellite data, Survey of India (SOI) topographic maps and extensive field survey. The height of terrace surfaces was determined using the data acquired through toposheets and hand-held global positioning systems (GPS). The data presented are based on the field observation on several outcrops, where a representative section has been discussed in detail. The morphological details are presented after careful examination of SOI topographic maps and observations made in the field. Sections were logged and sedimentary structures were described in the field itself. The fieldwork is under taken during 2009–10.

The river shows the evidences of channel movement by avulsion and the lineaments largely control these. Older palaeo-levees exist in the form of ridges 4–5 m high at Ter, Killari, Sastur and Makni villages along the Terna River floodplain. The abnormally greater thickness of sediments is recorded at Ter village, consisting of mounds of 12 to 15 m height from the bed level of the Terna River (Rajendran et al. 1996; Sukhija et al. 2006). In the field these are marked by a curvilinear deposition of Paleolithic sites on the silty or sandy over bank deposits. They occur as irregular patches and can be related to the older course of the river. Several lineaments run NE-SW, NW-SE, E-W and WNW-ESE directions (Fig. 3), which control the basement structure in the study area. The lineament map is prepared using the basin map prepared in Arc GIS 9.3 and the lineaments are incorporated from the lineament maps of Arya et al. (1995) and Srivastava et al. (1997). The geology of area is illustrated in Fig. 4, litho-sections were logged (Fig. 5) and sedimentary structures were described.

3.3 Radiocarbon Analysis

For the present study seven charcoal samples were collected from different locations including two samples from Ter area, one each from Duta and Makhani and three samples from Killari villages locations, depth and ages with Figures referred are given in Table 4. Ages of charcoal fragments were estimated by radiocarbon dating method following liquid scintillation spectrometry (Yadava & Ramesh 1999). Benzene was synthesized from sample carbon in three radiochemical steps: 1) under dynamic vacuum sample carbon was first combusted to carbon dioxide 2) it was reacted with lithium metal to get acetylene 3) finally benzene was catalytically synthesized from acetylene. Residual radiocarbon activity of the sample benzene was measured by liquid scintillation counter (LKB-QUANTULUS). All the estimated ages reported here was calibrated using online version (http://www.calib.qub.ac.uk) of the programme Calib 6.1 (Stuiver

Tab. 1: Lava flows in the Terna River basin (Modified after GSDA, 1973–74).

Sr. No.	Flow No.	Lithology of the flow	Altitude range (m)
1	IX	Highly jointed compact basalt flow (fine grained massive and moderately weathered)	746.00 to 722.00
2	VIII	Jointed compact basalt flow (fine grained massive and moderately weathered)	722.00 to 685.00
3	VII	Highly weathered vesicular amygdaloidal basalt flow	685.00 to 675.00
4	VI	Jointed compact basalt flow (fine grained massive, grey to dark grey coloured and poorly weathered)	675.00 to 643.00
5	--	Red bole bed	643.00 to 642.00
6	V	Compact basalt flow (fine grained massive and moderately weathered)	642.00 to 634.00
7	--	Red bole bed	634.00 to 633.00
8	IV	Highly weathered vesicular amygdaloidal basalt flow	633.00 to 621.00
9	III	Jointed compact basalt flow (fine grained massive, dark grey coloured and highly to moderately weathered)	621.00 to 580.00
10	II	Poorly weathered vesicular amygdaloidal basalt flow	580.00 to 569.00
11	I	Jointed compact basalt flow (fine grained massive, dark brownish coloured and poorly weathered)	569.00 to 551.00

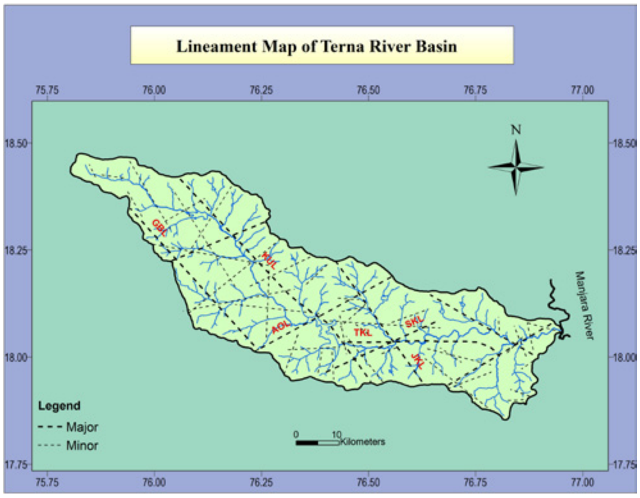

Fig. 3: Lineament map of the study area showing lineaments occurring along NE-SW, NW-SE, E-W and WNW-ESE directions (Modified after Arya *et al. 1995 and* Srivastava *et al 1997). TKL – Terna-Killari Lineament, SKL – Sastur-Killari Lineament, AOL – Ausa-Osmanabad Lineament, JKL – Jawalganala-Killari Lineament, KUL – Kallam-Umarga Lineament, GBL – Ghod-Bhima Lineament.*

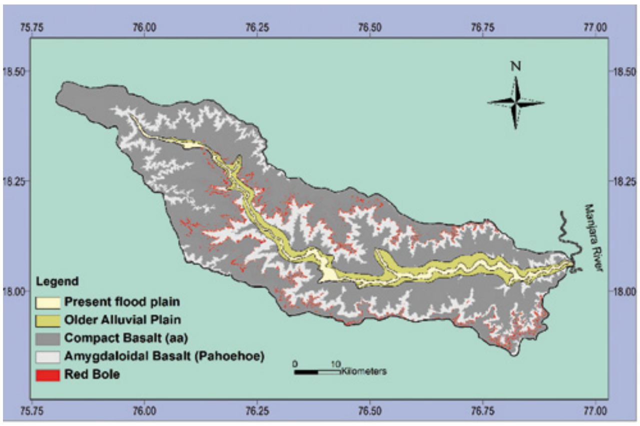

Fig. 4: Geological Map of the Terna River Basin.

& Reimer 1993). In most of the cases estimated radiocarbon ages when calibrated results into several age ranges with varying relative area (or probability). For simplifying the discussion, here we consider only those ranges which have high corresponding relative area (> 0.89, given in column 5, Tab. 4). For further details on the procedure refer the Stuiver & Reimer (1993).

4 Results

4.1 Quaternary Sediments

In the Deccan Peninsular India Quaternary deposits are primarily fluvial. They are confined to very narrow belts along rivers with not much recognizable landscape features, except for the sediments recognized along Tapi and Purna rivers (Ghatak & Ghatak 2008; Tiwari et al. 1996; Tiwari & Bhai 1997b; Tiwari & Bhai 1998; Tiwari 1999; Tiwari 2001; Tiwari et al. 2010). These deposits are often discontinuous, generally unfossiliferous and lack suitable material for radiometric dating, further more; the deposits lack proper preservation of pollen and proper sedimentological record.

The lithology of the Terna valley of older alluvium consists of dark grey sand and silts with grey brown clay and at some places development of calcretes suggests that the Older Quaternary Alluvial deposits are of Upper Pleistocene age. Lithostratigraphically the Quaternary deposits of the Terna River basin have been divided into three informal formations including (i) dark grey silt formation – Late Holocene, (ii) Light grey silt formations – Early Holocene, (iii) Dark grayish brown silt formation – Late Pleistocene. There are two formations of Holocene age including early and late Holocene, which are equivalent of the Ramnagar and Bauras formations of Narmada alluvium (Tiwari & Bhai 1997). The Quaternary sediments observed in the area are present floodplain (To), older floodplain (T1) and pediplain (T2). The fine clay and silt formations in the lower reaches reflect that the streams are of low gradient and more sinuosity. In this area monsoon is the most dominating parameter controlling the behavior of the river. The highly seasonal rainfall results in the highly seasonal discharge in the river. Most of the geomorphic work is done during the flood events that occurred during individual storm at the end of monsoon. A large part of the alluvial record is therefore produced during flood and this is well illustrated by the fineness in the sediments (Bull 1991; Nanson & Tooth 1999; Schumm 1978; Mishra et al. 2003).

4.2 Morphostratigraphy of Terna River Sediments

Alluvial plain of the Terna River shows 3 terraces namely, T0, T1, T2 in increasing order of elevations (Tab. 2). These terraces were described as suggested by Tiwari & Bhai (1997b) and Babar et al. 2010 with reference to the soil types and soil characteristics.

The lithostratigraphic formations have been identified on the basis of nature of sediments, sedimentary structures and pedogenic characters. Thus we have four lithostratigraphic formations.

The dark grey sand and silt with grew brown clay formation is correlated with upper Hirdepur formation of late (upper) Pleistocene age (13 ka to 25 ka). The Gray Sand and Silt formation and dark grey silt formation is correlated with the Ramnagar formation of late Holocene age (2 ka to 5 ka) (Tiwari 1999).

4.3 Lithologs of sediments exposed along Terna River

The lithologs of the Quaternary sediments are studied from the source area of the Terna River at Terkheda to the confluence with Manjra river at Aurad Shahjani. The important localities of the litholog studied are Yermala, Rui, Ter, Borgaon, Ujni, Makhni, Duta, Sastur, Sawari, Gunjarga, Aurad Shahjani and Wanjarkheda (Fig. 5).

At Yermala highly jointed compact basalt (aa type) lava flow is exposed on the right bank, while on left bank there is exposure of 6 m thick sediment consisting of the grey clayey soil followed by pebbly gravel, sandy silt and sandy gravel (Fig. 5 a). The litholog (Fig. 5 b) at Rui is 5.7 m thick and consist of grey clayey soil followed by sandy silt, clay, sandy silt and pebbly gravel along with compact basalt at bottom. The surprising element in the Terna River basin is the thickness of about 15 m of Quaternary sediment at Ter. The river bluff

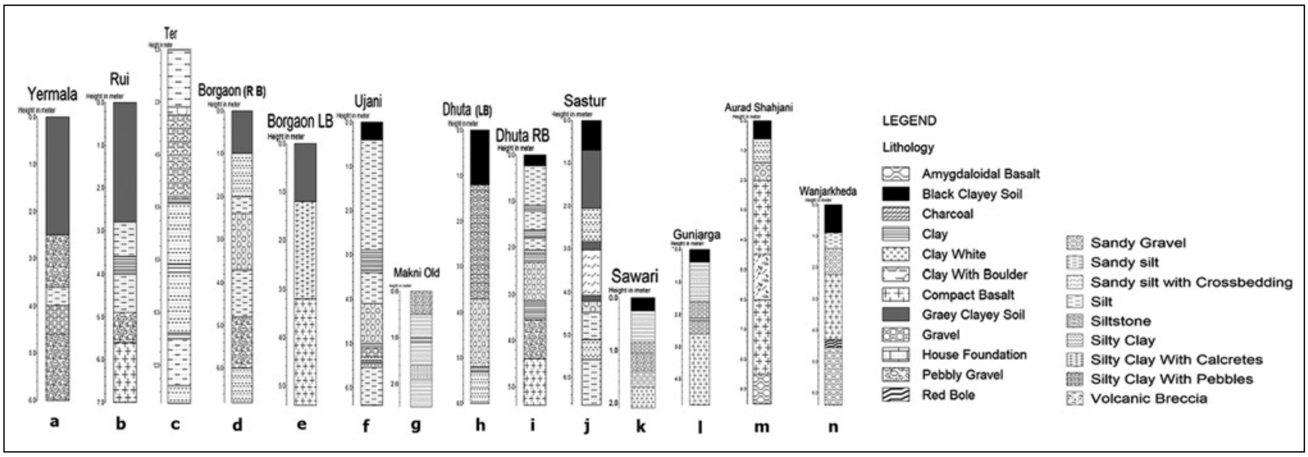

Fig. 5: Lithologs of the sediments observed along the Terna Valley. a to n indicates the exposed lithologs at different localities.

at Ter village (Fig. 5 c) in Latur District is occurring on the western bank of the Terna River.

The Quaternary sediments section occurring along the right bank of Terna valley at Borgaon (Fig. 5 d) is 6.8 m thick. The section shows the silty clay at the base resting over the present level of the floodplain followed by grayish black pebbly gravel, light grey sandy silt, dark grey sandy gravel followed by grey sandy silt and grey silty clay intercalated with thin clay layers and grey clayey soil at the top (Fig. 5 d). There is less thickness of sediment on the left bank of the Terna River i.e. 3.2 m consisting of top most grey clayey soil followed by grey silty clay with calcretes and jointed compact basalt at the base (Fig. 5 e).

The Quaternary sediment found along the Terna valley at Ujni (Fig. 5 f) is 6.4 m thick. The section shows sandy silt at the base followed by alternate layers of silt and clay, pebbly gravel layer, grey black clay, sandy gravel, sandy silt, clay and major sandy silt layers in upward succession. The topmost layer is the black clayey soil. The major sequence of this section is the thick massive grey sandy silt.

The litho-section along left bank of the Terna River is occurring at Makni village (Fig. 5 g) and having total thickness of 2.4 m. The section consists of silty clay at the base, which is followed by the dark sandy layer, silty clay, grey brown clayey soil, light grey silty clay in upward succession with top most grey black silty clay with pebbles.

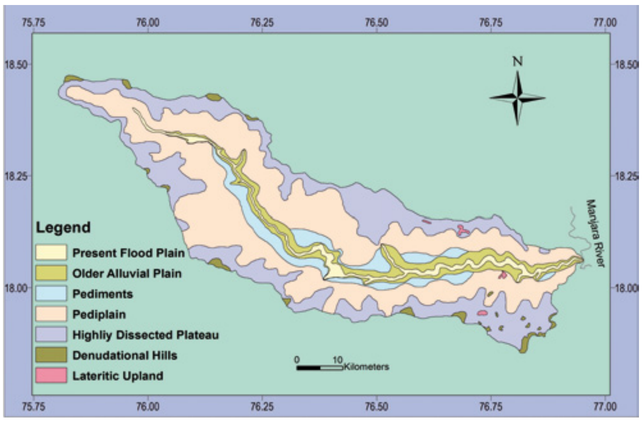

Fig. 6: Geomorphic surfaces of Terna River Basin.

The two exposures of Litho-section at Dhuta are found at left bank (Fig. 5 h) and right bank (Fig. 5 i). The left bank section (6 m thick, Fig. 5 h) shows sandy silt (0.60 m) at the base followed by alternate layers of clay (0.20 m), sandy gravel layer (1.50 m), and silty clay with pebble (2.50 m) as a major layer in upward succession. The topmost layer is the black clayey soil (1.20 m). The sedimentary section observed along the right bank of Terna valley at Dhuta village (Fig. 5 i) is 6.8 m thick. This section is developed as the compact basalt at the base followed by pebbly gravel (0.85 m) grayish brown clay (0.40), sandy gravel (0.82 m), alternating layers of silty clay and light grey clay and black clayey soil (0.24 m) at the top.

The Quaternary sediments along Terna valley at Sastur (old) village (Fig. 5 j) is 6.8 m thick. This section is developed as the grey brown sandy silt at the base followed by grayish brown clay bed, grey brown sandy silt layer, sandy gravel, which is overlain by Grey clayey soil. Above clayey soil bed there is a layer of Sandy silt showing cross bedding structure followed by light grey clayey, silt, light grey clayey and black clayey soil.

The litho-section along left bank of the Terna River is occurring at Sawari village (Fig. 5 k) and having total thickness of 2.0 m with 0.4 m jointed compact basalt exposed at the base. The Quaternary sediment (1.6 m thick) exposed consists of gravel at the base, which is followed by the silty clay with gravel, sandy gravel, silty clay and black clayey soil at the top. The similar section is also visible at Gunjarga (Fig. 5 l) with jointed compact basalt at the base and Quaternary sediments of 2.5 m thickness.

The sedimentary sections at Aurad Shahjani (2 m thick, Fig. 5 m) and Wanjarkheda (2.2 m thick, Fig. 5 n) near confluence of Terna River with the Manjra river, show the similarity in the Quaternary sediment such as the black clayey soil at the top followed by silt and then sandy gravel with jointed compact basalt at the base. The exposure of basalt flows show differences in these two areas. At Aurad Shahjani there is exposure of Amygdaloidal basalt at the base followed by jointed compact basalt flow, volcanic breccia and jointed compact basalt, whereas at Wanjarkheda there are two lava flows including Amygdaloidal basalt flow at the base followed by jointed compact basalt flow and both are separated by red bole bed.

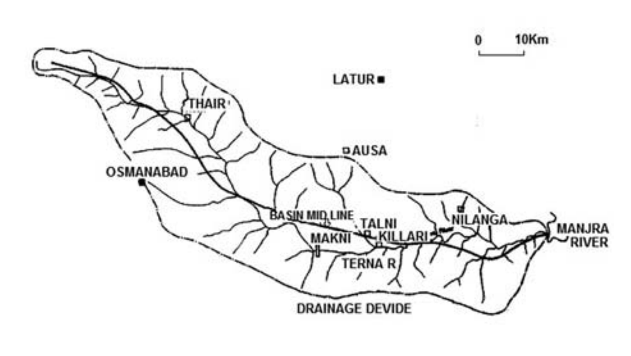

Fig. 7: Drainage morphology of Terna River basin showing the migration of river course (after Chetty *2006).*

5 Geomorphic Characteristics

The Terna River basin has been divided in to six geomorphic surfaces: present floodplain, older alluvial plain, pediplains, highly dissected plateau, denudational hills and lateritic upland (Fig. 6) based on the topographic features, morphological characteristics and IRS P6 LISS-III satellite imagery. Towards the source of the basin, i.e. in western and northwestern part, the area is characterised by the moderately steep gradient, rocky upland with deep and narrow valleys and moderately steep longitudinal profile. Such features are due to Quaternary tectonic uplift as found in upland Maharashtra (Powar 1993; Radhakrishna 1993; Rajaguru et al. 1993). The uplifted terrace, escarpment, deep grooves, dissected colluvium and a general youthful topography are believed to be indicative of tectonic uplift during the Quaternary.

The middle zone of the basin corresponds to shallow and gently sloping pediplain. The Quaternary sediments directly overlie Deccan Basalt in the pediplain zone. The third zone towards the confluence with Manjra River includes the area of Pleistocene-Holocene alluvial deposits. The development of gullies and badlands in this zone suggests active denudation processes, which may be attributed to the Quaternary tectonic uplift.

Based on geomorphic characteristics of Terna River and locations of archaeological sites, complex surface deformational features and the shallow focal depth consideration, it is suggested that block rotation tectonics about the vertical axis seems to have played a crucial role in causing such a deadly earthquake of magnitude 6.3 (Chetty 2006). The block structure of the basalts also shows considerable influence on the behaviour of seismic waves across the block boundaries. The seismic energy might have been channelled along the boundaries and interfaces amongst different compositional flows. It is inferred that the block rotation model for the basement tectonics could be responsible for the continued tectonic activity in Eastern Dharwar Craton (EDC) and in turn the inherited structural fabric and reactivation tectonics in the overlying Deccan traps.

Two major sets of lineaments trending NW-SE and ENE-WSW are inferred from satellite data (Chetty & Rao 1994) revealing a well-developed mosaic of block structure (Fig. 3), similar to that described in the EDC. Orientation of structural and lineament fabrics in the Latur region mimics those of the adjacent pre-Deccan basement regions. It is most probably the reflection of structural inheritance of the basement. The mechanism for this transmission is probably related to movements along the reactivated ancient structures in the basement exerting profound control in generating fractures and small-scale displacements in the overlying basalts. While some of the inferred lineaments terminate against the east-flowing Terna River course, sinistral strike-slip displacements could be seen along the ENE–WSW lineaments. Gravity maps (Mishra et al. 1998) exhibit many localized gravity highs and lows of 3–5 mgal coinciding with the major NW–SE striking lineaments in the region. The geomorphic features associated with Terna River (Chetty & Rao 1994) indicate that it follows a tectonically active lineament. Further, several archaeological sites along NW–SE trending lineament are also observed along the Terna River. Rajendran & Rajendran (1998) inferred that one ancient earthquake of AD 450 had occurred around one such archaeological site near Ter.

Examination of the Terna basin and its morphology reveals the shift and migration of the river course in an alternating changing fashion (Fig. 7). Migration of the river course is inferred based on the imaginary midline drawn on the basis of the symmetry of the river basin. In the northwestern part, the river course is east-west, and the migration is towards south. The migration direction changes in accordance with the change in direction of the river course. Further, it is also evident that the lineament fabric pattern influenced the river course (Fig. 3). Interestingly, the location of the archaeological site at Ter, lies at the intersection of two major lineaments. The topographic profile along the Terna River (Fig. 8) shows steep gradient until the river takes an eastward direction, 10 km west of Killari. The gradient becomes zero near Killari and further east. There is a gradual change not only in the gradient, but also in the regional topography from ~ 700 m in the northwest to 560 m in the east. The main shock and aftershock activities are restricted to the region of lower elevation. Considerable influence of the lineament fabric on the topography, drainage pattern as well as on the river gradient is evident.

Any tectonic deformation that changes the slope of a river valley will result in corresponding changes in sinuosity so as to maintain an equilibrium channel slope (Keller & Pin-

Tab. 2: Morphostratigraphy of Terna Alluvium.

Terrace	Origin	Soil Type	Soil Characteristics	Av. Elevation m amsl
T0	Depositional	Entisol (I)	Dark Gray Sand and Silt	574.0
T1	Erosional	Inceptisol (II)	Gray Sand and Silt	582.5
T2	Depositional	Vertisol (III)	Dark Gray Sand and Silt with Gray Brown Clay	591.0

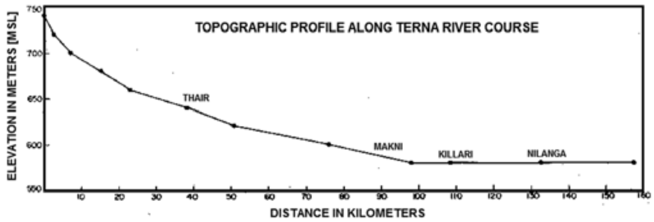

Fig. 8: Topographic profile along Terna River showing steep gradient before becoming flat (after Chetty 2006).

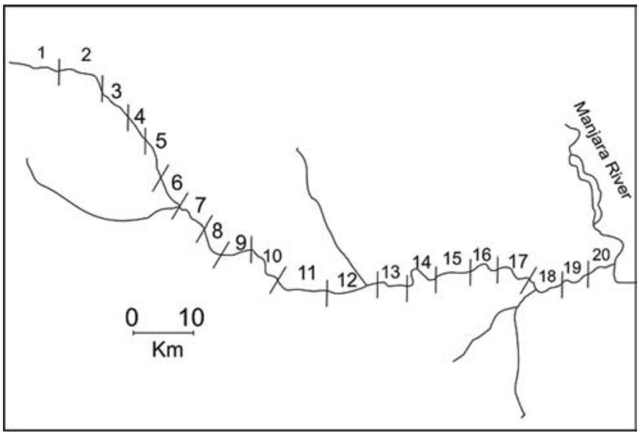

Fig. 9: Map showing segments studied for sinuosity.

ter 1996). Thus sinuosity parameters can be used to deduce the role of tectonism development of channel morphology (eg. Gomez & Marron 1991; Rhea 1993; Rachna Raj et al. 1999). The sinuosity of a meandering stream is the result of topography and hydraulic factors, which can be expressed by a ratio called the index of sinuosity (Muller 1968). The river channel is divided into number of segments (Fig. 9) as suggested by Muller (1968) and Friend & Sinha (1988) for determination of sinuosity parameter. The measurements of channel length (CL), valley length (VL) and the shortest distance between the source and the mouth of the river (AL), i.e. air lengths are used for calculation of channel index, CI=CL/AL and valley index VI= VL/AL. The standard sinuosity index is calculated using SSI= CL/VL, hydrological sinuosity index HIS=% equivalence of CI-VI/CI-1 and topographical sinuosity index TSI=% equivalent of VI-1/CI-1. The standard sinuosity index (SSI) for the Terna River channel varies from 1.002 to 1.76 (Tab. 3). The increase in SSI in the lower reaches of the basin is accompanied by lower values of Hydraulic Sinuosity Index (HSI), which suggests that the hydraulic factor is not responsible for increase in SSI in the lower reaches. The values of HSI vary from 2.13 to 85.58 (Tab. 3).

Low values of HSI and correspondingly higher values of Topographic sinuosity index (TSI) in the upland areas and pediment zone suggests that the meandering streams do not belong to the initial denudation cycle (Friend & Sinha 1988), but the present area has been rejuvenated there by indicating the role of tectonism. As the river progresses in the cycle of erosion, the role of hydraulics increases and the role of topography decrease (Rachana Raj et al. 1999). This is observed in the lower reaches of the area; where there is a relative increase occurs in HSI though it remains substantially lower than TSI (Tab. 3).

6 Discussion

The Peninsular Shield of India was supposed to be seismically stable, but the 1993 Latur earthquake indicated the seismic vulnerability of the area. The deep drilling in the Deccan Volcanic Province in basalt flows, on both side of 1993 rupture zone provided the evidence of a fault and displacement of about 6 m, at a depth of 220 m (Gupta et al. 1998). Down dip slickenlines on the steep dipping slickenside surface in the drill cores confirm dip slip nature of the fault. However the observed displacement is too much to account for a single earthquake of Mw 6.1, hence they suggested repeated seismicity in the area. Rajendran & Rajendran (1999) suggested the reactivation of the pre-existing fault and evidence of earlier seismicity in the area by trenching in the rupture zone of 1993 earthquake near Killari. While Sukhija et al. (2006) find the wide spread geological evidence of a large paleoseismic event near the Meizoseismal area of the 1993 Latur earthquake at Ter on Terna River and Halki and Shivoor on Manjra river.

The deformational structures in the sediments observed are flexures, warps, buckle folds and vertical offset in the sediments (Fig. 10). These structures are earlier studied by Rajendran (1997), Rajendran & Rajendran (1999) and Sukhija et al. (2006). The variation in the individual layers that belong to the same deforming mass can be explained by strain partitioning, which depends on the bulk properties of the rock (Hatcher 1995). Because of partitioning of mechanical behaviour, the stiffer and more competent rocks are expected to show variation is shapes and wavelength of folds. From the style of deposition it is clear that buckling of the sediment strata has formed the structures in section at Ter. Here the buckling must have been accompanied by flexural slip between the layers (Fig. 11).

The vertical offset of marker horizons at the northern part of the section (Fig. 12) indicates a displacement of 10–15 cm. The up thrown block is on the southern side of the offset plane. These features are observed in the 15–20 cm thick inter layers of whitish clay in blackish clay.

Detailed assessment of morphological and morphometric characteristics have confirmed the role of neotectonism in the evolution of Terna River basin. The relative degree of tectonic activity is also manifested by the anomalous behaviour of the streams such as right angled turn of the streams, convergence and divergence of stream, streams flowing parallel to main river for a considerable distance and offset drainage (Babar et al. 2000). All the parameters in general, suggest an increasing degree of tectonic activity from lower reaches towards the upland source region. The alignments of the significant morphological features and stream orientations have provided information about the linear tectonic elements. The anomalous behaviour of streams indicates that subsurface faulting may be controlling the drainage patterns of the basin (Kaplay et al. 2004).

The orientations of the stream channels in the upland zone and in the middle reaches of the Terna River basin have been guided by lineaments, which are the indicators of recent tectonic activity. Higher order stream channels reflect the general NW-SE and E-W trend. These trends have been reactivated in recent times as shown by displaced Quaternary deposits (Rajendran & Rajendran 1999). The offset in the sedimentary section at Ter (Fig. 10, 11 and 12) reflects only a fraction of movement in the basement fault below the basalt flows. On the basis of seismogenic features exposed in the sedimentary sections including flexures, warps, buckle fold and vertical offsets. Rajendran (1997) marked the signatures of pre-existing earthquake (~1500 year ago). These deformational features may be the result of reactivation of NW-SE trending fault. These older tectonic directions, although active until very recent times, had conditioned the drainage network in an earlier period. The TSI values indicate rejuvenation of the area leading to the dominating effect of topography on the sinuosity of the river channels. The break in slope in the long profile is also indication of the Quaternary tectonic uplift of the area. Additional support for the neotectonic activity in the upland zone and middle zone of the Terna River basin is provided by valley floor ratios and longitudinal profile. Morphometric analysis has thus been useful in delineating areas with differing levels of tectonic activity in the basin (Babar et al. 2009).

The reactivation of pre-existing basement structures was proposed by Chetty & Rao (1994), while Kayal (2000) opined that the Latur earthquake could be due to interactions of two shallow crustal faults. However, Chetty (2006), proposed an alternative explanation in terms of block rotation tectonics as a plausible mechanism for the Latur earthquake. According to him based on the structural fabric in the EDC, as derived from satellite data as well as aerial photos, and the unusual shapes, sizes and geometry of mafic dykes and distinct fault systems, block rotation tectonics with clockwise rotations were inferred from the deformational system of the EDC. Block rotation is a significant mode of deformation in the earth's crust (Freund 1970; Kissel & Laj 1989; McKenzie & Jackson 1986). The Latur region lies in the proximity of the Kurudwadi lineament, earlier described by Brahmam & Negi (1973) as a subtrappean rift on the basis of gravity anomalies. Based on geomorphic studies using satellite data and aerial photos, Peshwa & Kale (1997) concluded that this is a Precambrian deep crustal-scale shear zone comprising an array of NW-SW trending faults along which dextral sense of movements have taken place, even during the Quaternary period. This is evident from the parallelism of the drainage network with the Kurudwadi lineament, suggesting the control of basement structures in their development. These movements based on the presence of sheared segments of the Archean gneisses were also responsible for the secondary development of east-west trending faults identified by gravity studies. Structural architecture of the Latur earthquake region presented in this study favours the block rotation model, which could be a part of dextral sense of shear along the NW-SE trending lineaments (Chetty 2006).

Fig. 12: Close-view of offset in middle of the section at Ter shown in Fig. 10.

Fig. 10: Quaternary Sediments at Ter showing folding in upward section and faulting (f – f) in the middle part. The upper layer is marked with the pottery layer showing the warping (Location is given in Fig. 2). Arrows indicate the direction of compression for folding and horizontal shortening.

Fig. 11: Photograph showing the flexure and horizontal shortening in the sediment at Ter village.

The seismicity associated with the Killari source is comparable to those in other cratons, such as Australia. Interestingly, location of historic earthquakes during AD 1201–1960 (Geological Survey of India) appears to be mostly confined to a 400-km-long NW corridor passing through Killari (Fig. 1). Spatial correlation of this corridor of activity with a structure inferred from a variety of data as well as the fault plane solution of the main event suggested reactivation of a NW-oriented fault (Rajendran & Rajendran 1999). Search for palaeoearthquakes in the vicinity of Killari led to the identification a deformation event dating to AD 350–450, at a location known as Ter, about 40 km northwest of Killari (Rajendran 1999). The data on age dating of the charcoal samples varies from AD 120 to AD 1671 and is given in Table 4. In the present study the calendar ages of the deformed section at Ter is found to be between AD 1151 to AD

Fig. 13: (a) Sediment section at Makni showing deformation in silty clay formation, (b) Sediment section at Dhuta along the left bank of Terna valley.

353, while for the deformed sections at Dhuta and Makhni the corresponding calendar ages are between AD 650 and AD 1183. Similarly, the deformed structure at Killari (Fig. 14) indicates the dates are between AD 1256 and AD 1454.

The litholog of Quaternary sediments occurring along the left bank of Terna valley at Dhuta (Fig. 5 h) is 6.0 m thick. The section shows the silt at the base resting over the present level of the floodplain followed by grayish black pebbly gravel, light grey sandy silt, dark grey sandy gravel followed by clay, grey sandy gravel and thick grey silty clay with pebble and black clayey soil at the top. The charcoal sample from this sediment (Fig. 13 b) has been dated and found the age of 1010 ± 110 year B.P. (Tab. 4). The sedimentary section on the right bank of the Terna River at Dhuta is 4.4 m thick and quite different than the left bank section. It consists of top most black clayey soil followed by alternate layers of sandy silt and clay, which is succeeded by sandy gravel, clay and pebbly gravel (with subangular pebbles) and jointed compact basalt occurs at the base (Fig. 5 i).

The epicenter of Latur 1993 earthquake old Killari village is located on the left bank of the Terna River and has an approximately 08 m thick layer of alluvium topped by the anthropogenic dump. The deposit is in the form of a mound

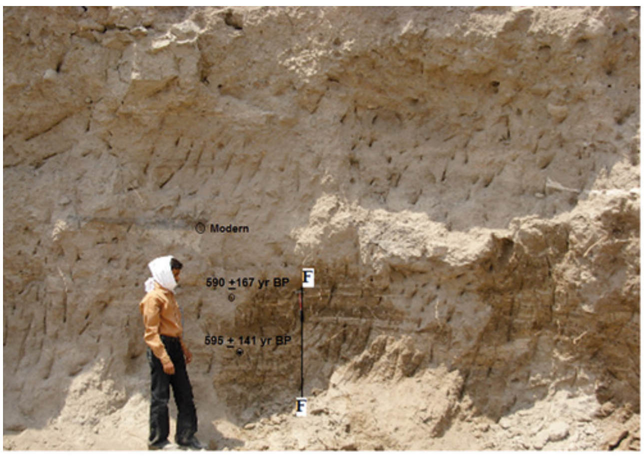

Fig. 14: Sediment section at Killari showing Fault at a depth of 6.20 m.

occurring below the ruins of the Killari-Latur earthquake of 1993. The trench is developed in this region because of the fact that the local people are excavating the soil for the purpose to use it as a fertilizer. The alluvium is highly dissected and now represented by irregular excavated mounds. Old Killari village was spread over these mounds, now nothing left except the Nilkantheshwar Temple and remains of earthquake affected dump.

The bluffs in general show evidences of deposition by river surges and are marked by alternating layers of coarse sands with cobbles and silty clay. The courser layers may have been deposited during the floods and fine sediments during the leaner seasons. Texturally the sediments can be classified as clay loam, sandy clay and silty clay loam. These types of sediments have been noted for major rivers in Maharashtra and are categorized as flood loams or diluvium (Rajaguru & Kale 1985). These rivers are noted for highly fluctuating discharge and active channel migration (Rajaguru et al. 1993). In the Ter village section, in a pit at the base of bluff, it is found that the alluvium extends > 2m below the present riverbed. This probably suggests that fluvial processes at Ter (Rajendran 1997) must have started much earlier, analogues to other rivers in Maharashtra and the same is the case of the Killari area.

The well-exposed vertical section of the mound along the left (southern) bank of the Terna River, was selected for paleoseismic investigation, this 8 m thick section (Fig. 10) extends in E-W trending arc measuring about 35 m long. This section mainly consist of dumped material like broken bricks, pottery, boulders etc for the top 1.5 m, followed by gray clays inter-bedded with varied mixtures of sand, silt and ash in the form of either thin layers or elongated lenses and wedges. The bedding is imperfect and is commonly marked by colour variations in individual layers. The fluvial/fluvio-lacustrine nature of bottom layers at the depth of 7.5 m is evidenced by the presence of load cast and scour and fill structures deposited in a high-energy environment. Pebble/stone beddings at that depth.

The site has been modified by human activity and artifacts like pottery, beads, idols, human bones and even large objects like earthen pots and a number of in situ wooden

Tab. 3: Sinuosity variation in Terna River channel.

Segment No.	CL (Km)	VL (Km)	AL (Km)	CI	VI	SSI	HSI	TSI
1	8.60	8.52	8.36	1.033	1.020	1.014	39.39	60.61
2	9.72	9.70	8.52	1.141	1.138	1.002	2. 13	97.87
3	6.48	6.35	5.76	1.125	1.102	1.020	18.40	81.60
4	5.28	5.04	4.68	1.128	1.077	1.048	39.84	60.16
5	7.32	6.97	6.25	1.171	1.115	1.050	32.75	67.25
6	6.49	6.44	5.76	1.127	1.118	1.008	56.52	43.48
7	5.76	5.54	5.08	1.134	1.091	1.040	32.75	67.25
8	6.00	5.78	5.40	1.111	1.087	1.022	21.62	78.38
9	5.64	5.62	5.40	1.044	1.041	1.004	6.82	93. 18
10	7.49	6.94	6.25	1.198	1.110	1.079	44.44	55.56
11	9.72	9.24	8.43	1.153	1.096	1.052	37.25	62.75
12	8.88	8.16	8.04	1.104	1.015	1.088	85.58	14.42
13	7.08	6.27	6.09	1.163	1.030	1.129	81.60	18.40
14	6.24	5.48	4.08	1.529	1.343	1.138	35. 16	64.84
15	8.16	7.23	6.25	1.306	1.157	1.129	48.69	51.31
16	6.61	5.66	5.16	1.281	1.097	1.168	65.48	34.52
17	6.27	5.60	5.25	1.194	1.067	1.119	65.45	34.55
18	7.32	6.25	5.52	1.326	1.132	1.171	59.51	40.49
19	5.28	4.49	4.20	1.255	1.069	1.176	69.33	30.67
20	4.44	3.79	3.46	1.284	1.094	1.174	66.90	33. 10

CL – Channel length, VL – Valley length, AL – Air length, CI – Channel index, VI – Valley index, SSI – Standard Sinuosity Index, HSI – Hydraulic sinuosity index, TSI – Topographic sinuosity index

posts used for construction of various structures are present in the section. There is a small wedge shaped burnt layer of 10 cm thickness is found in the section at the depth of 5.7 m from the ground surface.

The structural feature observed in this section is the northwest dipping normal fault trending N-S (Fig. 14). The observed fault appears to be a secondary manifestation of a deep-seated disturbance in the area. Surface faults are not reported in the region. Ancient faults are likely to be present below the Deccan trap volcanic cover and do not have any direct expression on the surface. Hence, it becomes necessary that geomorphic evidences indicating tectonic activity have to be linked with seismicity via drainage pattern, soft sediment deformation in alluvial and colluvial sediments (Chetty & Rao 1994). Thus in the absence of surface expression of fault and in view of the presence of several inferred faults in the region, it is thought reasonably to conclude that features observed in this section could the surface manifestation of a deep seated disturbance in the region.

The fault in the section is observed in the variegated clay/silty clay beds separated by feature less horizon of 3.5 m thick silty clay. The fault is about 6.2 m below the ground surface and the observed displacement is about 40–45 cm of the silty clay bed. The layers above these faulted sediments are undisturbed, while the clay horizon at the lower portion of the fault shows the severity of frictional and compressional forces acting simultaneously on it. It is attributed by the present study that the displacement along the fault and slickensided surfaces of the clay blocks as the surface manifestation of the tectonic disturbance. To assess the possible time of the faulting a number of charcoal samples were collected around the fault as well as from the rest of the section.

The well-exposed vertical Section of the mound along the left bank of the Terna River is 3.5 m thick section (Fig. 15 a) extends in a NE-SW trending arc measuring about 25 m long. This section mainly consist of dumped material like broken bricks, pottery, boulders etc for the top 0.35 m, followed by gray clays inter-bedded with varied mixtures of sand, silt and ash in the form of either thin layers or elongated lenses and wedges. The bedding is imperfect and is commonly marked by colour variations in individual layers. The sedimentary unit observed in the section on the whole has been warped at different scales. On a large scale, the entire section appeared to have been folded. Individually, the structures present in the section can be broadly categorized as flexures, warps and buckle fold (Fig. 15 a). The burn layer in the same section shows the folding (Fig. 15 b). The structures including warping and low amplitude folding of near surface beds of alternating clay and cohesion less sediment have been reported from the other earthquake prone areas (Audemard & de Santis 1991). It was observed that the structures at Killari are better developed on more competent layers such as the layers containing ceramics and pebble foundations made by settlers, although weak traces of folds can be discerned on the argillaceous layers as well, on cluster look. Variation in intensity of deformation observed in individual layers that belong to the same deforming mass can be explained by strain partitioning, which depends on the bulk properties of the rock (Hatcher 1995). From the style of the deformation, it is clear that the structure in the section at Killari have been formed by buckling of the sediment strata.

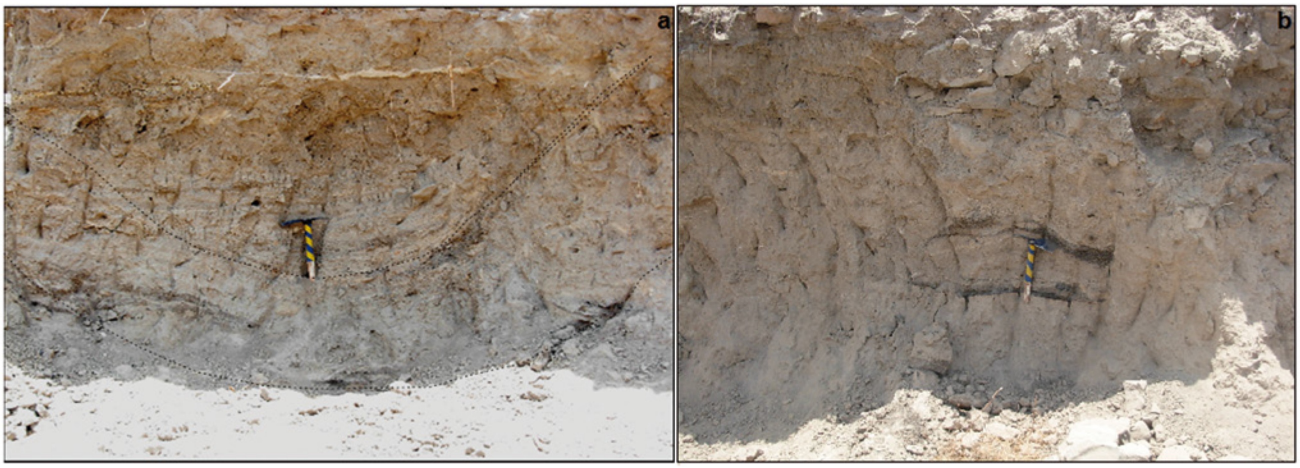

Fig 15: (a) Sediment section at Killari showing Folding in lower part at a depth of 3.87 m, (b) Sediment section at Killari showing Warp at a depth of 2.60 m in a burn layer.

7 Conclusion

The study of lithologs and the description of Quaternary sediments of Terna River basin indicate that there is significant amount of the coarse gravelly deposits along with silty deposits. These deposits are indicator of changes in the hydraulic conditions which are induced by climate or tectonics. The lithology of the Terna valley alluvium suggests that the Older Quaternary Alluvial deposits are of Upper Pleistocene age. Lithostratigraphically the Quaternary deposits of the Terna River basin have been divided into three informal formations including (i) dark grey silt formation – Late Holocene, (ii) Light grey silt formations – Early Holocene, (iii) Dark grayish brown silt formation – Late Pleistocene. The basin area has been divided in to six Quaternary geomorphic units including present floodplain, older alluvial plain, pediplains, highly dissected plateau, denudational hills and lateritic upland. The lineaments occur along NE-SW, NW-SE, E-W and WNW-ESE directions, which control the basement structure in the study area. The TSI values indicate rejuvenation of the area leading to the dominating effect of topography on the sinuosity of the river channels. The break in slope in the long profile is also indication of the Quaternary tectonic uplift of the area. The radiocarbon dating of charcoal samples indicate that the event of palaeoseismic activity might had taken place along the Terna valley from AD 971 to AD 1183 and at Ter it may be AD 1151 to AD 353.

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Tab. 4: Data on age dating of Quaternary Sediments.

Lab No.	Sample ID	Location	Depth cm	uncal ^{14}C Date year BP	Ref. Fig.
PRL-3154	TTE-1	Ter	284	570+/-340	Fig.10
PRL-3155	TTE-2	Ter	565	1,850+/-120	Fig.10, 12
PRL-3153	TMK-2	Makni	255	1,300+/-140	Fig. 13 (a)
PRL-3157	TDU-1	Dhuta	211	1,010+/-110	Fig. 13 (b)
PRL-3150	TKL-2	Killari	590	Modern	Fig. 14
KL-1	KL-1	Killari	610	590+/-167	Fig. 14
KL-2	KL-2	Killari	622	595+/-141	Fig. 14

* based ^{14}C half-life=5730 yr

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Reconstructing 2500 years of land use history on the Kemel Heath (Kemeler Heide), southern Rhenish Massif, Germany

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Abstract: The Kemel Heath (Kemeler Heide) in the Lower Taunus Mts. was used as a heath until the early 19th century. Today, it is the most densely wooded area of the German state Hesse (about 60 %). The history of the regional landscape and the land-use patterns of this area in the last 2500 years will be reconstructed by different methods and considering relicts, which have been preserved in the forest.

For reconstructing the former situation, three deserted agriculture areas with well recognizable field balks under forest were investigated for the first time by ^{14}C and OSL dating, sediment analysis and mapping. Furthermore, points of interest were the traces of Early Modern charcoal burning. For this purpose, we reconstructed the spectra of tree-species of the burned wood and dated it by ^{14}C. In addition, we dated the formation of two former slag heaps, of a medieval refuge castle, and calculated the sedimentation rate of a small colluvial filling of a slope depression that was deposited since the Roman times.

Regarding the results, there are clear traces of land use during the Iron Age and Roman Period, and strong impacts during the Middle Ages and the Early Modern Period. Thus, it is likely, that the deforestation in the investigated area was much higher during these periods than previously believed. Most of the field balks originate from the High Middle Ages. In contrast, during the Early Modern Period, the landscape was predominantly pastureland.

Die Rekonstruktion der Landnutzungsgeschichte während der letzten 2500 Jahre auf der Kemeler Heide im südlichen Rheinisches Schiefergebirge

Kurzfassung: Die Kemeler Heide im westlichen Hintertaunus ist heute Teil des größten zusammenhängenden Waldgebietes in Hessen mit einer Waldbedeckung von rund 60 %. Bis ins frühe 19. Jahrhundert wurde sie jedoch als Heide genutzt. Mit der vorliegenden Studie wird versucht, die regionale Landnutzungsgeschichte auf der Kemeler Heide mithilfe verschiedenartiger methodischer Ansätze zu rekonstruieren. Eine besondere Berücksichtigung erfahren dabei historische Relikte, die sich im Wald erhalten haben.

Zur Rekonstruktion früherer Landnutzungssysteme wurden hochmittelalterliche Ackerraine in drei verschiedenen Wüstungsfluren kartiert und im Hinblick auf ihre Sedimentzusammensetzung und ihr Alter untersucht. Die Datierung derartiger Ackerkolluvien erfolgte erstmals mit mehreren ^{14}C- und einer OSL-Datierung. Ein weiterer Schwerpunkt der Untersuchungen waren frühneuzeitliche Holzkohlemeilerplätze, anhand derer die Artenzusammensetzung der frühneuzeitlichen Wälder rekonstruiert werden konnte. Zusätzlich wurden auch zwei verschiedene Schlackenhalden als Hinterlassenschaften hochmittelalterlicher Eisenverhüttung datiert und die Ergebnisse mit den Sedimentationsraten einer kolluvialen Dellenfüllung verglichen.

Dabei konnte nachgewiesen werden, dass die anthropogene Landnutzung auf der Kemeler Heide spätestens während der Eisenzeit begann. Die stärksten Einflüsse erfolgten jedoch erst während des hohen Mittelalters und der frühen Neuzeit. Besonders im Hochmittelalter führte ausgedehnter Ackerbau dazu, dass der Waldanteil weitaus kleiner war als heute. Die meisten Ackerraine stammen daher aus dieser Periode. Während der Neuzeit wurde dagegen vermehrt Heidewirtschaft betrieben.

Keywords: *Field balks, charcoal burning, iron slag, deforestation, sedimentation rate, Rhenish Massif, Taunus Mts.*

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1 Introduction
1.1 Open questions

The Kemel Heath (*Kemeler Heide*; Fig. 1) is a historic area of around 60 km² in the Lower Taunus Mts. characterized by wide spread etchplains and deeply incised valleys at their edges (Hüser 1972). Today, about 60 % of the former heath area is forested (forest area of the community of Heidenrod; Hessisches Statistisches Landesamt 2008). However, historical reports confirm a large deforestation in the past (Ehmke 2003). So far, very little is known about the real proportion of the deforested area and land use intensity during different prehistoric and historic periods in this area.

In at least half of the wooded area of the Lower Taunus Mts., remains of former agriculture, such as field balks (Fig. 4) and clearance cairns are visible, which give evidence of an enhanced cropland area in the past. In most cases, the age of these relicts is quite unknown. In addition, there are frequently found kiln sites as relicts of former charcoal production for the local iron industry.

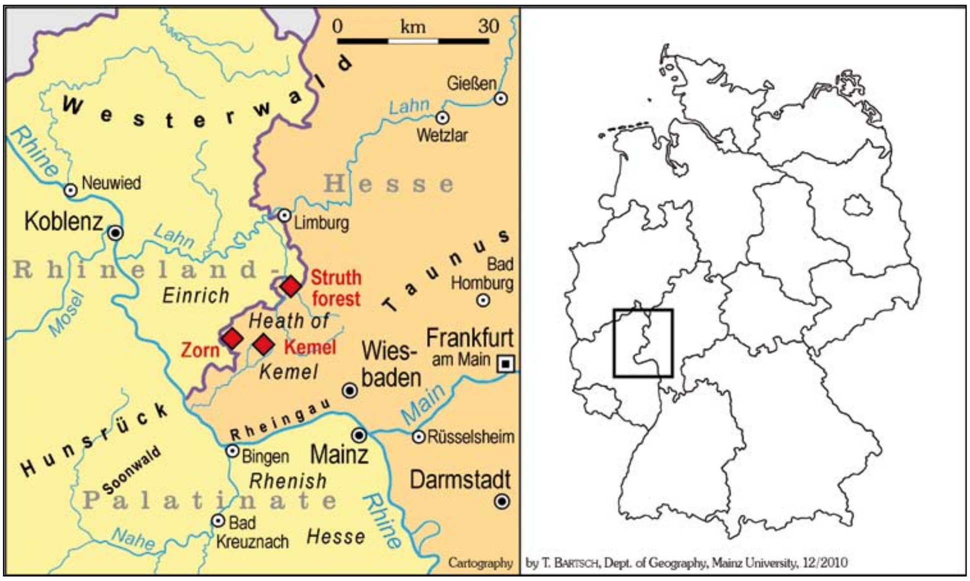

Fig. 1: Regional setting of the investigation area.

Abb. 1: Lage der Untersuchungsgebiete.

The special aim of this study is to create a chronological order of these remains for reconstructing the land-use distribution during specific periods and for comparing this with historical data: How old are the field balks and clearance cairns at selected locations, how are they structured and, in which way is it possible to date the concerning colluvial sediments? In which periods agricultural activities can be proven on the Kemel Heath? Furthermore, when the kiln sites were being used and which was the composition of the forests? Moreover, this study focuses on the consequences of former land-use. This was done by the investigation of the filling of a small valley the chronology of which was used to identify local soil erosion phases. The different sedimentation rates of the valley filling are given in mm/a.

We used a small valley filling for the chronological classification of local soil erosion phases (listed in mm/a), respectively by heavy rainfalls and a strengthened susceptibility on soil erosion by land-use.

Altogether, we investigated 7 forested locations in 3 different parts of the Kemel Heath by pedological analyses of selected soil profiles inside of field balks, datings of organic remains by AMS-^{14}C, one dating of Holocene colluvium by Optical Stimulated Luminescence (OSL) and by the anthracologic analyses of charcoals from kiln sites (Fig. 1):

a) 3 former field locations: In the Pfaffenwald forest (the term means "forest of pastors", in the past it was probably owned by the church), district of Heidenrod-Zorn (N 50°10'8.61", E 7°55'34.3"); in the Ohren-Forest (the term Ohren means the former presence of maple trees), district of Heidenrod-Niedermeilingen (N 50.1742°, E 7.9458°); in the Struth forest, district of Aarbergen-Kettenbach at the outside rim of the Kemel Heath (N 50.257°, E 8.065°).

b) 2 charcoal kiln sites: In the Pfaffenwald Forest, district of Heidenrod-Obermeilingen (N 50.167°, E 7.919°) and in the Reißbrühl forest, district of Heidenrod-Zorn (N 50.169°, E 7.930°; both are located on north-facing slopes which improves comparability).

c) 2 slag heaps: In the district of Zorn, five slag-heaps in the forests were found. Most of them are not located close to flowing waters, which may be a hint of a relatively high age (14th century and earlier; Geisthardt 1954). The first one we investigated is located a few meters beside the refuge castle (see below). Another one is located in the upper course of the small stream Rödelbach near Zorn (N 50.15829°, E 7.93905°).

d) 1 refuge castle in the district of Heidenrod- Zorn: The refuge castle *"Alte Schanz"* (N 50.1538°, E 7.9145°) is an 18m broad and 1.5-3m high, round earthwork in the Struthheck forest with a moat around. Von Cohausen (1879) assumed it to be a part of transition phase from Early to High Middle Ages.

e) The sediment sequence of the alluvial fan of a small valley near the village of Kemel, which included a Roman water well (N 50.1650°, E 8.0147°).

2 Regional setting and state of research

2.1 Physical conditions

The Kemel Heath belongs to the northern foreland of the Taunus mountain range. The altitudes range from 350 to 540m a.s.l. Its bedrock consists almost exclusively of mostly clayey Paleozoic metamorphs, which are completely weathered on the higher etchplain levels (Felix-Henningsen 1990). The bedrock on the slopes is comprehensively covered by loess-containing periglacial cover beds on the slopes typically of at least three different solifluction layers. This profile can be typically divided by the German concept of periglacial slope deposits into at least two, however, more often into three different solifluction layers (Semmel 1968, Kleber 1997, Völkel et al. 2002, Stolz & Grunert 2010, Semmel & Terhorst 2010). In the Kemel region, the cover-beds mostly consist of a loess containing upper layer and one or several basal layers, rich in debris. Additionally, at protected locations there are one or several loess containing intermediate layers. Common soil types on these sediments are cambisols, cambisol-luvisols, luvisols and podsolic cambisols. Due to the clayey weathered bedrock, several of these profiles show stagnic conditions. The annual precipitation rate is approx. 750 mm and the annual temperature is 8 °C (data from the climate station of Waldems-Steinfischbach, measuring period 1961-1990; DLR-RLP 2012).

2.2 Cultural history

Principally, the German Uplands can be divided into older and younger settled areas. The first ones, settled from prehistoric periods to the Early Middle Ages, are the loess-rich forelands of the mountain ranges, the tectonic depressions and the valleys of the large rivers, such as the Rhine (Born 1989). An exception is the village of Kemel. Its name without a suffix, respectively with the former suffix -aha, is of prehistoric age and probably of Celtic origin (Bach 1927). The settlement was located at the important "Hohe Straße" (*High Street*; connection between the old cities of Mainz and Koblenz) on the watershed between the Lahn and Rhine catchments. Along this probably prehistoric road, there are hundreds of grave mounds from the Iron Age. Due to their special construction some of them must be of Bronze Age. One example was been found in the district of Laufenselden (6 km S of Kemel; Kubach 1984). Basically, most of these grave mounds in the Lower Taunus are located along paths on the watersheds and typically arranged like cemeteries (Herrmann & Jockenhövel 1990, Behaghel 1949). However, prehistoric settlements are unknown in this area (Schwind 1984). Kubach (1984) believes that there is a finding gap in the Taunus Mts. Evidences of settlement activities for this period are located in the Rhine-Main lowland, the Basin of Neuwied and the adjacent early settled areas outside of the Taunus Mts.

Furthermore, the Romans had settled in the region between 10 and 260 AD and built up the Upper German-Raetian Limes. Its western part was probably constructed since 85 AD and crossed the prehistoric "Hohe Straße" near the village of Kemel (Baatz & Herrmann 2002). Roman forts on the Kemel Heath were located in Kemel and Holzhausen an der Haide. Several *villae rusticae* were only known from the southern Rhine-Main area.

From the Early Middle Ages (Merovingian Period) only some findings are known from the Basin of Nastätten (Neumayer 1993).

At the end of the 10th century AD, the region was ruled by the archbishops of Mainz, who started a colonization phase. In many upland regions of western Germany, the colonization of the High Middle Ages (*Hochmittelalterlicher Landesausbau*) started at the same time (Born 1989). Numerous settlements with the suffixes -roth, -schied and -hain are indicative of this period on the Kemel Heath (Bach 1927). In agriculture, the shifting cultivation including grassland and cropland phases (*Feld-Gras-Wechselwirtschaft*) was widely disseminated (Ehmke 2003, cf. Born 1989).

Bork et al. (1998) describe a change from the bread-eating to meat-eating people in Central Europe since the Late Middle Ages. Therefore, the term "heath" can be explained by a predominant use of pastureland, mostly for sheep in a sparsely wooded area during the Early Modern Period (since approx. 1500 AD; cf. Born 1989) up to the beginning of the 19th century. The main part of the heath was an area with common grazing rights for everyone's animals (*Allmende*). In most villages, there were many more sheep than inhabitants during the Early Modern Period (Stolz 2008).

The small town of Nastätten at the rim of the Kemel Heath was a center of wool weaving and textile fabrication since the 16th century, which was already mentioned in the 13th century (Spielmann 1926). During the 15th and 16th century, towels from Kemel heath and from the surrounding Nassau and Hesse territories were even traded by the powerful merchant family, Fugger, in Augsburg, Bavaria (Orth 1953).

At the same time, the region was also a center of iron production with a high consumption of charcoal.

The charcoal which was primarily needed for iron-smelting was produced in numerous charcoal kilns. For production, small round or oval leveled places on slopes or on plateaus were prepared by the charcoal-burners. On these places, the wood branches were stacked and covered by grass and earth material. As a result, inside of the kiln was a lack of oxygen, which prohibited quick burning. Only the volatile wood gases burned, which resulted the wood to become transformed into pure carbon. By the introduction of fossil coal after 1850 AD, charcoal burning was strongly declining; in the 20th century the profession became extinct (cf. Kortzfleisch 2008).

The main consumer of the charcoal from the Kemel Heath was the iron melt of Michelbach (10 km NNE of Kemel). It has been running since 1656. Charcoal has not been in use since 1856 (Stolz 2008, Geisthardt 1957). Maybe also the melt of Geroldstein in the Wisper valley (12 km SW of Kemel, worked from 1589 to 1634 AD) and the melt of Katzenelnbogen (14 km N of Kemel, worked from 1736 to 1840 AD; Geisthardt 1957, Herold 1974, Ehmke 2003) were consumers of the charcoal from the Kemel Heath. By 1677, the melt of Michelbach was forced to get its charcoal from the forests on the quartzite mountain range of the Taunus because of a severe lack of charcoal in its surroundings. In 1780, the iron melts of the Nassau-Idstein county employed 300–400 people only for the purpose of charcoal and wood transport (Geisthardt 1957: 169). The melt of Katzenelnbogen was temporarily shut down around 1810 because of the absence of charcoal (Herold 1974).

The few remaining forests were of great importance to the people. Harsh punishments were the consequences for the theft of wood, grass, green branches for cattle feed, leaf litter, charcoal, oak bark for tannery, and venison (Roedler 1910). The consequence of this overexploitation was in the neighbored Aar valley near Michelbach the formation and further development of more than 200 gully systems (Stolz & Grunert 2006, Stolz 2008). Similar situations are known from other parts of the Rhine-Main area (Moldenhauer et al. 2010, Semmel 1995, Bauer 1993).

Already in the beginning of the 18th century, landlords tried to draw a clear border between fields and forests by the enacting of special laws (Ehmke 2003, Kaltwasser 1991). Since 1815, the local Earls of Nassau (*Herzöge von Nassau*) started a reforestation campaign led by the forest scientist, Ludwig Hartig (Kuls 1951). In forest district of Bad Schwalbach (Kemel region) the forests increased from between 1816 and 1866 of 1670ha (Kaltwasser 1991). By 1926, only a few remains of the heath existed (Roedler 1926).

So far, there are no pollen data from the Kemel heath, but from the Usa Valley (40 km in the NE) and from the Lower Westerwald Mts. (40 km in the NW). Schmenkel (2001) evidenced in the Usa Valley a first significant increase of non-tree pollen during the iron age but sinking back in the Roman Period. However, cereal pollen could be first evidenced since Early Middle Ages. The largest proportion of non-tree pollen is proven for the High Middle Ages. Hildebrandt et al. (2001) confirm in the Westerwald a low level of beech pollen *(Fagus sylvatica)* and a moderate rise of grass pollen (*Poaceae)* during the High Middle Ages. This was followed by a reforestation phase during the Late Medieval destruction period from about 1320 AD, which affected especially the uplands in Central Europe (Hildebrandt 2004; Abel 1976). While this time the values of beech pollen in the Westerwald are more than tripled.

Concerning the intensity of soil erosion, Becker (2011) assumed a total rate of soil erosion since the Roman period of almost 1m, proven by the depth of an investigated Limes-ditch in the village of Kemel. Furthermore, three charcoal particles from the filling of the ditch were dated by ^{14}C to a period between 3rd and 6th century AD as indication for human activities and fire events.

In the neighboring upper course of the Aar valley, there is, however, no evidence of prehistoric or Roman floodplain deposits, although the Limes crosses the Aar valley in this area. First sedimentation could not have been proven until 1000 AD. In the lower course of the Aar the overbank fines are of earliest Bronze age, but the main part was deposited not until Early Modern period (Stolz 2011a; Stolz & Grunert 2008).

3 Results from other mountain areas

Increased soil erosion in Central Europe is primarily triggered by anthropogenic land-use or by climate (cf. Dikau et al. 2005; Bork et al. 1998). First anthropogenic triggered erosion events are known for the Neolithic Period in the early settled parts of Germany (Lang 2003; Dreibrodt 2010). However, the sedimentation of Holocene colluvia started during quite different periods, just like the temporal peaks of soil erosion and redeposit are varied (cf. Leopold & Völkel 2007, Wunderlich 2000, Dreibrodt & Bork 2005, Dotterweich et al. 2003, Bork et al. 1998, Semmel 1993, Bibus 1989). Including the results of Holocene German river activity, summarized by Hoffmann et al. (2008), the sediment fluxes until 2250 BC are mainly coupled to climate. Since a geomorphologic activity phase 1320–820 BC, the influence cannot clearly be related to climate but rather to anthropogenic influence.

In contrast, Mäckel et al. (2009) describe a very early beginning of anthropogenic influences on the landscape in low mountain ranges of the Central Black Forest and the Kaiserstuhl Mt. (Southwestern Germany; 250 km S of Kemel). By sedimentological investigations and pollen analyses, an anthropogenic influenced sedimentation of loamy river sediments and slope colluvia could be proven since Neolithic, even for the river valleys of the Black Forest. The highest sedimentation values in these valleys were detected during Iron Age and Late Middle Ages. These results become confirmed by Rösch & Tserendorj (2011) who detected a shrunken forest cover to less than 70% in the Northern Black Forest Mts. in the Iron Age. During the Roman period and the following Migration Period the forest cover rises again.

For the High Middle Ages Wolters (2007) described a clear rising of *Poaceae* pollen and a moderate shrinking of arboreal pollen for two spring mires near Johanniskreuz in the Palatinate Forest (95 km SSW of Kemel), a young settled region of SW-Germany in the Bunter Sandstone. Bork et al. (1998) assume that the biggest proportion of forest distribution in Germany within the last 1000 years is during this period. Mäckel et al. (2009) detected a gap in sedimentation within profiles of the Black Forest at the transition between Early and High Middle Ages followed by strong sedimentation of particular alluvial sediments from the Upper Rhine Rift to the watershed of the Black Forest.

However, very little is known about the real proportion of deforested areas and land use intensity during different periods in Central Europe. In many cases, there are indices for a stronger utilization of woods and forests during several historical periods (cf. Ludemann & Nelle 2002, Küster 2008). Another method reconstructing former forests is the anthracologic analysis of former kiln sites. Due to their investigations of charcoal samples from the Palatinate Forest Hildebrandt et al. (2007) described a strong overexploitation of the forests as consequence of charcoal burning and harvesting especially during the 18th century. In the central Black Forest, Ludemann (2008) indicated a main period of charcoal burning in the 16th and 17th century. Similar results are known from the Harz Mts. (northern Germany; Hillebrecht 1982).

Evidences indicating former cropping are field balks and clearance cairns known from different European mountain areas. In most cases, the age of these relicts is quite unknown. For some examples a formation during High Middle Ages is assumed (cf. Born 1961; Scharlau 1961).

4 Materials and methods

Many of the studied sites pits had to be dug with the help of an excavator. Several profiles were investigated according to the rules set by the German *Bodenkundliche Kartieranleitung* (Ad-hoc AG Boden 2005) and International Union of Soil Sciences (2006). Laboratory analyses were conducted according to Blume (2000). The parameters analyzed were grain size, pH, carbonate content, organic matter and heavy mineral content. The determination of heavy minerals was made by M. Guddat-Seipel, Bad Nauheim.

The presence of Laacher See tephra in the field was proven by rapid testing, which is a method that was employed by Sauer & Felix-Henningsen (2006): bringing the sample into contact with filter paper impregnated with a 0.1% solution of phenolphtaleine in ethanol and a 5% aqueous NaF solution.

Dateable fragments of charcoal were separated from colluvial sediments by an archaeobotanical elutriation procedure of five liters sediment per sample (c.f. Jacomet & Kreuz 1999). These samples were not taken in regular intervals, but rather with regard to genetic layers and soil horizons.

Another possibility to reconstruct former land-use is offered by the analysis of charcoal kiln sites. By this method it is possible to indicate the used and, in consequence, the availability of wood species in the surroundings of a kiln. The charcoals from two different kiln sites were taken by sieve with a mesh size of 10 mm. To avoid any contamination while sampling, the individual layers of charcoal containing soil sediment were removed in thin layers by a small spade and a spatula (Fig. 8). Because of the low sediment-thickness on the investigated kiln sites, the samples of at least 100 charcoal fragments were taken in only two different depths of the *Stübbewall* (Fig. 7). A further sample was taken on the surface in the center of the kiln site. Thereafter, the charcoals were identified under a reflected light microscope with a magnification range of 100 to 400x (Schweingruber 1990; cf. Hillebrecht 1982, Manske 1997, Hildebrandt et al. 2001, Ludemann & Nelle 2002, Hildebrandt et al. 2007, Kortzfleisch 2008, Ludemann 2008). For identification, it is necessary to look at the charcoal in radial, longitudinal and tangential sections. The relevant characteristics of wood anatomy are the distribution and the size of the pores, the vascular rays the presence of spiral thickenings and the pits inside of the pores. In most cases, it is only possible to identify the genus and not the precise tree species (Nelle & Schmidgall 2003). After the identification procedure, the fragments of every genus were counted to identify the number of units (the charcoals were not weighted; therefore, G/N values for the individual samples could not be calculated; cf. Nelle & Schmidgall 2003). Other parameters like the former diameter of the wood were not measured. Some of the charcoal fragments or woods were chosen for radiocarbon-dating at the Radiocarbon Laboratories of Erlangen University (Germany), Poznan University (Poland) and Beta Analytics (USA). With regard to these results, it should be noted that there are possible error sources. Fundamentally, charcoal fragments can be much older as the time of sediment production. Thus, it has to be considered that ^{14}C-datings give only minimum ages (*terminus post quem*). Furthermore, disturbances and a vertical displacement by past land-use or bioturbation are possible. However, the deeper the sediment is taken beneath the surface, the possibility of this error becomes smaller.

Additionally, one OSL (Optical Stimulated Luminescence) sample was dated at the Department of Geography of the Humboldt-University of Berlin. Within the interpretations of the results, it must be observed that ^{14}C and OSL ages are not exactly equivalent with historic data but rather only a statistical probability (cf. Geyh 2008).

5 Results

5.1 Former field balks in forests

Field balks on slopes and accumulations consisting of gathered stones and colluvium on flat ground (in some parts of Germany such small landforms are called *Ackerberge*; Fig. 4) occur frequently in the forests of the Kemel Heath. At three different locations, we opened pits by an excavator to a depth of 300 cm.

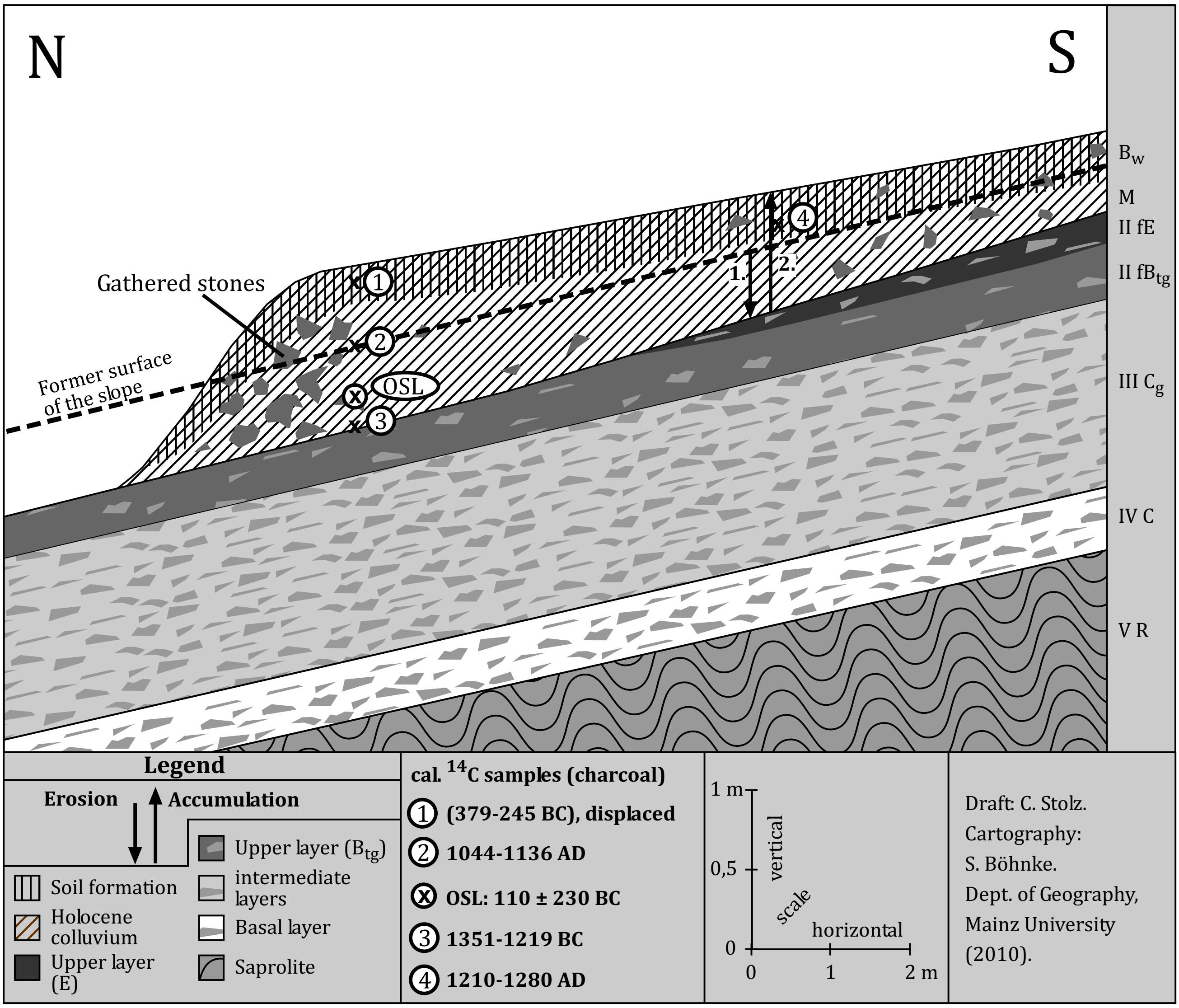

Fig. 2: Longitudinal section through the investigated field balk in the Pfaffenwald forest.

Abb. 2: Längsprofil durch den untersuchten Ackerrain im Pfaffenwald.

5.1.1 Deserted fields in Pfaffenwald forest near the village of Zorn (N 50°10'8.61", E 7°55'34.3")

In the Pfaffenwald forest (Fig. 5), there are two well visible, as well as some indistinct field balks, forming long-striped field terraces on a north-exposed, 4–9° inclined slope. The recent forest consists of tall, 160 year old beeches.

In one of the balks, a pit was dug and another one 10 m upwards on the field terrace, in which a stone cluster (slates and quartzites) became visible within the balk (Fig. 2). Down to a depth of 100 cm, a Holocene colluvium has been accumulated by former agriculture. Its texture is very homogenous and clayey, because of its origin as an eroded soil (Tab. 1). The slightly brighter color of the uppermost 30 cm layer indicates an initial soil formation, while the texture and other parameters are inconspicuous. Underneath, the clay content rises and the soil aggregates are covered by well visible clay films indicating an eroded fossil Bt horizon, 33 cm thick (Tab. 1). The typically high contents of the heavy minerals augite, brown hornblende and titanite indicate the pumice of the event of Lake Laach. Thus, this layer represents the upper layer, which was active latest during Younger Dryas (cf. Semmel 2002; Tab. 2). The Holocene colluvium also contains the Lake Laach heavy minerals because of its origin from the eroded upper layer upslope.

Below the upper layer, two individual intermediate layers are deposited, which have a high content of skeleton (stones) and loess-like sediments. The typical heavy minerals of loess-like garnet and green hornblende are detectable (cf. Semmel 2002). The basal layer below only consists of lo-

Tab. 1: Sedimentological data of the field balk profile in the Pfaffenwald forest.

Tab. 1: Sedimentologische Daten zum Profil innerhalb des Ackerrains im Pfaffenwald.

Horizon/layer	Depth	gS	mS	fS	ffS	gU	mU	fU	T	Skeleton content	pH	Loss on ignition	Color	Charcoal
	cm	%	%	%	%	%	%	%	%	%		%	Munsell	mg/L
1 Bw/M (colluvium)	10-30	11,45	6,37	4,43	3,63	18,05	17,46	12,18	26,42	26,73	3,97	3,75	2,5Y-4/4	4,53
1 M (colluvium)	30-83	6,86	6,99	4,97	3,79	19,28	18,53	12,22	27,36	34,28	3,97	2,72	2,5Y-4/4	463,57
1 M (colluvium)	83-100	7,00	7,21	5,32	3,98	18,47	17,78	11,56	28,67	25,52	3,91	2,77	5Y/R-4/6	42,87
2 fBtg (upper layer)	100-125	3,80	8,43	6,12	4,07	17,94	16,55	11,59	31,51	9,89	3,88	2,86	5Y/R-4/6	23,40
1 fBtg (upper layer)	125-133	6,86	6,79	3,45	3,19	21,07	20,08	11,77	26,79	18,84	3,85	2,39	10YR-4/4	n.a.
3 Cg (interm. layer)	133-175	20,87	11,30	3,84	4,74	14,12	11,97	10,27	22,89	55,08	3,85	2,62	10YR-5/6	0,00
3 Cg (interm. layer)	175-218	16,67	11,08	4,79	7,61	13,31	12,62	12,79	21,14	51,32	3,79	2,79	10YR-5/6	n.a.
4 C (basal layer)	218-244	23,48	14,85	5,05	4,81	10,32	10,87	12,92	17,71	42,14	3,79	2,99	10YR-6/6	n.a.
5 R (weathered slates)	244-280	22,17	14,63	5,18	5,52	11,73	11,56	13,81	15,39	62,61	4,04	2,63	10YR-6/4	n.a.
M = Holocene colluvium, gS = course sand, mS = middle sand, fS = fine sand, ffS = finest sand, gU = course silt, mU = middle silt, fU = fine silt, T = clay														

Tab. 2: Heavy mineral content of the field balk profile in the Pfaffenwald forest. (M = Holocene colluvium, UL = upper layer, IL = intermediate layer, BL = basal layer).

Tab. 2: Schwermineralgehalt des Ackerrain-Profils im Pfaffenwald (M = Kolluvium, UL = Hauptlage, IL = Mittellage, BL = Basislage).

Horizon/layer	Typical for	Bw/M	M	II fBtg, UL	II fBtg, UL	III Cg, IL	III Cg, IL
Depth [cm]	(Semmel 2002)	30-83	83-100	100-125	125-133	133-175	175-218
Augite	*pumice*	46	60	59	84	18	1
Epidote/zoisite		3	0	0	0	0	0
Garnet	*loess*	0	1	0	0	1	0
Green hornblende	*loess*	1	0	0	0	0	0
Brown hornblende	*pumice*	127	145	155	121	26	2
Titanite	*pumice*	42	34	32	25	2	0
Zircon		15	12	14	10	5	0
SUM		237	252	260	240	57	7

M = Holocene colluvium, UL = upper layer, IL = intermediate layer, BL = basal layer; analysis: M. Guddat-Seipel, Bad Nauheim.

cal debris. At a depth of 244 cm, the weathered Devonian bedrock is reached.

The colluvium was dated by three charcoals (^{14}C) and one OSL sample to the following ages (Tab. 5): 20 cm deep (cal. 379–245 BC, La Tène Period, Poz-36328), 53 cm depth (cal. 1044–1136 AD, High Middle Ages, Poz-36337), 83 cm (OSL: 110±230 BC, La Tène Period, HUB-0095) and 112.5 cm (cal. 1351–1219 BC, Bronze Age, Poz-36338).

A piece of charcoal from the field terrace above the balk was dated to: 28–50 cm, cal. 1210–1280 AD (Beta-294174; Fig. 2).

5.1.2 Deserted fields in the Ohren forest near the village of Niedermeilingen (N 50.1742°, E 7.9458°)

Likewise, in the Ohren forest traces of former agriculture were found. The location is relatively isolated at the rim of an old etchplain. There are several elongated low earthworks (*Ackerberge*, Fig. 4) with clearance cairns partly corresponding with each other by the right angle. They are formed by soil material fallen out during the turning of the plough at this place and also formed by clearance cairns.

A pit of 3m depth, dug by an excavator, revealed a structure consisting of gathered stones and a 71 cm thick layer of tarnished colored loess-like Holocene colluvium with an initial soil formation in the uppermost 30 cm (Fig. 3). Underneath follows the 29 cm thick remain of the upper layer with strong stagnic conditions and iron stains. The Btg horizon of a luvisol with visible clay films on the soil aggregates has been formed in both the upper and the intermediate layer (clay content 27–29 %). The colluvium consists of former material of the eroded upper layer. Both layers are containing typical heavy minerals of Lake Laach pumice (Tab. 4). In contrast, the only 25 cm thick intermediate layer below is nearly free of these minerals. The sandy basal layer (48 cm thick) is poor in skeleton (stone content) due to the bedrock of strongly weathered slates.

The colluvium only contains charcoal fragments (Tab. 3) which were dated at two different depths: 7–31 cm (cal. 919–999 AD, transition from Early to High Middle Ages, *Quercus spec.*, Poz-36339) and 50–71 cm (cal. 7–79 AD, early Roman Period, *Quercus spec.,* Poz-36340). Thus, the results are similar to those of Pfaffenwald forest (chapter 3.1).

5.1.3 Deserted fields in the Struth forest near the village of Kettenbach (N 50.257°, E 8.065°)

Eighteen km away from the two previously presented locations near Zorn, we investigated another wooded area at the rim of the Kemel Heath with deserted fields near the villages of Kettenbach and Hausen über Aar. There are two well visible field balks running parallel on a slightly inclined upper slope (3–9°; in western exposition; 265 m a.s.l.). A pit in the lowermost one revealed a 60 cm thick Holocene colluvium with an initial soil formation and charcoal content (Fig. 6). A covered luvisol had developed underneath. Its E-horizon located in the upper layer has been shortened by erosion. The Btg-horizon has been generated within the intermediate layer (clay content 33%). The whole profile is rich in loess and poor in skeleton (0–14%). Partly, the skeleton content consists of Oligocene-Miocene gravel (*Arenberger Fazies*; Müller 1973), which occurs on the adjacent plateau above.

A charcoal fragment taken from a depth of 45 cm, which was above the lowermost third of the colluvium, was dated at cal. 361–272 BC (La Tène Period; Erl-7539).

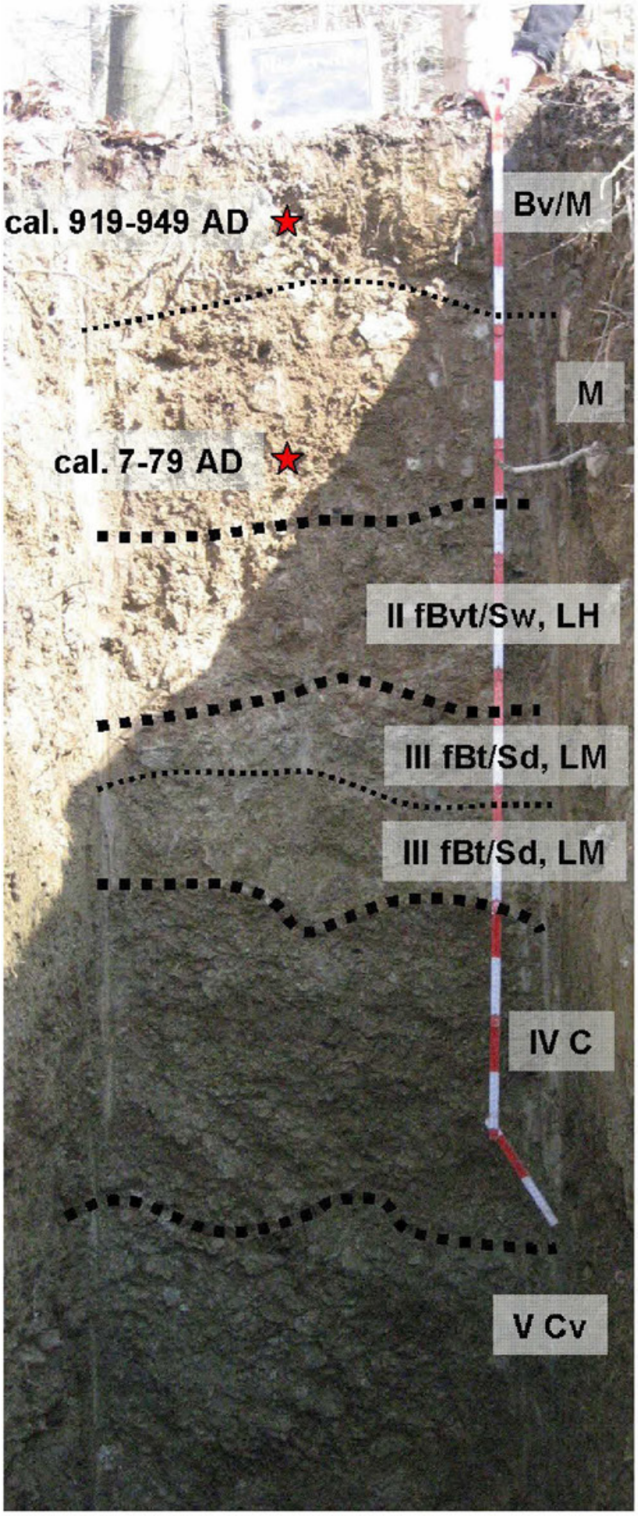

Fig. 3: The profile of Ohren forest with datings.

Abb. 3: Das Profil in der Waldabteilung Ohren mit Datierungen.

Table 3: Sedimentological data of the field balk-profile in Ohren forest.

Tab. 3: Sedimentologische Daten zum Profil innerhalb des Ackerberges in der Waldabteilung Ohren.

Horizon/layer	Depth	gS	mS	fS	ffS	gU	mU	fU	T	Skeleton content	pH	Loss on ignition	Colour	Charcoal
	cm	%	%	%	%	%	%	%	%	%		%	Munsell	mg/L
1 Bw/M (colluvium)	10-30	7,56	5,58	4,81	3,43	20,17	19,34	12,47	26,65	26,76	3,94	4,60	10YR-3/24	255,03
1 M (colluvium)	30-83	6,73	6,29	4,85	3,67	20,96	18,21	12,34	26,96	18,46	3,97	3,04	10YR-5/6	369,93
2 fBwtg (upper layer)	83-100	10,46	5,76	4,36	3,48	18,14	19,28	12,70	25,82	24,12	3,94	3,23	10YR-5/8	0,00
3 fBtg (interm. layer)	100-125	12,66	5,49	1,88	2,60	22,05	17,98	10,30	27,04	32,18	3,76	2,79	10YR-6/4	n.a.
3 fBtg (interm. layer)	125-133	7,79	5,21	1,86	2,71	26,11	17,26	9,58	29,47	15,55	3,77	2,56	10YR-5/6	0,00
4 C (basal layer)	133-175	19,41	9,02	4,14	8,24	16,76	13,44	11,75	17,25	2,30	3,66	2,13	10YR-4/6	n.a.
5 R (weathered slates)	175-218	37,04	16,21	5,35	6,38	11,01	6,44	4,69	12,89	71,97	3,81	2,64	10YR-4/4	n.a.
5 R (weathered slates)	218-244	37,11	18,00	5,69	6,97	9,19	7,06	5,31	10,69	72,21	3,70	3,16	10YR-2/1	n.a.

M = Holocene colluvium, gS = course sand, mS = middle sand, fS = fine sand, ffS = finest sand, gU = course silt, mU = middle silt, fU = fine silt, T = clay

Horizon/layer	Typical for	M	M	II fBwtg UL	III fBtg IL
Depth (cm)		31-50	50-71	71-100	100-115
Augite	pumice	77	88	38	4
Epidote/zoisite		0	1	4	12
Garnet	loess	0	0	1	2
Green hornblende	loess	0	1	2	4
Brown hornblende	pumice	130	159	75	8
Titanite	pumice	49	50	16	1
Zircon		12	18	8	4
SUM		268	317	144	35

M = Holocene colluvium, UL = upper layer, IL = intermediate layer, BL = basal layer; analysis: M. Guddat-Seipel, Bad Nauheim.

Tab. 4: Heavy mineral content of the field balk profile in Ohren forest. (M = Holocene colluvium, UL = upper layer, IL = intermediate layer, BL = basal layer).

Tab. 4: Schwermineralgehalt des Ackerrain-Profils im Pfaffenwald (M = Kolluvium, UL = Hauptlage, IL = Mittellage, BL = Basislage).

5.2 Charcoal kiln sites

For the detailed investigation of kiln sites the complete forested area of Zorn was mapped (40 kiln sites; 0.11 sites/ha).

To investigate the influences of historical charcoal burning, we chose two different kiln sites, one in the Pfaffenwald and another one in the nearby Reißbrühl forest. From each of them, we took 83-130 pieces of charcoal by sieving top-down at different depths below the plane of the kiln (*Meilerplatte*; see Fig. 7) and from the bordering rim (*Stübbewall*).

The determination of tree species resulted only three different types (*Fagus sylvatica*, *Quercus spec.* and *Betula pendula*; Fig. 9) but in several compositions. Five charcoals were radiocarbon dated.

5.2.1 Dating and determination of tree species

The first investigated kiln site (N 50.167°, E 7.919°; 430 m a.s.l.) is actually located on the low inclined, NNW exposed slope of a small valley in a nearly pure, old beech forest (*Galio odorati Fagetum*; cf. Ellenberg 1996, Ludemann & Nelle 2002). The investigated one belongs to a group of 4 kiln sites in an area 130 m wide. Due to the hillside location of the kiln, it is plausible that the used wood originates from the forested upper slope. It is plausible that the origination area of the wood is wider on the slope above the kiln, because it was easier to carry the wood downslope (cf. Hildebrandt et al. 2007). After 100 m, the slope is bounded by the edge to the open fields. It is furthermore noticeable in this forest that there are several former field balks around the kiln sites. These belong to the investigated former farmland of the Pfaffenwald forest. The investigated kiln site is located exactly on one of these former field terraces.

The spectra of species at different depths of the kiln site-sediment are very homogenous and show nearly the same result as today (90% beech and 10% oak). The 3 datings are

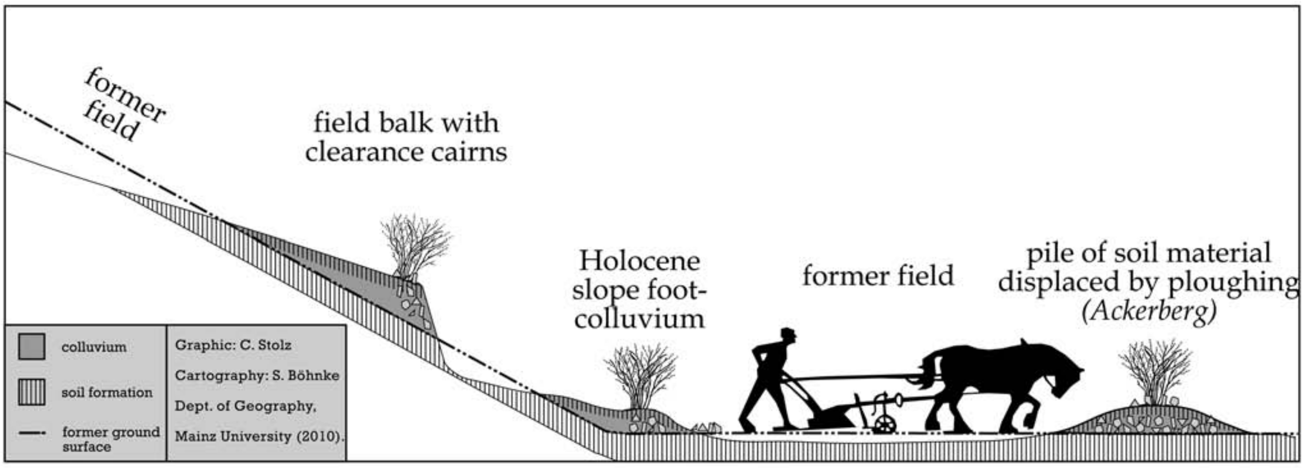

Fig. 4: Model of former field balks on a slope and an earthwork of soil material and gathered stones (Ackerberg) on a flat location under forest.
Abb. 4: Modell eines ehemaligen Ackerrains und eines Ackerberges, bestehend aus Bodenmaterial und Lesesteinen, in Hanglage.

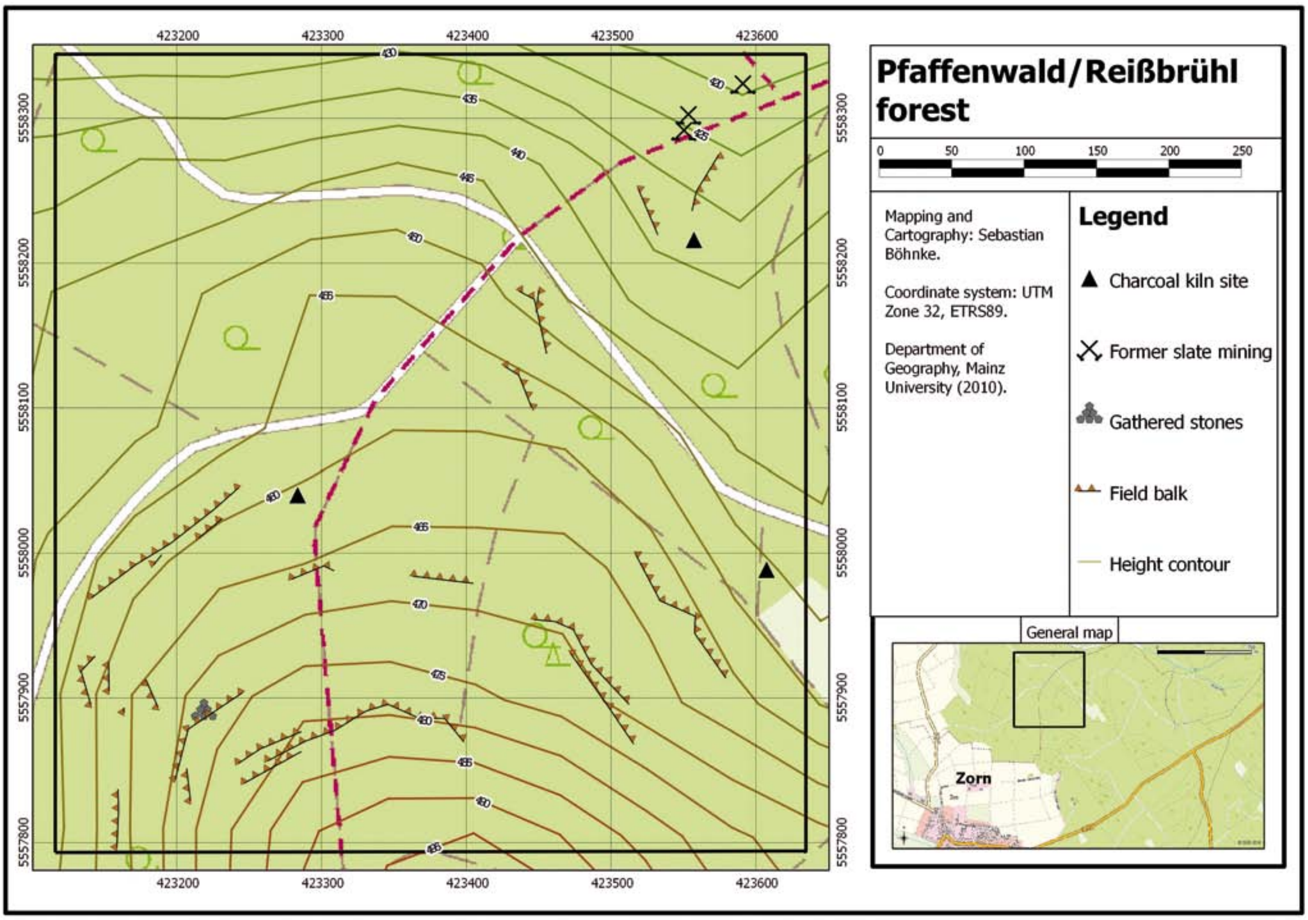

Fig. 5: Map section of the Zorn district with former field balks and charcoal kiln sites in forest.
Abb. 5: Kartenausschnitt der Gemarkung Zorn mit Ackerrainen und Meilerplätzen unter Wald.

indicative of the Early Modern Period: cal. 1712–1906 AD (*Fagus sylvatica*, Poz-36326), cal. 1654–1794 AD (*Quercus spec*., Poz-36322) and cal. 1476–1604 AD (*Fagus sylvatica*, Poz-36321). However, the dated samples were not layered stratigraphic, because the sediment cover was obviously mixed between the individual burning sessions of the kiln or beyond. Thus, a detailed reporting about the forest history in the surrounding of the kiln site is only possible to a limited extend. However, it could be demonstrated that the kiln was used from the 16th to the 19th century.

The second one (N 50.169°, E 7.930°; 427 m a.s.l.) is located on a moderate inclined, NNE exposed lower slope within a small, wet spring-depression, covered by a forest with up to 160 year old beeches (result of a dendrochronological count) and some oaks. In the surrounding area of 450 m, there is no further evidence of kiln sites. The forested slope above the kiln is with a distance of 500 m to the top of the hill much larger. Eventually, the origin area of the used wood could have been much larger, too.

The spectra of species in the sampling depths of 0–13 cm

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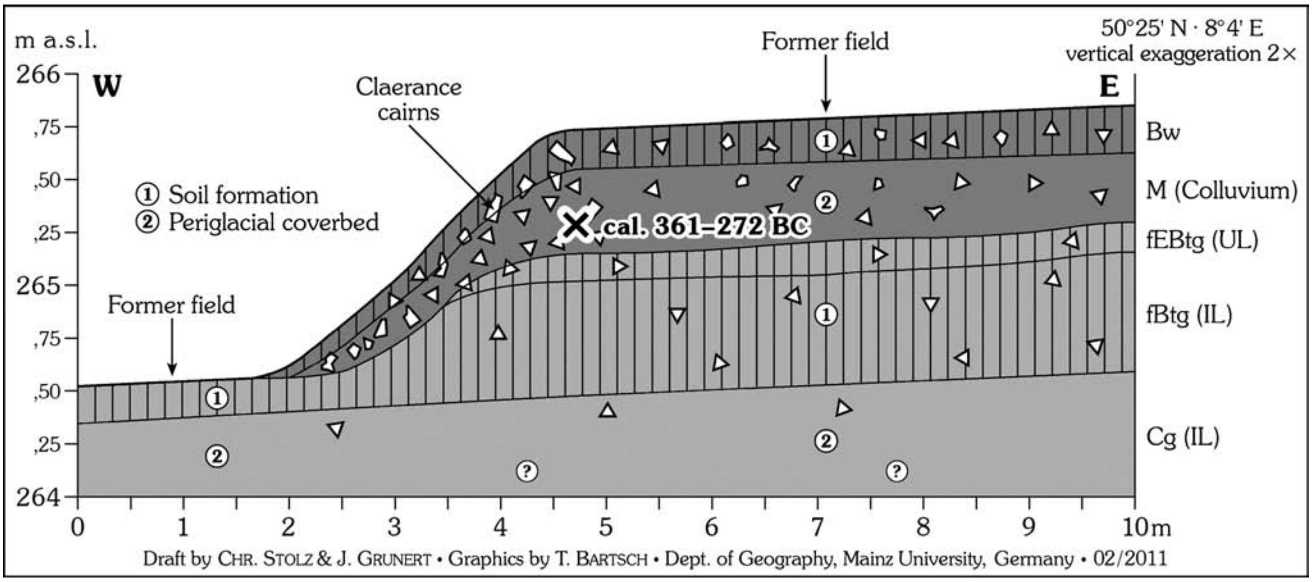

Fig. 6: Field balk in Struth forest, probably generated during Iron Age and later. For the parts of the profile marked with question marks there is no detailed information.

Abb. 6: Ein Ackerrain in der Waldabteilung Struth. Er geht vermutlich auf die Eisenzeit zurück. Für die mit Fragezeichen gekennzeichneten Profilbereiche liegen keine detaillierten Informationen vor.

and 13–25 cm of the *Stübbewall* were different (Fig. 8). The lower spectrum is dominated by beech (82%), followed by birch (13%, a tree which needs an exposure to the sunshine) and oak (5%). In the uppermost spectrum, beech is only 34% and birch is found in higher concentration (19%). Oak is the predominant species (47%) in this layer. Although the kiln site is located directly above the spring depression, there could not be proven any hydrophilic species like willows or alders. This indicates an origin of the used wood exclusively upslope of the kiln.

Dating of charcoal samples from the two presumably different layers gave exactly the same ages: cal. 1677–1921 AD (*Fagus sylvatica*, Poz-36323 and 36324). It is possible that the kiln was used for only a short period at the end of the 18th and beginning of the 19th century. However, the spectra of species in the different layers are quite different. Therefore, the lower charcoal sample could have been displaced by bioturbation or similar processes in the past.

5.3 The medieval refuge castle of Zorn with a neighboring slag heap

At the eastern rim of the castle (81° E), where it has been damaged by a modern stairway, we took a small portion of soil material from the lower part of the rampart (the permission from the local preservation authority was given). By the archaeobotanical elutriation procedure, a fragment of charcoal was eliminated, which was dated to cal. 900–970 AD (Poz-36343).

A charcoal sample of the slag heap, eliminated in the same way, was dated to cal. 1103–1203 AD (Poz-36341; High Middle Ages). An analysis of the slag by x-ray diffractometer resulted in residual iron contents of 36–49% and silicate contents of 9–25%.

A piece of charcoal of the other slag heap, which is located beside the small stream Rödelbach near Zorn (N 50.15829°, E 7.93905°) was dated to the similar age of cal. 1160–1260 AD (Beta-294172).

5.4 Calculation of sedimentation rates of a small alluvial fan near Kemel (N 50.1650°, E 8.0147°)

To calculate the local sedimentation rate from a single location on the Kemel Heath, we investigated an archaeological site, which included a Roman water well, stabilized by wood beams. It was located close to the former Roman castle of Kemel and, geomorphologically, on a small and flat alluvial fan, respectively a colluvial depression filling in the non-perennial upper course of the Aulbach and Wisper stream. The side walls of the well were supported by several sediment covered oak-wooden beams, which were dated dendrochronologically to 215 AD (information given by the Hessian Office of Monument Preservation).

The site was dug 384 cm into a sandy-silty, uppermost clayey, well-layered colluvial/alluvial sediment with a distinct content of skeleton (5–29%). Downwards 316 cm inside the well, the groundwater level was detected and the sediment is grey-reduced. Around 179 cm deep, it contains charcoal fragments; underneath, there are no organic remains. A charcoal fragment of 0–35 cm depth was dated to cal. 1160-1260 AD (High Middle Ages; Beta-294173), another piece of 35–55 cm to cal. 1080–1124 AD (High Middle Ages; Erl-8905) and a further piece of 140–163 cm to cal. 678–773 AD (Early Middle Ages; Erl-8906). Below this sample the top of the cover of the water-well was detected by archaeologists.

The soil samples downwards to 383 cm were analyzed concerning their content of heavy minerals. All samples contain the minerals augite, brown hornblende, titanite, green hornblende and garnet thus giving evidence of distinct contents of loess and the pumice of Lake Laach eruption of the Allerød Interstadial (cf. Stolz & Grunert 2006, Semmel 2002).

Summarized, the lower part of the profile (152–384 cm and deeper) had been already deposited when the Romans built the well in 215 AD. This can be assumed because of the lack of finds and charcoals in these sediments.

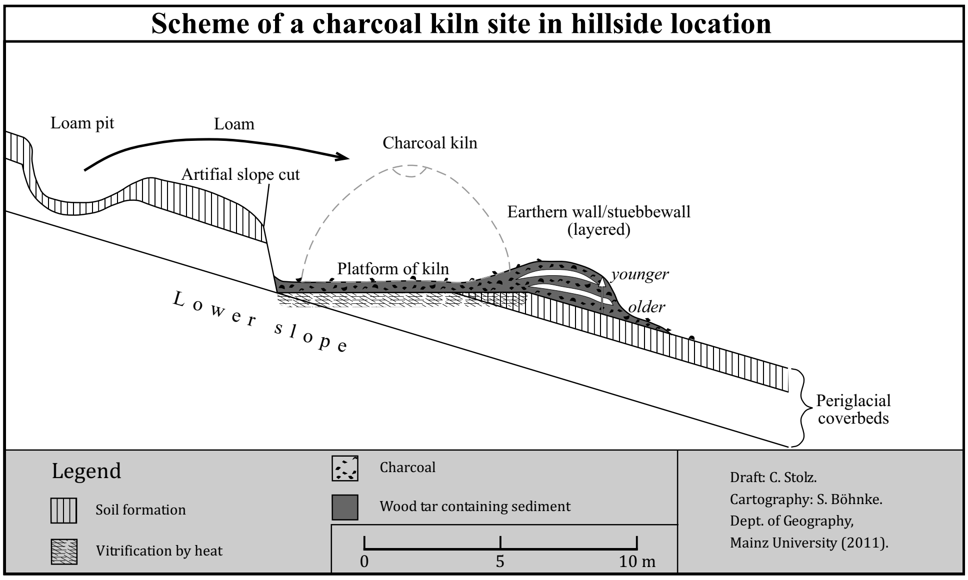

Fig. 7: Scheme of a charcoal kiln site in hillside location with loam pit.

Abb. 7: Schema eines Hangmeilerplatzes mit Lehmgrube zur Entnahme der Stübbe, mit der der Meiler verkleidet wird.

According to the ^{14}C-datings, it is possible to calculate approximate sedimentation rates for different periods. Therefore, it has to be considered that there are some uncertainties of this analysis concerning a subsequent displacement of charcoal samples: Between the 4th and 7th centuries AD, no traces of deposition have been found. After this period, the second part of the profile (152–45 cm depth) was deposited between approx. 725 and 1100 AD. This corresponds to a sedimentation rate of 2.8 mm/a (Fig. 10). The third part (45–20 cm depth) was deposited between approx. 1100 and 1210 (sedimentation rate of 2.3 mm/a). The upper part (20–0 cm depth) was deposited during the time after 1210 until today, corresponding with a much smaller rate of only 0.25 mm/a.

6 Discussion

The 16 datings of this study include 5 of Prehistoric and Roman periods (until 260 AD; cf. Baatz and Herrmann 2002), 1 of Early Middle Ages (approx. 400–1000 AD), 5 of High Middle Ages (approx. 1000–1320 AD) and 5 of Early Modern Age (approx. 1450–1850 AD; cf. Born 1989; Tab. 5). The presence of charcoal particles in the sediment archives concludes that in most cases, there is a certain amount of human influence on the landscape in the different periods. But it has to be considered that ^{14}C-datings give only minimum ages (*terminus post quem*). If a piece of charcoal is trapped within a sediment layer, the sediment must have been displaced after the formation time of the organic carbon of the charcoal.

6.1 Agriculture

This is proven by the uppermost sample (20 cm deep) within the field balk-profile in the Pfaffenwald forest, which was dated to an earlier time (cal. 379–245 BC, La Tène Period, Poz-36328) compared with the samples below. It could have been deposited for a long time (approx. 1300 years) on the soil surface upslope or within a colluvium when it was buried by the sediment. However, a vertical displacement by bioturbation could be also plausible. To clarify this, we additively used one OSL dating which confirmed the other

Fig. 8: Sampling of a charcoal kiln site in different layers.

Abb. 8: Die Beprobung eines Hangmeilerplatzes (Stübbewall) in verschiedenen Tiefen.

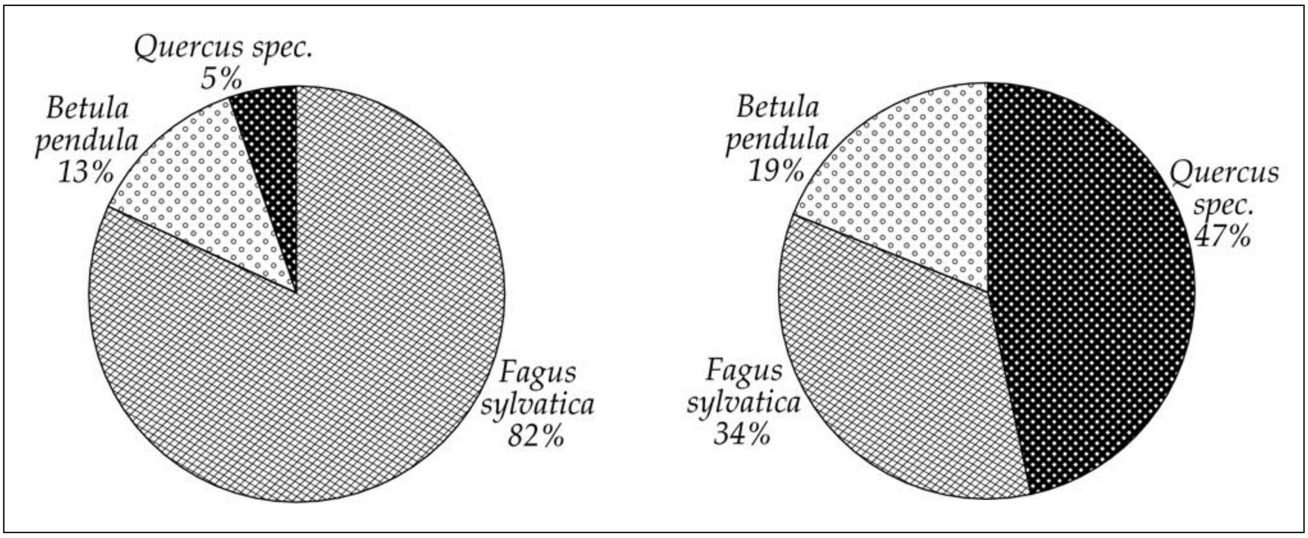

Fig. 9: Spectra of tree species at depths of 0–13 cm and 13–25 cm (kiln site 2).

Abb. 9: Baumartenspektrum in 0–13 und in 13–25 cm Tiefe (Meilerplatz 2).

[14]C-datings. Additionally, a piece of charcoal from the field terrace above the balk (28–50 cm deep) was dated to cal. 1210–1280 AD (Beta-294174). Unfortunately, archaeological findings were missing, which is typical for agrarian locations away from settlementsAs well as within the field balk-profiles in Pfaffenwald forest, in those of Ohren forest and Struth forest, we dated charcoals to prehistoric and Roman Periods. In consequence, at these locations the agriculture began before the arrival of the Romans (OSL: 340 BC-120 AD). Thus, the agriculture started during the Iron Age is proven for one location and it is likely so for the other two locations. In addition, based on the datings of Pfaffenwald and Ohren forests, it is also plausible that Roman agriculture existed for local supply of the border troops. Basically, we should know that fields were not so wide spread in that time as within Early Modern Periods or today. However, as non-favorable areas (North faced slopes) were cropped during that time and which are currently wooded. Thus, we also have to assume prehistoric agriculture in the actual cropped areas, too. Therefore, fields could have been more distributed than today. On the other hand, the forests must have been

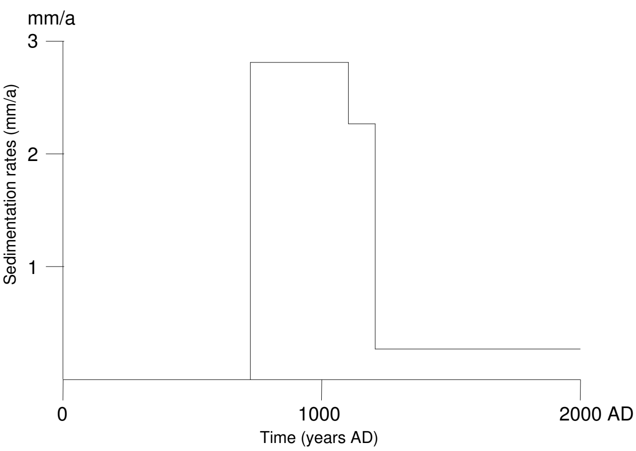

Fig. 10: Sedimentation rates in a small valley filling near Kemel.

Abb. 10: Unterschiedliche Sedimentationsraten im verfüllten Unterlauf eines Tälchens bei Kemel.

smaller. In contrast, prehistoric people might have preferred other locations than modern farmers, for example locations on plateaus or smooth slopes far away from recent settlements.

Within the field balks, dates from Early Middle Ages are missing. But the agriculture on these fields continued until the High Middle Ages. The covering of the terrace itself by colluvial sediments took place within the High Middle Ages. Maybe it was reactivated in this period. The presence of gathered stones within the colluvial sediments and the OSL age give a solid result. After the High Middle Ages, there is no further evidence of agriculture in the three investigated former field districts. Furthermore, due to the soil formation (luvisol) within the colluvium, a resting phase since High Middle Ages seems plausible (Fig. 2).

6.2 Charcoal burning

In the forests of the Zorn district (374ha), we found and mapped only 40 kiln sites (0.11 sites/ha). In a forest 15 km away near to the ironworks of Michelbach, we calculated 0.38–0.58 kiln sites per hectare (Stolz 2011b). In the mining region of the southern Harz Mts. von Kortzfleisch (2008) mapped 3.3 sites/ha. Therefore, charcoal burning was not so wide spread in the investigation area compared with other districts. However, in the cleared areas, charcoal burning was a frequent activity. Most of the mapped kiln sites have a diameter of 10–12 m. This indicates an origin within the period of massive charcoal burning of the 18th or 19th century in the Taunus Mts. (cf. Stolz 2011b, Hildebrandt et al. 2001). Basically, the investigated field balks in the Pfaffenwald forest must be older than the kiln sites, because kiln sites are located on the former field terraces using the leveled surface. By physical dating methods we could substantiate charcoal burning in the investigated area from the 16th century and increasing until the end of the 18th century.

The tree spectrum of the second investigated kiln site (Reißbrühl forest) taken from the lower layer of the charcoal containing sediment is dominated by *Quercus spec.* and *Betula pendula* (in total 66%). *Betula* is a typical pioneer

Tab. 5: ^{14}C and OSL datings of this study.

Tab. 5: Aufstellung der in der vorliegenden Studie enthaltnen ^{14}C- und OSL-Datierungen.

	No.	Type	Site	Depth [cm]	Material	^{14}C Age [a BP]	Calibration or OSL age [a BC/AD]
1	Poz-36328	^{14}C	Pfaffenwald forest	10-30	charcoal [field balk]	2260±35	379-245 BC
2	Poz-36337	^{14}C	Pfaffenwald forest	30-83	charcoal, hardwood [field balk]	940±30	1044-1136 AD
3	HUB-0095	OSL	Pfaffenwald forest	83	basal colluvium [field balk]	-	110±230 BC
4	Poz-36338	^{14}C	Pfaffenwald forest	100-125	charcoal, *Quercus* [field balk]	3015±35	1351-1219 BC
5	Beta-294174	^{14}C	Pfaffenwald forest	28-55	charcoal, hardwood [colluvium from terraced field]	790±30	1210-1280 AD
6	Poz-36339	^{14}C	Ohren forest	7-31	charcoal, *Quercus* [field balk]	1065±30	918-999 AD
7	Poz-36340	^{14}C	Ohren forest	50-71	charcoal, *Quercus* [field balk]	1950±35	7-79 AD
8	Erl-7539	^{14}C	Struth forest, Kettenbach	45	charcoal [field balk]	2205±56	361-272 BC
9	Poz-36321	^{14}C	Kiln site 1, Zorn	0-8	charcoal, *Fagus* [kiln site]	360±30	1476-1604 AD
10	Poz-36322	^{14}C	Kiln site 1, Zorn	8-17	charcoal, *Quercus* [kiln site]	220±30	1654-1794 AD
11	Poz-36326	^{14}C	Kiln site 1, Zorn	30-48	charcoal, *Fagus* [kiln site]	85±30	1712-1906 AD
12	Poz-36323	^{14}C	Kiln site 2	0-13	charcoal, *Fagus* [kiln site]	180±30	1677-1921 AD
13	Poz-36324	^{14}C	Kiln site 2	13-25	charcoal, *Fagus* [kiln site]	180±30	1677-1921 AD
14	Poz-36343	^{14}C	Refuge castle "Alte Schanz"	-	charcoal [artificial deposit]	1105±30	900-970 AD
15	Poz-36341	^{14}C	Slag heap near the refuge castle	0-20	charcoal, *Fagus* [slag heap]	865±30	1103-1203 AD
16	Beta-294172	^{14}C	Slag heap near the Rötelbach stream	0-15	charcoal, hardwood [slag heap]	850±30	1160-1260 AD
17	Erl-8905	^{14}C	Kemel, site with Roman water well	35-55	charcoal [colluvium]	953±37	1080-1125 AD
18	Beta-294173	^{14}C	Kemel, site with Roman water well	55-90	charcoal, half-ring porous hardwood [colluvium]	840±30	1160-1260 AD
19	Erl-8906	^{14}C	Kemel, site with Roman water well	140-163	charcoal [colluvium]	1274±44	678-773 AD

tree species (Ludemann & Nelle 2002). The species are an indication for a bright, cleared forest with a high content of *Quercus spec.*, which was an important kind of timber (Ellenberg et al. 1991). While degradation processes, *Quercus spec.* and *Carpinus betulus* can be favored (Ludemann 2008). In the layer above, the content of *Quercus* and *Betula* has fallen (18%) and *Fagus* is the dominating species; however, the content of birch is still 13%. Today, there grow only few birches. However, the so-called Little Ice Age (Glaser 2008) cannot had direct influences to the detected changes in vegetation, because oaks *(Quercus robur)* are, in contrast to the beeches, counted among the slightly more thermophile deciduous trees (Ellenberg et al. 1991), however with a wider ecological range.

An eventual wood selection by the charcoal burners should not be considered. Ludemann (2008) demonstrated that charcoal burners during the Early Modern Period in the Black Forest used all available species and all thicknesses of wood.

6.3 Refuge castle and iron smelting

The refuge castle of Zorn, the neighboring slag heap and a further slag heap in the Rödelbach valley originate almost simultaneously from the High Middle Ages (10th to 12th century AD). Based on these findings, we also have to assume a high wood consumption around these small melts. Furthermore, the relatively high ferrous content of the slags indicates a low yield of iron due to a still primitive technology of iron-smelting. Nearly the same age we detected for the use of the deserted fields. The slag heaps, which are 2 of 5 samples from within the Zorn district, the castle and the field terraces give evidence of a wide spread, decentralized, agriculture, iron production and forest clearing in the district of Zorn during the High Middle Ages (cf. Stolz & Grunert 2008, Geisthardt 1954).

The dating of a charcoal fragment from inside the rampart of the refuge castle has to be regarded as an approximate value for a possible real age of the refuge castle As the charcoal was located inside the rampart, it must be older or of the same age as the castle. In consequence, the castle must originate from the 10th century or from a younger period.

6.4 Sediment sequences of the alluvial fan of Kemel

Within the alluvial fan profile of Kemel we calculated the highest sedimentation rate for the High Middle Ages, too. Direct climatic influences as a triggering factor are unlikely; in the contrary, however the local influence of anthropogenic triggered soil erosion in the small catchment of the depression was high (Thiemeyer et al. 2005). It must be expected that the accumulation of alluvial sediments probably was discontinuous, which means that it could have probably happened during heavy precipitation events in a more or less cleared landscape. Thus, the calculated sedimentation rates have to be considered as limited reference values for the susceptibility of soil erosion within the catchment.

The only dated charcoal from the Early Middle Ages (7^{th}/8^{th} century) is taken from this alluvial fan which is an important fact. The charcoal has been deposited directly above the Roman well. Thus, soil erosion rates during the Migration Period must have been reduced. Otherwise, the finding of Early Medieval charcoal supports the opinion of Becker (2011) concerning an increased human influence in Kemel in this period. Most likely, during Early Modern times the distribution of fields did not reach the level of the High Middle Ages. Instead, heathlands became predominant in the landscape. Therefore, the sedimentation rate within the fan profile of Kemel was shrinking again.

7 Conclusions

The agriculture on the Kemel Heath probably started during the La Tène or Roman Period. This conforms to the results of regional palynologic investigations. Since this time – interrupted during the Migration Period – to the High Middle Ages agriculture was intensified and non-favorable areas were cultivated. The spreading of fields increased up to 13^{th} century and was shrinking again in favor of pastureland during the Early Modern Period. In the remaining cleared forests, charcoal burning was widespread, especially from 16^{th} to the end of the 18^{th} century. The consequences concerning forest degradation and changes in the composition of tree species could be evidenced due to the charcoals from two different kilns.

Of nearly the same age like the High Medieval agriculture are the slag relicts of decentralized iron melting and, probably, the refuge castle of Zorn next to the slags.

Although the sediment sequence of a small alluvial fan near Kemel is of very local character, we evidenced the highest deposition rate in its catchment for the High Middle Ages, too.

With this study, we evidenced the benefit of the application of several different methods to reconstruct former landscapes of different periods: Future studies have to investigate as much single relicts to get more representative information. Unfortunately, this was not possible due to limited financial resources.

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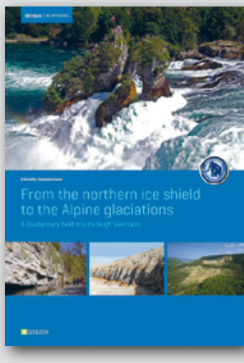

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