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COVER FIGURE

Cluster of moderately weathered, banded
argillite boulders (Fig. p. 48)

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Climato-stratigraphic subdivision of the Pleistocene in Schleswig-Holstein, Germany and adjoining areas

status and problems

Hans-Jürgen Stephan

Dedicated to Burchard Menke (1932–1993)

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Abstract:

In Schleswig-Holstein detailed petrographical and palynological studies were undertaken with samples from exposures and core drillings examined over the last four decades. Analyses of the gravel fraction and 'indicator rocks' of glacial deposits were used for stratigraphical interpretations and correlation. Glaciofluvial sediments were dated by TL and OSL. Combined with the stratigraphical information from organic deposits, a new climato-stratigraphic table of the Pleistocene for Schleswig-Holstein with a revised subdivision of the Middle and Late Pleistocene could be established. The pollen sequence of the oldest Pleistocene warm phase (Warmhörn-Thermomer) is published for the first time. The complete Early Pleistocene and lower Middle Pleistocene stratigraphical sequence is documented at Lieth and at Gorleben (Lower Saxony) in two continuous successions of organic beds developed during warm phases alternating with cold phase deposits. The uppermost part of the Lieth succession and the lowermost part of Gorleben overlap. The combined succession provides a unique reference for correlations through Europe.

No more than three cold stages involving glaciation are demonstrated to exist currently in Schleswig-Holstein, the Elsterian and the Saalian in the upper Middle Pleistocene, and the Weichselian in the Late Pleistocene. A possible pre-Elsterian glaciation is discussed. The Holsteinian is correlated with MIS 9e. The Saalian includes a lower part with non-glacial cold phases and warm phases, the 'Wacken-Warmzeit' (=Dömnitz) and the 'Leck-Warmzeit' correlated with MIS 7e and MIS 7c respectively, and a glacial upper part. During the Weichselian probably two phases of glaciation existed, the first in the early Middle Weichselian ('Ellund-Phase'; late MIS 4 or/and the early MIS 3), the second in the Upper Weichselian (MIS 2). The Weichselian glaciation of Schleswig-Holstein ends around 15 ka BP when huge stagnant and dead ice masses of the Young Baltic glacier advance ('Mecklenburg-Phase') melted.

Klimastratigraphische Gliederung des Pleistozäns in Schleswig-Holstein, Deutschland und angrenzenden Gebieten – Stand und Probleme

Kurzfassung:

In Schleswig-Holstein wurden in den zurückliegenden vier Jahrzehnten detaillierte Untersuchungen an Probenmaterial aus Aufschlüssen und Kernbohrungen durchgeführt. Analysen der Kiesfraktion und von 'Leitgeschieben' aus glazialen Ablagerungen wurden für die Interpretation und Korrelation genutzt. Glazifluviale Sedimente wurden TL- und OSL-datiert. Zusammen mit der stratigraphischen Information aus organischen Ablagerungen konnte eine neue klimostratigraphische Tabelle des Pleistozäns für Schleswig-Holstein mit revidierter Gliederung des Mittel- und Oberpleistozäns erstellt werden. Die Pollensequenz der ältesten pleistozänen Warmzeit (Warmhörn-Thermomer) wird erstmals veröffentlicht. Das Unterpleistozän (Altpleistozän) und das untere Mittelpleistozän sind bei Lieth und bei Gorleben (Niedersachsen) in zwei kontinuierlichen Schichtfolgen vollständig dokumentiert, abgelagert in einem Wechsel von warmen und kalten Klimaphasen. Der oberste Abschnitt der Abfolge von Lieth und der unterste der Abfolge von Gorleben überlappen. Die kombinierte Abfolge bietet ein einzigartiges Referenzprofil für Korrelationen innerhalb Europas.

Bisher wurden in Schleswig-Holstein nur drei Kaltzeiten mit Vergletscherungen nachgewiesen, Elster- und Saale-Kaltzeit im oberen Mittelpleistozän und die Weichsel-Kaltzeit im Oberpleistozän. Eine mögliche prä-elsterzeitliche Vergletscherung wird diskutiert. Die Holstein-Warmzeit wird mit MIS 9e korreliert. Das Saale umfasst einen unteren Abschnitt mit nicht-glaziären Kaltzeiten und mit Warmzeiten, der 'Wacken-Warmzeit' (=Dömnitz) und der 'Leck-Warmzeit', korreliert mit MIS 7e und MIS 7c, und einen glaziären oberen Abschnitt. Während des Weichsel gab es wahrscheinlich zwei Phasen mit Vergletscherung, die erste im frühen Mittelweichsel ('Ellund-Phase', spätes MIS 4 oder/und frühes MIS 3), die zweite im oberen Weichsel (MIS 2). Die weichselzeitliche Vergletscherung Schleswig-Holsteins endet um 15 ka BP, als großflächige stagnierende Eisflächen und Toteismassen des Jungbaltischen Gletschervorstoßes ('Mecklenburg-Phase') schmolzen.

Keywords:

Pleistocene, climato-stratigraphic table, correlations, Schleswig-Holstein

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1 Introduction

The paper outlines the evidence for the Pleistocene of Schleswig-Holstein, Germany. It presents a stratigraphic scheme that is based on a correlation with the Quater-

nary stratigraphy of the Netherlands and the Lower Rhine area and has been revised as a consequence of recent OSL and $^{230}\text{U}/\text{Th}$ dates of deposits from Schleswig-Holstein and Lower Saxony. The oldest part of the Pleistocene sequence in Schleswig-Holstein is the 'Lieth-Serie' Beds (MENKE 1970,

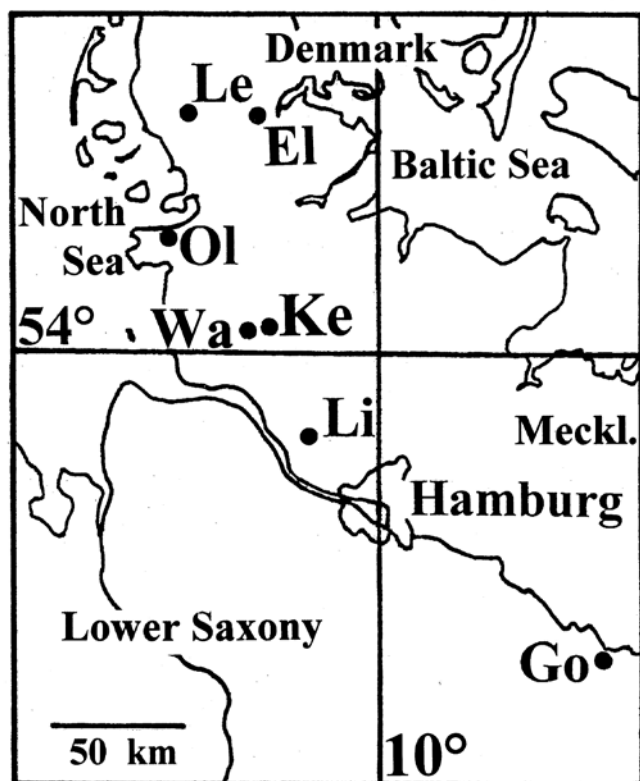


Fig. 1: Location map. El = Ellund, Go = Gorleben, Ke = Keller, Le = Leck, Li = Lieth, Ol = Oldenswort, Wa = Wacken.

Abb. 1: Lagekarte.

1975) (Fig. 2, 3). In the neighbourhood (northeastern Lower Saxony), according to palynological investigations on cores from the top of the Gorleben salt dome (Fig. 1) by MÜLLER (1986, 1992), the upper part of the 'Lieth-Serie' (Fig. 3: 'Dannesch-Komplex') is found again, however, there as the lowest part of the very long sequence of Gorleben. The 'Dannesch-Komplex' is followed by the pre-Elsterian middle part of the Gorleben succession, here named 'Gorleben-Complex'. Evidence is presented for Elsterian, Saalian and Weichselian glaciations (cf. BENDA 1995; EHLERS et al. 2004) which appear to be characterized by a cyclic pattern of ice-flow directions. It is suggested that ice advanced first from a northerly or northeasterly direction, to be succeeded by an ice stream from the east (e.g. WENNBERG 1949; WOLDSTEDT & DUPHORN 1974; EISSMANN & MÜLLER 1979). Northeast and eastward sourced ice reached North Germany in all glaciations, but there is evidence for ice from the north only during the Elsterian Glaciation. Each glaciation is represented by several ice advances (Fig. 3, blue bars at the right), which more or less differ in source area, flow direction, extent and composition of their deposits. German terms are given in quotation marks or – for better understanding – occasionally added in paraphrases behind the English terms. For cold phases (German: Kaltzeit) often the neutral term 'kryomer' is used (cf. LÜTTIG 1965, MENKE 1980, MÜLLER 1992: kryomere), for warm phases the term 'thermomer' (German: Warmzeit). The term 'interglacial' is only used in citations, because none of the described thermomers seem to have separated glaciations.

All important sites mentioned in the text are given in Fig. 1.

2 Early Pleistocene [Lower Pleistocene]

The evidence for the 'Lieth-Serie' is recorded in a former limestone pit at Lieth in Holstein. The depositional site is a sinkhole on top of a salt dome and the sediments consist of a succession of lignite-like organic beds separated by sandy units. Palynological and stratigraphic investigations were carried out on the organic units by MENKE (1970, 1975) revealing gyttjas and a boreal *Sphagnum*-Ericales-peat. MENKE correlated the 'Lieth-Serie' with the Quaternary sequence of the Netherlands (ZAGWIJN 1960, 1963) and suggested that it is part of the Early Pleistocene (Fig. 2) and named it 'Ältest-quartär' beginning at c. 2.6 Ma BP. [This age has only recently been formally fixed as the base of the Pleistocene by the IUGS (GIBBARD et al. 2010)]. Later MENKE (cf. STEPHAN & MENKE 1993) additionally compared it with Early Quaternary deposits of the Lower Rhine (URBAN 1978) and palynological investigations published by ZAGWIJN & DE JONG (1984). A full description of the stratigraphic members within this succession is given briefly by Behre in LITT et al. (2007).

The lowest part of the 'Lieth-Serie' consists of fluvial sands with some organic material in thin layers or streaks. In its upper portion organic material is more frequent and silty muds occur with pollen assemblages providing evidence for a warm phase, named the 'Warmhörn-Thermomer'. At Lieth these deposits are disturbed by salt-tectonic processes and are incomplete, but in a borehole from Oldenswort in western Schleswig-Holstein that succession was found well preserved, documented as a 40 m thick sequence (STEPHAN & MENKE 1993). At this site, glacial Elsterian deposits cover the sediments of the 'Warmhörn-Thermomer', however, only their uppermost part has been eroded by the Elsterian glacier. The main elements of the pollen spectra are depicted in Fig. 4 and it can be seen that the typical Pliocene flora disappears at the base of the Quaternary. The break begins with the 'Kaltenhörn-Kaltzeit' which is the first cold phase (Kryomer) of the Lieth succession and the overlying organic beds reveal a stepwise decrease of 'exotic' pollen (Tertiary relics). STEPHAN & MENKE (1993) used this vegetational evidence to derive a temperature curve (Fig. 2). The coldest phase of the Early Pleistocene seems to have been the 'Ekholt cold phase' ('Ekholt-Kaltzeit'). During this and other cold phases coarse sandy fluvial sediments were deposited over large areas of Northwest Germany, presumably in a subarctic environment. These deposits are exposed on the island of Sylt where they are known as the upper part of the 'Kaolinsand'. This unit contains sand blocks that could only have been transported in a frozen state and are considered to be indicators for very cold winters. Likewise very angular quartzite blocks within the 'Kaolinsand' indicate transport without edge abrasion and therefore are considered to have been transported by ice floes (EHLERS 1987, HACHT 1987). Similarities exist with a probably fluvial lag sediment exposed in the Rehburger end moraine in Lower Saxony containing up to boulder-sized Scandinavian erratics that are interpreted as belonging to the 'Complex of Hattem' (EHLERS, MEYER & STEPHAN 1984). Evidence of a glaciation in North Germany at that stage has not been found.

The first warm phase of the 'Gorleben Complex', the 'Osterholz-Warmzeit', still belongs to the Early Pleistocene (cf. LITT et al. 2007: supplement 1).

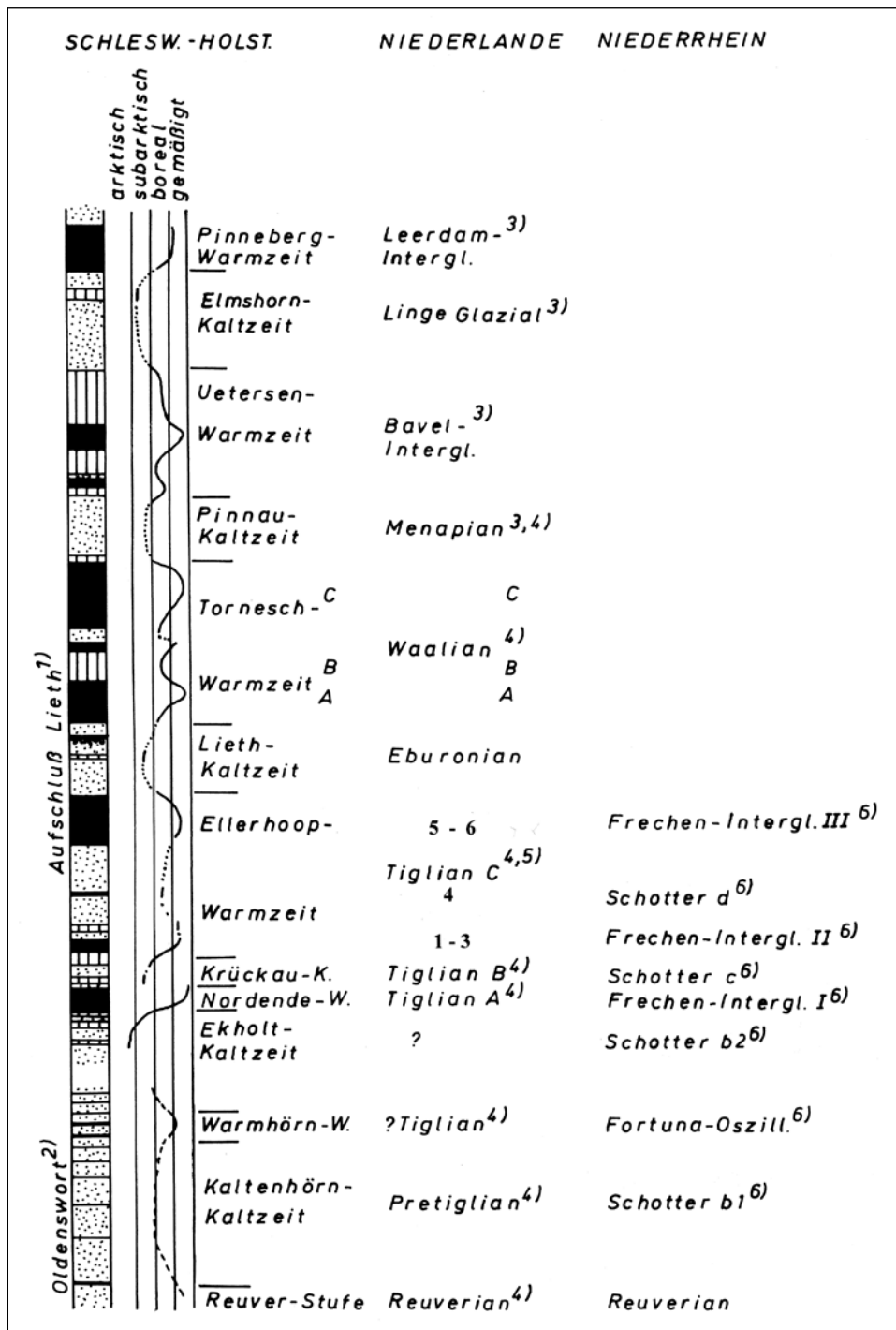


Fig. 2: The Early Pleistocene ("Ältestquartär"), correlations and temperature curve. Original drawing by MENKE (STEPHAN & MENKE 1993, Abb. 1). Petrographical column: Black = brown coal-like mud und peat; dotted = sand; vertical signature = silty/sandy brown coal (silt in cold phases). 1) MENKE 1975, 2) STREMMER & MENKE 1980, 3) ZAGWIJN & DE JONG 1984, 4) ZAGWIJN 1960, 5) ZAGWIJN 1963, 6) URBAN 1978.

Abb. 2: Das Frühpleistozän („Ältestquartär“), Korrelationen und Temperaturkurve. Originalzeichnung von MENKE (STEPHAN & MENKE 1993, Abb. 1). Petrographie-Säule: Schwarz = Braunkohleartige Gytja und Torf; gepunktet = Sand; senkrecht liniert = siltige/sandige Braunkohle (Silt in Kaltzeiten).

Sediments of the 'Gorleben-Complex' are not yet known from Schleswig-Holstein. They were preserved in a sinkhole formed by subrosion, quite similar to the conditions at Lieth. There, sediments of the 'Gorleben Complex' are underlain by deposits of the 'Dannesch Complex' (= Bavelium) comprising the upper part of the Lieth sequence, including the Uetersen warm phase, Elmshorn cold phase, Pinneberg warm phase and Dorst cold phase.

Problems

The age of the 'Lieth-Serie' could not be determined by physical methods. Magnetostratigraphic investigations failed, because magnetic grains in the sands between the organic beds are extremely rare. The correlation with the Early Pleis-

tocene stratigraphy of the Netherlands as depicted in Fig. 2 and 3 seems questionable in some parts. In contrast to the Lieth succession, the Dutch organic layers represent shorter sequences with the organic beds showing hardly complete vegetational development, and have not yet been found in superposition.

In Fig. 3 the transition between the magnetostratigraphic Matuyama and Brunhes chronozones is used as the boundary between the Early Pleistocene and the Middle Pleistocene, as proposed by HEAD, PILLANS & FARQUHAR (2008). In this case at least the first warm phase ('Osterholz warm phase') and eventually also the second ('Hunteburg warm phase') of the 'Gorleben Complex' would be regarded as Early Pleistocene, as both show a reverse magnetization. LITT et al.

Age Ma		NW-Europe	Schleswig-Holstein	Glaciations
HOLOCENE				
0,0117	Late Pleistocene	Weichselium (Complex)	Weichsel-Complex	Weichsel - cold phase s.s.
0,115		Eemium		Eem - warm phase
0,22	Middle Pleistocene	Saaliium (Complex)	Saale-Complex	Saale - cold phase s.s.
				Leck - warm phase
				Bokelrehm - cold phase
				Wacken - warm phase
0,33	Pleistocene	Holsteinium		Mehlbek - cold phase
				Holstein - warm phase
				Elster - cold phase
0,78	Brunhes	Cromerium (Complex)		Gorleben - Complex with 5 stronger warm phases
	Matuyama	Bavelium (Complex)	Danesch-Complex	Dorst - cold phase
				Pinneberg - warm phase
				Elmshorn - cold phase
	Early Pleistocene	Menapium		Uetersen - warm phase
		Waalium (Complex)		Pinnau - cold phase
1,8	Gelasium	Eburonium	Barmstedt-Complex	Tornesch - warm phase
				Lieth - cold phase
		Tiglium (Complex)		Ellerhoop - warm phase
				Krückau - cold phase
				Nordende - warm phase
2,6	Pliocene	Praetiglium (Complex)	Appen-Complex	Ekholt - cold phase
				Warmhörn - warm phase
				Kaltenhörn - cold phase
				Reuver Brunssum Suster Garding

(Age axis not to scale)

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Last update : July 2010

Fig. 3: Climato-stratigraphical table of the Pleistocene in Schleswig-Holstein. Last updated September 2010. Layout influenced by the 'Stratigraphische Tabelle von Deutschland (STD)' (DSK 2000) and LITT et al. (2005). Vertical signature: Unit not found in Schleswig-Holstein so far. Abb. 3: Klimastratigraphische Tafel des Pleistozäns in Schleswig-Holstein. Zuletzt überarbeitet im September 2010. Gestaltung angelehnt an die 'Stratigraphische Tabelle von Deutschland (STD)' (DSK 2000) und LITT et al. (2005). Senkrechte Signatur: In Schleswig-Holstein bisher nicht nachgewiesen.

(2007) include the second warm phase in the Middle Pleistocene and suggest correlation with the Lishi-Event with a reverse magnetization within the Brunhes chronozone, but discuss an alternative inclusion in the Early Pleistocene. From a vegetational point of view, the base of the cold phase succeeding the 'Osterholz warm phase' would represent a convincing boundary. After this cold phase the Tertiary relic *Eucommia* did not return (HAHNE et al. 1994).

3 Middle Pleistocene

The Middle Pleistocene is divided into an upper succession with glaciations and a lower non-glacial succession represented by the main part of the 'Gorleben Complex' (early Middle Pleistocene) in Fig. 3. The sediments of this complex were deposited during five major warm phases of interglacial rank and 7 shorter and weaker warm phases of interstadial character, interrupted by phases of cold but non-glacial

climate (MÜLLER 1992). Two of the warm phases have not been given names until now. The upper Middle Pleistocene comprises the deposits of the Elsterian, the Holsteinian and the Saalian.

Problems

The correlation of the early Middle Pleistocene of the well-documented 'Gorleben Complex' with the units of the so-called 'Cromer Complex' in Western Europe (ZAGWIJN, VAN MONTFRANS & ZANDSTRA 1971; ZAGWIJN 1996) is still in discussion (e.g. URBAN in STEPHAN et al. 2011). It seems to be not clear so far how to combine the 12 thermomeres of the 'Gorleben Complex', especially the 5 warm phases of 'interglacial' rank with the warm events of the 'Gorleben Complex'. The number of known Cromerian warm stages in Western Europe is obviously incomplete yet. The same is true regarding the early Middle Pleistocene deposits of East Germany (e.g. EISSMANN 2004).

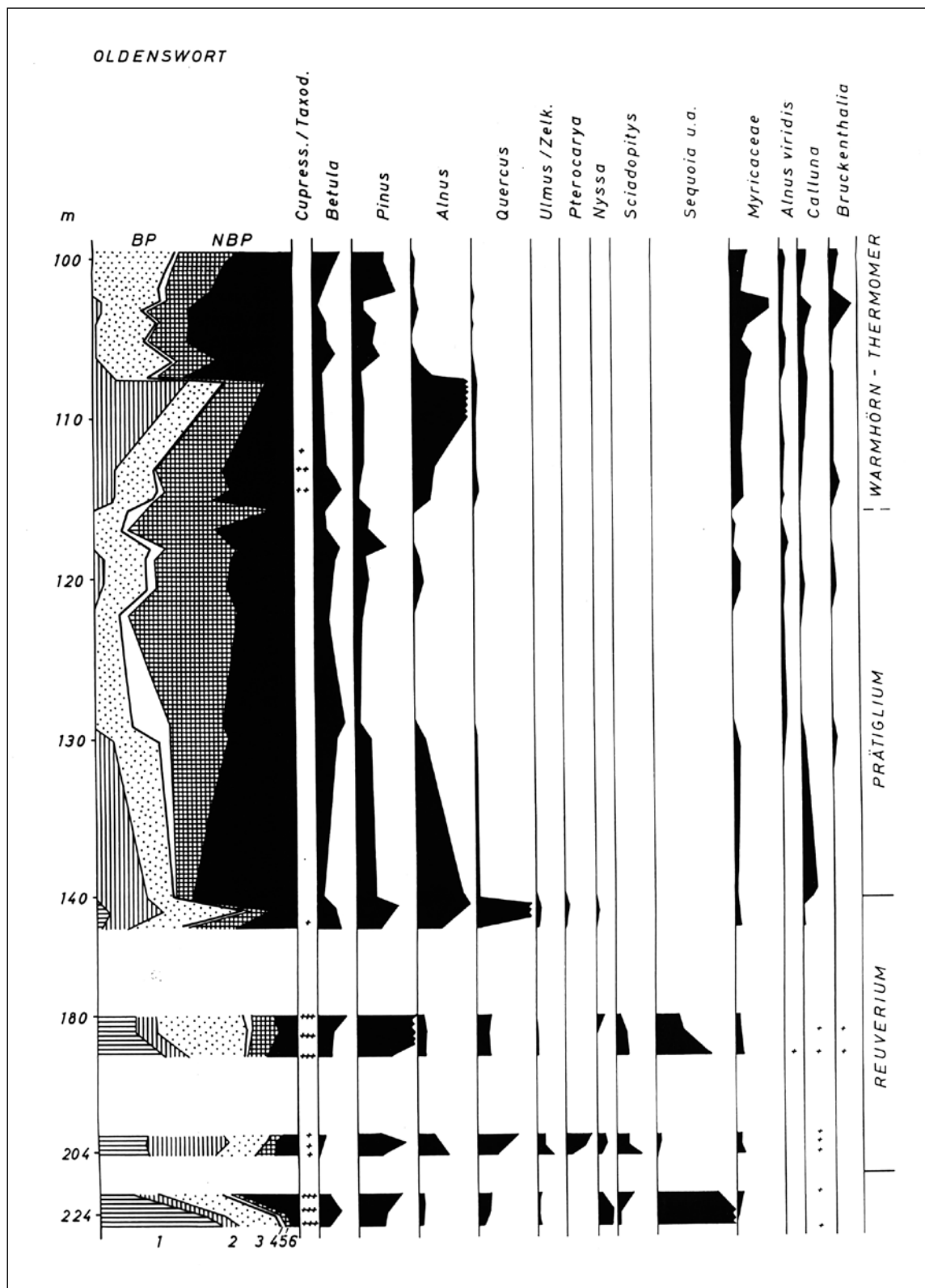


Fig. 4: Simplified pollen diagram (selected species) of the 'Warmhorn warm phase' ('Warmhorn-Warmzeit') (Lower Pleistocene) above Pliocene. Oldenswort drilling. Counts and original drawing by Menke (without year). BP = pollen of trees and shrubs: 1 = *Sequoia* and other typical Pliocene species, 2 = other thermophile species, 3 = boreal species, including *Myricaceae*. NBP = non-arboreal pollen: 4 = herbs, 5 = *Poaceae*, *Cyperaceae*, 6 = *Ericales*.

Abb. 4: Vereinfachtes Pollendiagramm (ausgewählte Arten) der 'Warmhorn-Warmzeit' (Unter-Pleistozän) über Pliozän. Bohrung Oldenswort. Zählung und Originalzeichnung von Menke (ohne Jahresangabe). BP = Pollen von Bäumen und Sträuchern: 1 = *Sequoia* und andere typisch pliozäne Arten, 2 = andere wärmeliebende Arten, 3 = Boreale Arten, einschließlich *Myricaceae*. NBP = Nichtbaumpollen: 4 = Kräuter, 5 = *Poaceae*, *Cyperaceae*, 6 = *Ericales*.



Fig. 5: The Lower Pleistocene Lieth succession in the Lieth open cast mine; dark bands = brown coal-like organic beds, white bands = sands (cf. EHLERS 1983, Fig. 243). In front: Helmut Stremme (with megaphone) and Burchard Menke. Photograph: H.-J. Stephan 1970.

Abb. 5: Die unterpleistozäne Lieth-Abfolge in der Grube Lieth; dunkle Lagen = braunkohlenartige organische Ablagerungen, weiße Lagen = Sande (vgl. EHLERS 1983, Fig. 243). Vorne Helmut Stremme (mit Megaphon) und Burchard Menke. Foto: H.-J. Stephan 1970.

Elsterian (Elster Complex; Elster-Kaltzeit)

Deposits of the Elsterian are found close to the surface in a relatively narrow fringe in eastern Germany, in other regions in few elevated areas with only a thin cover of younger sediments and in some 'geological windows', or in disturbed positions where Elsterian beds were thrust upwards by younger glaciations. In most areas, however, Elsterian deposits were more or less eroded by younger glaciations. They are then most likely found in depressions and on the shoulders of deeply incised tunnel valleys, sometimes completely filling the valleys. Based on the well-investigated areas of Saxony, it was believed for a long time that there were two larger ice advances during the Elsterian (LITT et al. 2007). However, in Schleswig-Holstein evidence was found recently for three different till sheets, so that now a minimum of three glacier advances is proved (cf. KRONBORG 1986 for Denmark). During the youngest of these, a 'Baltic ice stream' formed, with a source area in eastern Fennoscandia (STEPHAN 2007: Goldelund-Formation). The Elsterian 'red till' described from eastern Lower Saxony (e.g. MEYER 1976) probably belongs to this advance. Between the youngest glacier advance and the preceding glaciations a short period of subarctic climate with a regional decay of the ice sheets seems probable (STEPHAN et al. 2011).

At the end of the Elster cold phase, glaciolacustrine 'Lauenburg Clay' or equivalent mostly fine-grained deposits are found widespread in North Germany, especially on top of the infill of tunnel valleys. Within this late glacial deposit an early and short ingress of the sea is documented by an arctic macrofauna in deep tunnel valleys of southwestern Holstein (HINSCH 1993) and in western Mecklenburg by arctic species of foraminifera (MÜLLER & OBST 2008). After this event, probably due to fast isostatic uplift, the marine influence ceased. A second transgression occurred shortly before the onset of the Holsteinian warm stage, but still under arctic or subarctic conditions.

Due to its complex development the Elsterian is now often referred to as 'Elster Complex' (Fig. 3). The Elsterian is correlated with MIS 10 (LITT et al. 2007). Its age is indirectly

determined by dating deposits of the youngest thermomere of the 'Gorleben Complex' ('Rhume-' or 'Bilshausen-Warmzeit'; MÜLLER 1992) and of the Holsteinian (see below).

Problems

No older sediments of glacial origin than Elsterian have been reliably found in North Germany so far. VINX, GRUBE & GRUBE (1997), however, discuss an older till, found at the base of the glacial sequence at Lieth. They regard this local till as a 'pre-Elster-1 Till', because they interpret the overlying clayey till as Elsterian-1. However, it has to be kept in mind that Elsterian tills are extremely variable in composition, resulting from mostly strong reworking and incorporation of material from the local to regional subsurface. In North Germany it is mainly reworked Tertiary material that makes up the clayey, silty, and sandy till matrix. So, for instance, the existence of an extremely sandy till of the second glacier advance in northwestern Lower Saxony (MEYER 1976, 1987) does not exclude a clayey till of the same glaciation phase deposited at Lieth. Both the first and the second till and also the so-called 'pre-Elster-1 till' contain pebbles and blocks from western Scandinavian source areas with a distinctive component of rocks from the Oslo region.

It can, however, not be ruled out that North Germany was reached by a pre-Elsterian glaciation. In Central Jutland (Denmark) ANDERSEN (1965) found 'pre-Harreskovian' glacial deposits including a clayey till. As the 'Harreskov-Thermomere' – according to Urban (in STEPHAN et al. 2011) – correlates with the second thermomere of the 'Gorleben Complex' (Hunteburg warm phase) at least north of the Danish/German borderline an early Middle Pleistocene glaciation is proved.

Hitherto no evidence for such old glaciation was found in Schleswig-Holstein. However, erosion by Elsterian glaciers and meltwater was obviously very strong. It seems possible that deposits of an older glaciation were more or less completely eroded during that period, perhaps with the exception of some local remnants in protected depressions. It is, for instance, not clear how the sediments below the organic

bed of Surheide that has been correlated with the Ferdinandovian by BEHRE (2004) must be interpreted. According to Urban (in STEPHAN et al. 2011) the vegetational history of the really warm 'Surheide-Thermomer' was found to be similar to the 'Hunteburg-Thermomer'. If this correlation is correct and if the underlying sediments are glacial, as believed, one must deduce a pre-Elsterian glaciation extending at least into northwestern Lower Saxony. It might have reached North Germany as early as the end of the Early Pleistocene or the onset of the Middle Pleistocene, i.e. much earlier than ever thought previously. Behre alternatively discusses the inclusion of the Ferdinandovian into the 'Elster Complex' which would result in a much broader definition of the term Elsterian.

The solution of this problem depends on a detailed investigation of the petrographical composition and origin of the sediments below the organic bed of Surheide.

Hitherto it is not known whether the two Elsterian tills of East Germany (WANSA in LITT et al. 2007) can be correlated with the tills found in Northwest Germany. It is also unknown, how far the individual ice advances reached. The late Elsterian 'Baltic ice' must have covered at least northwestern and southern Lower Saxony where LÜTTIG (e.g. 1999) and Meyer (LÜTTIG & MEYER 2002) found Elsterian tills with the majority of erratics from more easterly source areas ('Elster-E-tills') than deduced for other Elsterian tills.

Holsteinian (Holstein warm phase; Holstein-Warmzeit)

The 'classical' Holsteinian is well documented in North Germany, and several detailed investigations of marine deposits and non-marine organic beds have been published. Its name was introduced by GEIKIE (1894: 441). He called the marine deposits in southwestern Schleswig-Holstein 'Holstein bed'. Its lower and upper boundaries are clearly defined by INQUA and the Subcommission on European Quaternary Stratigraphy (SEQS) (JERZ & LINKE 1987). Type region is the Lower Elbe area. According to $^{230}\text{U}/\text{Th}$ -dating (GEYH & MÜLLER 2005), it is synchronous with Marine Isotope Stage 9 (MIS 9), which is supported by ages of Infrared-Radiofluorescence (IR-RF)-dated early Saalian sands (KRBETSCHKE, DEGERING & ALEXOWSKY 2008; KRBETSCHKE & STEPHAN 2010) with ages between 141 ka BP and c. 300 ka. IR-RF-dated 'flaser sands' of the Wacken succession in western Holstein lying between Holsteinian marine beds and sands of the Mehlbek cold phase yielded an age of 296 ± 21 ka (KRBETSCHKE & STEPHAN 2010).

A brief definition and description of the Holsteinian is given by Litt (in LITT et al. 2007).

Problems

The age of the Holsteinian is increasingly debated. Several scientists (e.g. KOUTSODENDRIS et al. 2010; SARNTHEIN, STREMMER & MANGINI 1986, publishing U/Th- and ESR-ages of shells), especially from western Europe, believe this strong thermomer belongs to MIS 11 (cf. TURNER 1996, EHLERS 2010). The question arises whether both, the 'Holsteinian' from western Europe and the Holsteinian from North Germany, in fact describe different warm phases.

At Schöningen (brown coal mine) in Lower Saxony organic deposits overlying late Elsterian sediments were palynologically investigated by URBAN (e.g. 1995, 2007; URBAN et

al. 2011). Beside clear Holsteinian sequences Urban described profiles with different palynological successions, one of them displaying a distinct 'interglacial' character. She named it the 'Reinsdorf Interglacial' and interpreted it as a thermomer younger than the Holstein warm phase. Contrary to her several scientists correlate this organic bed with the Holsteinian (e.g. MEYER 2012), an interpretation that seems to be no longer completely rejected by URBAN (URBAN & SIERRALTA 2012). However, during an excursion of the *Subkommission Quartär der deutschen Stratigraphische Kommission* (DSK) in 2011 to the Schöningen mine Urban presented a carefully prepared section with the Holsteinian bed at the base and the Reinsdorf horizon several metres above.

The age of this warm phase is strongly debated. $^{230}\text{U}/\text{Th}$ data (e.g. SIERRALTA et al. 2012) seem to fix this thermomer approximately in the range of MIS 9 thus intensifying the debate about the age of the 'classic Holsteinian'. Regarding the weakness of the climatic optima of MIS 9a or 9c in the δD record of ice cores (e.g. JOUZEL et al. 2007), a connection of the Reinsdorf as a clearly warm thermomer with them seems to be unlikely.

Hitherto deposits of the Reinsdorf thermomer and succeeding interstadials are unknown in Schleswig-Holstein.

Saalian (Saale Complex)

The Saalian can be roughly divided into a non-glacial lower and a glacial upper part, the 'Saale cold phase s.s.' (STEPHAN 2005: 'Glaziäre Saale-Subgruppe'). However, at closer inspection the Saalian comprises a complex succession of cold and warm phases as depicted in Fig. 3. The 'Wacken warm phase' and the 'Leck warm phase' are recently recognized as members of the Saalian. In the Wacken clay pit a continuous stratigraphical sequence was exposed, reaching from late Elsterian glaciolacustrine Lauenburg Clay via marine Holsteinian clay, fine grained sands with ripple marks ('flaser sands'), several metres thick deposits of the Mehlbek kryomer (so-called 'white sands' coarsening upwards and including a thick kerkoboloid-layer in its upper part), to peat and gyttja of the Wacken warm phase (Fig. 6, 7) (MENKE 1968; DÜCKER 1969). The organic bed is overlain by fine-grained white sands very similar to the underlying periglacial sands, probably deposited in a succeeding cold phase, named 'Bokelrehm cold phase' in Fig. 3. (This new name was derived from the village of Bokelrehm located c. 2 km NW of the Wacken clay pit). Upwards they rapidly become coarser and are discordantly cut by meltwater(?) sands.

According to IR-RF dating by KRBETSCHKE, the Wacken warm phase seems to correlate with MIS 7e (KRBETSCHKE & STEPHAN 2010).

The Leck succession was described by STEPHAN (2006b), results of different studies were published by STEPHAN et al. (2011). In the drillings north of Leck (North Frisia) a c. 9 m thick fluvial sand was found containing streaks or layers with organic sand or silt and a thicker bed with a sandy-silty mud of the 'Leck-Warmzeit' s.s.. The vegetational development of the Leck warm phase is different from that of the Eemian, Holsteinian, the Wacken and also the Reinsdorf succession as described by URBAN (e.g. 1995, 2007); URBAN, SIERRALTA & FRECHEN (2011). It is believed to correlate with MIS 7c (STEPHAN et al. 2011). An organic bed in the clay pit Nachtigall near Höxter, Northrhine-Westfalia, with very similar pollen assemblages (cf. also KLEINMANN et al. 2011)



Fig. 6: Wacken clay pit. Periglacial deposits of the 'Mehlbek cold phase' ('Mehlbek-Kaltzeit') displaying a horizon with strong involutions (kerkobiloids), at the right overlain by the organic bed of the 'Wacken warm phase' ('Wacken-Warmzeit'). Photograph: H.-J. Stephan 1978.

Abb. 6: Tongrube Wacken. Periglaziale Ablagerungen der 'Mehlbek-Kaltzeit' mit einem Horizont mit starker Verbrodelung (Kerkobiloiden). Rechts Überlagerung durch die organischen Ablagerungen der Wacken-Warmzeit. Photo: H.-J. Stephan 1978.



Fig. 7: Wacken clay pit. The organic bed of the 'Wacken warm phase' (peat beneath gyttja) below sands of the 'Bokelrehm cold phase' (new term) overlying sands of the 'Mehlbek cold phase'. Photograph: H.-J. Stephan 1978.

Abb. 7: Tongrube Wacken. Die organischen Ablagerungen der Wacken-Warmzeit (Torf unter Gytja) unter Sanden der Bokelrehm-Kaltzeit (neuer Begriff) und Sande der Mehlbek-Kaltzeit überlagernd. Foto: H.-J. Stephan 1978.

show $^{230}\text{U}/\text{Th}$ ages ranging from $227 \pm 9/-8$ ka to $201 \pm 15/-13$ ka (KLEINMANN, LEPPER & WAAS 2011).

The two Saalian glaciation phases found in Saxony can probably both be correlated with the first Saalian glacial phase of North Germany, the Drenthe Stadium (LITT et al. 2007). Between the Drenthe Stadium and the succeeding Warthe Stadium an ice-free phase occurred in North Germany, either due to a weak amelioration of climate, or to a decrease of snowfalls in the southern Scandinavian highlands. It is not known where the active ice margin was located during the 'interphase'. At least in the westernmost areas of NW Germany, for instance in Dithmarsia, the ice sheet melted down and the landscape became partly ice-free. The land surface was altered by periglacial processes, and features typical for an arctic environment formed. The climate obviously never reached interstadial conditions. Neither soils nor organic beds have been found anywhere (cf. STEPHAN 1980; MEYER 1987, 2005; MÜLLER 2004). The bleached loam in the Saalian Drenthe till of the 'Rotes Kliff' (red cliff) sec-

tion on the island of Sylt was originally interpreted as an 'interglacial' and later as an interstadial soil (FELIX-HENNINGSSEN & URBAN 1982; FELIX-HENNINGSSEN 1983), covered by a Warthian till. STEPHAN & MENKE (1993) questioned this interpretation. Most likely the older Saalian (Drenthian) till was bleached during the Eemian and the Early Weichselian interstadials and then, during the cold phases of MIS 4 and MIS 3 (see chapter Weichselian), covered by a fresher sheet of periglacially reworked (not bleached) till material from nearby. Both, the Drenthian till and its periglacial cover then underwent further weathering during the Middle and Late Weichselian interstadials and the Holocene.

In Schleswig-Holstein, south of Kiel, meltwater deposits overlain by a Saalian basal till yielded OSL ages between 164 ± 27 ka and 199 ± 19 ka (PREUSSER 1999) and TL ages between c. 127 ± 19 ka and 131.4 ± 19.7 ka BP (MARKS et al. 1995). According to recent IF-RF-dating by KRBETSCHKE, DEGERING & ALEXOWSKY (2008), sandy sediments of the glacial upper part of the Saalian in Central and East Germany are no older than

150 ka BP and no younger than 130 ka BP, thus having been deposited within a period of only 20 ka.

Problems

The subdivision of the glacial Saalian is not at all clear. According to WOLDSTEDT (1926, 1954) glaciers reached North Germany during two different large stadials, the 'Drenthe-Stadium' and the 'Warthe-Stadium'. However, in Northwest Germany three different widespread Saalian till sheets exist and even more till facies are to be found. For many decades glacial geologists tried to find out methods for associating Saalian deposits with either the first or the second stadium. Petrographical analyses of tills, especially gravel analyses and counts of Scandinavian 'indicator rocks' ('Leitgeschiebe') were used. Deposits of the Drenthe Stadium were supposed to be characterized by a 'Swedish-dominated' drift, deposits of the Warthe Stadium by a stronger 'Baltic' influence (rocks from East Fennoscandia and the base of the eastern Baltic Sea), due to (climate-induced?) fundamental changes in the main source areas and stream directions of the ice sheet. Consequently, all Saalian deposits not characterized by a 'Baltic' composition, were associated with the Drenthe Stadium (LÜTTIG 1957, 1991, 1999; LÜTTIG & MEYER 2002; MEYER 1976, 2005). However, STEPHAN (1980, 1998) pointed out that both stadia showed a cyclic development with an earlier drift characterized by mainly Swedish rocks and a later drift characterized by relatively frequent Baltic- or East Baltic components. A 'Baltic to East Baltic facies' is revealed by the so-called 'red Drenthe till' in Northwest Germany overlying a succession with a Swedish-dominated composition and also by the typical Warthian drift (in SE-Holstein and NE-Lower Saxony often again with a 'red till'; EHLERS 1992) overlying Swedish-dominated deposits. Using this model STEPHAN (1980) attributes all tills younger than the 'red Drenthe till' to the Warthian, including the so-called 'Kreidemoräne', rich in flint and often also in chalk, which corresponds to the description of WOLDSTEDT (1954) but not of WOLDSTEDT (1926). MEYER (2005) disregards those results and refers to that till as 'Drenthe-2 Till' or the 'younger Drenthe Till', in agreement with LÜTTIG (e.g. 1991, 1999).

To avoid confusion and in agreement with the stratigraphical rules in Schleswig-Holstein (STEPHAN 1995) and partly in Hamburg (GRUBE 1981) regional terms are applied. In publications focussing on larger national or international correlation, often the neutral terms Saalian 'older', 'middle' and 'younger advance' (or till) have been used (e.g. KABEL 1982, STEPHAN 1982, 1987, 1998, EHLERS 1990) (see LithoLex at www.bgr.bund.de/litholex).

The 'Kreidemoräne' is often regarded as the typical facies of the middle Saalian ice advance in NW-Germany deposited by a 'Northeast Ice' that had derived abundant Cretaceous material from the southern Danish islands. It is normally used as the stratigraphically defining horizon. However, lately this facies has also been found in older Saalian (for instance in Dithmarsia) and – in northern Schleswig-Holstein and in neighbouring Denmark – in younger Saalian deposits. There till rich or extremely rich in chalk was deposited by an 'East Ice'. Furthermore, it became evident that middle Saalian tills in southwestern Schleswig-Holstein and northwestern Lower Saxony accumulated much chalk and flint from glacial erosion on the tops of salt domes in south-

western and central Holstein. At least, a 'Kreidemoräne' of Weichselian age has lately been detected at the outermost Weichselian margin in southern Holstein (hills of 'Kisdorfer Wohld') (STEPHAN 2011). Much fresh chalk and unweathered flint without abrasion (observed in the fine gravel fraction) prove the till as a 'local till' that had incorporated Cretaceous material from the top of a salt dome nearby. These hills, interpreted as a Saalian endmoraine previously, now can be understood as a Weichselian push moraine. Similarities exist with chalk-rich deposits of the oldest Weichselian glacier advance in Mecklenburg during the Warnow stadial (RÜHBERG et al. 1995, MÜLLER 2004).

Thus incorporation of abundant chalk and flint in tills is not restricted to one stratigraphical horizon in large parts of Northwest Germany and can no longer be used there for stratigraphical correlations. The only exception might be East Holstein situated east of the belt of high-lying salt domes.

In Mecklenburg only two widespread Saalian till horizons are known, and none of them is a 'Kreidemoräne' (MÜLLER 2004). In Brandenburg tills rich in chalk and flint have been associated with the 'younger stadium of the Saalian' by LIPPSTREU, BROSE & MARCINEK (1995). However, until now no overlying Eemian deposits have been found, and this till facies might eventually be determined to be of Weichselian age.

It is still controversially discussed how far the youngest Saalian ice sheet reached in Schleswig-Holstein. Due to finds of 'East Ice' material in fine gravel samples from northwestern Schleswig-Holstein, STEPHAN (1998) drew the ice margin through the island of Sylt towards the northwest, following RICHTER (1937). This line seems to correspond with the extent of the 'Warthian' in Jutland (HOUMARK-NIELSEN 2007). However, this concept is rejected by LÜTTIG (2009). Relying on counts of indicator rocks in northernmost Schleswig-Holstein, he interprets all deposits in the northwest as older Saalian (Drenthian facies or his lately defined North Frisian facies).

Also in southern Schleswig-Holstein the outermost limit of the younger Saalian ice advance is still unknown.

During the early decay of the Saalian ice shield, temporary re-advances in Schleswig-Holstein terminated increasingly in the east. Their deposits are progressively 'Baltic-dominated' probably due to increased ice supply via the recent Baltic Sea Basin. Extremely Baltic to East Baltic ice advanced through this depression presumably caused by a faster melting of West Swedish ice that had blocked westward drainage of the eastern ice masses previously (cf. the Weichselian 'Young Baltic Advance'; STEPHAN 2001). Baltic and finally East Baltic ice reached Mecklenburg, north-eastern Lower Saxony, and Schleswig-Holstein. Baltic ice even reached Jutland, time-transgressive from southeast towards northwest and north. The northernmost 'red tills' of East Baltic origin are found southwest of Kiel. Further to the north reddish tills of Baltic origin occur, characterized by abundant reddish-violet sandstones and quartzites derived from the sedimentary rocks covering the crystalline basement of East Fennoscandia (Jotnian and Eocambrian sandstones). Both facies overlap in Central Holstein.

However, it must be emphasized that the younger Saalian ice also deposited tills with a facies similar to Drenthian tills besides Baltic and East Baltic facies (c.f. KABEL 1982). One reason is the fact that the ice sheet eroded and incorporated



Fig. 8: Eemian and Early Weichselian soils in the sand pit of Hof Keller, type section of the Keller Interstadial. From bottom to top: Distinct Eemian podsol (Keller I), Brörup podsol (Keller II), relic of the Odderade podsol (Keller III), and a thin and weak podsol of the Keller Interstadial (Keller IV), in the right half bipartited. Photograph: H.-J. Stephan 1976.

Abb. 8: Eem- und früh-weichselzeitliche Böden in der Sandgrube Hof Keller, Typ-profil des Keller-Interstadials. Von unten nach oben: Ausgeprägter Podsol des Eem (Keller I), Podsol des Brörup (Keller II), Erosionsrest des Odderade-Podsols (Keller III) und ein dünner, schwach entwickelter Podsol des Keller-Interstadials (Keller IV), in der rechten Hälfte zweigeteilt. Foto: H.-J. Stephan 1976.

much older glacial material resulting in 'mixed facies'. They are difficult to interpret by counts of erratics or by gravel analyses. The degree of mixing probably increased the further to the west or south the ice sheet advanced.

Furthermore, a primary mixing could already have occurred when 'tributary ice streams' flowing down the Swedish mainland towards southeast (cf. drawing by STEPHAN in EHLERS, MEYER & STEPHAN 1984: Fig.11) added their drift to the 'Baltic ice stream' flowing through the Baltic Basin. As a result the northwestern ice band of the Baltic ice stream transported material with a composition similar to older Saalian tills. South of Sweden it turned towards west and even northwest (now as the northern band). Till deposited by this stream band would erroneously not be interpreted as deposit of the younger Saalian ice ('Warthe-Till') by counting indicator rocks.

4 Late Pleistocene [Upper Pleistocene]

Eemian (Eem warm phase, Eem-Warmzeit)

Eemian deposits were found and investigated in numerous sites in Schleswig-Holstein (e.g. MENKE 1976) and other parts of North Germany and Europe. The Eemian vegetational and climatic development seems well known. The correlation with MIS 5e is nowhere called in question. For a short description see LITT et al. (2007).

In North Germany the marine transgression flooded all river valleys along the coasts of the Baltic Sea and the North Sea (cf. HÖFLE, MERKT & MÜLLER 1985). Remarkably, Eemian deposits lie deeper than one would expect considering that a warm thermomer with higher than recent temperatures should be accompanied by a higher sea level. South of Lübeck a typical Eemian beach sediment was found in a drilling at c. -28 m below the recent sea level (NN). This means a subsidence of that region by c. 0.25 mm p.a. on average since 120 ka, which agrees with the epirogenetic sinking of the Lower Elbe region in geological time.

In land areas not affected by the transgression, Eemian

peats or gyttjas developed in many depressions, later covered either by periglacial sediments or – in the 'young morainic landscape' – by glacial Weichselian deposits. Eemian soils are frequently found (cf. Figs. 8, 9), providing a good base for the stratigraphical separation of Middle and Late Pleistocene deposits.

Weichselian (Weichsel Complex, Weichsel-Kaltzeit)

The Weichselian is subdivided into the Lower (or Early) Weichselian, usually correlated with MIS 5d to 5a, the Middle Weichselian, correlated to MIS 4 and 3, and the Upper (or Late) Weichselian (MIS 2). The lower part is characterized by cold phases (stadials) alternating with interstadials. In contrast to conventional stratigraphic tables (DSK 2002, LITT et al. 2007), the Eemian in Fig. 3 is followed at least by three Lower Weichselian interstadial horizons (cf. Fig. 8, 9, 10). Short descriptions of the first 2 interstadials (Brörup, Odderade) were given by BEHRE (in LITT et al. 2007). For more detailed descriptions see e.g. MENKE (1975) and AVERDIECK (1967). The third interstadial (Keller) was described by MENKE (1980). In exposures in Schleswig-Holstein frost cracks, thin convoluted layers, and horizons with ventifacts prove that the Lower Weichselian cold phases saw at least temporary subarctic to arctic conditions. In an exposure in the northernmost Schleswig-Holstein (Osterbylund) in the stadial sediment separating the interstadial Brörup- and Odderade-soils, a mass flow deposit ('Fließerde') with stones and even blocks derived from a thrust Saalian till body has recently been observed (cf. Fig 9), proving a temporary moist arctic climate.

The lower Weichselian interstadials precede a very cold arctic phase between c. 72 and 60 ka BP (JOUZEL et al. 2007, Wolff et al. 2010). Its base is thought to mark the beginning of the Middle Weichselian (MIS 4) (see below). In 1994 in northern Schleswig-Holstein first hints were found for a glaciation probably taking place during this phase. TL dating of meltwater deposits yielded ages between 59.4 ± 8.9 ka and 52.5 ± 7.9 ka (MARKS et al. 1995). Later PREUSSER (1999) pub-



Fig. 9: Osterbylund sand pit WNW of Ellund (Fig.1). At the base (right) Saalian till transformed to an Eemian gley (Ee), overlain by Early Weichselian sediments with a thick podsol of the Brörup Interstadial (Br), a podsol of the Odderade (Od), and a very weak podsol of the Keller Interstadial (Ke) at the top. Between the Brörup podsol and the Odderade podsol a mass flow deposit of the Rederstall Stadial is visible in the centre and further left. Photograph: H.-J. Stephan 2012.

Abb. 9: Sandgrube Osterbylund WNW von Ellund (s. Abb. 1). Rechts unten saalezeitlicher Till, zu einem eemzeitlichen Gley (Ee) verwittert, überlagert von früh-weichselzeitlichen Sedimenten mit einem mächtigen Podsol des Brörup (Br), einem Podsol des Odderade (Od) und einem sehr schwachen Podsol des Keller-Interstadials (Ke) (z. T. verwaschen). Zwischen Brörup- und Odderade-Podsol kann man in der Bildmitte und links davon eine Fließerde erkennen. Foto: H.-J. Stephan 2012

lished OSL-dates between 92 ± 24 ka and 61 ± 16 ka. The consistency of ages pointing to the lower Weichselian or early Middle Weichselian is the main reason for the assumption of such an early glaciation. The cold phase was named 'Ellund-Phase' by STEPHAN (2003) (cf. STEPHAN 2006a: Ellund-Formation). It is in rough agreement with the postulated age of the so-called 'Old Baltic ice advance' in Denmark (PETERSEN & KRONBORG 1991) or the synonymous 'Ristinge Glaciation' (HOUMARK-NIELSEN 2007). This glaciation was probably induced by the arctic 'Ellund-Phase'. HOUMARK-NIELSEN (2010) using OSL dates of meltwater deposits believes that glaciation to have taken place not earlier than 54 ka ago in the north-western Baltic region, that means during the early phase of the succeeding weak warming.

Recently a Lower Weichselian glaciation has also been deduced by MÜLLER (2004: 'Warnow-Stadial', 2007: 'Warnow-Formation'), based on core samples from northwest Mecklenburg. There, marine deposits or an organic horizon ('Saßnitz-Interstadial') overlay the glacial deposits. According to Müller (2004, 2007) it was the strongest of all Weichselian glaciations terminating in most cases south of the LGM-line drawn by EHLERS et al. (2004). Its till (qw0-till) is often rich in flint and chalk (RÜHBERG et al. 1995) and is thought to have been commonly mistaken as the 'Saale III till' in older publications (e.g. CEPEK et al. 1975).

After the 'Ellund-Phase' the Middle Weichselian continued as a mainly cold period with several slightly milder interruptions. It lasted c. 30 ka.

The upper Weichselian comprises the pleniglacial with glacier advances during the 'Brandenburg-Phase', 'Frankfurt-Phase', 'Pommern-Phase' and 'Mecklenburg-Phase' (LITT et al. 2007; cf. LithoLex data bank) and the phase of late Weichselian interstadials. It ends at 11.7 ka BP, based on correlation with the NGRIP ice core (RASMUSSEN et al. 2006; WALKER et al. 2009). During the 'Mecklenburg-Phase', which is synonymous with the 'Young Baltic advance' in Schleswig-Holstein and Denmark (STEPHAN 2001), again a

'Baltic ice stream' originated, called the 'Young Baltic ice advance' (STEPHAN 2001), similar to the end phases of Elsterian and Saalian glaciations (cf. last paragraph of chapter Saalian). Ice masses, especially from an ice dome in the northeast of Scandinavia and from tributary ice streams from Central Sweden, flowed along the recent Baltic Sea Basin turning west and then northwest towards the Kattegat and Skagerrak, very likely due to the melting of former blocking ice in the west. This was probably augmented by a simultaneous isostatic uplift of the southern areas of Pomerania and North Germany in response to deglaciation. 'Baltic ice' covered the northern Mecklenburg and the east of Schleswig-Holstein. In tills of this phase layers or streaks of 'red till' occur, characterized by many clasts of reddish or violet sandstones or quartzites (most are Jotnian sandstones) and Palaeozoic limestones. 'Baltic crystalline' (e.g. Åland rocks or Baltic porphyries) is rather rare in deposits of the Young Baltic ice advance in Schleswig-Holstein. Probably a northwestern to northern band of the Baltic ice stream reached its coastal regions (cf. EHLERS, STEPHAN & MEYER 1984: Fig. 11) predominantly depositing a 'Svecobaltic facies' (STEPHAN 2003) that was mainly derived from eastern Sweden, the adjacent subground of the recent Baltic Sea and the islands of Öland and Gotland.

In the western Baltic area, the Young Baltic ice advance probably ended at about 15 ka BP, when kame sediments TL- and OSL-dated by PREUSSER (1999) were deposited between stagnant and melting ice masses, approximately contemporaneous with the end of the last pleniglacial cold phase that was followed by rapid warming (STUIVER, GROOTES & BRAZIUNAS 1995).

Problems

Dating of the Weichselian deposits in Schleswig-Holstein is still uncertain. It is, for instance, debatable whether TL dates and OSL dates for the assumed early Middle Weichselian glaciation are reliable or not. It cannot be ruled out that

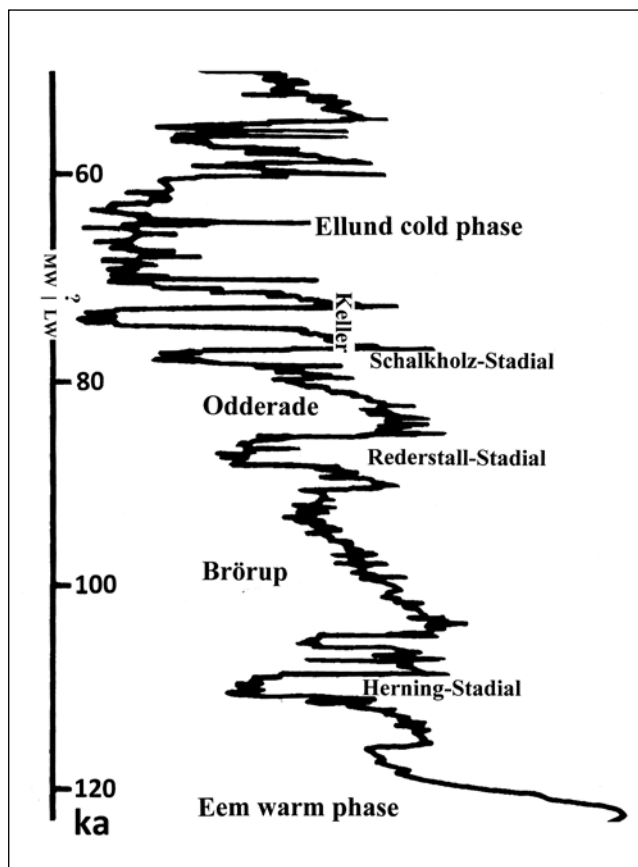


Fig. 10: Stratigraphical table of the Lower Weichselian of Schleswig-Holstein correlated with the $\delta^{18}\text{O}$ record (simplified curve) of NGRIP ice core. MW / LW: Transition Lower Weichselian - Middle Weichselian? The period Odderade - Keller corresponds with stage MIS 5a (cf. BOND et al. 1993).

Abb. 10: Stratigraphische Tabelle des Unterweichsel von Schleswig-Holstein, korreliert mit der vereinfachten $\delta^{18}\text{O}$ -Kurve des NGRIP Eiskerns. MW / LW: Übergang Unterweichsel - Mittelweichsel? Der Abschnitt Odderade - Keller entspricht dem marinen Isotopen-Stadium 5a (MIS 5a) (vgl. BOND et al. 1993).

measured grains were not completely bleached by sunlight before deposition, resulting in an overestimation of their age.

The position of the 'Keller-Interstadial' could be debated. MENKE (1980) proposed to include the Keller into the Lower Weichselian. He assumed that it correlates with the 'Oerel-Interstadial' of Lower Saxony that was found above a Lower Weichselian sequence including peats of the Brörup and Odderade. The Oerel peat was ^{14}C -dated (BEHRE & VAN DER PLICHT 1992) revealing an age between 53.5 and 57.7 ka. Behre (in LITT et al. 2007) associated the Oerel to the Middle Weichselian. This discrepancy could have several reasons:

1. Correlation of the Keller interstadial and the Oerel interstadial is wrong. However, the continuous sequences with Lower Weichselian interstadial soils (or sometimes organic layers) above clear organic Eemian in Schleswig-Holstein and the continuous sequence with organic horizons above organic Eemian at Oerel in Lower Saxony (BEHRE & LADE 1986) plead for synchronism of these successions.

2. The inclusion of the Keller interstadial in the Lower Weichselian could be wrong. However, in the four known depressions in Schleswig-Holstein where the complete succession Eemian - Brörup - Odderade - Keller separated by stadials was found (cf. Fig. 8, 9) continuous deposition of

periglacial niveofluviatile to eolian sediments was observed in the stadial beds. Intensive cryoturbations as indications for a strong change of climatic conditions which must be expected during a glacial event as the succeeding cold Ellund phase and glaciation occur first above the Keller horizon. Therefore, the association of the Keller interstadial to the Lower Weichselian seems to be plausible.

3. The inclusion of the Oerel interstadial in the Middle Weichselian could be wrong. This connection was supported by the ^{14}C -dating of the peat of the Oerel bed. The question arises whether conventional radiocarbon dates of more than 50 ka published by BEHRE & VAN DER PLICHT (1992) are reliable or not. Important arguments for the assumption that the radiocarbon dates yield too young ages are the climatic records observed in the ice cores of Greenland and Antarctica (STUIVER & GROOTES 2000, GROOTES et al. 2001, JOUZEL et al. 2007, WOLFF et al. 2010). In the NGRIP ice core of Greenland the uppermost part of the Odderade thermomer has an age around 80 ka BP (cf. Fig. 10), quite different to the ^{14}C -age of c. 60.8 ka BP of the analogous part of the Odderade peat from the Oerel drilling (BEHRE & VAN DER PLICHT 1992).

In Schleswig-Holstein, in three depressions with the Lower Weichselian sequence one soil of the 'Keller-Interstadial' was observed hitherto above the Odderade soil, succeeding the 'Schalkholz-Stadial' (cf. Fig. 10). In $\delta^{18}\text{O}$ records of Weichselian successions the attributive thermomer is marked as Dansgaard-Oeschger Event 20 and part of the Lower Weichselian (cf. WOLFF et al. 2010). Special problems arise, however, because Menke at the type locality of Hof Keller used the name 'Keller-Interstadial' for a short complex of two separated thin and weak soil horizons. It is the only site where two post-Odderade soils have been found. Probably they document weathering during the weak thermomers of the Dansgaard-Oeschger Events 19 and 20. Unfortunately WOLFF et al. (2010) and others place the transition Lower/Middle Weichselian between these events. As a consequence the older soil would be part of the Lower Weichselian (MIS 5), the younger soil part of the Middle Weichselian (MIS 4). Dansgaard-Oeschger Events 19 and 20 are interrupted by a short but very cold phase (perhaps caused or at least strongly influenced by the Lake Toba 'Mega-eruption'?). On the one hand this sharp interruption could be well used for the establishment of a boundary. On the other hand, however, in marine environments climatic records are less detailed. There the strong drop of temperature and the transition to a long-lasting cold period (Ellund cold phase) are commonly found at c. 70 ka BP (e.g. JOUZEL et al. 2007). The same is true for the $\delta^{18}\text{O}$ record of Taylor Dome in Antarctica (GROOTES et al. 2001). The onset of the Ellund cold phase after the Dansgaard-Oeschger Event 19 therefore seems to fit better with the beginning of the Middle Weichselian and MIS 4 (cf. BOND et al. 1993).

It cannot be ruled out that the two thin and weak soil horizons of Keller in fact correlate with both the Oerel and the Glinde interstadial of Lower Saxony. The thin organic bed of the 'Glinde-Interstadial' would then represent the last weak thermomer of the Lower Weichselian.

The age of the first upper Weichselian glacier advances cannot be finally determined in North Germany. It might be younger than 24 ka BP (cf. STEPHAN in LITT et al. 2007). That suggests a delay between the strong cooling around 26

ka BP (STUIVER & GROOTES 2000), the climatic onset of the Upper Weichselian (or even around 30 ka BP, cf. WOLFF et al. 2010), and the advance of the ice sheet. LÜTHGENS et al. (2010), however, according to new OSL dates of fluvioglacial sediments in north-eastern Germany suggest a first ice advance already during early MIS 2.

The Brandenburg and the Frankfurt Advances were originally morphostratigraphically defined by WOLDSTEDT (1926, 1928). It is unclear whether they are independent advances or whether the latter was just a readvance of the former without deposition of an individual till sheet. The complicated petrographical and stratigraphical situation was briefly discussed by STEPHAN, MÜLLER & VON BÜLOW (2008). LÜTHGENS, BÖSE & KRBETSCHKE (2009) found evidence for a complex origin of the so-called Frankfurt end moraine. They question its origin as the terminal ridge of the 'Frankfurt(Oder) Phase'. A similar debate concerns the glacier advances during the 'Pommern-Phase' and the 'Mecklenburg-Phase'. Some scientists are of the opinion that both advances belong to the same glaciation phase (e.g. SCHULZ 2003; LÜTTIG 2005). Theoretically all glacier advances of the Weichselian Pleniglacial might be combined into only two larger glacial events separated by a phase of gradual melting ('interphase').

The extent of the Weichselian ice advances in W-Mecklenburg and Schleswig-Holstein is not yet clear. The morphological connection of end moraines from Pomerania and Brandenburg to the west is partly debatable. In northern East Holstein the end moraine-lines are interrupted by a landscape dominated by Saalian uplands and strong Weichselian erosion. Stratigraphical correlations using counts of indicator rocks (TGZ-method) by LÜTTIG (e.g. 2004, 2005) provide no solution. Lüttig concludes that the Brandenburg Advance did not reach western Mecklenburg and Schleswig-Holstein and that the Pomeranian Advance ended in East Holstein, and that no evidence for the Mecklenburg Advance exists. So in eastern Central Holstein, in the Schleswig district, and further to the north, all Weichselian deposits should belong to the 'Frankfurt(Oder) Phase'. However, at least the distribution of two 'Young Baltic' tills in Denmark (HOUMARK-NIELSEN 2007) and their continuation southwards (STEPHAN 2001) contradict this view.

The dating of Weichselian end moraines or deposits results in different ages, depending on the methods used. The youngest controversy arose regarding the age of the Pomeranian end moraine. ^{10}Be exposure dates of erratic blocks by RINTERKNECHT et al. (e.g. 2006) yielded an average age of 14.6 ± 0.3 ka BP. This seems to be much too young (cf. MÜLLER, STEPHAN & VON BÜLOW 2009). The postulated age lies in the time of the Late Weichselian strong warming (STUIVER, GROOTES & BRAZIUNAS 1995; RASMUSSEN et al. 2006). The main reason might be that the investigated blocks did not reach their final position at the time of the formation of the end moraine due to long-lasting melt of stagnant and buried ice and continuing periglacial movements (MÜLLER, STEPHAN & VON BÜLOW 2009; STEPHAN & MÜLLER 2007; cf. HOUMARK-NIELSEN et al. 2012). LÜTHGENS, BÖSE & KRBETSCHKE (2009) found an OSL age about 20 ka BP for Pomeranian outwash sediments in north-eastern Brandenburg, an age that seems to be much more convincing.

The first Late Weichselian climatic amelioration is often correlated with the Bölling Interstadial. However, USINGER

(1995) made it clear that the Bölling is synonymous to the lower part of the Alleröd and not identical with the first Late Weichselian interstadial that occurred earlier (*Hippophae* maximum). The latter was defined by MENKE in BOCK et al. (1985) and named 'Meiendorf-Interstadial'. Calibrated ^{14}C dates suggest an age of more than 14 ka BP. This corresponds to the first intensive Late Weichselian warming between 14.7 ka and 14.2 ka BP (STUIVER, GROOTES & BRAZIUNAS 1995; RASMUSSEN et al. 2006).

5 Conclusions

By evaluating all information about the stratigraphic position of Pleistocene deposits in Schleswig-Holstein and adjoining areas including detailed petrographical studies of the sediments, palynological investigations, and dating by physical methods it has been possible to establish a reliable stratigraphical table, based on the regional climatic development. This table is valid not only for Schleswig-Holstein but for North Germany in general and can provide an update for the Pleistocene in the stratigraphical table for Germany (STD), (LITT et al. 2005, 2007: Beilage 1). It can be used as an actual base for chronostratigraphical correlations of Pleistocene successions throughout Europe.

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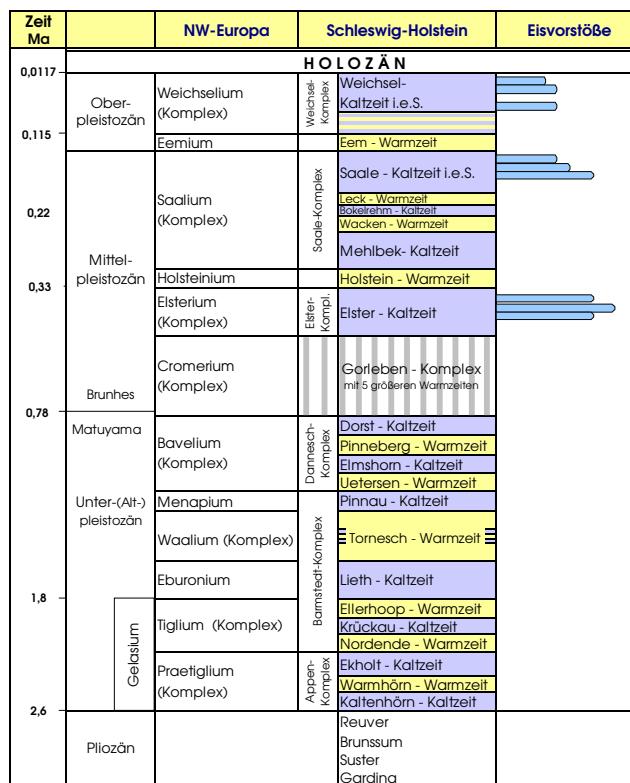
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Klima- und stratigraphische Gliederung des Pleistozäns in Schleswig-Holstein



(Zeitachse nicht maßstäblich)

Entwurf: H.-J. Stephan Juli 2003
Digitale Bearbeitung: M. Jagusch / C. Verdieck
Überarbeitet: Juli 2010

Appendix Fig. 11: Original German version of the climato-stratigraphical table of Fig. 3.

Anhang Abb. 11: Die originale deutsche Version der klima- und stratigraphischen Tafel von Abb. 3.

Erwägungen zur Korrelation mittelpleistozäner Relikte des Rheingletschers mit der Nordschweizer Stratigraphie

Oskar Keller

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Kurzfassung:

Für die Nordschweiz wird in PREUSSER, GRAF, KELLER, KRAYSS & SCHLÜCHTER (2011) dargelegt, dass vor den bekannten Glazialen Würm = Birrfeld und Riss = Beringen zwei (eventuell drei) weitere mittelpleistozäne Glaziale einzuschieben sind: Habsburg und Möhlin. In dieser Arbeit werden an ausgewählten Schlüsselstellen in Süddeutschland litho- und morphostratigraphische Befunde ausgewertet, die es ermöglichen diese zwei neu eingeführten Glaziale auch auf den Rheingletscher zu übertragen.

Dem Zeitraum der Deckenschotter-Eiszeiten schliesst sich die Periode der MPR (Mittelpleistozäne Reorganisation) mit einer markanten fluvialen Ausräumung des Bodenseegebiets an. Es wird gezeigt, dass die Gletscher der nachfolgenden mittelpleistozänen Becken-Eiszeiten das übertiefte Bodenseebecken ausschürften. Die älteste dieser Eiszeiten wird als Größtes Rheinisches Glazial (GRG) = Möhlin bezeichnet und entspricht dem von ELLWANGER (2003) eingeführten Hosskirch. Ihm können im nördlichen Rheingletschergebiet alle äussersten glaziären Bildungen zugewiesen werden. Im Nordwesten wurden diese früher als risszeitlich interpretiert, im Nordosten hingegen als mindelzeitlich. Das GRG ist nachweisbar älter als Riss, aber jünger als die Jüngeren Deckenschotter. Mindel im Nordosten des Rheingletschers würde daher zur Becken-Eiszeit.

Das Habsburg-Glazial erreichte in der Nordschweiz ungefähr Ausmaße wie Würm. Im Rheingletschergebiet konnten in der entsprechenden Bandbreite, aber auch ausserhalb Zeugen und/oder Indizien gefunden werden, die dieses Glazial wahrscheinlich machen. Bisher ist diese eiszeitliche Großvergletscherung im nördlichen Rheingletschergebiet nicht erkannt worden, weil ihre Hinterlassenschaften von den nachfolgenden Glazialen Riss und Würm grobenteils zerstört wurden.

Für die chronostratigraphische Einordnung des Größten Rheinischen Glazials GRG sowie des Habsburg-Glazials können zwischengeschaltete Interglaziale als Zeitmarken beigezogen werden: IG Unterpfaufenwald – GRG – IG Holstein – Habsburg – IG Meikirch – Riss – IG Eem. Demgemäss ergibt sich folgende zeitliche Ordnung für die Glaziale: GRG ± 350 ka BP (MIS 10), Habsburg ± 250 ka BP (MIS 8), Riss ± 150 ka BP (MIS 6).

Considerations on the correlation of relics of the Rhine Glacier in the Middle Pleistocene with the stratigraphy of Northern Switzerland

Abstract:

In the glaciation history of Switzerland (PREUSSER, GRAF, KELLER, KRAYSS & SCHLÜCHTER 2011) is shown that two (possibly three) older glacials had happened before the well known glacials Würm = Birrfeld and Riss = Beringen. These are the Habsburg- and the Möhlin-Glacials. In this paper selected key regions in Southern Germany are analysed litho- and morphostratigraphically. This analysis enables us to find the two older glacials proved also at the Rhine Glacier.

The period of the „Deckenschotter“-glaciations is followed by a time of striking fluvial erosion (MPR = Middle Pleistocene Reorganisation) in the region of Lake Constance. It is demonstrated that the glaciers of the subsequent Middle Pleistocene glacials exarated the overdeepened Lake Constance basin. The oldest of these „Becken-Glacials“ is named „Größtes Rheinisches Glazial“ (GRG) = Möhlin in the northern part of Switzerland. It corresponds to the Hosskirch introduced by ELLWANGER (2003). In the northern Rhine Glacier region the most external glacial deposits belong to this glaciation. Earlier in the northwestern part these deposits are interpreted as Riss-Glacial, but in the northeast as Mindel-Glacial. The GRG evidently must be older than Riss, but younger than the „Younger Deckenschotter“. Therefore Mindel in the northeastern region would be a Becken-Glacial.

In the northern part of Switzerland the Habsburg-Glacial nearly reached the extension of the Würm-Glacial. In the region of the Rhine Glacier references and/or indications of the Habsburg-Glacial could be found in a band width corresponding to the Würm, but outside it as well. In the northern region of the Rhine Glacier this vast glaciation has not been discovered up to now because the subsequent glacials Riss and Würm mainly destroyed its relics.

To classify the „Greatest Rhine Glacial“ GRG as well as the Habsburg-Glacial chronologically interpolated interglacials with time marks can be applied: IG Unterpfaufenwald – GRG – IG Holstein – Habsburg – IG Meikirch – Riss – IG Eem. Accordingly a temporal classification for the glacials results as follows: (GRG) ± 350 ka BP (MIS 10), Habsburg ± 250 ka BP (MIS 8), Riss ± 150 ka BP (MIS 6).

Keywords:

Alpine foreland, Rhine Glacier, chronostratigraphy, glacial deposits, glacial drainage

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zweitletzte und, vor kurzem neu eingeführt (ELLWANGER in LGRB 2003), Hosskirch als drittletzte und grösste Vergletscherung betrachtet werden.

Das gesamte rheinische hydrographische System von der Aare bis zum Alpenrhein bildet bezüglich der Topographie eine Einheit, charakterisiert durch tief ausgeschürfte Auslastäler aus den Alpen und stark übertiefte Vorlandbecken. Unzweifelhaft ist das Bodenseebecken, wenn auch besonders ausgedehnt, Teil dieser glazial ausgeformten Großregion. In Anbetracht des vorstehend erwähnten Kenntnisstandes zu den mittel- und spätpleistozänen Eiszeiten in der Nordschweiz liegt es auf der Hand zu versuchen diese Erkenntnisse auch im nördlichen Rheingletscherraum anzuwenden. Gemäss LGRB (2003), VILLINGER (2011) dürfte eine Entsprechung des nordschweizerischen Birrfeld mit Würm, Beringen mit Riss und Möhlin mit Hosskirch anzunehmen sein.

Bereits in den 1990er Jahren wurde für den Rheingletscher im Bodenseeraum eine größte Vergletscherung nach der Deckenschotterzeit postuliert (KELLER 1994, KELLER & KRAYSS 1999), für die damals die Bezeichnung GHV (Grösste Helvetische Vergletscherung) eingeführt worden ist. Sie wird von GRAF (2009a) und KELLER & KRAYSS (2010) aufgrund ihrer äussersten Frontlage als Möhlin-Glazial bezeichnet und als Entsprechung zu Hosskirch gemäss ELLWANGER (2003) und ELLWANGER et al. (2011a und b) betrachtet. Im vorliegenden Artikel wird sie für den Rheingletscher in ihren Ausmaßen, ihrem Frontverlauf, ihren proglazialen Relikten und in ihrer chronostratigraphischen Stellung diskutiert.

Zwischen die größte Vergletscherung = Möhlin und Riss = Beringen schiebt sich in der Nordschweiz das Habsburg-Glazial ein, das aufgrund der dortigen Untersuchungen deutlich geringere Ausmaße aufwies als jene beiden Vergletscherungen (GRAF 2009a, KELLER & KRAYSS 2010). Diese Habsburg-Vorlandvergletscherung wurde bisher im Rheingletschergebiet in Oberschwaben nicht identifiziert. Für sie sollen in dieser Arbeit Argumente beigebracht werden, die eine solche Vergletscherung als wahrscheinlich erscheinen lassen.

Die vorliegende Publikation beruht auf gemeinsamen Vorarbeiten und Begehungen mit Edgar Krayss bis Sommer 2011. Nun hat sich E. Krayss per Ende 2011 altershalber von der Weiterbearbeitung des Projekts „Rheingletscher in Oberschwaben“ zurückgezogen. Damit übernimmt der Autor O. Keller die alleinige Verantwortung für diese Veröffentlichung.

2 Terminologie

Mit der Publikation von PENCK & BRÜCKNER (1901/09, 396) lag ein Kärtchen im Massstab 1 : 700 000 vor, in welchem der äusserste Eisrand des Rheingletschers nachgezeichnet ist. Das Moränen-Gebiet zwischen der Aitrach und dem Ristal wurde der Mindel-Eiszeit zugeordnet, dasjenige von dort bis zum Hegau der Riss-Eiszeit. Und so blieb es fast hundert Jahre.

Im Zuge der Neuordnung der Eiszeitgliederung im Schweizer Mittelland postulierte SCHLÜCHTER (1988, Fig. 1) vor dem bisherigen Riss eine viertletzte Eiszeit als Most Extensive Glaciation (MEG). In der Arbeit von GRAF (2009a, Tab. 19) figuriert diese Grösste Eiszeit unter dem Namen Möhlin-Eiszeit.

In ihrem Beitrag in den Erläuterungen zur Geologischen Übersichtskarte des Kantons Thurgau führten KELLER &

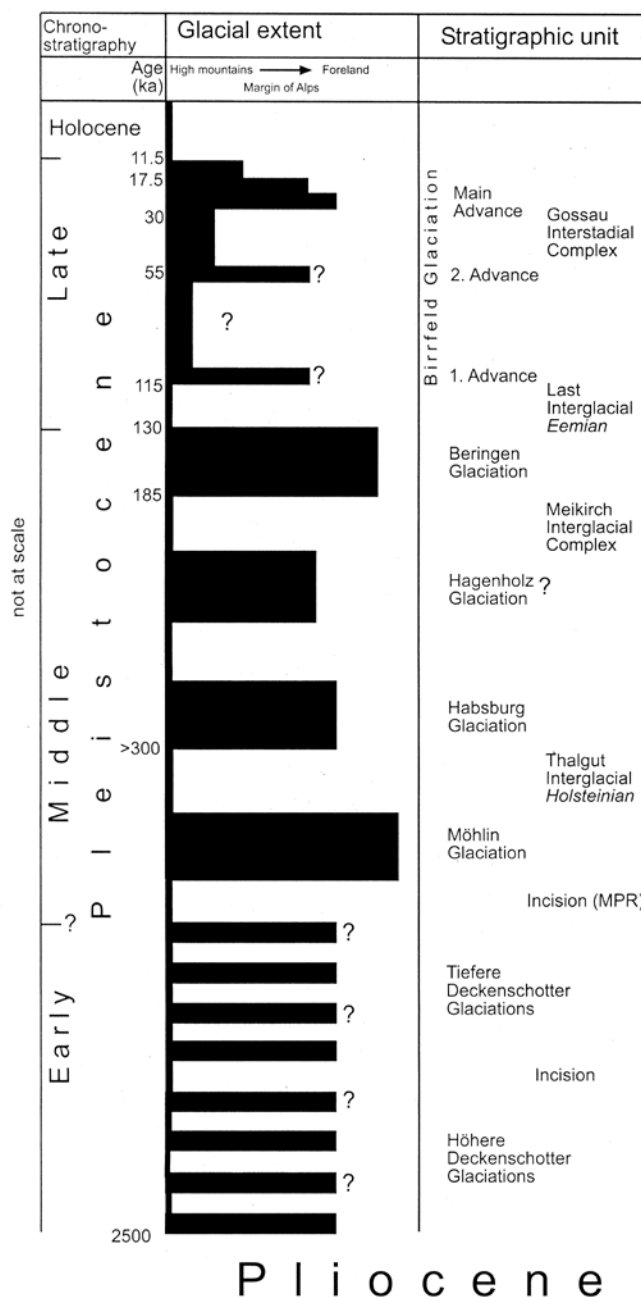


Abb. 2: Stratigraphisches Schema der Vergletscherungsgeschichte der Schweiz (nach PREUSSER et al. 2011).

Fig. 2: Stratigraphy scheme of the glaciation history of Switzerland (after PREUSSER et al. 2011).

KRAYSS (1999) vor der Riss-Eiszeit eine Grösste Helvetische Vergletscherung (GHV) ein, gültig für die Vorlandvergletscherungen der Nordschweiz und des Rheingletschers.

Ellwanger verwendete für die größte Vergletscherung verschiedene Bezeichnungen: Älteres Riss (ELLWANGER 1995), MEG (ELLWANGER in LGRB 2003, 2005). VILLINGER (2011) bezeichnet die „drittletzte/grösste Vergletscherung im Rheingletschergebiet“ mit dem Namen Hosskirch. Dieser bezieht sich topographisch einerseits auf einen früher als Riss angesprochenen Geländestreifen zwischen Messkirch und Riedlingen und andererseits auf den externsten Gletscherrand vom Raum Biberach bis zum Aitrachtal, womit sämtliche äussersten Glazialrelikte der gleichen Eiszeit zugewiesen werden. Im Gelände-Abschnitt Riss-Aitrach wurden bisher die

Glaziale Dynamik (Raum Schaffhausen)

Referenzen: Schindler 1985, Hofmann 1994,
Graf 2009, Keller & Krayss 2010

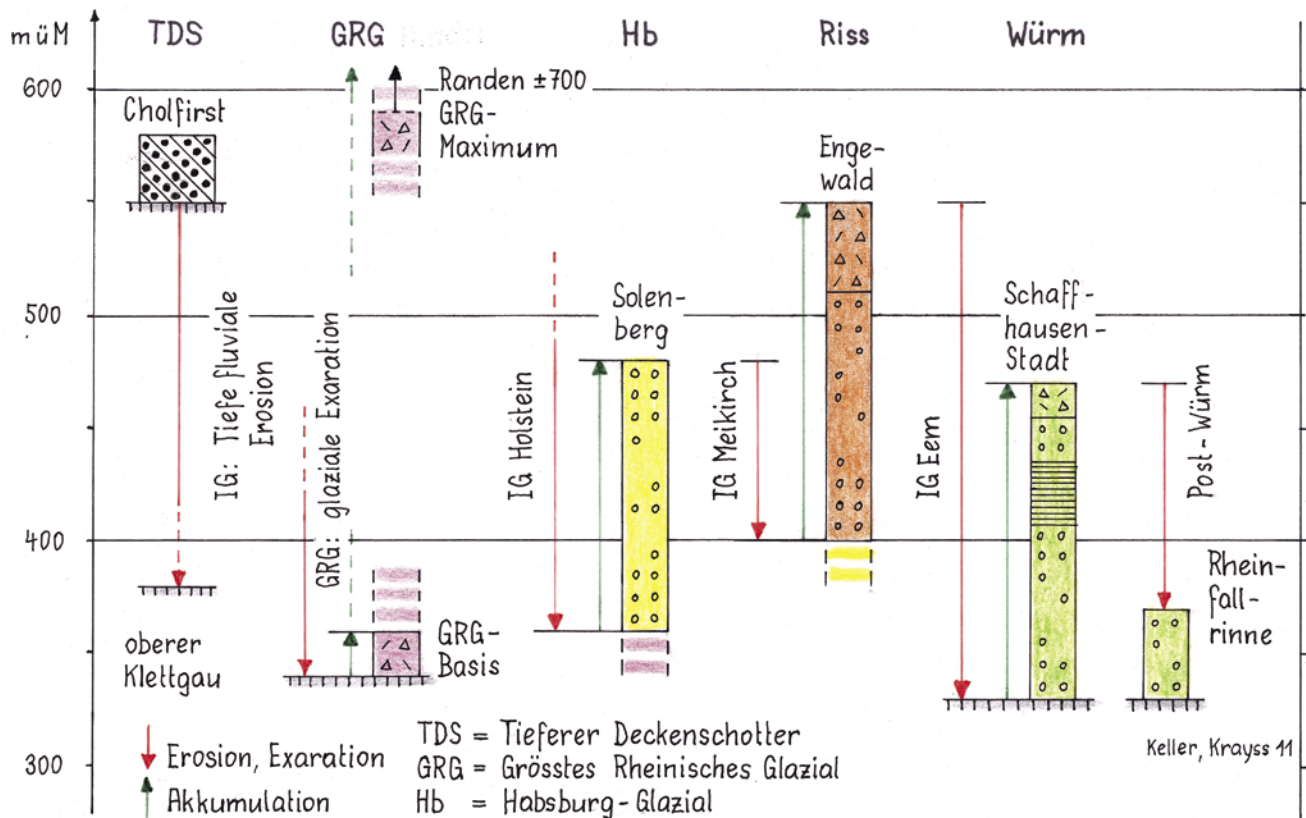


Abb. 3: Morphologische Dynamik der Eiszeiten mit Phasen der Akkumulation und Erosion.

Fig. 3: Morphological dynamics of the glacials shown by phases of accumulation and erosion.

entsprechenden Elemente Mindel zugeordnet. Das Bayerische Landesamt für Umwelt, Geologischer Dienst sowie auch DOPPLER et al. (2011) halten an den Bezeichnungen Mindel und Haslach für die größte Ausdehnung des Bayern zugewandten Rheingletschers fest.

Mit den Begriffen MEG, Möhlin, GHV, Hosskirch und Mindel liegen also im letzten Vierteljahrhundert mehrere Namen für die ausgedehnteste Vergletscherung von der Nordschweiz bis zur Iller vor. Aktuell wird im süddeutschen Raum im nördlichen Rheingletschergebiet für die größte Vergletscherung der Name „Hosskirch“ verwendet, in der Nordschweiz, inklusive Rheingletscher südlich und westlich des Bodensees, hingegen „Möhlin“. Das bedeutet, dass für dieselbe Vorlandvergletscherung zwei Bezeichnungen im Gebrauch sind. Es wird daher in dieser Arbeit hierfür die neutrale Sammelbezeichnung „Größtes Rheinisches Glazial GRG“ verwendet.

Zwischen dem Größten Rheinischen Glazial = Möhlin und dem Riss-Glazial = Beringen ist in der Nordschweiz eine weitere Großvergletscherung des Vorlandes, das Habsburg-Glazial erkannt worden (GRAF 2009b, KELLER & KRAYSS 2010). Für diese Vergletscherung, deren Nachweis hier für das nördliche Rheingletschergebiet angestrebt wird, findet vorläufig die Bezeichnung „Habsburg-Glazial“ Verwendung. Somit wird in der vorliegenden Arbeit für das Mittel- und Spätpleistozän hinsichtlich der Eiszeitengliederung folgende

Terminologie verwendet:

Nordschweiz:	Rheingletschergebiet:
Birrfeld	Würm
Beringen	Riss
Habsburg	Habsburg
Möhlin	GRG (Größtes Rheinisches Glazial)
	= Hosskirch

3 Der Rheingletscher im Mittel- und Spätpleistozän

3.1 Geländestruktur

Zwischen Berner Aaretal und Bodensee ist das Nordschweizer Alpenvorland durch ein tief reichendes fluviales Entwässerungssystem, das auf die niedrig gelegene Oberrheinische Tiefebene ausgerichtet ist, gekennzeichnet. Die Umlenkung des Alpenrheins weg von der Donau nach Westen zum Oberrhein (KELLER 2009) führte zu einer massiven fluvialen Durchtalung des Hochrhein- und des Aaresystems. Die mächtigen Eisströme der anschliessenden größten Vergletscherung schürften in der Folge die meisten übertieften Becken der Nord- und Ostschweiz aus, darunter insbesondere auch das ausgedehnte Bodenseebecken. Dieses nimmt eine Sonderstellung ein, indem die Entwässerung zwar nach Westen gerichtet war, aber während den Maximalständen der Vergletscherungen auch nordwärts in das hoch liegende danubische System übergreif.

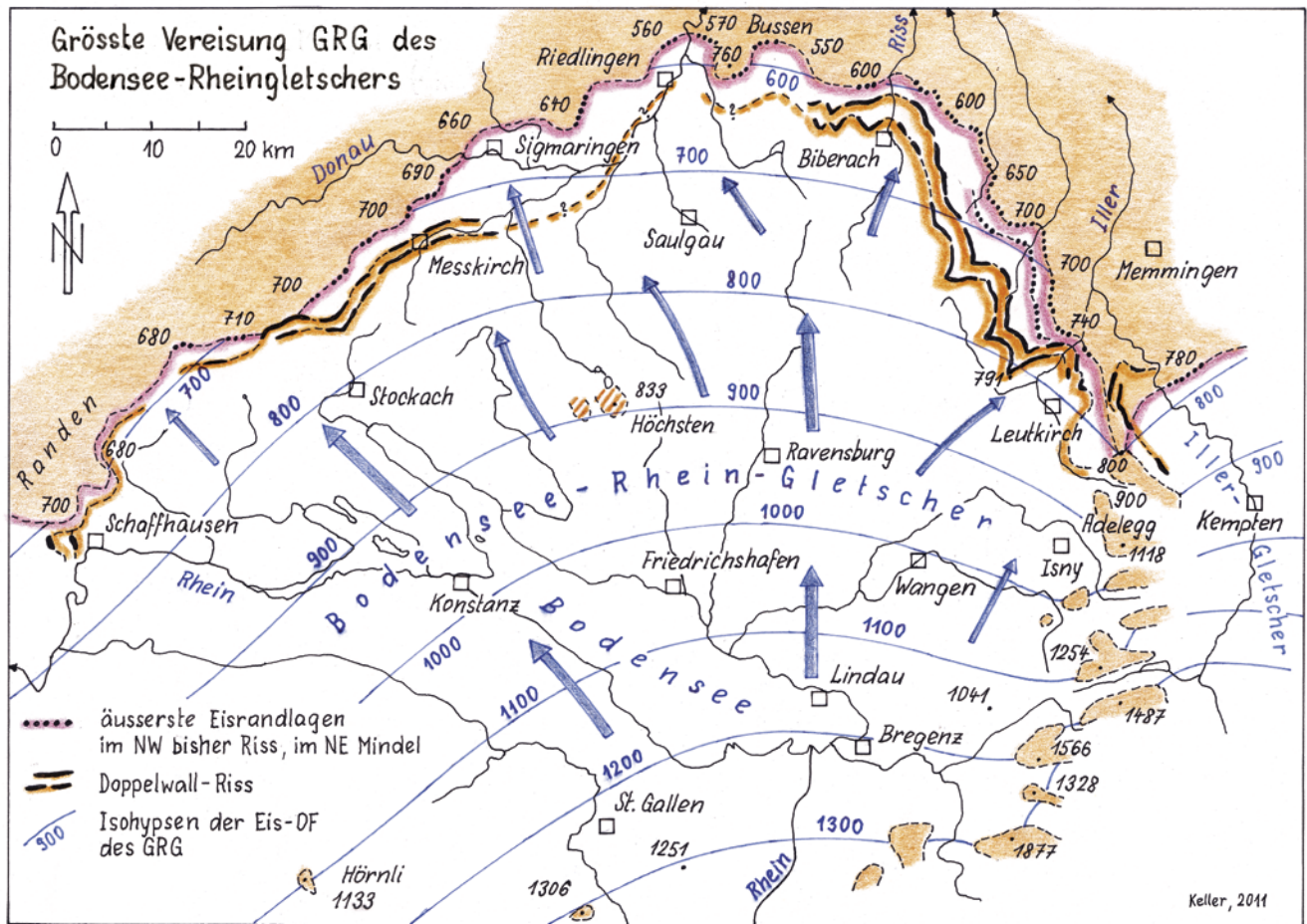


Abb. 4: Maximale Eisausdehnung des Grössten Rheinischen Glazials (GRG) im Vorland.

Fig. 4: Maximal ice extension of the Greatest Rhine-Glacial (GRG) in the foreland.

Eiszeiten, deren Eisströme den nördlichen Rand des Bodenseebeckens, die Wasserscheide zur Donau, überschritten, hinterliessen zur Donau hin glaziale Sedimente, den direkten Nachweis einer Gletscherpräsenz. Erreichten sie aber nicht diese Mindestausmaße, so konnten nur deren Schmelzwasserströme allenfalls kurzfristig Richtung Donau glazifluviale Sedimente ablagern. Lithostratigraphisch sind daher im Nord- und Ostsektor des Rheingletschers ausserhalb des Bodensee-Beckenrandes Eiszeiten mit geringeren Vergletscherausmaßen mit Hilfe glaziärer Sedimente nicht fassbar. Auch die Morphostratigraphie mit Terrasseneinschachtelungen und glazialen Serien ist kaum anwendbar.

3.2 Glaziologie

Vorlandvergletscherungen im Alpenrhein- und Bodensee-raum stellen sich dann ein, wenn im inneralpinen Rhein-Einzugsgebiet die regionale Schneegrenze von den interglazialen Hochlagen auf weniger als 1500 m ü. M. absinkt. Dadurch vergletschern die Alpentäler, deren Eisströme sich vereinigen und durch die Auslastäler ins Vorland vorstoßen. Bei Schneegrenzen um 1000 m ü. M., wie sie beim Rheingletscher im letzten Glazial auftraten (KELLER & KRAYSS 2005), baut sich ein das ganze Bodenseebecken bis zur Donau-Wasserscheide oder darüber hinaus erfüllender Eiskörper auf.

Bedingt durch die in den jüngeren Glazialen bereits niedrig gelegene Diffluenz von Sargans zweigt jeweils ein Ast

des Rheingletschers westwärts ab, was zusammen mit dem Linthgletscher zu einem weiteren Vorlandgletscher im Limmat- und Glattal-Gebiet führt. Die Ausmaße dieses Linth-Rheingletschers machten allerdings gegenüber dem Bodensee-Rheingletscher nur ca. 1/5 des eisbedeckten Gebietes aus (KELLER & KRAYSS 2005).

Rekonstruktionen und Berechnungen für das letzteiszeitliche Maximum des Bodensee-Rheingletschers ergaben für das Vorland ein vergletschertes Areal von 7200 km² und ein Eisvolumen von 2400 km³. Die älteren mittelpleistozänen Vergletscherungen erreichten vergleichbare Ausmaße (Kapitel 4.4).

Der Aufbau der Vorlandvereisung im letzten Glazial nahm nur rund 5000 Jahre in Anspruch, das Rückschmelzen bis in die Alpentäler erfolgte in einem ebenfalls kurzen Zeitrahmen von ca. 6500 Jahren (KELLER & KRAYSS 1993, 2005). Es ist davon auszugehen, dass auch die mittelpleistozänen Eiszeiten für Auf- und Abbau ihrer Vorlandgletscher einen ähnlichen Zeitbedarf hatten.

Außer der größten Vergletscherung, die am Randen hoch hinaufreichte (HOFMANN 1994, KELLER & KRAYSS 2010), stiessen die Gletscher stets bis in den Raum Schaffhausen-Klettgau vor, von wo aus sie zum Hochrhein entwässerten. Über das Mittel- und Spätpleistozän hinweg ergibt sich ein charakteristisches Wechselspiel zwischen Akkumulation glazialer Sedimente und interglazialer Erosion (Abb. 3). Die kaltzeitliche Schüttung der Tieferen Deckenschotter erfolgte auf

einem noch hoch liegenden Niveau um 500 m ü. M. auf dem Cholfirst südlich Schaffhausen. In der anschließenden Interglazialzeit führte die Umlenkung des Alpenrheins Richtung Westen (KELLER 2009) zu einer besonders ausgeprägten fluvialen Erosion, die MPR (Mittelpleistozäne Reorganisation) (SCHLÜCHTER 1988), die im oberen Klettgau das Flussniveau um rund 200 Höhenmeter absenkte. Während den weiteren interglazialen Phasen war die Tiefenerosion stets markant, sodass die Entwässerungsbahnen des Rheins niedrig lagen. Sie verliefen anfangs durch den Klettgau, nach dem Riss durch das noch heute benutzte Rheintal. Die in jedem Glazial anfallenden Schotter und Feinsedimente wurden an den Gletscherfronten zu mächtigen Sedimentstapeln akkumuliert, bevor sie wieder interglazial mindestens teilweise erodiert wurden (GRAF 2009a, KELLER & KRAYSS 2010).

4 Zum Größten Rheinischen Glazial (GRG)

Aus den morphostratigraphischen Untersuchungen zu den jüngeren Eiszeiten in der Nordschweiz geht hervor, dass die ausgedehnteste Vergletscherung nicht Riss sein kann, sondern deutlich älter ist (GRAF 2009a, KELLER & KRAYSS 2010, PREUSSER et al. 2011). Es ist zudem anzunehmen, dass dieses Glazial aufgrund von Paläoböden, Pollensequenzen und pflanzlichen Makroresten, die das Holstein-Interglazial widerspiegeln und im Hangenden über den Sedimenten dieser grössten Vergletscherung auftreten, älter sein muss als dieses Interglazial.

Im nördlichen Bodensee-Rheingletschergebiet wurde bis vor wenigen Jahren von einer einzigen Vergletscherung, die die externsten Glazialrelikte hinterlassen hätte, abgesehen. Hier hingegen wird von einer solchen Vergletscherung ausgegangen, die gemäß LGRB (2003, 2005), ELLWANGER et al. (2011a), VILLINGER (2011) als Hosskirch bezeichnet wird. In der Karte der Größten Rheinischen Vergletscherung GRG (Abb. 4) bilden alle äußersten glazialen Zeugen den Eisrand.

Der Außensaum der größten Vorland-Vereisung des Rheingletschers setzt im Westen dort ein, wo seit dem Endabschnitt der Deckenschotter-Eiszeiten die Erosionsbasis des gesamten Rheingletschergebiets lag, an der Klettgauer-Pforte bei Schaffhausen am Hochrhein (PENCK 1939). Die Peripherie dieser Vergletscherung schwingt sich in weitem Bogen über den Hegau und das oberschwäbische Donautal hinüber ins Illergebiet bis zum Bergland der Adelegg. Seit PENCK & BRÜCKNER (1909) wurde bis vor kurzem der nord-westliche Randstreifen dieser äussersten Vergletscherung bis zum Bussen ausnahmslos der Riss-Eiszeit zugeordnet. Ab dem Raum Biberach südostwärts bis zur Aitrach und zur Adelegg wurde hingegen postuliert, dass die äußersten Moränen in der Mindel-Eiszeit gebildet worden seien (Abb. 5). Erst Ellwanger (2003) verknüpfte alle äußersten glazialen Bildungen zu einem einzigen Eisrand unter der von ihm eingeführten Bezeichnung MEG.

4.1 Nordwestliches Rheingletschergebiet

Angesprochen ist der Geländestreifen von Messkirch bis Riedlingen beidseits der Donau, respektive von der Homburg bis zum Bussen. Detaillierte und auf eine Grosszahl von Bohrungen gestützte Daten aus dem Gebiet von Messkirch finden sich bei WERNER (1975, Beil. 8). Er stellt sämtli-

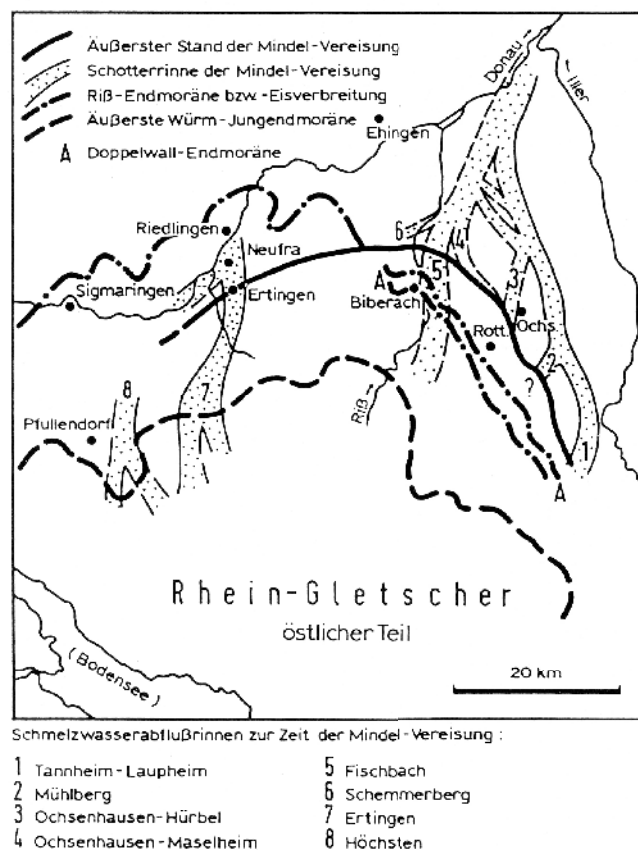


Abb. 5: Nordrand der Mindel- und der Riss-Vergletscherung nach Haag 1982. Beachte die Überkreuzung der Gletscherfronten.

Fig. 5: Northern margin of the Mindel- and the Riss-Glaciation according to Haag 1982. Notice the crossing-over of the glacier frontiers.

che glazialen Bildungen in die Riss-Eiszeit. Hinsichtlich der frühen „risszeitlichen“ Gletscherbewegungen unterscheidet er einen ersten Vorstoß auf so genannte Messkircher Stände und, nach einem Eisrückzug, den Hauptvorstoß zum äussersten Stand bei Vilsingen. Nach einer weiteren Rückzugsphase verbunden mit der Anlage des heutigen Ablachtals sowie der Entstehung des heutigen Donautals kam es zum Vorstoß auf den von ihm definierten Waldbühlstand (Abb. 6). Ausgewiesen durch eine Abfolge von Endmoränenkuppen zwischen Gallmannsweil und Messkirch wird diese Eisrandlage mit dem Biberacher Riss-Doppelwall korreliert (WERNER 1994). Gemäß diesem Szenario soll die bedeutende morphologische Umgestaltung des Donau-Ablach-Systems in einem Interstadial zwischen Zungen-Riss und Doppelwall-Riss sensu SCHREINER (1985, Tab. 6) erfolgt sein.

Auf die Interpretation von WERNER (1975) stützen sich die umfassenden Untersuchungen von VILLINGER (1985, 166-174, Abb. 4) zur Fluss- und Landschaftsgeschichte im Raum Sigmaringen-Riedlingen. Zwar wird dort in Abb. 4 ein Mindel-Maximum zwischen Scheer und Kanzach angedeutet, die grossen Ereignisse

- der Donau-Staubildungen
- der Verschüttung der altpleistozänen Donaurinnen,
- der Anlage eines neuen breiten Donaulaufes

bleiben jedoch auch hier der älteren bis jüngeren Riss-Eiszeit zugeordnet.

Die von SCHREINER (1980) detailliert untersuchten Ertinger Schotter (Abb. 8) wurden aufgrund des hangenden Pa-

Quartärgeologisches Sammelprofil Raum Messkirch

Nicht massstäblich, horizontal ca. 10 km, stark überhöht, Höhen korrekt

Grundlagen: Werner 1975; Geol. Karte und Erläuterungen Blatt Messkirch 1994

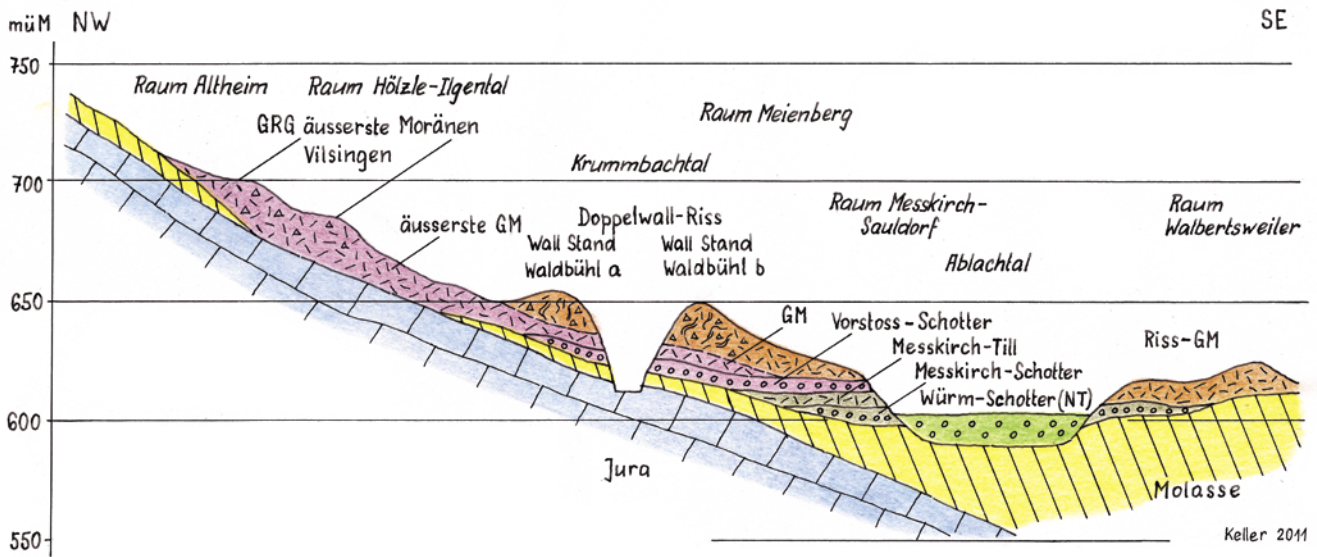


Abb. 6: Quartärgeologisches Sammelprofil der äußersten Randzone im Nordwesten des Rheingletschers bei Messkirch.

Fig. 6: Quaternary geological composite section of the most external margin in the northwest of the Rhine Glacier near Messkirch.

läobodens von Neufra und der abdeckenden Riss-Moräne als älter erkannt und demzufolge als mindelzeitlicher Deckenschotter angesehen. In der paläopedologischen Untersuchung des Bodens von Neufra (BIBUS & KÖSEL 2001) wird als Altersstellung eine mögliche mehrphasige Bildung während mehr als einem Interglazial vor dem Doppelwall-Riss in Betracht gezogen.

Der Schritt zur Erkenntnis, dass im äußersten, glazial überfahrenen Gebiet zwischen Messkirch und Riedlingen eine ältere Eiszeit vorliegt, beruht auf Beobachtungen in zwei Kiesgruben: Krauchenwies/Bittelschiess und Datthausen (ELLWANGER et al. 2011a, 404–407). In Krauchenwies findet sich unter mächtigen Abfolgen von Diamikten und Schottern eine Feinsedimentlage, in der pollenanalytisch ein Holstein-Interglazial nachgewiesen wurde (Abb. 12 in Kap. 5.2). An der Lokalität Datthausen stiess eine Bohrung unter einem 45 m mächtigen Stapel alpiner Schotter mit Rissmoränen im Top auf Rheingletscher-Till über Molassefels (Abb. 7). Diese Eiszeit älter als Riss bezeichnet ELLWANGER (2003) mit dem Namen MEG.

Diskussion und Interpretation

Aufgrund der Befunde in den Aufschlüssen Krauchenwies/Bittelschiess und Datthausen hat die Vergletscherung des GRG, gleichzusetzen mit „Hoskirch“, die älter als Riss einzustufen ist, zwischen Sigmaringen und Riedlingen die Donau überschritten und das äußerste Band glaziärer Ablagerungen hinterlassen (Abb. 4). Südwestlich Sigmaringen ziehen sich die äußersten Glazialrelikte in einem Band von flachen Kuppen hin, die abschnittsweise undeutlich in zwei Reihen angeordnet sind (Abb. 6). Bei der Homburg über der Hegau-Niederung schieben sich markantere, unzweifelhafte Riss-Wälle bis zur Außengrenze der glazialen Bildungen vor. Von hier an bleibt Riss deutlich intern und ist wie bei Biberach zu ei-

nem Doppelwallsystem entwickelt, das sich von Honstetten über Gallmannsweil und Messkirch bis gegen Krauchenwies als Waldbühl-Eisrandlage (WERNER 1975, 1994) verfolgen lässt (Abb. 6). Zwischen Mengen und Datthausen sind keine morphologischen Anhaltspunkte für eine Riss-Randlage auszumachen.

Von diesen Erkenntnissen ausgehend würden die von VILLINGER (1985) der Riss-Eiszeit zugeschriebenen Verschüttungen, Staubbildungen und Laufverlegungen der Donau im Raum Sigmaringen-Riedlingen in die Zeit des GRG fallen.

Das Profil des Ertinger Schotterbandes (SCHREINER 1980) kann zwanglos bis Datthausen verlängert werden, womit dort die liegenden Schotter dem Ertinger Schotter zuzuordnen wären (Abb. 8). Das Paket der hangenden Schotter würde damit als Vorstoß-Schotter der Riss-Eiszeit zu deuten sein, wobei die auflagernde Moräne durch den Rissgletscher abgelagert worden wäre. Der basale Diamikt in der Bohrung von Datthausen (Abb. 7) dürfte gemäß seiner nördlichen Lage dem GRG = Hoskirch angehören, was auch ELLWANGER et al. (2011a) annehmen. Dies legt nahe, dass der Ertinger Schotter im Größten Rheinischen Glazial beim etappenweisen Rückschmelzen abgelagert wurde.

Diese Vorstellung kann durch die den Strang der Ertinger Schotter kreuzende Mottschieser Rinne untermauert werden. Diese verläuft in SW-NE-Richtung von Pfullendorf Richtung Federsee und folgt einer Molasse-Störung. Sie weist ein geringes Gefälle von ca. 1‰ auf, was aus GK 25, Blatt 8021, Pfullendorf hervorgeht (VILLINGER 2003). Es ergibt sich, dass das S-N verlaufende Ertinger Schotterband westlich Saulgau von der Mottschieser Rinne, die jüngere Deckenschotter führt, gekreuzt wird. Ihre Rinnenbasis liegt östlich Pfullendorf auf 630 m ü. M., womit sie den Ertinger Schotter auf rund 615 m queren dürfte. Das Top der Ertinger Schotter reicht knapp bis auf dieses Niveau, die Schotterbasis

befindet sich auf ca. 590 m (Abb. 8). Daraus darf geschlossen werden, dass der Ertinger Schotter nicht mit den Deckenschottern der Mottschieser Rinne zu verbinden ist, sondern eine jüngere Schüttung darstellt, wofür das GRG in Frage kommt.

Im Weiteren verläuft das nach Süden als Extrapolation verlängerte Band der Ertinger Schotter um 50 m tiefer als der jüngere Deckenschotter von Königseggwald (Abb. 8). Demzufolge wären die Ertinger Schotter keine Deckenschotter, sie müssen jünger sein als diese, aber älter als Riss, was ihre Zuordnung zum GRG annehmen lässt.

In der Nordschweiz wird die größte Vergletscherung als Möhlin-Glazial bezeichnet (GRAF 2009a, KELLER & KRAYSS 2010, PREUSSER et al. 2011) und vor dem Holstein-Interglazial positioniert. Diese Eiszeit entspricht gemäß ihrer räumlichen Ausbreitung dem Größten Rheinischen Glazial im hier diskutierten nordwestlichen Gebiet des Rheingletschers. Aufgrund des Nachweises des Holstein-Interglazials in der Sedimentabfolge von Bittelschiess steht einer Gleichsetzung Möhlin = Hosskirch = Größtes Rheinisches Glazial (GRG) nichts im Wege.

4.2 Zur Bildung der tiefen Becken

Die frühpleistozänen Deckenschotter des westlichen Bodenseeraums sind durch eine Periode erhöhter Eintiefung von den mittel- und spätpleistozänen Ablagerungen getrennt (SCHREINER 1974, 1992; GRAF 1993, 2009b; KELLER & KRAYSS 1999, 2010). Neben Absenkungen im Oberrheingraben spielte die Umlenkung des Alpenrheins von der Donau zum Oberrhein die Hauptrolle, die im Anschluss an die Zeit der Tieferen Deckenschotter erfolgte (KELLER 2009). Sie brachte dem Hochrhein ab der Aare-Mündung nahezu die doppelte Wassermenge und damit verstärkte Erosionskraft. Für das Flusssystem des Alpenrheins bis zur Aare setzte eine äußerst intensive fluviale Ausräumung des Bodensee-Hochlandes mit dem Resultat einer tiefen Durchtalung ein.

Waren die Gletscher der Deckenschotter-Eiszeiten im Bodensee-Vorland noch auf das hoch liegende, frühpleistozäne danubische Entwässerungsnetz ausgerichtet, so folgten nun die mittelpleistozänen Eisströme den vorgezeichneten tiefen Flusstälern. Durch subglaziale Erosion verbreiterten sie die Talzüge und schürften übertiefte Becken und Tröge aus, weshalb diese Glaziale im Rhein- und Aareraum als Becken-Eiszeiten bezeichnet werden (KELLER & KRAYSS 1999; PREUSSER et al. 2011).

Häufig findet sich in den tiefen Becken über dem Felsboden Moräne. In den hangenden, warmzeitlichen Sedimenten konnte mehrfach das Holstein-Interglazial direkt nachgewiesen oder zumindest wahrscheinlich gemacht werden. Alpennah betrifft dies beispielsweise das Aaretal südöstlich Bern mit dem Thalgut (SCHLÜCHTER 1989; WELTEN 1982) oder das Linthbecken mit dem Günterstall-Delta (JEANNET 1923; SCHINDLER 2004; KELLER & KRAYSS 2010). Alpenfern bildet das Singener Becken den westlichsten Teil des Bodenseebeckens. Die dem Felsgrund aufsitzenden Diamikte gehören ebenfalls der größten Vergletscherung an (SZENKLER & BOCK 1999), indem typische Pollen für Samerberg-Holstein nachgewiesen wurden. Die Becken selbst müssen daher vorwiegend in der ersten und grössten Becken-Eiszeit, dem

Kiesgrube Maucher bei Datthausen Aufschlussprofil und Bohrprofil

Grundlagen: Ellwanger et al. 2011a

Feldaufnahmen Keller und Krayss

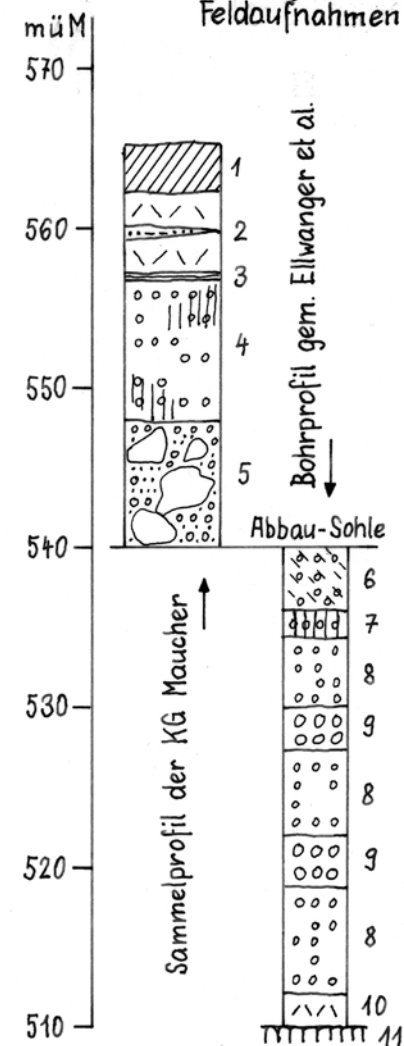


Abb. 7: Quartärgeologisches Profil Datthausen (Kiesgrube und Bohrung) im äußersten Norden des Rheingletschergebietes. Auswertung siehe Tabelle.

Fig. 7: Quaternary geological section of Datthausen (gravel pit and drilling) in the most northern part of the Rhine Glacier. Analysis see table.

Größten Rheinischen Glazial GRG = Möhlin = Hosskirch, ausgeschürft worden sein.

Auch die an die Hauptbecken nach außen ins weitere Vorland vorgeschobenen Neben- und Zweigbecken sind überwiegend im Größten Rheinischen Glazial von den von den Eiskörpern der Hauptbecken ausstrahlenden Gletscherzungen ausgeschürft worden. Belegt wird dies in den Becken von Hosskirch oder Wattenweiler durch Funde von Pollen, die auf das Holstein-Interglazial schliessen lassen (ELLWANGER et al. 1995; BLUDAU in ELLWANGER et al. 1999).

4.3 Raum Risstal (Biberach) – Aitrachtal

Ab Biberach galt bisher für den äußersten Eisrand das Konzept von PENCK & BRÜCKNER (1909), das diesen Mindel zuwies. Der Geländestreifen dieser größten Vergletscherung

Auswertung Profil Datthausen [Abb. 7]

	Lithologie	Genetische Interpretation	Glazialchronologie gemäß ELLWANGER et al. 2011a	Glazialchronologie gemäß Autor
1	Verwitterungszone Bodenbildung	Ältere Verwitterung und Paläoboden, bis 3 m mächtig		Eem-Interglazial und rezenten Boden
2	Moräne Rheingletscher, im E zweigeteilt	Eisüberfahung, Gletscher-Vorstoß mit Oszillation	Riss-Eiszeit, Maximalvorstoß Außenwall-Riss	Riss-Glazial, Maximalvorstoß bis Datthausen
3	siltig-tonige Feinsedimente	lokale Stausedimente		Tümpel vor der Riss-Eisfront
4	Kies mit alpinem Material, teils verkittet	Vorstoß-Schotter, örtlich durch Kalkkrusten verfestigt	Riss-Eiszeit, Hochterrassenschotter	Riss-Gletschervormarsch, Vorstoß-Schotter
5	Grobblock-Horizont aus Oberjura & Tertiär, kiesreiche Matrix	Katastrophen-Ereignis: Bergsturz und See-Ausbruch? Flutwelle aus der Alb	Prä-Riss?	Prä-Riss, interglaziales Ereignis oder während Habsburg-Glazial?
6	Kiese mit alpinen Geröllern, Entkalkung	Entkalkung: Hinweis auf Sedimentationspause		Interglazial? Entkalkung als Unter- bruch in der Ablagerung
7	verkittete Kiese, alpin zurückschmelzenden	Schotter eines etappenweise	Riss-Hochterrasse oder älter in Etappen während	GRG Mindel-Schotter, Schüttung
8	Kiese mit alpinem Material	Gletschers		dem Abschmelzen
9	Grobkiese mit Alpin und Oberjura	Begrenzte Wiedervorstöße angezeigt durch Grobmaterial		
10	Till Rheingletscher mindestens Datthausen	Gletschervorstoß bis oder Älteres Riss	Hosskirch-Vorstoß? äußerste Glazialgrenze	GRG: Gletscher-Vorstoß bis
11	Untere Süßwassermolasse			GRG Mindel-Glazialwanne

ist dadurch gekennzeichnet, dass er intern durch den äußeren Moränenwall des Doppelwall-Riss und das Zungenriss morphologisch eindeutig überfahren ist (SCHREINER & EBEL 1981; HAAG 1982 Abb. 10). Als bedeutende Typuslokalität innerhalb des Mindel-Geländestreifens gelten die zum Illertal gerichteten Querprofile von Hauerz und Seibranz (SCHREINER 1996, Fig. 2–5). Hier demonstrierten SCHREINER & EBEL (1981) die Einführung der Haslach-Eiszeit, einer im Penck'schen System neuen Eiszeit zwischen Günz und Mindel. In der Bohrung von Unterpfaufenwald belegt eine pollenanalytisch gesicherte Interglazialschicht (GÖTTLICH & WERNER 1974) ein älteres Interglazial als Holstein (BLUDAU gemäss SCHREINER 1992). Sie trennt die Mindelmoräne im Hangenden von der Haslachmoräne im Liegenden. Dieses Interglazial dürfte dem Cromer Komplex angehören (BIBUS et al. 1996). Unter dieser finden sich nach einer Erosionsdiskordanz und einem Paläoboden Schotter, die als Zeiler Schotter (= Günz) angesprochen werden. Daraus ergab sich die Eiszeitenabfolge Günz – Haslach – Mindel – Riss, wobei Günz als Ältere, Haslach und Mindel als Jüngere Deckenschotter verstanden werden.

Zur Entwässerung des Mindel-Glazials im nordöstlichen Rheingletschergebiet liegt die fundierte Untersuchung von HAAG (1982) vor. In seiner Abb. 1 lassen sich zwei Typen von Schottersträngen erkennen. Ein System (1) verweist offensichtlich auf die Eisrandentwässerung während dem Maximalstand der Mindel-Vergletscherung. Schotterstränge des Typs (2) hingegen haben ihren Ursprung am Südrand des oberschwäbischen Plateaus und beziehen sich somit auf in-

terne Stände der Vergletscherung. Dem Höchststand entsprechen (1) die Schemmerhofer Schotter, der Ochsenhausen-Maselheimer und -Hürbeler Schotter sowie sehr ausgeprägt der Tannheim-Laupheimer Schotter. Zum Typ (2) gehören der Fischbachschotter sowie der Mühlberger Schotter. Von allen diesen Abflussbahnen finden sich bei HAAG (1982, Abb. 4) detaillierte Längsprofile. Speziell auf den Typ (2) verweisen die Ertinger Schotter (SCHREINER 1980), die sich aus dem Raum Saulgau-Hosskirch bis zur Donau bei Datthausen erstrecken (Abb. 8 in Kap. 4.1).

Über die Verhältnisse im oberen Aitrachtal gibt die GK Leutkirch-Ost (WENNINGER 1994, Abb.1) einen guten Einblick. Offensichtlich reichte die Mindel-Vergletscherung als ansgedehnteste Vereisung bis Altmannshofen und südlich bis auf die Linie der heutigen Hofser Ach, wo sie auf die Ausläufer des Adelegg-Berglandes stieß. Demnach lag bei Altmannshofen auch das Gletschertor, an dem die Tannheim-Laupheimer Schotter ihren Ursprung hatten (Abb. 9).

Diskussion und Interpretation

Im Raum Ristal-Aitrachtal ist gemäß den dargelegten Befunden die größte Vergletscherung mit den äußersten Glazialzeugen älter als Riss. Im Nordosten des Rheingletschergebiets sowie im Nordwesten ist aufgrund der bisherigen Erkenntnisse und Darlegungen nur eine einzige Vergletscherung ausgedehnter als Riss nachweisbar. Für den Nordwesten konnte glaubhaft gemacht werden (Kap. 4.1), daß es sich dabei um das Größte Rheinische Glazial GRG = Hosskirch handelt, das sich vor Riss, aber nachfolgend zu den Decken-

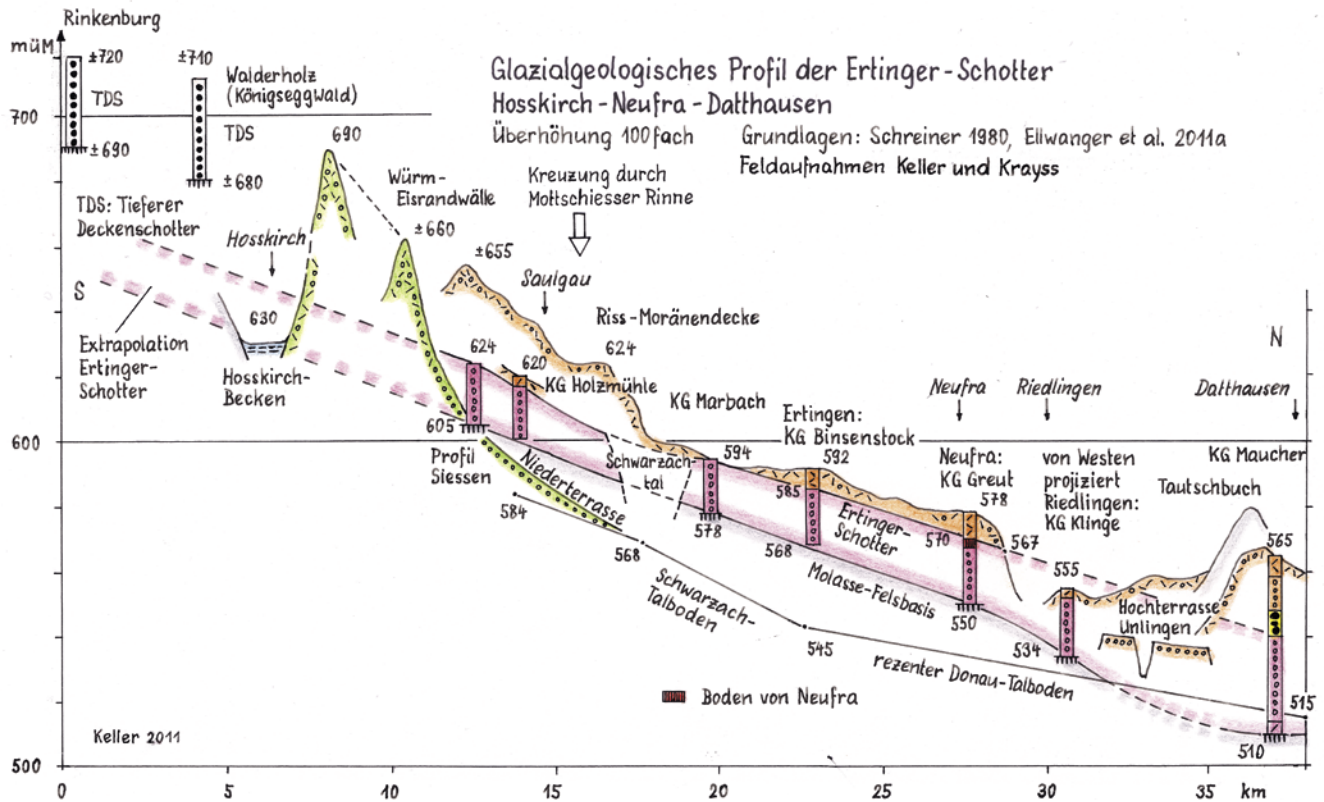


Abb. 8: Glacialgeologisches Längsprofil der Ertinger Schotter zwischen Hosskirch und Dalthausen (nördlicher Rheingletscher), ergänzt nach Schreiner 1980.
 Fig. 8: Glacial geological longitudinal profile of the Ertinger gravel between Hosskirch and Dalthausen (northern Rhine Glacier), completed after Schreiner 1980.

schottern einschiebt. Falls diese Befunde akzeptiert werden, ergeben sich daraus für den Nordosten des Rheingletschers schwerwiegende Konsequenzen:

1. Es muss wohl davon ausgegangen werden, dass auch im Nordosten die äussersten Vergletscherungszeugen dem GRG = Hosskirch angehören. Sie wurden bisher Mindel zugewiesen, das damit hier zum GRG wird.

2. SCHREINER & EBEL (1981) sowie HAAG (1982) konnten zeigen, dass der Hauptstrang der Entwässerung, der Tannheim-Laupheimer-Schotter im Süden im Aitrachtal mit Moräne verzahnt ist, was die Schotter als Mindel-Schüttungen ausweist. Wenn die Mindel-Moränenwälle dem GRG zugeordnet werden, so ist daraus zu schließen, dass die zugehörigen bisherigen Mindel-Schotterstränge zu GRG-Sandern werden.

3. Da das GRG eine Becken-Eiszeit ist (Kap. 4.2), die nachfolgend nach den Deckenschottern einzustufen ist, so muss hier im Nordosten Mindel zur Becken-Eiszeit werden, womit es jünger wäre als die Deckenschotter.

4. Die bisherigen Mindel-Schotterstränge im Nordosten des Rheingletschers würden zu Schüttungen des GRG. Sie wären damit keine Deckenschotter, da diese ihrerseits älter sind als die Becken-Eiszeiten. Die jüngeren Deckenschotter würden sich somit auf Haslach beschränken.

Welche Folgen sich aus diesen Interpretationen für den östlich anschliessenden Illergletscher ergeben würden, kann in diesem Zusammenhang nicht abgeschätzt werden.

4.4 Ausmasse der Vorlandvergletscherung

Für den Versuch der paläoglazialen, kartographischen Re-

konstruktion des Vorlandgletschers im Größten Rheinischen Glazial GRG (Abb. 4) wurden Eisrandmarken am Außensaum des Gletschers und höchste glaziale Spuren in den Voralpen verwendet. Die alpenwärtige Begrenzung des Vorlandgletschers ist festgelegt durch die dem tektonischen Alpenrand vorgelagerten hohen Molasse-Berge, die das Eisstromnetz noch überragten. Im Rheintal ist das Alpentor von Oberriet-Koblach massgebend. Das Isohypsenbild beruht auf diesen Marken sowie auf Analogieschlüssen zum Rhein-Linth-Gletscher während der Würm-Eiszeit (KELLER & KRAYSS 1982, 1993, 2005). Für die Vorlandvergletscherung des GRG ergibt sich ein plausibles Bild, das den Eisflussrichtungen aus den Alpen heraus und der Geometrie des Eiskörpers gerecht wird.

Methodisch standen als Grundlage die Berechnungen und Abschätzungen zur Würm-Vergletscherung zur Verfügung (KELLER & KRAYSS 1993), die von BENZ (2003) durch digitale Datenerfassung und Modellierung im GIS bestätigt worden sind. Für die Volumenberechnung wurden für 260 Punkte in einem 5x5 km² Raster die Höhendifferenzen zum Eiskörper des Würm-Maximums bestimmt. Bei Rasterpunkten, die in ein dem GRG zugewiesenes Becken fielen, wurden die Höhen des Felsbodens für die Volumenberechnung mit einbezogen, dies unter der Annahme, dass die Beckenausäumung sich im Wesentlichen im GRG abgespielt hat (Kap. 4.2). Die über den Würm-Eisrand hinaus reichenden Areale wurden mit einem Flächenraster von 2,5 x 2,5 km² ausplanimetriert. Zusätzlich musste das Eisvolumen ausserhalb der Würm-Grenze bestimmt werden.

Auf diesen Grundlagen basierend konnten Fläche und Volumen des Vorlandgletschers bestimmt werden. Resultate:

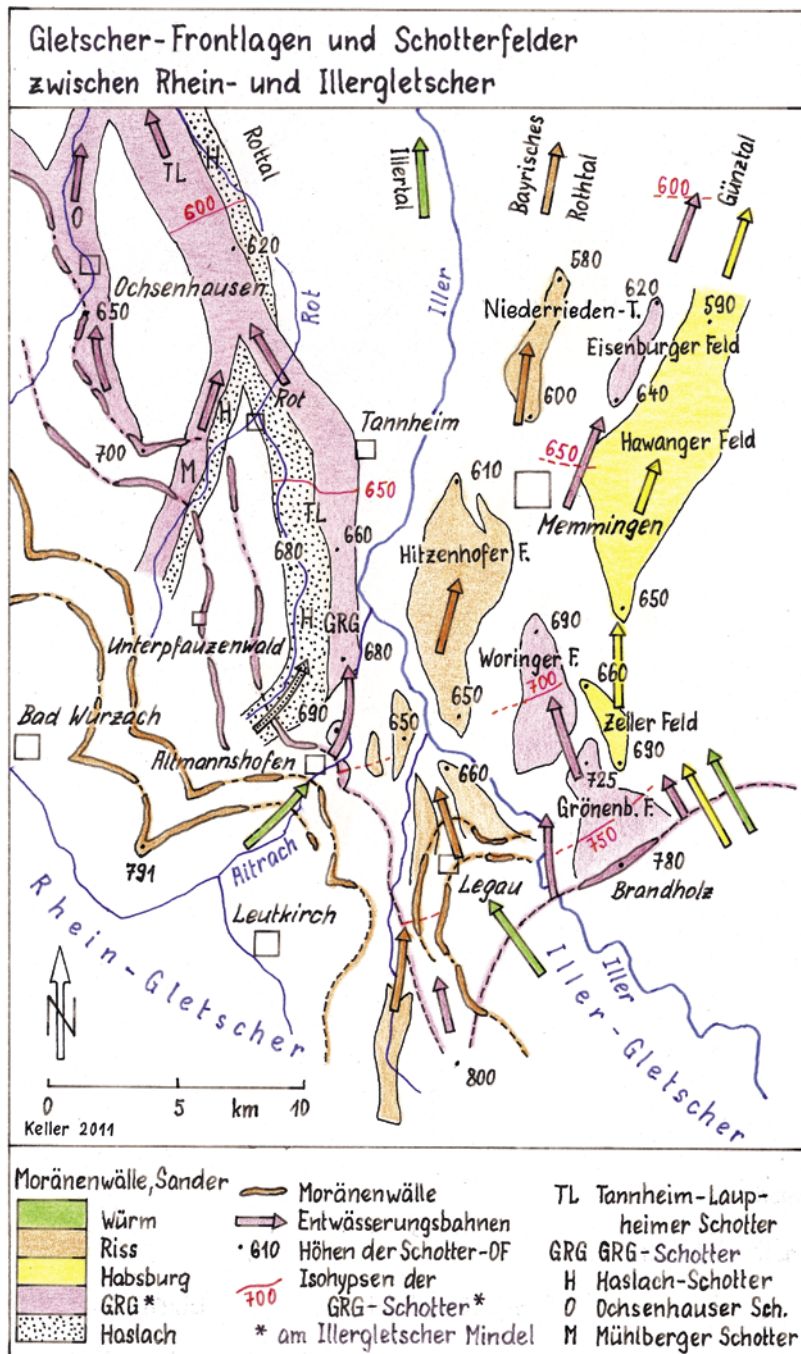


Abb. 9: Gletscherfronten, Sander und Entwässerung zwischen dem nordöstlichen Rheingletscher und dem Illergletscher.

Fig. 9: Frontiers of glaciers, outwash plains and drainage between the northeastern Rhine Glacier and the Iller Glacier.

Vorlandvergletscherung
des Größten Rheinischen
Glazials = Hosskirch

Zum Vergleich die
Vorlandvergletscherung
des Würm-Maximums

Vergletschertes Areal

9000 km² (125% zu Würm)

7200 km²

Volumen des Eiskörpers

3300 km³ (137% zu Würm)

2400 km³

Mittlere Eismächtigkeit

370 m

335 m

5 Modell eines Habsburg-Glazials

Für die Nordschweiz konnte plausibel gezeigt werden, dass sich zwischen das Größte Rheinische Glazial (GRG) = Möhlin und die Riss-Eiszeit = Beringen eine Vorlandvergletscherung einschleibt, die in etwa die Ausmaße der Würm-Vergletsche-

rung erreichte und von deren Frontpositionen aus Schmelzwasserströme die Hochterrassen schütteten. Diese als Habsburg-Glazial bezeichnete drittletzte Eiszeit ist in den übertieften Becken in den Sedimentstapeln mehrfach zu erkennen (GRAF 2009a, KELLER & KRAYSS 2010, PREUSSER et al. 2011).

Damit stellt sich die Frage, ob nicht auch im übertieften Bodenseebecken und in seinen Randgebieten das Habsburg-Glazial erkennbar ist. In diesem Sinne wird eine Modellvorstellung entwickelt und anschließend versucht, Indizien, die für dieses Glazial sprechen, beizubringen.

5.1 Ausdehnung und Entwässerung

In der Karte der Abb. 10 sind zu Vergleichszwecken die in der Nordschweiz anhand der Ansatzstellen der Hochterrassen rekonstruierten Zungenlagen der Habsburg-Vergletscherung (KELLER & KRAYSS 2010) für den Reuss- und den Linthglet-

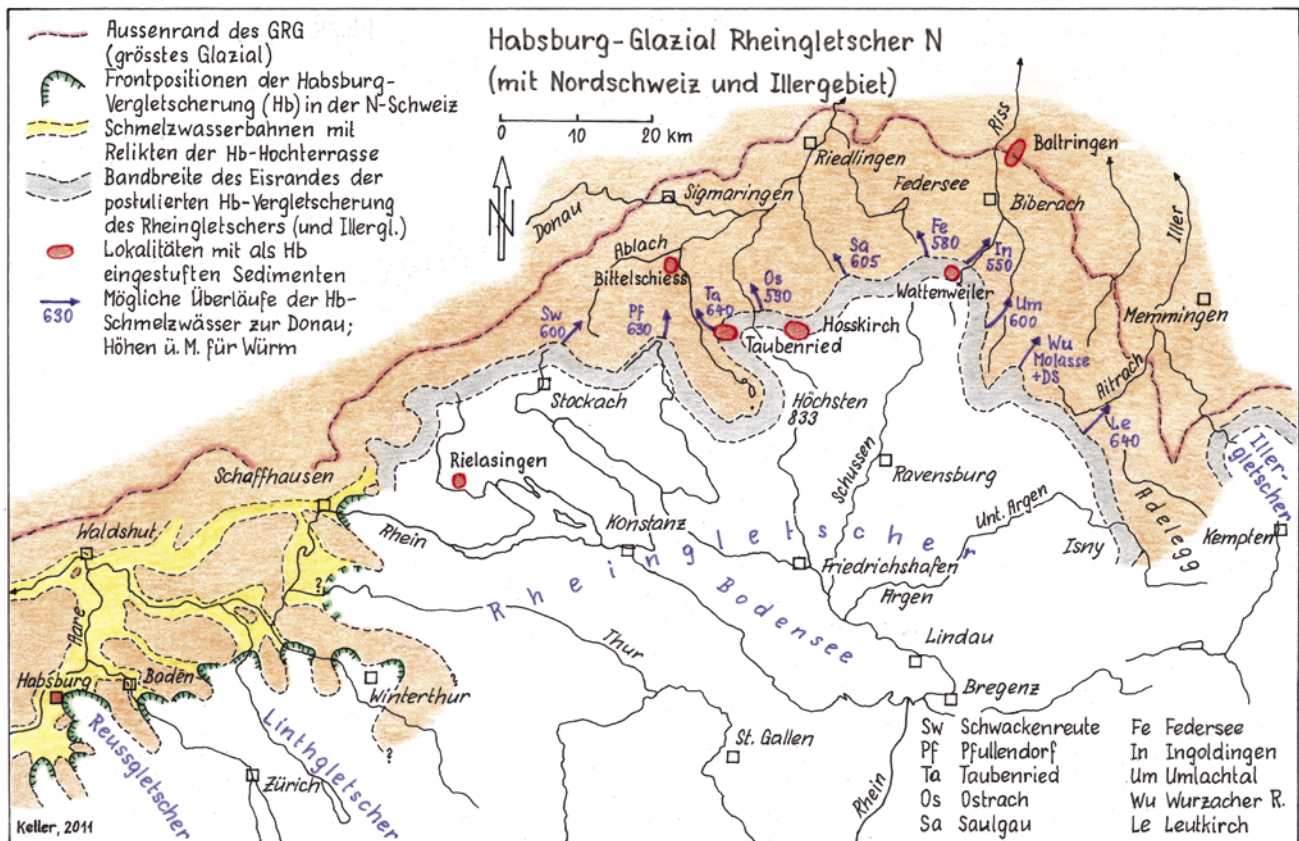


Abb. 10: Geschätzte Eisausdehnung der postulierten Habsburg-Vergletscherung.

Fig. 10: Estimated ice extension of the postulated Habsburg glaciation.

scher, bei Schaffhausen auch für den westlichen Rheingletscher, eingetragen. In Erkenntnis, dass dieses Frontenbild sehr gut mit der Würm-Vergletscherung vergleichbar ist, ist im nördlichen Rheingletschergebiet modellmäßig und basierend auf der Würmausdehnung ein Bandstreifen eingezeichnet, in dem allenfalls Indizien für eine Habsburg-Vergletscherung zu erwarten sind. Beim Illergletscher ist in der Karte Abb. 10 ebenfalls das für eine Habsburg-Vergletscherung in Frage kommende Band für den Frontbereich dargestellt.

Im Gegensatz zur Nordschweiz, wo die Entwässerung nach Westen vorgegeben ist, müssen beim nördlichen Rheingletscher die Eisfronten die Wasserscheide zur Donau mindestens erreichen, damit Schmelzwässer nach Norden abfließen können. Dabei ist auch zu bedenken, dass in einem Glazial dieser Ausdehnung nur während den Hochständen Richtung Donau Schotter durch Schmelzwasserströme abgelagert werden können. Die für eine Habsburg-Vergletscherung in Frage kommenden Überlaufschwelen sind in der Abb. 10 eingefügt.

Die nach Norden gerichteten Überläufe der würmeiszeitlichen Vergletscherung wurden, ausgenommen Ostrach und Riss bei Ingoldingen, nur im Würm-Maximum als der Donau zustrebende Entwässerungswege benutzt. Es sind in der Anlage ältere Rinnen, die vermutlich zum Teil schon im GRG funktionierten. Die in der Karte der Abb. 10 eingetragenen Schwellenhöhen sind würmzeitlich und dürften im GRG noch um Einiges höher gelegen haben, wie es das Südende der Ertinger Schotter nahe legt (Abb. 8).

Im Osten existiert auch noch heute aus dem Bodensee-becken heraus nur der Überlauf der Aitrach zur Iller, der

während der Würm-Eiszeit aktiv war. Hoch liegende risszeitliche Moränenwälle nördlich und südlich der Aitrach bei Altmannshofen deuten an, dass der Abfluss zur Iller risszeitlich noch hoch lag und erst würmzeitlich tief ausgestaltet worden ist. Somit steht diese Schmelzwasserbahn für ein allfälliges Habsburg-Glazial außer Betracht.

Im mit 150 m sehr tiefen Wurzacher Becken, das sich weit nach Nordosten erstreckt, sind in den Beckensedimenten Würm und Riss nachgewiesen GRÜGER & SCHREINER (1993). Es dürfte daher im Zuge älterer Glaziale ausgeschürft worden sein. Es scheidet aber, da nordöstlich anschließend Höhenrücken mit Deckenschottern vorhanden sind, als Entwässerungsweg für eine angenommene Habsburg-Vergletscherung zum Vorneherein aus.

Ausgehend vom Austritt der Unteren Argen aus dem Adelegg-Bergland kann eine hypothetische Randentwässerung der Habsburg-Vergletscherung entlang dem Band des Modells (Abb. 10) angenommen werden. Sie setzt in dem in Molasse angelegten Tal der Unteren Argen an, das einen flachen Talboden mit geringer Lockermaterialbedeckung bei 4‰ Gefälle aufweist. Bis zum Riss-Überlauf bei Ingoldingen beträgt das Gefälle 3,5‰. Die Fortsetzung Risstalabwärts zum Niveau der Baltringer Hochterrasse reduziert sich auf 2,5‰. Werte dieser Größenordnung weisen auch würmzeitlich gesicherte Täler auf. Dieses Randsystem ist so ausgeprägt, dass es zeitlich weiter als Würm in die älteren Becken-Eiszeiten zurückreichen dürfte, dabei aber noch etwas höher gelegen hätte.

Aus dem Modell abgeleitet kann zusammenfassend für eine angenommene Habsburg-Eiszeit höchstens mit sechs

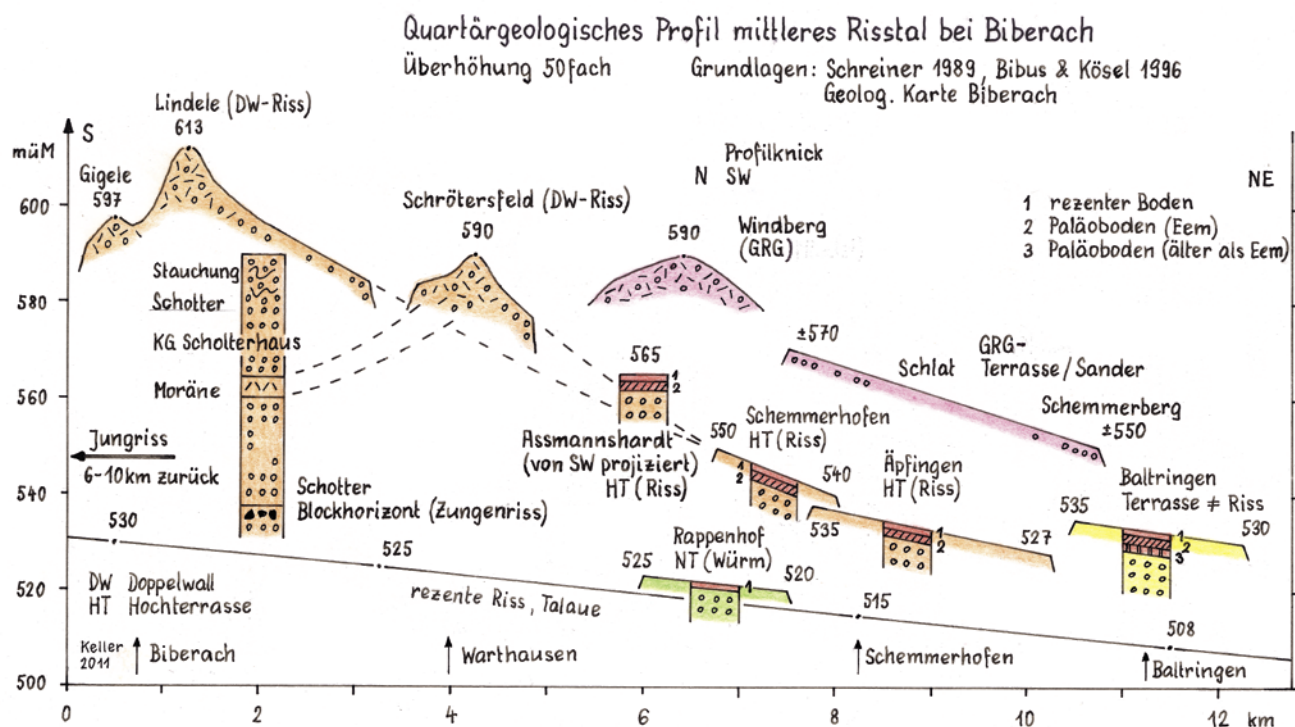


Abb. 11: Profil durch die Frontpositionen der GRG- und der Riss-Vergletscherung im Zusammenhang mit den Sanderresten im Raum Biberach (teilweise nach SCHREINER 1989).

Fig. 11: Longitudinal profile through the front positions of the GRG- and the Riss-glaciation in context with the relics of outwash plains in the region of Biberach (partly after SCHREINER 1989).

möglichen Schmelzwasser-Überläufen, alle zur Donau nach Norden ausgerichtet, gerechnet werden (Abb. 10).

5.2 Mögliche Relikte im Donau-wärtigen Bereich

Um einer vorerst modellmäßig postulierten Habsburg-Eiszeit auf die Spur zu kommen, sind einerseits periglazial mögliche Aufschotterungen und/oder Terrassenrelikte zu diskutieren. Andererseits muss intraglacial im beckenseitigen Randbereich der anzunehmenden Habsburg-Vergletscherung (Bandstreifen in Abb. 10) oder im Inneren des Bodenseebeckens nach Indizien gesucht werden. In diesem Sinne wurde versucht die zur Verfügung stehenden quartärgeologischen Profile, bearbeiteten Aufschlüsse und Bohrungen zusammen mit den vorhandenen Bodenbestimmungen und Pollenanalysen neu zu interpretieren. Dabei ist festzuhalten, dass die bisherigen sorgfältigen Bearbeitungen als Grundlage verwendet und als solche in keiner Weise angezweifelt werden.

Risstal-Biberach

Seit Penck (PENCK & BRÜCKNER 1909) werden die Frontmoränen bei und südöstlich von Biberach mit der Hochterrasse verknüpft und gelten als Riss-Glazial. Im Gebiet Biberach im Norden und der Aitrach im Süden sind drei Riss-Eisrandlagen gefunden worden (SCHREINER & HAAG 1982, SCHREINER 1985, 1989). Detaillierte Untersuchungen haben SCHREINER (1992) zur Unterscheidung von Zungenriss, Doppelwall-Riss und Jungriss veranlasst, die er durch Interstadiale voneinander abgetrennt hat. In seiner späteren Publikation (SCHREINER 1996) bleibt offen, ob es sich eventuell um Interglaziale handeln könnte, die aber nicht belegt sind. Stratigraphisch wird dem Zungenriss (Älteres Riss) keine Terrasse zugeord-

net, dem Doppelwall-Riss (Mittleres Riss) die obere Hochterrasse: Assmannshardt, Schemmerhofen, Baltringen, dem Jungriss die Untere Hochterrasse: Äpfingen oder 13 m-Terrasse (Abb. 11). Im Gegensatz dazu wurden früher (PENCK & BRÜCKNER 1909) die Frontmoränen des Doppelwall-Riss mit der Unteren Hochterrasse von Äpfingen verknüpft.

Paläopedologische Forschungen (BIBUS & KÖSEL 1996, 2001) erbrachten auf der 13 m-Terrasse von Äpfingen (Untere Hochterrasse) einen warmzeitlichen Paläoboden, weshalb die Schotter der vorletzten Eiszeit, dem als eigenständiges Glazial betrachteten Jungriss, zugewiesen wurden. Hingegen wird die Baltringer Terrasse mit zwei Paläoböden als Obere Hochterrasse ins Doppelwall-Riss gestellt, das daher durch ein Interglazial vom Jungriss zu unterscheiden wäre. Miara (MIARA et al. 1996) fand auf der Baltringer Terrasse einen Horizont mit intensiver Kiesverwitterung und davon getrennt auch einen fossilen Boden. Dementsprechend nimmt er für die Schüttung der Terrasse die drittletzte Eiszeit an.

Diskussion und Interpretation

Die paläopedologischen Untersuchungen von BIBUS & KÖSEL (1996) ergaben für die Deckschichten und Bodenbildungen zwischen der Unteren Hochterrasse von Äpfingen und den Oberen Hochterrassen von Assmannshardt und Schemmerhofen keine Unterschiede, sodass anzunehmen ist, dass alle derselben Eiszeit angehören. Gemäß der Höhenlage können auch morphostratigraphisch alle drei Terrassen als Hochterrassen dem Doppelwall-Riss zugeordnet werden (Abb. 11). Demnach dürfte die Terrasse von Äpfingen nicht mehr mit dem Jungriss zu verknüpfen sein. Dabei wird davon ausgegangen, dass Jungriss keine eigenständige Vergletscherung darstellt. Dieses bekundet wohl eine interne, stadiale

Quartärgeologisches Sammelprofil der Kiesgrube Bittelschiess/Krauchenwies südlich Sigmaringen

nicht massstäblich Grundlage Bibus & Kösel 1996

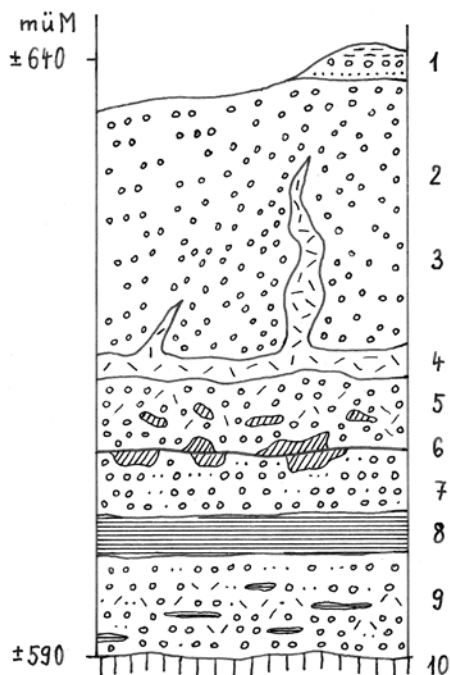


Abb. 12: Geologische Abfolge in der Kiesgrube Bittelschiess im Randbereich der Riss-Vergletscherung südöstlich Sigmaringen (umgezeichnet nach BIBUS & KÖSEL 1996). Auswertung siehe Tabelle.

Fig. 12: Geological record of the Bittelschiess gravel pit near the margin of the Riss glaciation southeast of Sigmaringen (re-drawn after BIBUS & KÖSEL 1996). Analysis see table.

Eisrandlage des Doppelwall-Riss. Riss wird in dieser Arbeit generell als ein einziges Grossglazial aufgefasst, indem zwischen Zungenriss und Doppelwall-Riss sowie zwischen Doppelwall-Riss und Jungriss Interstadiale angenommen werden (SCHREINER 1992). Von Interglazialen, wozu keine Nachweise vorhanden sind, wird abgesehen.

Wie Schemmerhofen und Assmannshardt wird die Baltringer Terrasse (Abb. 11) von SCHREINER (1996) morphostratigraphisch dem Doppelwall-Riss als Obere Hochterrasse zugeordnet. Da wie oben dargelegt die Äpfinger Terrasse (13 m-Terrasse) als Obere Hochterrasse aufgefasst werden kann, wird die Baltringer Terrasse um einen Glazialzyklus älter. Diese Auffassung wird durch die Untersuchungen von BIBUS & KÖSEL (1996, 2001) sowie von MIARA et al. (1996) bestätigt, indem zwei fossile Böden, repektive ein fossiler Boden und eine Kiesverwitterung übereinander bestimmt werden konnten. Bei Baltringen kann unter Annahme von zwei Paläoböden der obere als Eem-zeitlich, der untere als um ein Interglazial älter angesehen werden. Der Schotterkörper müsste dementsprechend eine Eiszeit älter sein als Riss, wobei aber dazu keine Eisrandlage im Risstal bekannt ist. Das ältere Riss scheidet als stadialer Eisrand von Riss und nicht eigenständiges Glazial aus.

Auffallend ist, dass gemäss der geologischen Übersichtskarte des Iller-Riss-Gebiets (WEIDENBACH et al. 1988) nördlich von Baltringen keine oberen Hochterrasen vorkommen, hingegen die 13 m-Terrasse des Jungriss bis Ulm deutlich zu verfolgen ist. Würde diese Terrasse dem Doppelwall-Riss zugeordnet, wie oben dargelegt, ergäbe sich eine der Mächtigkeit und Bedeutung des Doppelwall-Riss entsprechende Sander-Schüttung. Baltringen wäre dann ein Rest der nicht vollständig ausgeräumten oberen Hochterrasse eines um eine Eiszeit älteren Glazials.

Die zugehörige Vergletscherung hat offenbar die Über-

Auswertung Profil KG Bittelschiess [Abb. 12]

Lithologie	Genetische Interpretation	Glazialchronologie gemäß BIBUS & KÖSEL 1996 ELLWANGER et al. 2011a	Glazialchronologie gemäß Autor
1 Rinnenkies, Feinsand, Silt	Toteisloch-Füllung, Formumkehr		Riss-Spätglazial
2 Schotter, z. T. gestaucht	Eisrandnahe Schotter, glazial-tektonisch gestört	Doppelwall-Riss [Innenwall]	Riss-Glazial, jüngerer Teil [Innenwall-Riss?]
3 in Kiese eingearbeitete Diamikte	durch Eisdruck hochgepresste Moräne	Paulter Schwankung	
4 Moräne, Diamikt	basale Moräne, Eisüberfahung	Doppelwall-Riss [Außenwall]	Riss-Glazial, älterer Teil [Außenwall-Riss?]
5 Schotter mit Bodenfetzen und Diamikten	eisrandnah, eingearbeitete Bodenteile und Diamikte		Vorstoß Riss-Gletscher
6 Diskordanz mit Paläobodenresten	Bodenrelikt, diskordant abgesichert		IG Meikirch
7 Schotter und Sande	Proglaziale, eisferne Schüttung	Älteres Riss	Schotter Habsburg-Glazial
8 Warvenartige Beckentone	Beckensedimente, warmzeitl. Flora	IG Holstein	IG Holstein
9 Schotter mit eingearbeiteten Diamikten und Lehmen	Basale Ablagerungen bei Gletscherpräsenz	Hoskirch-Glazial	GRG Mindel [Größtes Rheinisches Glazial]
10 Molasse, USM	Glazial bearbeiteter Felsboden		Felsüberprägung im GRG Mindel

laufschwelle aus dem Bodenseebecken zur Donau nicht überschritten, hingegen könnten Schmelzwässer nach Norden durch das Rissstal abgeflossen sein (Kap. 5.1). Dabei wäre es zur Schüttung eines Hochterrassensystems gekommen, dessen von späterer Erosion verschonter Rest sich in der Hochterrasse von Baltringen erhalten hat (Abb. 11). Diese drittletzte Eiszeit hätte gemäß den Erkenntnissen im Raum Schaffhausen (GRAF 2009a, KELLER & KRAYSS 2010) in etwa die Ausmaße von Würm erreicht und könnte als die diskutierte Habsburg-Vergletscherung angesehen werden. Da sie den Beckenrand nicht überschritt und ihre glazialen Bildungen von den nachfolgenden Riss-Gletschern weitflächig überfahren wurden, konnte sie im nördlichen und östlichen Rheingletschergebiet bisher nirgends direkt erkannt und nachgewiesen werden. In der Nordschweiz hingegen ist sie fassbar und in ihren Frontlagen rekonstruierbar (Abb. 10). Ihr nachgeschaltet ist das Interglazial von Meikirch anzunehmen (PREUSSER et al. 2005), dem der untere Boden von Baltringen zugewiesen werden könnte.

Im Biberach-Rissgebiet müsste gemäß diesen Erkenntnissen und Überlegungen folgende Glazial-Gliederung angenommen werden (Abb. 11):

<u>Bisher</u>	<u>Neu</u>	<u>Terrassenstratigraphie</u>	<u>Interglaziale</u>
Würm	Würm	Niederterrasse	Eem: Paläoböden von Äpfingen und Schemmerhofen, oberer von Baltringen
Riss: Jungriess Doppelwall-Riss Zungenriess	Riss	Untere Hochterrasse: 13 m-Terrasse Äpfingen sowie Obere Hochterr., z. B. Schemmerhofen	Meikirch: Unterer Paläoboden von Baltringen
---	Habsburg	Obere Hochterrasse: Terrasse von Baltringen	

Krauchenwies/Bittelschiess

Die große Kiesgrube Bittelschiess bietet seit Jahrzehnten Einblicke in eine sehr differenzierte Sedimentabfolge bei einer Abbautiefe von über 50 m. Sie wurde insbesondere von ELLWANGER (1990), ELLWANGER et al. (1995, 1999) und BIBUS & KÖSEL (1996, 2001) eingehend untersucht. In ihren Publikationen kamen die Bearbeiter zum Schluss, dass sich in den Ablagerungen die Mehrphasigkeit von Riss, untergliedert durch Interglaziale oder Interstadiale manifestiere. Aus Untersuchungen von Bludau (schriftl. Mitt. in BIBUS & KÖSEL 1996) geht hervor, dass das kräftige Band der Beckentone ein vollständiges Interglazial umfasst, gekennzeichnet durch *Pterocarya* und *Buxus*.

Diskussion und Interpretation

Das in BIBUS & KÖSEL (1996) in ihrer Abb. 10 dargestellte Gesamtprofil kann aufgrund der Einführung des bisher im nördlichen Bodenseeraum nicht festgestellten Habsburg-Glazials sowie des Meikirch-Interglazials (KELLER & KRAYSS 2010) neu interpretiert werden (Abb. 12).

Das in den Beckentonen identifizierte Interglazial ist mit Holstein gleichzusetzen, weshalb die liegenden Kiese und Diamikte älter sein müssen, das heißt sie sind nach den Erkenntnissen in der Nordschweiz dem GRG zuzuordnen. Auch ELLWANGER et al. (2011a) vertreten die Zuordnung zum größten Glazial, das sie als Hosskirch bezeichnen (Kap. 4.1).

Davon ausgehend sind die über den Beckentonen liegenden Schotter als jünger zu betrachten, wobei auffällt, dass in diesem Paket keine Diamikte vorhanden sind, also wohl keine Eisüberfahrung erfolgt ist (Abb. 12). Die in der Nordschweiz erkannte Habsburg-Vergletscherung, die auf das Holstein-Interglazial folgte (KELLER & KRAYSS 2010), erreichte höchstens die räumlichen Ausmaße wie Würm. Daher kann auch für Bittelschiess damit gerechnet werden, dass in diesem Glazial höchstens Schmelzwässer aus dem Raum Pfullendorf zur Donau übergelaufen sind und den besagten Schotter zur Ablagerung gebracht haben (Abb. 10).

Entlang der darüber folgenden Diskordanz finden sich Reste eines Paläobodens, der ein nächstes Interglazial ausweist. Dieses dürfte das Meikirch-Interglazial repräsentieren, nicht aber das Eem, denn die hangenden, mächtigen Schotter, die zum Teil Moräne enthalten, können in dieser geographi-

schen und topographischen Lage nicht Würm sein. Die bis 40 m umfassenden oberen Schotter mit teils hochgepresster Moräne sind dem Riss-Glazial zuzuordnen, deren Gletscher bis an die Donau vorstießen.

Differenziertere Erklärungen zu den Sedimentabfolgen und Bewertungen der komplexen Verhältnisse je nach Standort innerhalb des mächtigen Aufschlusses geben ELLWANGER et al. (2011a).

5.3 Intraglacialbereich des nördlichen Bodenseebeckens

Hosskircher Becken

Aus dem hauptsächlich nordgerichteten Schussenbecken zweigt das Nebenbecken von Hosskirch nach Nordwesten ab und zielt Richtung Ostrach, dem Übergang und Überlauf zum Donaugebiet. Etliche Tiefbohrungen wurden zur Erkundung der Beckenstruktur und der Sedimentfüllung niedergebracht. Anlass war die Suche nach geeigneten Zonen für Deponiestandorte. Die Auswertungen im Hinblick auf die

Lithologie und Chronologie der Quartärsedimente sind publiziert in ELLWANGER (1990, 1995), ELLWANGER et al. (1995, 1999). Es ergab sich eine mehrphasige, teils diamiktische, teils kiesig-sandige Abfolge mit eingeschalteten warmzeitlichen Bildungen. Nachgewiesen wurden aufgrund der Pollenanalyse von Bludau (zitiert in ELLWANGER et al. 1995) eine Buchen-Warmzeit, ein weiteres Interglazial, das jünger als Holstein, aber älter als Eem ist, sowie das Holstein-Interglazial. ELLWANGER konnte auch zeigen, dass das Hosskirch-Becken quer zu einem Strang der jüngeren, tieferen Deckenschotter angelegt ist und somit als solches jünger sein muss. In seiner jüngsten Publikation korrigiert Ellwanger (ELLWANGER et al. 2011a) die Pfefferbichl-Warmzeit in Eem und das ältere Riss in Hosskirch (Abb. 13, Auswertung).

Diskussion und Interpretation

Der Vergleich des Hosskirch-Beckens mit den Nordschweizer Becken (KELLER & KRAYSS 2010) ergibt einen gleichartigen lithostratigraphischen Aufbau, der eine Korrelation mit der dort erkannten Chronostratigraphie ermöglicht. Im Hosskirch-Becken, dessen Sedimente sehr genau untersucht worden sind (ELLWANGER 1990, 1995, ELLWANGER et al. 2011a), sind die Indizien für die postulierte Habsburg-Vergletscherung besonders überzeugend. Die Basis-Diamikte direkt über dem Beckenboden sind der Größten Vergletscherung zuzuordnen, die im Zuge der Beckenausschürfung zum Absatz kamen, was gemäss ELLWANGER et al. (2011b) „Hosskirch“ entspricht. Die darüber anschließenden Feinsedimente enthalten nach Bludau (zitiert in ELLWANGER et al. 2011a) das Holstein-Interglazial. Im Hangenden folgt ein Diamikt, der das Habsburg-Glazial widerspiegeln dürfte, dessen Gletscher das Hosskirch-Becken zwar noch überfahren, jedoch das Donau-Hochgebiet nicht mehr erreicht haben (Abb. 10). In den überlagernden Feinsedimenten findet sich gemäss Bludau ein nicht klar definierbares Interglazial, das das Meikirch-Interglazial repräsentieren dürfte. Im Hangenden folgt Moräne, die das gesamte Riss-Glazial widerspiegelt, indem Riss im Gegensatz zu ELLWANGER et al. (1999) als eine einzige Großvergletscherung aufgefasst wird (siehe Abschnitt Risstal-Biberach). Zwischen die liegende und die hangende, oberste Moräne greifen Feinsedimente ein, in denen durch Bludau die Buchen-Warmzeit festgestellt wurde. Sie kann als eine post-Eem-Warmzeit stellvertretend für das Eem angesehen werden, das hier in der typischen Ausprägung offenbar fehlt. Die abschließenden, die Oberfläche bildenden Moränen sind Teil der würmzeitlichen Gletscherfront.

Taubenried bei Pfullendorf

Ausstrahlend vom Pfrunger Ried erstreckt sich östlich Pfullendorf ein kleines Zweigbecken nach Nordwesten, das Taubenried. Es unterschneidet den Außenmoränenkranz der Würm-Maximalvergletscherung. Dieses Becken ist in seiner Tiefe und in seiner Sedimentfüllung anhand von Bohrungen, die von ELLWANGER (1990) ausgewertet worden sind, bekannt geworden. ELLWANGER erkennt darin drei glaziale Sequenzen, die er dem Doppelwall-Riss, dem Jungriss und dem Würm zuweist, wobei er die Beckenausräumung im Unteren Riss sieht (Abb. 14, Auswertung).

Diskussion und Interpretation

Das wenig westlich von Hosskirch gelegene Taubenried-Be-

Quartärgeologisches Profil durch das Hosskirch-Becken südöstlich Saulgau

aufgezeichnet nach Ellwanger 1995, 2011a
Pollenanalyse Bludau

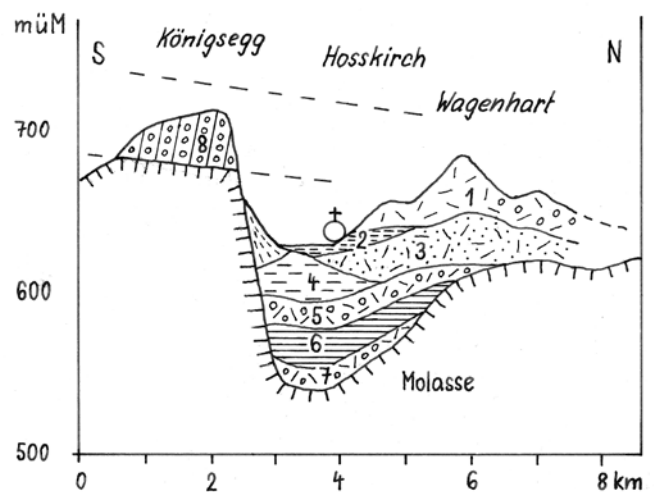


Abb. 13: Geologischer Querschnitt durch das Becken von Hosskirch bei Saulgau (nach ELLWANGER 2011a). Auswertung siehe Tabelle.

Fig. 13: Geological cross section through the Hosskirch basin near Saulgau (based on ELLWANGER 2011a). Analysis see table.

cken kann entsprechend zu dort interpretiert werden, wobei allerdings infolge Mangels an Zeitmarken (Bodenbildungen, Pollenanalyse) Spielraum für Zuweisungen offenbleibt. Die Beckenausräumung ist wie vielenorts im Zusammenhang mit der Größten Vergletscherung GRG zu sehen. Die basalen Sande können als vermutlich eisfreie Periode dem Holstein-Interglazial entsprechen. Darüber anschließende Schotter (Vorstoßschotter) werden von Moräne überlagert. Diese Sequenz könnte im Habsburg-Glazial eingestuft werden, da in dieser mit Hosskirch vergleichbaren geographischen Lage diese Vergletscherung das Becken noch überfahren haben müsste (Abb. 10).

Die über 20 m mächtigen, hangenden OSM-Sande scheinen, wie es von ELLWANGER (1990) angenommen wird, eine bedeutende Zeitspanne mit Eisfreiheit auszuweisen, was dem Meikirch-Interglazial entsprechen würde. Darüber abgelagerte Diamikte und Schotter enthalten wahrscheinlich die gesamte Riss-Eiszeit, deren Eisrandlagen (Doppelwall-Riss, aber auch Jungriss) weiter nördlich zu finden sind. In der östlicheren Bohrung angetroffene Verwitterungslehme sind entsprechend Ellwanger Indizien für das nächst jüngere Interglazial Eem. Sie werden überdeckt von Seesedimenten als Übergang zum folgenden Glazial, das durch Moräne belegt ist, die ihrerseits als Würm-Ablagerungen anzusprechen sind. Die westliche Bohrung liegt bereits ausserhalb der würmmaximalen Frontwälle und durchstösst die Sanderkiese der Niederterrasse.

Wattenweiler östlich Schussenried

Unterstützt von einer Profildarstellung beschreibt ELLWANGER (1995) eine Bohrung bei Wattenweiler, die knapp innerhalb der Wälle des Würm-Maximums abgeteuft wurde. Er stellt über einem markanten Hiatus zuerst Schotter und da-

Auswertung Profil Hosskirch (Abb. 13)

	Lithologie	Genetische Interpretation	Glazialchronologie gemäß ELLWANGER 1990 (2011a)	Glazialchronologie gemäß Autor
1	Moräne und Kiese	Frontaler Moränenwall	Würm-Eiszeit, LGM	Würm-Glazial
2	Feinsedimente mit warmzeitlicher Flora	Interglazial; gemäß Bludau „Buchen-Warmzeit“	IG „Buchen-Warmzeit“	IG [Eem fehlt]
3	Diamikte, Grundmoräne	Moränendecke, Eisüberfahung	Jungriss-Vergletscherung	gesamtes Riss-Glazial
4	Feinsande und Silte, warmzeitliche Flora	Älteres Interglazial, jünger als Holstein gemäß Bludau	Warmzeit Typ „Pfefferbichl“ [oder Eem-Interglazial?]	IG Meikirch
5	Konsolidierte Diamikte	Moräne eines Gletschervorstoßes	Doppelwall-Riss	Habsburg-Glazial
6	Feinsedimente mit warmzeitlicher Flora	Älteres Interglazial vom Typ Holstein	Samerberg-Holstein-Warmzeit	IG Holstein
7	stark konsolidierte Diamikte	Vergletscherung mit Beckenbildung und basaler Moräne	Älteres Riss [Hosskirch]	GRG [Größtes Rheinisches Glazial]
8	Verkittete Schotter	Tieferer Deckenschotter	Mindel-Deckenschotter	Jüngstes DS-Glazial

rüber mehrere Diamiktserien fest, die er gesamthaft einem Eisvorstoß der frühen Würm-Eiszeit zuordnet. Darüber lagern Feinsedimente mit einer warmzeitlichen Flora (Buchenwaldzeit gemäß Bludau zitiert in ELLWANGER 1995). Erst im Hangenden nimmt Ellwanger das Hochwürm an, repräsentiert durch Diamikte, Kiese und Sande (Abb. 15).

Unter dem Hiatus finden sich Diamikte unterlagert von Feinsedimenten, in denen Bludau eine Warmzeit vom Typ Samerberg-Holstein nachweisen konnte. Schliesslich folgen im Liegenden um die 30 m Diamikte bis zur anstehenden Molasse. Diese gesamte Sequenz positioniert Ellwanger im älteren und im Doppelwall-Riss, wobei er Riss in zwei eigenständige Eiszeiten, getrennt durch ein Interglazial, unterteilt.

Diskussion und Interpretation

Für die Beckenbildung und die basalen Diamikte ist die Vergletscherung des GRG = Hosskirch anzunehmen, folgt doch darüber das Holstein-Interglazial (Abb. 15). Die hangenden Diamikte sind dann Zeugen eines nächst jüngeren Glazials, das als Habsburg anzusprechen wäre, ausgehend von der Erkenntnis, dass diese Vergletscherung ausdehnungsmässig in etwa dem Würm entspricht und damit Wattenweiler noch erreicht haben müsste (Abb. 10). Der Hiatus dürfte dem vor dem nächsten glazialen Zyklus fehlenden Interglazial Meikirch entsprechen. Die Schotter und Diamikte über dem Hiatus sind somit ins Riss zu stellen, das mehrere Eisrandlagen mit wohl größeren Oszillationen aufweist, die sich in den verschiedenen Diamikten äußern.

Es sind zwar ein, eventuell zwei Frühwürm-Gletschervorstöße bekannt (SCHLÜCHTER et al. 1987, WINISTORF 1987, KELLER & KRAYSS 1998, PREUSSER 1999), die aber bedeutend geringere Ausmasse hatten als die Hochwürm-Vergletscherung. Wattenweiler in Frontnähe des Würm-Maximums wurde jedenfalls nicht erreicht. Die von ELLWANGER (1995) dem älteren Würm zugeschriebenen Moränenablagerungen über dem Hiatus sind demzufolge als risszeitlich anzusehen. Die in den darüber folgenden Feinsedimenten festgestellte Buchenwarmzeit ersetzt hier das fehlende typische Eem.

In den hangenden Sedimenten inklusive Diamikte ist die Würm-Vergletscherung dokumentiert.

Singener Becken

Das Singener Becken, die Tiefzone des Hegaus, ist von mittel- und spätpleistozänen Sedimenten erfüllt und umgeben von eisrandlichen Ablagerungen und Wallformen, die überwiegend der Würm-Eiszeit angehören (ERB 1934, SCHREINER 1974). Unter Einbezug von Bohrungen gibt SCHREINER (1992, Abb. 93) eine Übersicht zu den spätpleistozänen, würmzeitlichen Ablagerungen am und im Singener Becken.

Mit einer Serie von Bohrungen wurden in den 1990er Jahren die Teilbecken und deren Füllungen des gesamten Singener Beckens erforscht. SZENKLER & BOCK (1999) untersuchten die Bohrprofile und zeigten auf, dass das Großbecken sich in zahlreiche Teilbecken aufgliedert. Als typisches Beispiel wurde die Kernbohrung 1/1997 Rielasingen-Worblingen litho- und chronostratigraphisch interpretiert (Abb. 16). Es wurden dabei vier Glaziale aufgrund der Diamiktlagen ausgeschieden und vier Eiszeiten zugeordnet, wobei Doppelwall-Riss und Jungriss als eigenständige Glaziale aufgefasst werden. In einer anderen Bohrung wurden basal umgelagerte Pollen eines älteren Interglazials (süddeutsches Cromer) gefunden. In den Basissedimenten einer weiteren Bohrung konnten für die Warmzeit vom Typ Samerberg-Holstein charakteristische Pollen identifiziert werden (SZENKLER & BOCK 1999, mündl. Mitt. Bludau).

Diskussion und Interpretation

In Übereinstimmung mit SZENKLER & BOCK (1999) wird die Beckenbildung zusammen mit dem basalen Till in die Größte Vergletscherung = Hosskirch gestellt, gestützt auf die Pollenfunde (mündl. Mitt. Bludau). Hingegen können die zwei nahe beieinander liegenden, über einem Sand-Kies-Horizont folgenden Moränenhorizonte (Till) zwanglos in die Habsburg-Eiszeit eingestuft werden, umso mehr als darüber eine durch Sande und Silte ausgewiesene Seenphase anschließt, die für das Spät- und Post-Habsburg charakteristisch ist

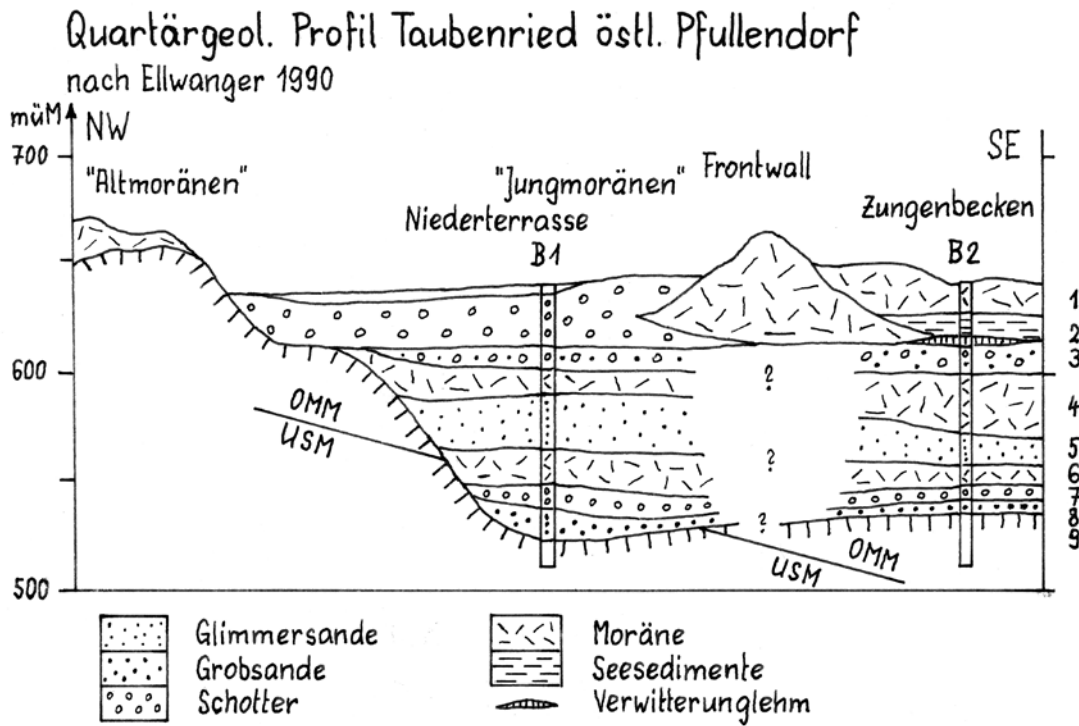


Abb. 14: Geologischer Querschnitt durch das Taubenried-Becken östlich Pfullendorf (nach ELLWANGER 1990). Auswertung siehe Tabelle.
Fig. 14: Geological cross section through the Taubenried basin in the east of Pfullendorf (based on ELLWANGER 1990). Analysis see table.

Auswertung Profil Taubenried (Abb. 14)

	Lithologie	Genetische Interpretation	Glazialchronologie gemäss ELLWANGER 1990	Glazialchronologie gemäss Autor
1	Diamikt, Moräne, im W Kiese	Grund- und Obermoräne, Gletscherfront, im W Sander	Würm-Eiszeit, Maximalstand	Würm-Glazial, Würm-Maximum
2	Feinsedimente, Verwitterungslehm	Stillwasser-Sedimente, basal Verwitterung warmzeitlich	Warmzeit, Eem	IG Eem
3	Grobsande, Kiese	Aufschüttung nach Eisfreigabe	Rückschmelz-Sequenz	Spätriss
4	Diamikt	Grundmoräne einer Eisüberfahung	Jungriss-Vergletscherung	Riss-Glazial
5	Glimmersande	umgelagerte Sande, eisfreie Periode	eisfrei, Warmzeit	IG Meikirch
6	Diamikt	Grundmoräne, Eisüberfahung	Mittlerriss-Vergletscherung	Habsburg-Glazial
7	Kiese	Vorstossschotter	Vorstoss Riss-Gletscher	Vorstoss Habsburg-Gletscher
8	Sande und Grobsande	Stillwasser-Sedimente	eisfreie Periode	Warmzeit, IG Holstein
9	Molasse-Felsbett	Exariertes Glazialbecken	Beckenbildung im Älteren Riss	Beckenausschürfung im GRG

(KELLER & KRAYSS 2010). Damit sind die mehrfachen Diamikte des mittleren Profilabschnittes nicht nur Jungriss, sondern dürften das gesamte Riss umfassen. Dabei wird davon ausgegangen, dass, wie oben gezeigt, der gesamte Riss-Komplex ein einziges Glazial darstellt. Abschließend folgen noch unzweifelhafte Würm-Moränen.

5.4 Ausblick zum Illergletscher

Wenn das Habsburg-Glazial in der Nordschweiz als gesichert

gelten kann (GRAF 2009a, KELLER & KRAYSS 2010) und nun auch im nördlichen Rheingletschergebiet eine Reihe Indizien dafür spricht, so müssen auch beim östlichen Nachbarn des Rheingletschers, beim Illergletscher, Hinweise für ein Glazial, das sich zwischen das größte Glazial, hier Mindel, und die Riss-Eiszeit einschleibt, zu finden sein.

Penck (PENCK & BRÜCKNER 1909) hat im Illergletscher-Vorland aufgrund der Höhenlagen der Schotterfelder Riss und Würm morphostratigraphisch erfasst und definiert. Dabei erkannte er zwar die ungleichen Niveaus des Hitzenho-

Bohrprofil Wattenweiler östl. Schussenried umgezeichneter Auszug aus Ellwanger 1995

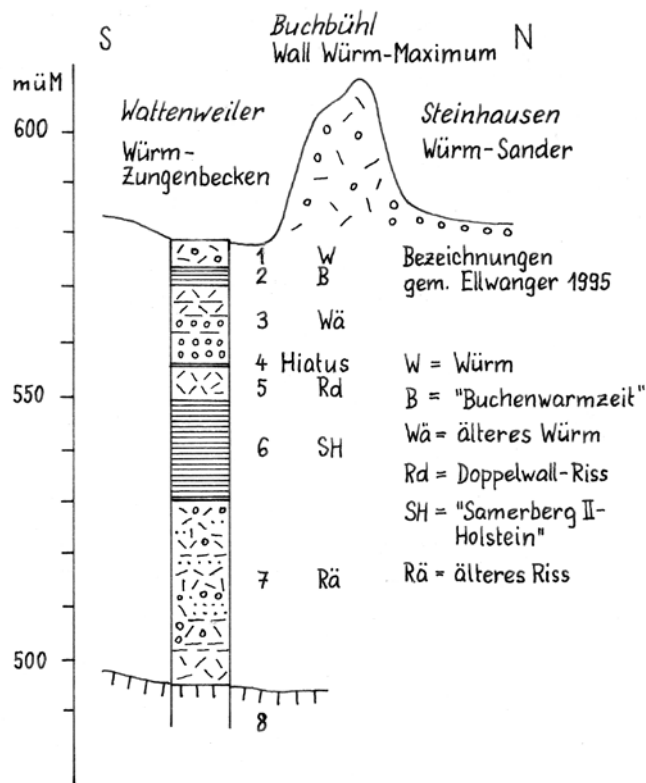


Abb. 15: Geologisches Profil der Bohrung Wattenweiler östlich Schussenried (umgezeichnet nach ELLWANGER et al. 1995). Auswertung siehe Tabelle.

Fig. 15: Geological profile of the Wattenweiler drilling in the east of Schussenried (re-drawn after ELLWANGER et al. 1995). Analysis see table.

fer Feldes und des Zeller-Hawanger Feldes (Abb. 17) sowie die verschiedenen Abflussbahnen aus dem Vergletscherungsraum über diese Schotterfelder. Er wies aber beide Systeme derselben Eiszeit, der Riss-Eiszeit zu. RÖGNER (1998) unterschied im Raum Memmingen vier Felder, die er entsprechend auch vier Eiszeiten zuordnete.

Auf dem Hitzenhofer Feld (Abb. 17) wurde von BIBUS (1995) bei drei Höfen je ein fossiler Boden erkannt, wobei beim Hof Baltes unter dem ersten fossilen Boden eine direkt folgende Kiesverwitterung als unsicherer zweiter fossiler Boden beschrieben wird. Hingegen weist das Hawanger Feld nach BIBUS an zwei Stellen je zwei fossile Böden übereinander auf.

Diskussion und Interpretation

Aufgrund von Penck's Konzept der treppenweisen Eintiefung und Einschachtelung der jüngeren Schotterfelder jeder Eiszeit in die älteren müsste dem Zeller-Hawanger Feld eine eigene Eiszeit zugewiesen werden (Vergleiche auch ELLWANGER 1988).

Aufgrund der Befunde bei den Nordschweizer Gletschern sowie beim Rheingletscher kann die Aussengrenze einer Habsburg-Vergletscherung auch beim Illergletscher maximal die Ausdehnung von Würm erreicht haben. Darauf basierend erklärt sich das Fehlen eines Endmoränensystems, an dem das Zeller-Hawanger Feld wurzelt, zwanglos, wenn von einer Vergletscherung ausgegangen wird, die intern im Illerbecken endete und deren Zeugen von den nachfolgenden Riss- und Würmgletschern ausgemerzt oder eingedeckt worden sind (Abb. 9).

Für das Hitzenhofer Feld ergibt sich gemäß BIBUS (1995) an drei Lokalitäten je ein sicherer fossiler Boden, was für die Bildung der Terrassenkiese in der zweitletzten Eiszeit Riss

Auswertung Bohrung Wattenweiler (Abb. 15)

Lithologie	Genetische Interpretation	Glazialchronologie gemäß ELLWANGER 1995	Glazialchronologie gemäß Autor
1 Diamikte, darin eingebettet Kiese und Sande	Eisrandnahe Grund- und Ober-Moräne mit Schottern und Sanden	Würm-Eiszeit, nahe Würm-Maximum	Würm-Glazial: Würm-Maximum
2 Feinsedimente mit Warmzeitlicher Flora	Interglaziale Seesedimente mit „Buchenwarmzeit“	Buchenwarmzeit als Frühwürm-Interstadial	Warmzeit; Eem fehlt
3 Schotter, darüber mehrere Diamikt-Horizonte	Vorstoßschotter, überlagert mit Moränen und Schotterlagen; Eisüberfahung	Eisvorstoß im frühen Würm: „Älteres Würm“	Riss-Glazial: gesamtes Riss mit Interstadialen
4 Hiatus, Diskordanz	durch Eisvorstoß abgesicherte Sedimente? Vermutlich Warmzeit	Hiatus	Diskordanz. Vermutlich IG Meikirch erodiert
5 Diamikte, stark konsolidiert	Grundmoräne einer Gletscher-Überfahung; konsolidiert infolge Überlagerung und Eisdruck	Gletscher-Vorstoß von Doppelwall-Riss als eigenständiges Glazial	Habsburg-Glazial
6 Feinsedimente, Silte, Ton	Mächtige Seesedimente, Beckenfüllung; Interglaziale Flora vom Typ Holstein mit <i>Pterocarya</i>	Holstein- Interglazial, trennt Älteres Riss vom Doppelwall-Riss	IG Holstein
7 Diamikte, stark konsolidiert, mit Einbettung von Sanden, Silten, Kieslagen	Wechselagerung von Grundmoräne und kiesigen Horizonten; durch Überlagerung konsolidiert	Gletscher-Vorstoß des Älteren Riss als eigenständiges Glazial	Basale Glazialsedimente des GRG
8 Molassefels der USM	Exariertes glaziales Becken	Becken-Ausschürfung im Älteren Riss	Beckenausschürfung durch GRG-Gletscher

spricht. Die Schüttung des Hawanger Feldes mit zwei fossilen Böden übereinander ist in die drittletzte Eiszeit zu legen, nach bisheriger Meinung die Riss 1-Vergletscherung. Nach der hier vertretenen Ansicht, die nur von einer einzigen Riss-Eiszeit ausgeht (siehe Ristal-Biberach, Kap. 5.2), kann dieses drittletzte Glazial der postulierten Habsburg-Vergletscherung entsprechen, die sich mit dem unteren fossilen Boden als Interglazial-Zeuge klar von Riss absetzt. Damit ist das Zeller-Hawanger Feld nicht in das gleiche Glazial einzustufen wie das Riss-zeitliche Hitzenhofer Feld, vielmehr ist anzunehmen, dass es der nächst älteren Eiszeit, dem Habsburg-Glazial mit dem Status eines eigenständigen Glazials, angehört.

6 Charakteristik der Glaziale im nördlichen Rheingletschergebiet

Die folgende Kurzbeschreibung geht von den vorstehend dargelegten Befunden und Indizien für vier mittel- und spätpleistozäne Glaziale aus. Einerseits wird ein Größtes Rheinisches Glazial GRG, nach ELLWANGER (2003) MEG, nach VILLINGER (2011) Hosskirch, das älter als Riss, aber jünger als die jüngeren Deckenschotter einzustufen ist, angenommen (Kap. 4). Andererseits wird bei Akzeptanz der Indizien ein selbständiges Habsburg-Glazial eingeführt, das sich zwischen GRG = Hosskirch und Riss einschleibt (Kap. 5).

Größtes Rheinisches Glazial GRG = Hosskirch

Eiszeit,

- deren Gletscher in einer „Eisüberflutung“ die größte Ausdehnung aller Vergletscherungen erreichten;
- die für die Ausschürfung des Bodensee-Hauptbeckens und der meisten Nebenbecken verantwortlich ist;
- die auf breiter Front die Donau zwischen Sigmaringen und Riedlingen noch überschritt und zwischen Bussen, Biberach und Aitrach die äußersten Wallmoränen schuf;
- die die einstigen Donauläufe von Sigmaringen bis Riedlingen verschüttete und den heutigen südlicheren Lauf induzierte.

Habsburg-Glazial

Eiszeit,

- deren Gletscherfront die Wasserscheide Rhein-Donau knapp erreichte, aber nicht mehr überfuhr;
- deren frontale Eisrandbildungen durch die nachfolgenden Eiszeiten überfahren und/oder zerstört wurden;
- deren glaziale Sedimente in den Bodensee-nahen Becken teilweise erhalten blieben oder später ausgeräumt wurden;
- deren Schmelzwässer nebst der Rheinachse nach Westen höchstens während der Maximalvereisung wenige Abflussbahnen zur Donau fanden.

Riss-Glazial

Eiszeit,

- deren Eisströme über die Rhein-Donau-Wasserscheide hinweg sich ausbreiteten, zwischen Mengen und Riedlingen bis an die Donau reichten und im Raum Riss-Aitrach markante parallele Wälle (Doppelwall-Riss) hinterliessen;
- die innerhalb ihrer Außengrenze umfangreiche glaziale Sedimente zur Ablagerung brachte;
- die verantwortlich ist für den Durchbruch der Aitrach aus dem Leutkircher Becken zur Iller.

Bohrprofil Rielasingen-Worblingen umgezeichnet nach Szenkler & Bock 1999

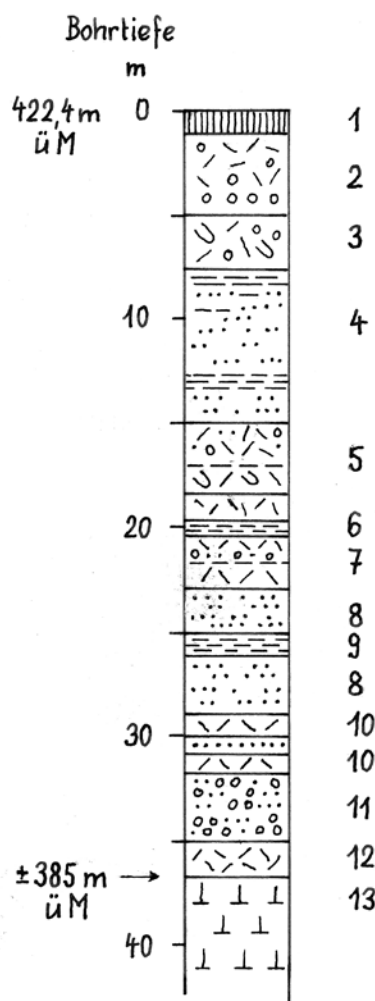


Abb. 16: Geologische Abfolge in der Bohrung Rielasingen-Worblingen, westliches Bodenseebecken (umgezeichnet nach SZENKLER & BOCK 1999). Auswertung siehe Tabelle.

Fig. 16: Geological record of the Rielasingen-Worblingen drilling, western part of the Lake Constance basin (re-drawn after SZENKLER & BOCK 1999). Analysis see table.

- deren Sander im Norden (Donau) und im Nordosten (Riss-Iller) als Hochterrasse in zahlreichen Relikten erhalten sind.

Würm-Glazial

Eiszeit,

- deren äußerste Zungen bis auf die Wasserscheide Rhein-Donau vorrückten und dort teils mächtige Frontmoränenwälle anhäuften;
- die sich in den Becken in gut erhaltenen Diamikten manifestiert;
- deren Schmelzwässer über zahlreiche Schwellen zwischen Homburg (westlich Messkirch) und Adelegg nordwärts zur Donau und ostwärts zur Iller abflossen;
- in deren Entwässerungsbahnen die Niederterrassen eingeschüttet wurden;
- deren zentraler Eisstrom das Bodenseebecken großenteils ausräumte und beim Abschmelzen den Bodensee entstehen liess.

Auswertung Bohrprofil Rielasingen-Worblingen (Abb. 16)

	Lithologie	Genetische Interpretation	Glazialchronologie gemäß SZENKLER & BOCK 1999	Glazialchronologie gemäß Autor
1	Anthropogen			
2	Diamikt und Schotter	Gletscher-Überfahung begrenztes Rückschmelzen	Last Glacial Termination	Würm-Glazial [Stein a/Rh-Stadial?]
3	Diamikt mit Umlagerung	Gletscher-Vorstoß und Überfahung	Last Glacial Maximum	Würm-Glazial [Maximum]
4	Feinsedimente, Sande	See-Sedimente	eisfreie Periode	IG Eem, Seenphase
5	Diamikt	Gletscher-Vorstoß	Jungriss	gesamtes Riss-Glazial
6	Diamikt-Umlagerung	mit Oszillationen		
7	Ton, Silt			
8	Diamikte			
9	Sande	Stillwassersedimente	eisfreie Periode	IG Meikirch? [ausgeprägte Seenphase]
10	tonige Feinsedimente	Seesedimente		
11	Sande			
12	Till	Gletscher-Vorstoß mit Oszillation	Doppelwall-Riss	Habsburg-Glazial
13	Sande und Kiese	Rückschmelz-Sedimente	eisfrei	IG Holstein? [Hinweise aus anderer Bohrung]
14	Till	Moräne, Eisüberfahung	Most Extensive Glaciation	GRG [Größtes Rheinisches Glazial]
15	OSM	Molasse-Felsbett		Beckenbildung im GRG

7 Schlussfolgerungen und Chronostratigraphie

Aus den in den vorstehenden Kapiteln dargelegten Befunden ergeben sich gute Hinweise darauf, dass im nördlichen Bodenseegebiet für das Mittel- und Spätpleistozän nebst den bekannten Eiszeiten Riss und Würm höchst wahrscheinlich zwei weitere Glaziale anzunehmen sind. Der Zeitraum dieser vier Glaziale beginnt nach der Ablagerung der Tieferen Deckenschotter mit der Umlenkung des Alpenrheins von der Donau zum Oberrhein und der damit verbundenen starken und tiefen Durchtalung im Bodenseeraum und in der Nordschweiz. Die daran anschließenden Eiszeiten führten zur charakteristischen Beckenbildung und zu übertiefen Trögen im Randalpengebiet und im Vorland. Sie können deshalb im Gegensatz zu den Deckenschotter-Eiszeiten als Becken-Eiszeiten bezeichnet werden. Die erste und älteste dieser Eiszeiten und gleichzeitig diejenige mit der größten Ausdehnung ist das Größte Rheinische Glazial GRG, für das in LGRB (2003) die Bezeichnung Hosskirch eingeführt wurde. Für das nächstfolgende Glazial, das im nördlichen Bodenseeraum bis jetzt unbekannt war, in der Nordschweiz aber nachgewiesen ist, das Habsburg-Glazial (GRAF 2009a; KELLER & KRAYSS 2010), konnte eine Reihe von Indizien gefunden werden. Erst daran anschließend folgen die Riss- und die Würm-Eiszeit.

Die Untersuchungen und Nachforschungen führten zur Erkenntnis, dass im gesamten nördlichen Bodenseeraum eine und dieselbe Vergletscherung die äußerste Grenze der Glazialrelikte hinterlassen hat, die postulierte Vergletscherung des **Größten Rheinischen Glazials GRG** (Kapitel 4).

Sie muss älter sein als Riss, denn es finden sich an einigen Stellen Schotter und glaziale Diamikte liegend unter Ablagerungen der Riss-Eiszeit. Im Nordwesten des Bodenseebeckens wurde bis vor kurzem Riss (Riss nach traditioneller Auffassung) als größte Vergletscherung betrachtet. In Übereinstimmung mit ELLWANGER (2003) und ELLWANGER et al. (2011a) kann gezeigt werden, dass die Riss-Vergletscherungsgrenze, morphologisch und sedimentär begründet, zwischen Homburg (NW Stockach)–Messkirch–Riedlingen deutlich intern der über die Donau hinausreichenden, äußersten Vergletscherungszeugen des GRG liegt.

Der mehrfache Nachweis des Holstein-Interglazials in Sedimenten im Hangenden derjenigen des GRG = Hosskirch, aber im Liegenden der Riss-Ablagerungen, respektive der als Habsburg angesprochenen Sedimente, ermöglicht die Trennung der Glaziale.

Im Nordosten und Osten des Bodenseebeckens sind die externsten Moränendecken und niedrigen Moränenwälle sowie die dazugehörigen Entwässerungsbahnen bisher in die Mindel-Eiszeit gestellt worden. Da diese Vergletscherungszeugen wie im Nordwesten eine ausgedehntere und ältere Eiszeit als Riss belegen, ist auch hier von der Zugehörigkeit zum Größten Rheinischen Glazial GRG auszugehen. Sie weisen zudem dieselben morphologischen Charakteristika auf wie jene. Im Weiteren ergibt die Eiskörper-Rekonstruktion des GRG (Abb. 4) ein plausibles dreidimensionales Bild.

Das Postulat von GRG = Hosskirch im östlichen Rheingletschergebiet als ausgedehnteste Vergletscherung dürfte hier „Mindel“ ablösen mit gravierenden Konsequenzen:

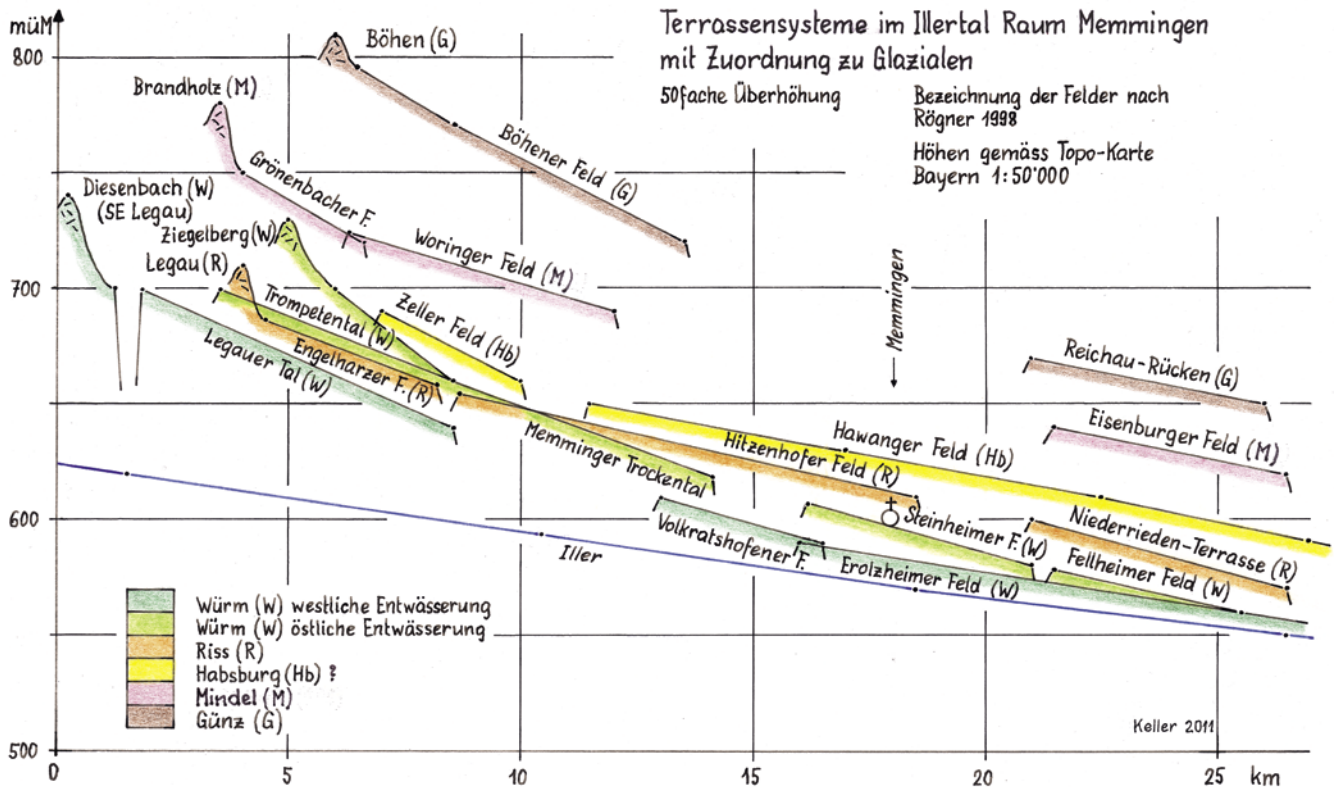


Abb. 17: Abfolge der Sander-Terrassen des Illergletschers im Raum Memmingen; S-N-Projektion.

Fig. 17: Succession of the outwash terraces of the Iller Glacier in the region of Memmingen; S-N-projection.

1. Im Nordosten und Osten des Rheingletscher-Ausbreitungsbereichs würden die als Mindel bekannten Moränen nicht mehr in die Zeit der Deckenschotter fallen, sondern sie wären dem Größten Rheinischen Glazial GRG = Hosskirch als älteste Becken-Eiszeit zuzuweisen.
2. Die bisher mit Mindel verknüpften Schotterstränge zwischen Aitrach und Riss würden zu Sandern des GRG und wären damit nicht mehr Deckenschotter im traditionellen Sinn.
3. Die Haslach-zeitlichen Ablagerungen blieben weiterhin Deckenschotter und zwar wie bisher jüngere Deckenschotter.

Neu ergibt sich folgende zeitlich relative Entwicklung für das Rheingletschergebiet:#

jünger	Größtes Rheinisches Glazial GRG = Hosskirch: Ausschürfung der meisten tiefen Becken, ausgedehnteste Vergletscherung, im NW ehemals Riss, im NE bisher Mindel.
	Umlenkung des Alpenrheins und tiefe fluviale Durchtalung
	Jüngerer, (tieferer) Deckenschotter = Haslach-Eiszeit
älter	(Mindel entfällt als Deckenschotter)

Ausgehend vom Nachweis des Habsburg-Glazials in der Nordschweiz (KELLER & KRAYSS 2010; PREUSSER et al. 2011) wurde für das nördliche Bodenseebecken das „Modell eines

Habsburg-Glazials“ aufgestellt mit dem Ziel, stichhaltige Hinweise für diese Eiszeit zu finden (Kapitel 5). Da die räumlichen Ausmasse dieser Vergletscherung in der Nordschweiz in etwa mit denjenigen der Würm-Eiszeit übereinstimmen, wurde für den nördlichen Bodenseeraum eine entsprechende Bandbreite für ein solches Glazial angenommen (Abb. 10), in welchem vorrangig Anzeichen für diese Eiszeit gesucht wurden. Morphologisch in Erscheinung tretende Eisrandbildungen sind nicht zu finden, denn die nachfolgende, weiter reichende Riss-Vergletscherung hat diese überfahren und zerstört oder eingedeckt. Hingegen ergab die Auswertung der bearbeiteten und diskutierten Schlüsselstellen gute Indizien für die angenommene Habsburg-Vergletscherung. Dabei wird von der Akzeptanz einer einzigen Riss-Eiszeit ausgegangen, in der Älteres Riss und Jungriss keine eigenständigen Glaziale sind.

Aus lithostratigraphischen Befunden resultiert die chronologische Einordnung des postulierten Habsburg-Glazials nach dem Größten Rheinischen Glazial GRG = Hosskirch, aber vor der Riss-Eiszeit. Abgesichert wird dies durch die stratigraphische Lage von Paläoböden sowie durch den Nachweis von pollenanalytisch erfassten Interglazialen. Im Liegenden der als Habsburg angesprochenen Sedimente findet sich verschiedentlich das Holstein-Interglazial, im Hangenden ein nicht genauer definierbares Interglazial, das dem Meikirch-Interglazial entsprechen dürfte.

Nachdem zahlreiche lithostratigraphische und/oder morphostratigraphische Indizien und Argumente vorliegen, die im ganzen Rheingletschergebiet das GRG = Hosskirch als ausgedehnteste Vergletscherung und als älteste Becken-Eiszeit ausweisen, die Existenz des Habsburg-Glazials als

Chronostratigraphie des Mittel- und Spätpleistozäns für die Nordschweiz und das Rheingletschergebiet

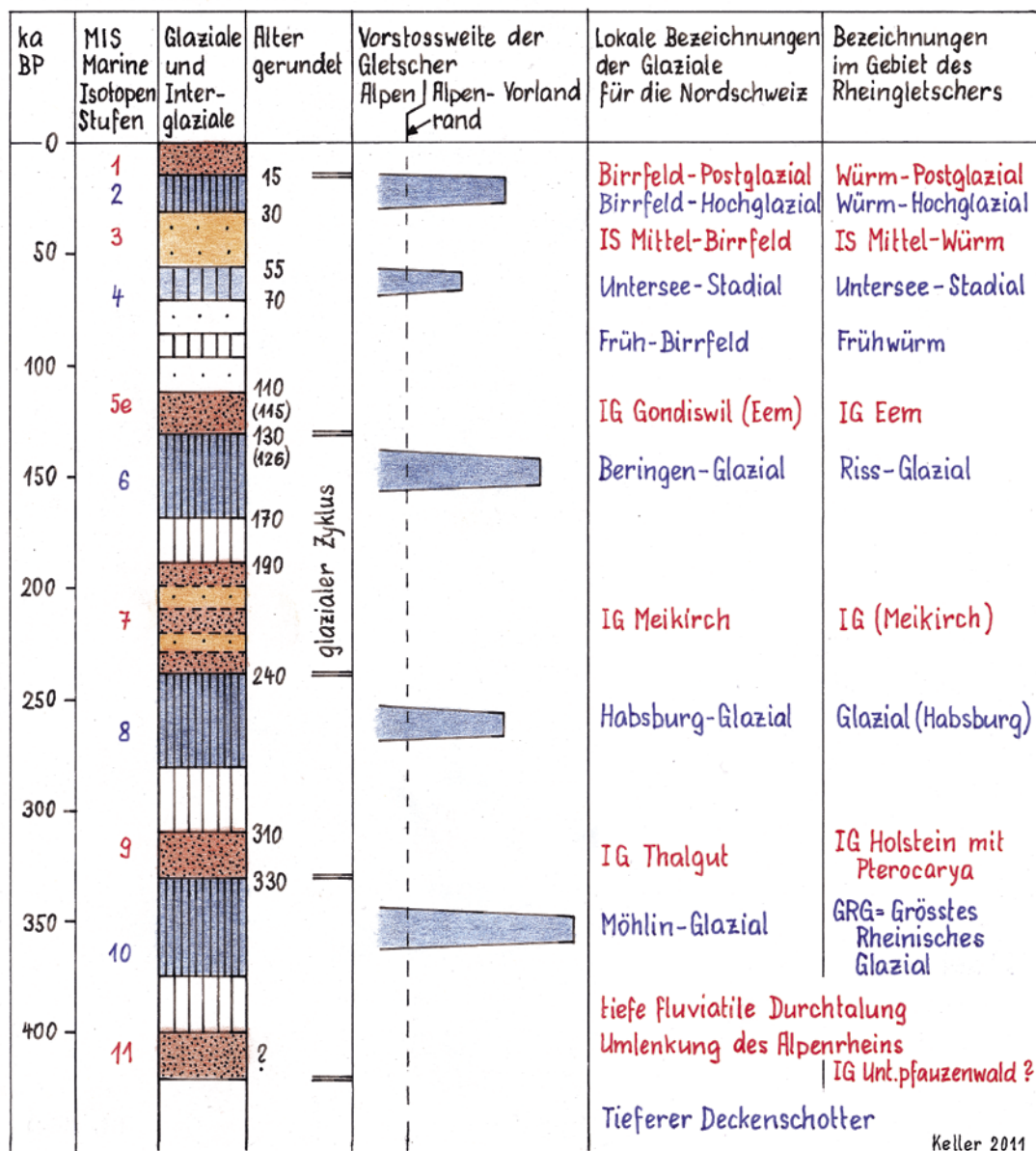


Abb. 18: Chronostratigraphie des Mittel- und Spätpleistozäns der Nordschweiz und des Rheingletschers. Marine Isotopenstufen und Zeitmarken nach LITT et al. (2007) und PREUSSER (2010).

Fig. 18: Chronostratigraphy of the Middle and the Late Pleistocene of Northern Switzerland and of the Rhine Glacier. Marine isotope stages and time marks after LITT et al. (2007) and PREUSSER (2010).

wahrscheinlich erscheinen lassen und die Einstufung dieser Eiszeiten vor der Riss-Eiszeit erfordern, fehlt noch eine Chronostratigraphie. Ausschlaggebend für eine zeitlich absolute Einstufung sind die die Glaziale trennenden Interglaziale. Zeitlich fixiert ist seit längerem das Eem mit 126–110 ka BP, während für das Holstein, noch mit Unsicherheiten behaftet, 320–310 ka BP anzunehmen sind (LITT et al. 2002, 2007). Beide Warmzeiten sind in den Sedimentabfolgen des nördlichen Rheingletschergebiets verschiedentlich nachgewiesen. Das dazwischen liegende Interglazial, das ebenfalls erfasst ist, dürfte dem Meikirch-Interglazial entsprechen, das mit 240–185 ka BP datiert ist (PREUSSER 2010).

Mit diesen Zeitmarken können die Eiszeiten des Mittel- und Spätpleistozäns im nördlichen Bodenseeraum in diesel-

be chronostratigraphische Abfolge eingefügt werden, wie sie in KELLER & KRAYSS (2010) für die Nordschweiz aufgestellt wurde. Sie wird hier in der Abb. 18 nochmals aufgeführt. Demnach erfolgten die Großvergletscherungen der Becken-Eiszeiten während den als Kaltzeiten ausgewiesenen marinen Isotopenstufen: das Grösste Rheinische Glazial GRG in der MIS 10, Habsburg in der MIS 8, Riss in der MIS 6 und das Würm-Hochglazial in der MIS 2.

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Ice-Rafted Erratics and Bergmounds from Pleistocene Outburst Floods, Rattlesnake Mountain, Washington, USA

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Abstract: Exotic ice-rafted debris from the breakup of ice-dammed glacial lakes Missoula and Columbia is common in slackwater areas along the 1,100-km route for outburst floods in the northwestern US. A detailed analysis was performed at Rattlesnake Mountain, which lay beyond the limit of the former ice sheet, where an exceptionally high concentration of ice-rafted debris exists midway along the floods' path. Here floodwaters temporarily rose to 380 m elevation (forming short-lived Lake Lewis) behind the first substantial hydraulic constriction for the outburst floods near Wallula Gap. Within the 60 km² study area more than 2,100 erratic isolates and clusters, as well as bergmounds were recorded. Three quarters of erratic boulders are of an exotic granitic composition, which stand in stark contrast to dark Columbia River basalt, the sole bedrock in the region. Other exotics include Proterozoic quartzite and argillite as well as gneiss, diorite, schist and gabbro, all once in direct contact with the Cordilleran Ice Sheet to the north. Most ice-rafted debris is concentrated between 200 and 300 m elevation. Far fewer erratics and bergmounds lie above 300 m elevation because of the preponderance of less-than-maximum floods. Plus, larger deep-rooted icebergs were forced to ground farther away from the ancient shorelines of transient Lake Lewis. As floodwaters moved across the uneven surface of Rattlesnake Mountain, many erratic-bearing icebergs congregated into pre-existing gullies that trend crosswise to flood flow.

Eisverfrachtete Findlinge und Bergmounds aus Ausbruchsflutwellen im Pleistozän, Rattlesnake Mountain, Washington, USA

Kurzfassung: Eisverfrachteter Schutt findet sich häufig im Stauwasserbereich der 1.100 km weit reichenden gigantischen Ausbruchsflutwellen aus den eiszeitlichen Missoula- und Columbia-Seen im Nordwesten der USA. Eine detaillierte Analyse erfolgte am Rattlesnake Mountain. Dort ist eine außergewöhnlich hohe Konzentration dieses Schutts in der Mitte des Gerinnebettbodens zu finden. Zeitweilig bildete der Flaschenhals des Wallula Gap die erste wesentliche hydraulische Verengung und ließ die Gletscherflut vorübergehend auf 380 m Seehöhe ansteigen, wodurch kurzzeitig der Lewis See gebildet wurde. Auf einer Fläche von 60 km² wurden mehr als 2.100 Findlinge, Ansammlungen von erratischen Felsbrocken und Bergmounds registriert. Im Gegensatz zum lokal vorkommenden dunklen Columbia River Basalt bestehen drei Viertel der Findlinge aus granitartigem Material. Auch Schutt von dem ehemals im Norden verlaufenden Eisgebirgszug wie sedimentärer Quarzit aus dem Proterozoikum, Tonschiefer, Gneisgestein, Diorit, Schiefer und Paulitfels wurden hier gefunden. Ein Großteil des eis-verfrachteten Schutts befindet sich auf 200–300 m Seehöhe. Weit weniger Findlinge und Bergmounds sind über 300 m Seehöhe anzutreffen, da es überwiegend Fluten von sub-maximalen Ausmaßen gab. Außerdem liefen größere Eisberge aufgrund ihres Tiefgangs weit von der Küstenlinie des vorübergehend bestehenden Lake Lewis auf Grund. Bei der Flutbewegung über den unebenen Untergrund des Rattlesnake Mountain blieben viele Findlinge in bereits vorhandenen, quer zur Flutrichtung verlaufenden Wasserrinnen hängen.

Keywords: *ice-rafted debris, erratic, bergmound, Missoula floods, Wallula Gap, Lake Lewis, glacial Lake Missoula, Wisconsin Glaciation, Columbia River basalt*

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1 Introduction and Background

Characteristics and distribution patterns of ice-rafted debris (erratics and bergmounds) can provide insight into the flow dynamics as well as possible source(s), timing, and frequency of Ice Age flood events. Erratics in the Pacific Northwest have long been recognized in areas downstream of the maximum extent of glacial ice (BRETZ 1919, 1923a, 1923b, 1930, 1969; ALLISON 1933, 1935; BRETZ et al. 1956; FECHT & TALLMAN 1978; MINERVINI et al. 2003). Multiple cataclysmic outburst floods, mostly from ice-dammed glacial Lake Missoula (Figure 1), are now generally accepted as the source for erratics beyond the limits of the Cordilleran

Ice Sheet (BRETZ et al. 1956; BRETZ 1969; BAKER 1978; BAKER & BUNKER 1985; ALLEN et al. 2009; SMITH 1993; WAITT 1980, 1985). Other sources for Ice Age floods in the area include at least one flood from ice-dammed glacial Lake Columbia (ATWATER 1987; WAITT 1994; WAITT et al. 2009), Lake Bonnevill (O'CONNOR 1993) and possibly from one or more subglacial outbursts from beneath the Cordilleran Ice Sheet itself (SHAW et al. 1999). The Cordilleran Ice Sheet never advanced south into the mid-Columbia Basin where much ice-rafted debris came to rest. Therefore, the only plausible explanation for high-elevation erratic debris beyond the ice front is from floating icebergs carried during outburst-flood events.

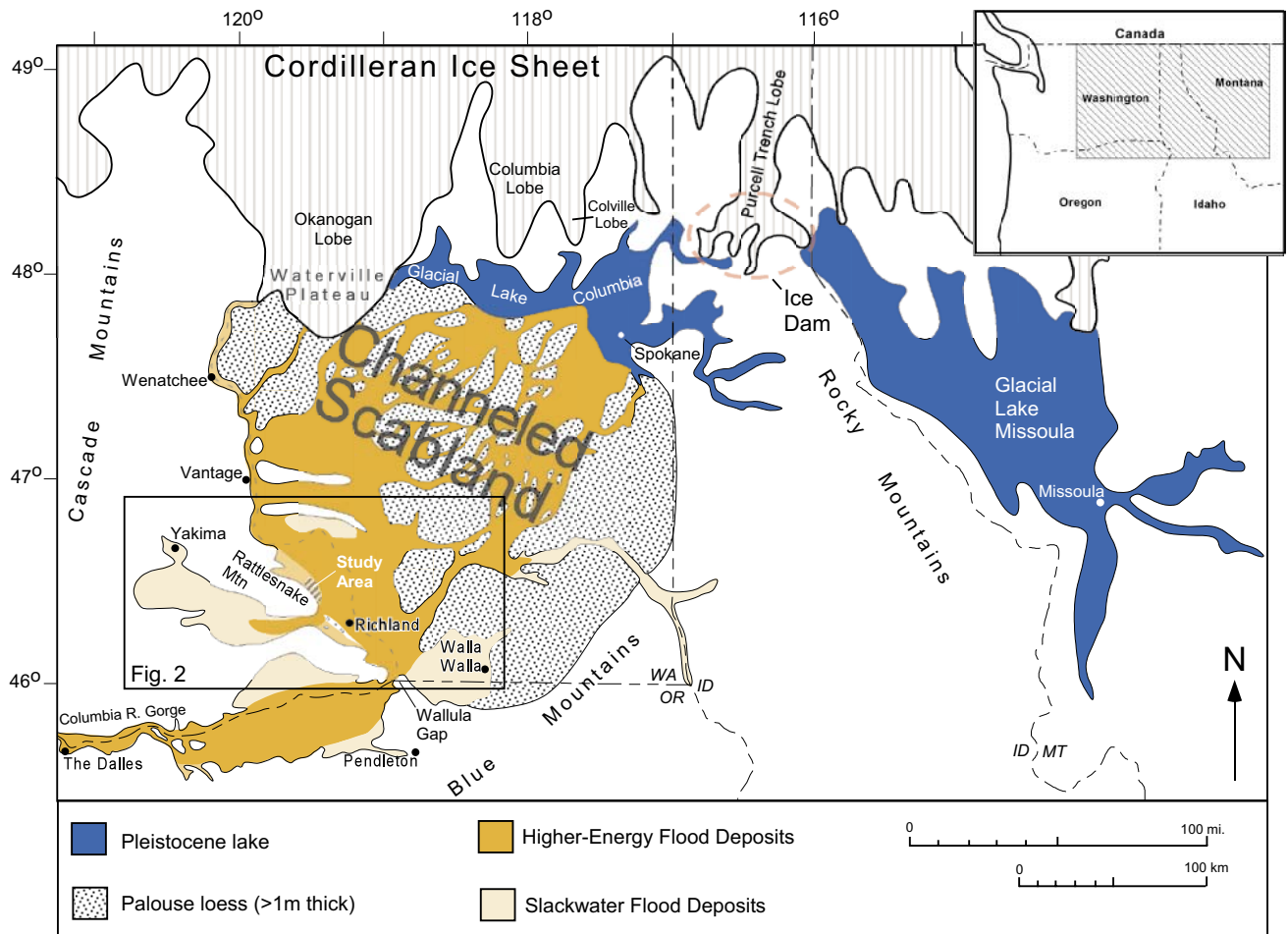


Fig. 1: Ice Age glacial and flood features over a portion of the northwestern United States.

Abb. 1: Eigenschaften von Gletschern und Gletscherläufen im Nordwesten der USA.

Although the largest Missoula floods drained within several days, up to three weeks were required for the floodwaters to completely pass through several hydraulic constrictions along route (WAITT et al. 2009; DENLINGER & O'CONNELL 2010). The first major constriction for the outburst floods was at Wallula Gap where, after spreading out across a 160 km-wide tract of the Channeled Scabland, floodwaters were forced through a single, narrow opening only 3 km wide (BJORNSTAD et al. 2007). During the largest floods Wallula Gap and the Columbia River Gorge downstream could transmit only a portion (10 ± 2.5 million m^3/sec) of all the floodwater entering the Pasco Basin (O'CONNOR & BAKER 1992; BENITO & O'CONNOR 2003). Subsequently, a huge temporary lake (Lake Lewis shown in Figure 2) backed up to a maximum elevation of ~ 380 m (1,250 feet) behind this bedrock constriction (BRETZ 1923a). Ice-rafted debris floating in Lake Lewis was sequestered in slackwater areas along basin margins and backflooded valleys (Figures 2 and 3). The stranded icebergs eventually melted, forever leaving behind their payloads of exotic detritus (Figures 4 and 5).

Maximum heights for the floods are indicated by strandlines, scarped hills, ice-rafted erratics, and divide crossings (BAKER 1978). The upper limit of the ice-rafted debris descends downstream in a stair-step fashion as floodwaters passed through a series of constrictions en route (BAKER 1978; BENITO & O'CONNOR 2003). Water-surface profiles of floods were relatively flat within basins but steepened in

bedrock-constrained reaches between basins like Wallula Gap and the Columbia River Gorge (Figure 1).

The earliest floods occurred during one of many previous glacial cycles $\geq 780,000$ yr ago in the early Pleistocene (PATTON & BAKER 1978; BJORNSTAD et al. 2001; PLUHAR et al. 2006; BJORNSTAD 2006). The last period of flooding occurred during the Late Wisconsin Glaciation between 20,000–15,000 cal years BP (O'CONNOR & BENITO 2009) when as many as 100 discrete flood events may have occurred (ATWATER 1986; WAITT et al. 2009). Many of these floods were proportionally small (a few million m^3/sec) relative to discharge for the largest (>17.5 million m^3/sec) Late Wisconsin outburst flood(s) from glacial Lake Missoula (O'CONNOR & BAKER 1992).

The majority of erratics, which consist of light-colored plutonic and metamorphic rocks, stand out in sharp contrast to the dark, Miocene, Columbia River Basalt Group – the only bedrock native to southeastern Washington State (Figure 6). Most ice-rafted debris appears derived from the breakup of the Purcell Trench Lobe of the Cordilleran Ice Sheet that temporarily dammed glacial Lake Missoula (Figure 1). Ice-rafted debris may also be associated with the final breakup of the Okanogan Lobe that blocked glacial Lake Columbia until the end of the Wisconsin Glaciation.

The detailed study of the nature and distribution of ice-rafted erratics and bergmounds provides valuable data sets to test various hypotheses regarding the history and rela-

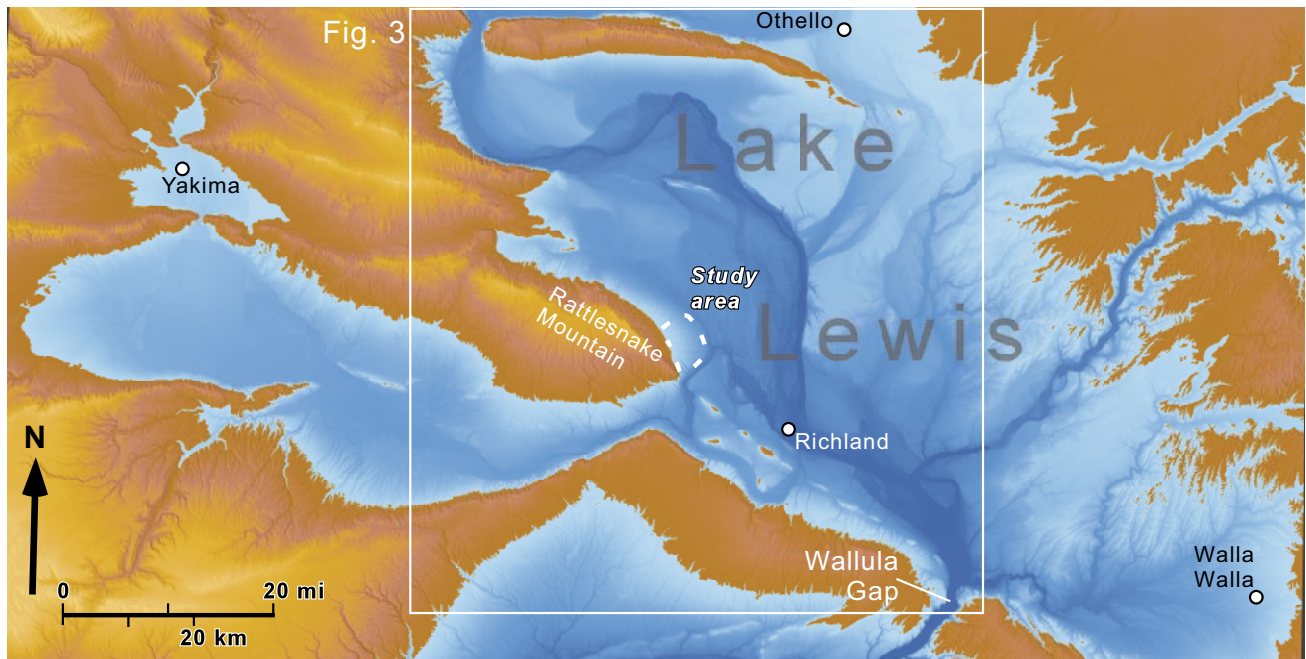


Fig. 2: Maximum extent (up to 380 m elevation) of back flooding (Lake Lewis in blue) behind hydraulic constriction at Wallula Gap during the largest outburst flood(s).

Abb. 2: Maximale Ausdehnung der Rückstaufut (bis zu 380 m Seehöhe) hinter der hydraulischen Verengung des Wallula Gap (blaue Markierung: Lewis-See).

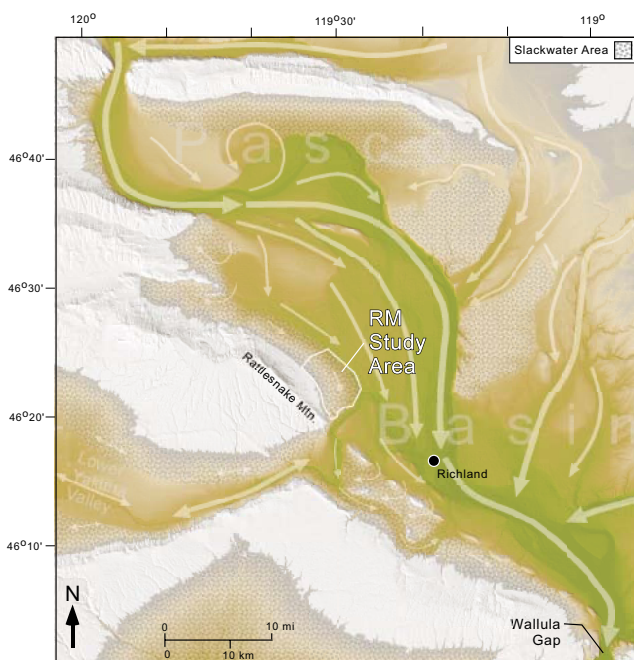


Fig. 3: Shaded-relief map showing variable flow of floodwaters through the Pasco Basin. White areas were above maximum flood level (>380 m elevation). Sizes of arrows are approximately proportional to flow velocity. Stippled pattern signifies slackwater areas that received the greatest quantities of ice-rafted debris. The Rattlesnake Mountain study area is located at center.

Abb. 3: Schattierte Reliefkarte zur Darstellung der Flussvariablen der Fluten im Pasco-Becken. Weiße Bereiche zeigen maximalen Wasserstand (> 380 m Seehöhe). Das Pfeilmaß ändert sich proportional zur Strömungsgeschwindigkeit. Diagonale Linien bezeichnen Stauwasserbereiche mit den größten Mengen von eisverfrachtetem Schutt. Das Untersuchungsgebiet Rattlesnake Mountain befindet sich im Zentrum.

tive sizes of Ice Age floods. Only a few previous studies exist for erratics emplaced by outburst floods. These include a study by ALLISON (1935) within the Portland Basin and later expanded upon by MINERVINI et al. (2003). Another study by KARLSON (2006) was performed in the Ginkgo Petrified Forest State Park, located 80 km northwest of Rattlesnake Mountain near Vantage. The present study gathered an exhaustive set of data including lithology, size, roundness, shape, and weathering characteristics of more than 2,000 individual erratics, along with their locations and elevations (BJORNSTAD et al. 2007).

2 Study Area

A comprehensive analysis of ice-rafted debris was performed in a 60 km² area along the north flank of Rattlesnake Mountain – the tallest segment of an elongated northwest-southeast trending anticlinal ridge (Yakima Fold) of Columbia River basalt (Figure 6) that rises to 1,075 m elevation. During the largest outburst floods, water rose to within 700 m of the summit, making Rattlesnake Mountain a long peninsular land body during flooding (Figure 2). The study area is located along the low-relief northeastern flank of the ridge where floodwater flowed from the northwest to the southeast (Figure 3). The upper third of the study area consists of a gentle (~2°), northeast-dipping surface (Iowa Flats) that reflects the surface on the underlying basalt bedrock. Only a few metres of relief exist within small gullies across Iowa Flats. Downslope of Iowa Flats ephemeral streams are more entrenched into the basalt displaying as much as 25 m relief between the tops and bottoms of gullies. Although basalt may be exposed at the bottoms of gullies, inter-fluvial areas are covered with up to several metres of mostly massive, fine-grained deposits of Quaternary-age loess, slope-wash, or slackwater-flood sediment.

Most of the study area lies within a long-protected ecology preserve, now part of the Hanford Reach National Monument managed by the United States Fish and Wildlife Service (BJORNSTAD 2006; BJORNSTAD et al. 2007). Closed to the public for the last 60 yr, this area is ideally suited to the study of ice-rafted debris because it has escaped widespread anthropogenic disturbances common to other parts of the region. Furthermore, light-colored erratic boulders stand out in stark contrast to indigenous black Columbia River basalt and the pale-brown blanket of fine-grained Quaternary sediments. These attributes, along with the low, sparse, shrub-steppe vegetation facilitate the identification and mapping of ice-rafted debris. Last, the gentle yet uneven slopes provide an opportunity to examine subtle changes in flood-flow dynamics reflected in the distribution of ice-rafted debris.

3 Ice-Rafted Debris

BRETZ (1923a: 605) accurately noted: “A widespread submergence of the lower Columbia Valley is known to have occurred during the Wisconsin glaciation. It is recorded by berg-floated erratic boulders, some of great size, scattered widely in the Columbia Valley below the present altitude of

about 1,250 feet [380 m] above tide.” Bretz and others, however, categorized ice-rafted debris into two classes: erratics or bergmounds. The present study distinguishes an important third category of ice-rafted debris – erratic clusters. Examples of the three types of ice-rafted debris are shown in Figure 4. Isolated erratics (Figure 4A) generally consist of a solitary boulder while clusters (Figure 4B and 5) represent a close grouping of several or more erratics. Bergmounds (Figure 4C) consist of distinct, low-relief mounds covered with ice-rafted debris, up to 30 m or more in diameter and several meters tall (BRETZ 1930; ALLISON 1933; FECHT & TALLMAN 1978; CHAMNESS 1993). In general, bergmounds are distinguished from erratic clusters by the presence of a conical mound displaying recognizable topographic relief. Individual clusters and bergmounds may contain a single rock type, but more are represented by multiple lithologies (ALLISON 1935; FECHT & TALLMAN 1978). The areal distribution for all three different types of ice-rafted debris is shown over a portion of Rattlesnake Mountain on Figure 7.

4 Methods

The 60-km² Rattlesnake Mountain study area was field map-



Fig. 4: Types of ice-rafted debris on Rattlesnake Mountain. (A) Isolated granodiorite erratic boulder at (183 m elevation) in basalt-floored gully. (B) Cluster of granodiorite erratics at 335 m elevation. (C) Pair of bergmounds (~245 m elevation); notice granitic erratic atop left bergmount (encircled). Looking north into the expansive Pasco Basin.

Abb. 4: Typen von eisverfrachtetem Schutt am Rattlesnake Mountain. (A) Vereinzelte erratische Gesteinsblöcke aus Granodiorit in basaltgrundiger Schluchtrinne (auf 183 m Seehöhe). (B) Anhäufung von erratischen Gesteinsblöcken aus Granodiorit Findlingen (auf 335 m Seehöhe). (C) Pärchen von Bergmounds (~245 m Seehöhe); man beachte den Granitfindling (eingekreist) oben auf dem linken Hügel. Blick nach Norden in das weitläufige Pasco-Becken.



Fig. 5: Cluster of moderately weathered, banded argillite boulders at 244 m elevation on Iowa Flats along the north flank of Rattlesnake Mountain. The largest erratic boulder is about three metres long. Argillite is derived from the Proterozoic Belt Supergroup once in contact with the ice dam for glacial Lake Missoula in northern Idaho ~650 km to the northeast. Note that the erratics in this cluster are partially buried in an unknown quantity of Quaternary loess, slopewash and slackwater flood deposits.

Abb. 5: Haufen von mäßig verwitterten Gesteinsbrocken aus gestreiftem Argilit (auf 244 m Seehöhe) auf den Iowa-Untiefen entlang der Nord-Flanke des Rattlesnake Mountain. Der größte Findling links der Bildmitte ist etwa drei Meter lang. Argilit (Tonschiefer) entstammt dem Gürtel der Übergruppe aus dem Proterozoikum, der einst mit dem Eisdamm des ca. 650 km nordöstlich entfernten Gletschers von Lake Missoula in Nord-Idaho in Verbindung stand. Man beachte, dass die Findlingsblöcke dieser Gruppe teilweise unter unbestimmten Mengen von Löss-, Ausspülungs- und Stauwasser-Ablagerungen aus dem Quartär begraben sind.

ped in a series of roughly parallel to diagonal, up-down transects. The transects traversed gullies as well as the inter-fluvial areas, usually spaced 100 m or more apart. All observed ice-rafted debris $>0.093 \text{ m}^2$ maximum exposed cross sectional area was located using a hand-held GPS receiver unit. An arbitrary cut-off value of 0.093 m^2 was chosen because that was near the minimum size readily observed in the field. Depending on location, erratics can be partially to almost totally buried by Quaternary-age fine-grained loess, slopewash, or slackwater flood deposits. For this reason it was not possible to determine the true volume of most erratics and the size reported here represents only the maximum exposed cross-sectional area observed in the field. For data manageability, because many erratic clusters and berg-mounds contain dozens or more individual erratics, specific information was recorded only for the largest erratic in each cluster or bergmound grouping.

Horizontal accuracy of handheld GPS units is generally within 3 m while the measured vertical accuracy is much less precise. For this reason, instead of the elevation recorded by the GPS, elevations were obtained from a topographic map after location coordinates were entered into National Geographic's TOPO!® program. Because there is wide spacing of contour lines in the generally low relief of the study area, recorded elevations are believed to be accurate to within at least half of the contour interval (12.2 m), or about 6 m vertical distance.

In addition to location and elevation, data were collected on lithology, roundness, shape, and surface characteristics, such as degree of weathering, wind polish, and striations. Because rates of weathering and roundness vary with lithology, these parameters were compared for only the dominant rock type (i.e., granitic). Granitic erratics classified as un-

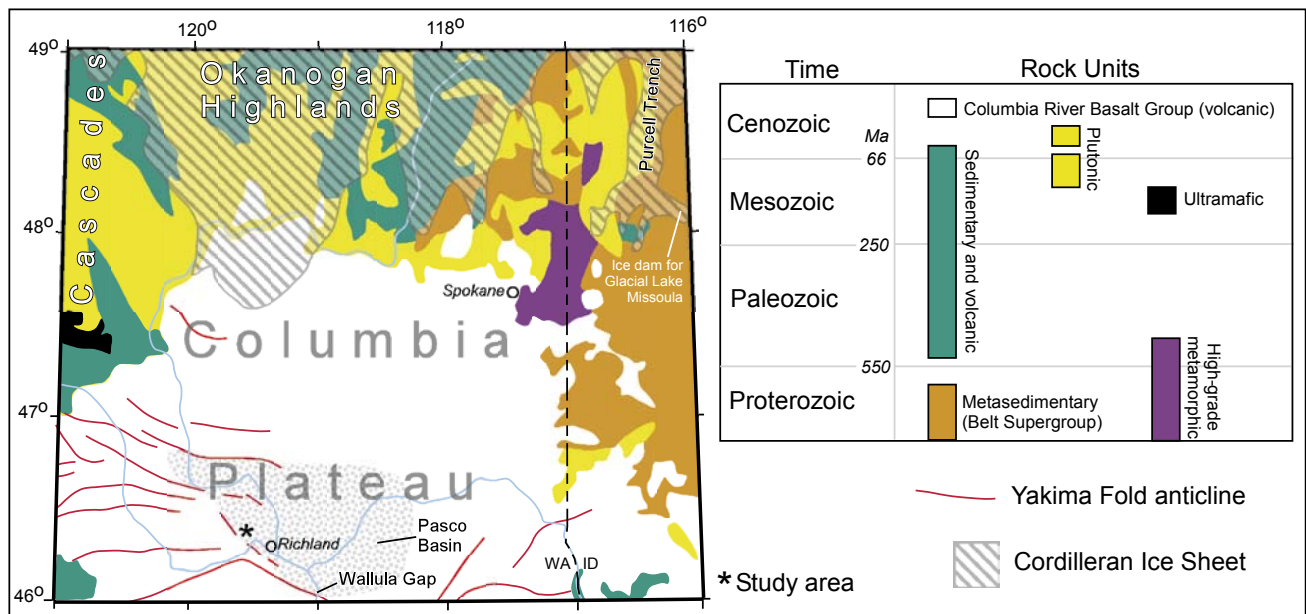


Fig. 6: Distribution of major rock units in eastern Washington and northern Idaho. Most ice-rafted erratics found on Rattlesnake Mountain can be traced to granitic (plutonic) and Belt Supergroup metasedimentary rocks once in contact with the Cordilleran Ice Sheet north and east of the Columbia Plateau. To the south, Yakima Folds strongly influenced the flow of floodwater into and out of the Pasco Basin.

Abb. 6: Verteilung größerer Felseinheiten im östlichen Washington und Nord-Idaho. Der Ursprung der meisten eisverfrachteten Findlingen auf Rattlesnake Mountain kann zu den granitartigen (plutonischen) und metasedimentären Felsen des Gürtels der Übergruppe zurückverfolgt werden, die einst in Verbindung mit dem Cordillerischen Eisschild nördlich und östlich des Columbia Plateau standen. Im Süden hatten die Falten der Yakima Senke starken Einfluß auf den Fluss der Flutwasser in das Pasco-Becken hinein und hinaus.

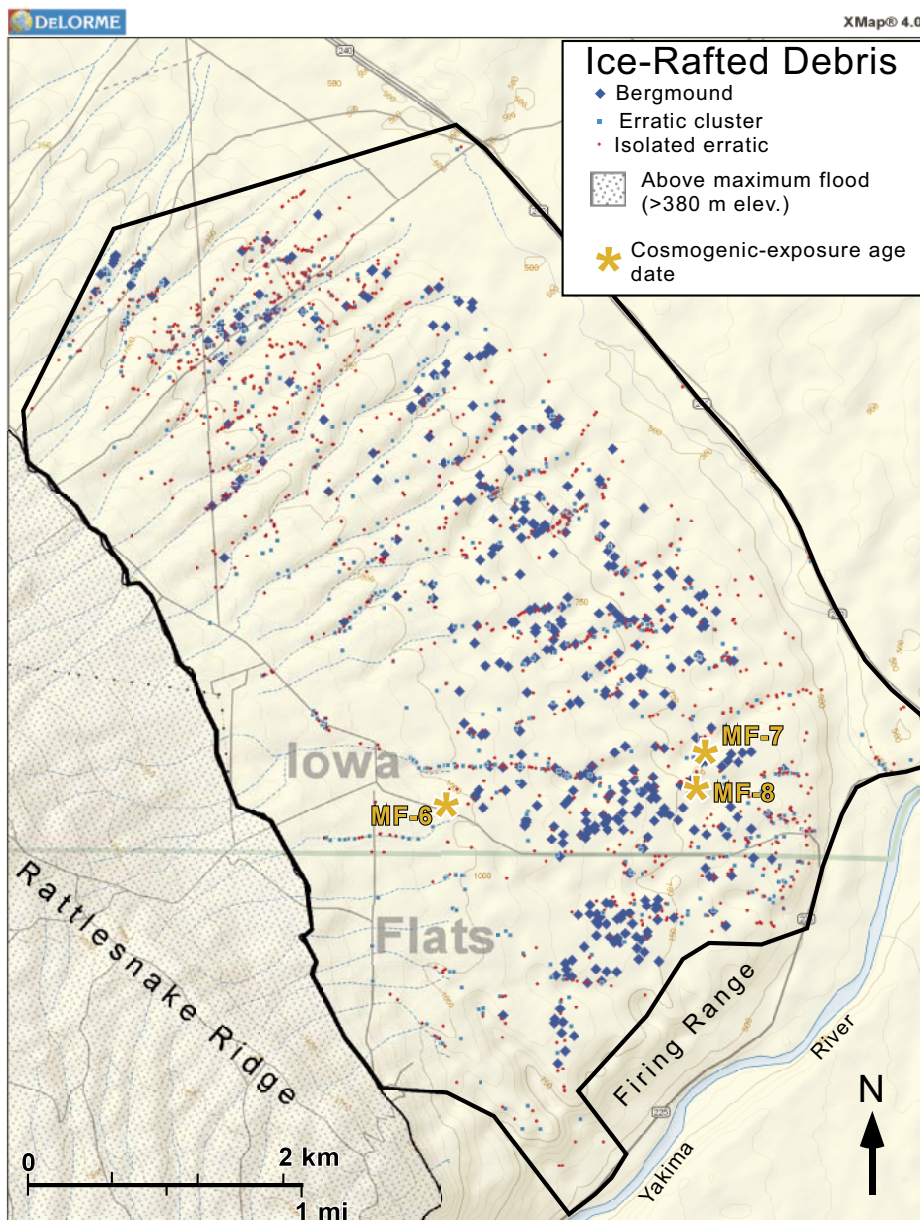


Fig. 7: Distribution of the three different types of ice-rafted debris across the study area. Details on the three erratics sampled for cosmogenic-nuclide exposure dating are listed in Table 1.

Abb. 7: Drei verschiedene Typen von eisverfrachteten Schutt sind ungleichmäßig im Untersuchungsgebiet verteilt. Einzelinformationen zu drei Findlingen, die für die Radionuklidmessung des Betrachtungsalters ausgewählt wurden in Tabelle 1 aufgelistet.

weathered are bright white and show no surficial oxidation or mineral decomposition. In contrast, strongly weathered granitic erratics displayed advanced stages of mineral decomposition and surface roughness, often with deep reddish brown or dark varnished surfaces. Slightly weathered to moderately weathered erratics displayed intermediate stages of roughness, oxidation and mineral decomposition (Figure 8).

Cosmogenic ^{36}Cl exposure-age dates were obtained from three large, dispersed, ice-rafted erratics reported in KESZTHELYI et al. (2009) and represented in Figure 7 and Table 1. Two of the analyses were performed on granitic boulders; the third was performed on an ice-rafted boulder of basalt. Because basalt is the only indigenous rock type and completely underlies the study area and crops out in the steep slopes of Rattlesnake Mountain to the southwest, special care is needed to positively identify basalt erratics. Some might argue that basalt does not qualify as a true “erratic” since it makes up the underlying bedrock at Rattlesnake Mountain, however, in this paper basalt is included as an erratic when it clearly has an ice-rafted origin. Accordingly, only basalt

boulders that were: 1) well beyond the limits where rocks could have rolled or moved down from higher elevations via mass movement, and (2) had several metres or more of fine-textured post-basalt sediment between basalt boulders and underlying bedrock were classified as erratics.

5 Results and Observations

On Rattlesnake Mountain, isolated erratics (52% of all ice-rafted debris) far outnumber erratic clusters (29%), which exceed bergmounds (19%). However, because many erratics smaller than a metre may be concealed beneath a cover of vegetation, the true proportions of isolated erratics and clusters may be greater than that indicated above. The highest concentration of ice-rafted debris lies near 210 m with a general decrease above and below this elevation (Figure 9). As previously noted by BRETZ (1930) bergmounds are mostly concentrated below 260 m elevation, and generally absent above 300 m (Figures 7 and 9).

As discussed above, mapping of erratics on Rattlesnake Mountain was limited to only those with an exposed cross-

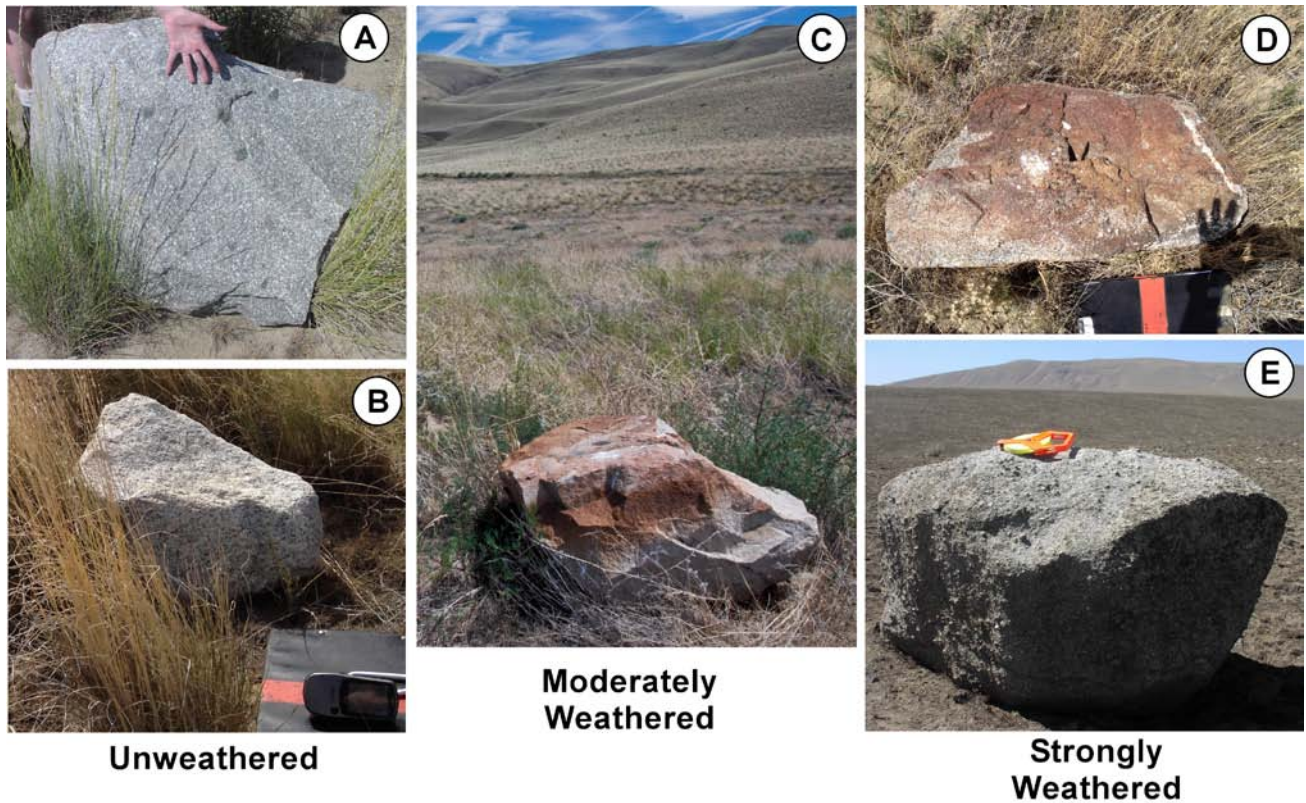


Fig. 8: Stages of weathering generally indicated by degree of surface roughness and amount of iron oxidation. All erratics are granitic except for the diorite boulder at upper left (A). Note advanced stage of surface roughness in photo E whereby the more-stable feldspar phenocrysts stand out in relief due to differential weathering of mineral grains.

Abb. 8: Verwitterungsphasen werden im Allgemeinen durch den Grad der Oberflächenrauheit und den Umfang der Eisen-Oxidation gekennzeichnet. Alle Findlinge sind Granite – mit Ausnahme des Diorit-Findlingsbrockens oben links (A). Man beachte das fortgeschrittene Stadium der Oberflächenrauheit in Foto E, Aufgrund unterschiedlichen Witterungsverhaltens der Mineralbestandteile treten die stabileren Feldspat-Phänokristalle reliefartig hervor.

sectional area $\geq 0.093 \text{ m}^2$. The majority (60%) were $< 0.37 \text{ m}^2$, while less than 1% were larger than 6.0 m^2 (Figure 10A). The largest erratic (12.5 m^2) found in the study area occurred as an isolate at a relatively high elevation (310 m).

The distribution of erratics by lithology is shown in Figure 11. More than 95% of erratics consist of rocks that are very different from indigenous, dark-colored basalt (Figure

12). The majority (~75%) are light-colored granitic, crystalline rocks (Figures 10B) like those shown in Figures 4A, 4B, and 8. However, the number of mapped erratics may be biased in favor of light-colored granitic rocks because they are more visible in the field. Most granitic erratics consist of granodiorite but also include true granites. Granitic rocks are 10 times more abundant than the next most common

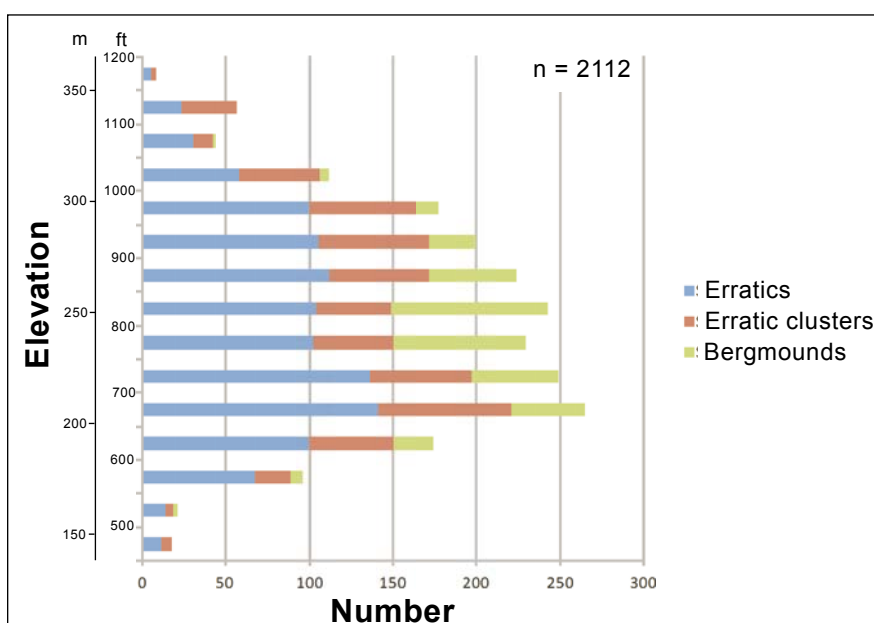


Fig 9: Concentration of ice-rafted debris vs. elevation.

Abb. 9: Konzentration von eisverfrachtetem Schutts in Bezug zur Höhenlage.

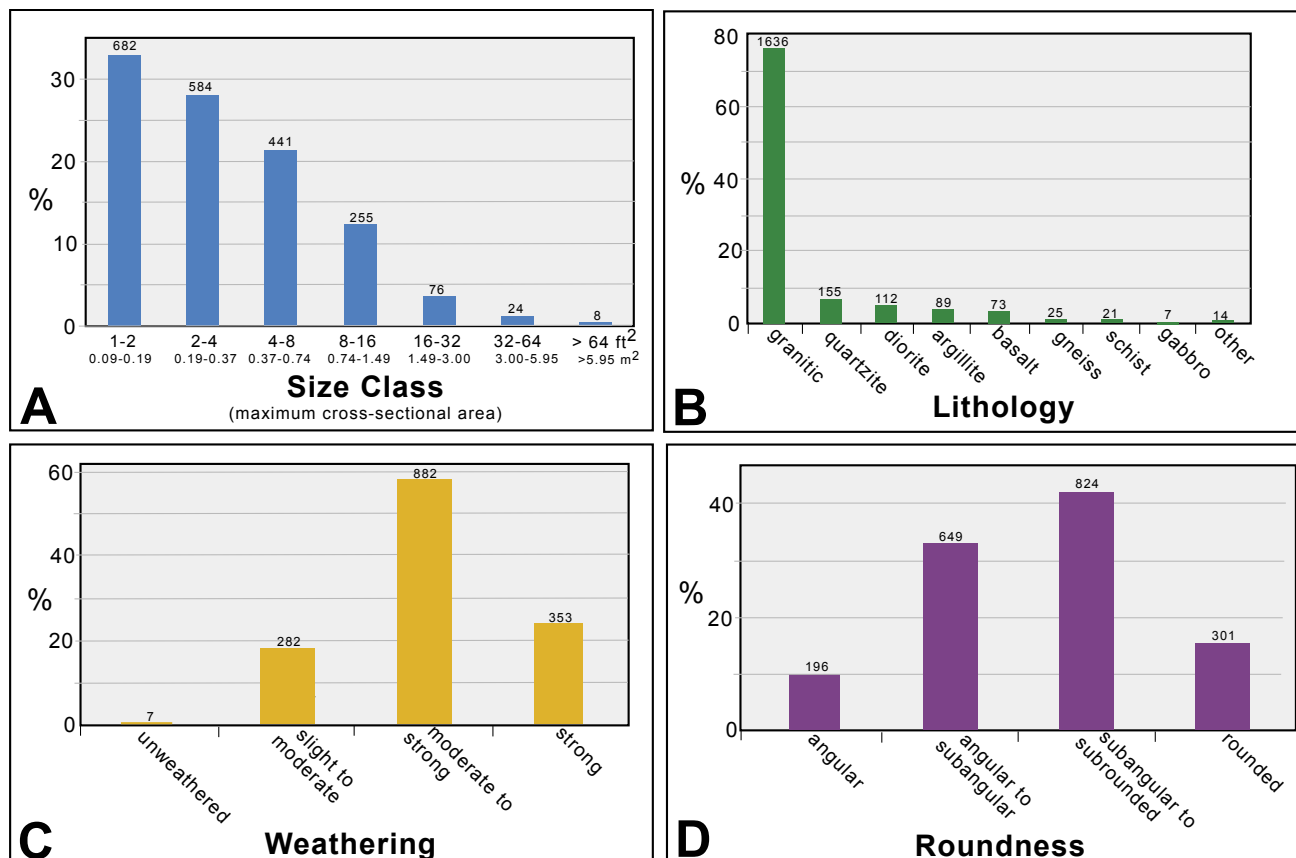


Fig. 10: Frequency distributions of Rattlesnake Mountain erratics by: (A) size, (B) lithology, (C) weathering, and (D) roundness. Weathering and roundness pertain only to erratics of granitic composition for reasons explained in the text.

Abb. 10: Dichte der Findlingsverteilung am Rattlesnake Mountain unterteilt in: (A) Größe, (B) Lithologie, (C) Verwitterung und (D) Rundheit. Verwitterungsgrad und Rundheit beziehen sich ausschließlich auf granitartige Findlinge, (Begründung im Text).

rock type within the study area – quartzite. Quartzites occur in a wide variety of colors, including pastel shades of blue, green, yellow, pink, and orange as well as white and gray. Diorite, an intrusive igneous rock containing about equal amounts of speckled light and dark minerals, is the third most common rock type. Next most abundant is argillite, an extremely hard, siliceous, dull gray to reddish brown rock that weathers and parts evenly along distinctive bedding planes (Figures 5 and 12). The fifth most common erratic type is Columbia River basalt. Other, less-common erratic lithologies include gneiss, schist, and gabbro.

The most notable surface characteristic on erratics is degree of weathering, which includes wind polish, mineral

decomposition, and oxidation rinds. Many strongly weathered granitic erratics show either a reddish-brown oxidative coating or occasionally a spalled, high-relief surface (Figure 8E). Because different lithologies weather at different rates, a comparison of weathering reported here was only performed on only the most common rock type (granitic), which makes up three quarters of all erratics (Figure 10B). More than 80% of the granitic erratics appear moderately to strongly weathered (Figure 10C). Striations and grooves, presumably derived from glacial abrasion at their places of origin, were also occasionally observed on surfaces of ice-raftered boulders.

The surface roundness of erratic clasts varies from angu-

Table 1: Characteristics and ³⁶Cl-exposure ages for three, scattered, ice-raftered erratics on Rattlesnake Mountain. See Figure 7 for relative locations within the study area.

Tabelle 1: Charakteristika und Ergebnisse der ³⁶Cl- Altersbestimmung von drei verstreut liegenden, eisverfrachteten Findlingen am Rattlesnake Mountain. Jeweilige Position im Untersuchungsgebiet: siehe Abbildung 7.

Field ID	Sample No.*	Cl-36 Age*	Latitude (north)	Longitude (west)	Elevation in m (ft)	Maximum Exposed Area in m ² (ft ²)	Roundness	Lithology	Shape	Weathering
E-223	M F-6	16,930	46.37878	-119.50037	310 (1,016)	12.5 (135)	subrounded	granite	elongate	strong
E-276	M F-7	16,170	46.38400	-119.46353	209 (686)	5.1 (55)	rounded	granite	equant	strong
M ERYK1	M F-8	16,740	46.38093	-119.46473	214 (702)	1.1 (12)	subrounded	basalt	irregular	moderate

* from Keszei et al. (2009)

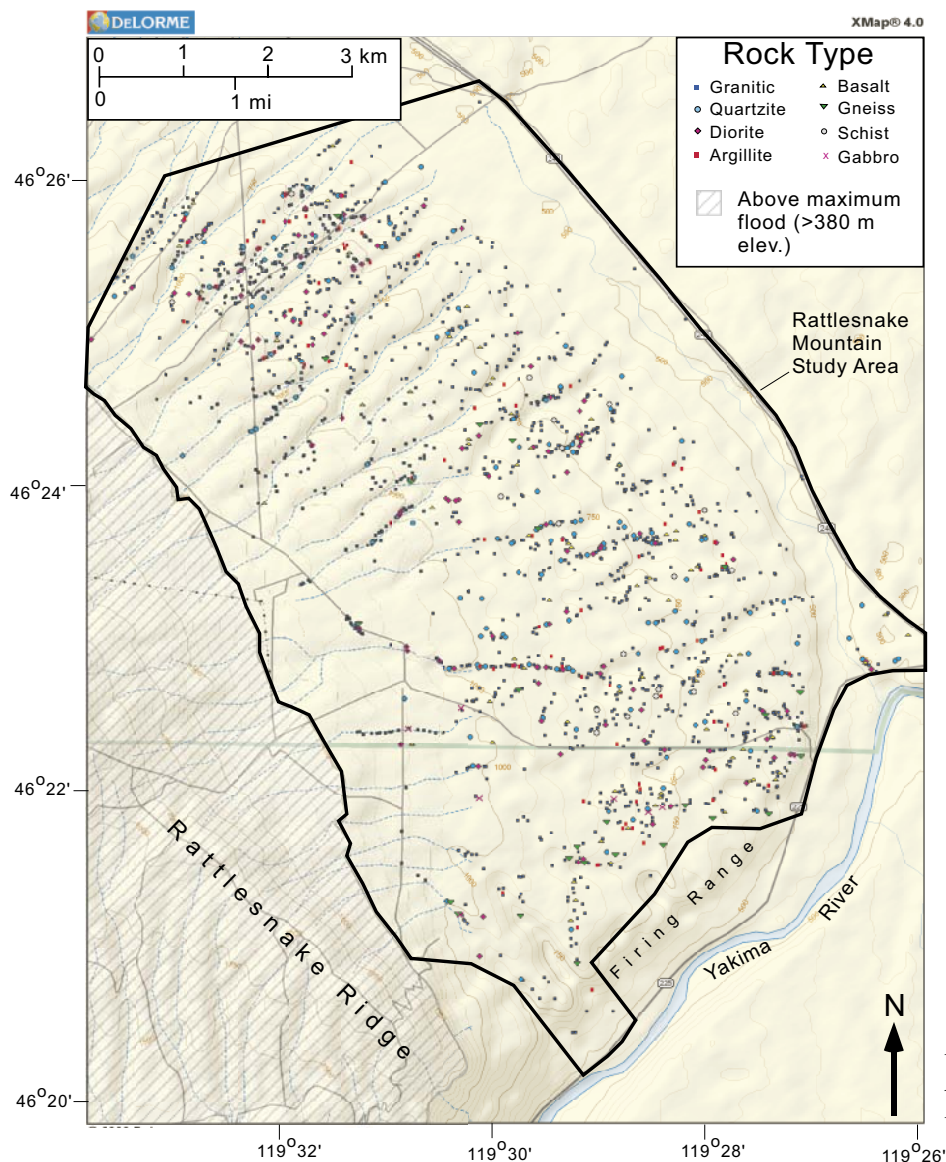


Fig. 11: Distribution of erratics by lithology.
Abb. 11: Verteilung von Findlingen gemäss Lithologie.

lar to rounded (Figure 10D). However, the majority of erratics were classified as subangular to subrounded. Shapes of angular to subangular boulders, argillites for example, often part along parallel bedding planes, leaving them flat to tabular, or rectangular. Plutonic crystalline rocks, on the other hand, are overall more equi-dimensional to elongate, often forming a pyramid shape. In general, the shape of over one-third of all the erratics is irregular.

Cosmogenic-exposure ages on three Rattlesnake Mountain erratics (Table 1) fall into a relatively narrow range between 16,000 to 17,000 cal yrs BP (KESZTHELYI et al. 2009). These ages fall within the range (15,000 to 20,000 calendar years) for Late Wisconsin Missoula floods reported elsewhere (O'CONNOR & BENITO 2009).

The distribution patterns of ice-rafted debris offer important clues to the way the debris was deposited. Summarized here:

- 1.) The highest observed erratic found within the study area lies at an elevation of ~362 m.
- 2.) Most ice-rafted debris within the study area is distributed along a broad band at the base of Rattlesnake Mountain

that lies between 200 to 300 m elevation (Figures 7 and 9). Fewer erratics and very few bergmounds lie above 300 m, despite the fact that the largest floods rose up to 80 m higher.

- 3.) Many isolated erratics and erratic clusters are aligned with gullies (Figure 7). At higher elevations erratics lie at the bottoms of low-relief gullies. In contrast, at lower elevations erratics are usually perched onto the north (up-current) sides of the higher-relief gullies.
- 4.) A higher concentration of erratics lies along roadways.
- 5.) Lithology of erratics versus elevation appears random (Figure 11), as does the distribution of erratic size (Figure 13).

6 Discussion

Concentrations of Ice-Rafted Debris

Different processes are likely responsible for erratics, erratic clusters and bergmounds. Isolated erratic boulders represent either dropstones from free-floating icebergs, or from



Fig. 12: Erratic lithologies on Rattlesnake Mountain.

Abb. 12: Lithologie der Findlinge vom Rattlesnake Mountain.

smaller, relatively “clean”, grounded icebergs. Erratic clusters likely melted out of small to moderate-sized, grounded icebergs, while bergmounds are probably associated with especially large, sediment-laden and grounded icebergs (BRETZ 1930).

The results of this study support BRETZ et al. (1956) who believed the north flank of Rattlesnake Mountain displayed the greatest concentration of berg-rafter debris anywhere within the area impacted by outburst floods. This appears to apply to both erratics as well as bergmounds, which are rarely reported outside the Pasco Basin (ALLISON 1933). More bergmounds and erratics accumulated here due to floodwaters that backflooded and pooled behind Wallula Gap, the first major constriction along the route of the floods beyond the Channeled Scabland. In the Pasco Basin, flow expansion and rising floodwaters behind Wallula Gap caused a temporary slowing of floodwaters (BJORNSTAD 2006). The slowing currents naturally directed icebergs to the quieter margins of

the basin like Rattlesnake Mountain and backflooded valleys such as the Yakima and Walla Walla valleys (Figures 1 and 2), especially during the waxing stages of flooding (BAKER et al. 1991). Far fewer erratics occur in unrestricted, higher-gradient reaches of the floods, like Wallula Gap, because of higher flow velocities that kept icebergs moving through the system.

Iceberg-deposited erratics occur downstream of Wallula Gap but are much less concentrated than on Rattlesnake Mountain. In the backflooded Willamette Valley (south of Portland, Oregon), for example, only about 1.9 erratics/km² were identified by MINERVINI et al. (2003), compared to almost 38.6 erratics+bergmounds/km² within the Rattlesnake Mountain study area. However, it is acknowledged that the concentration of erratics may be somewhat biased in favor of Rattlesnake Mountain because rafted debris is more visible in the semi-arid, low shrub-steppe of eastern Washington, compared to the more humid and densely vegetated

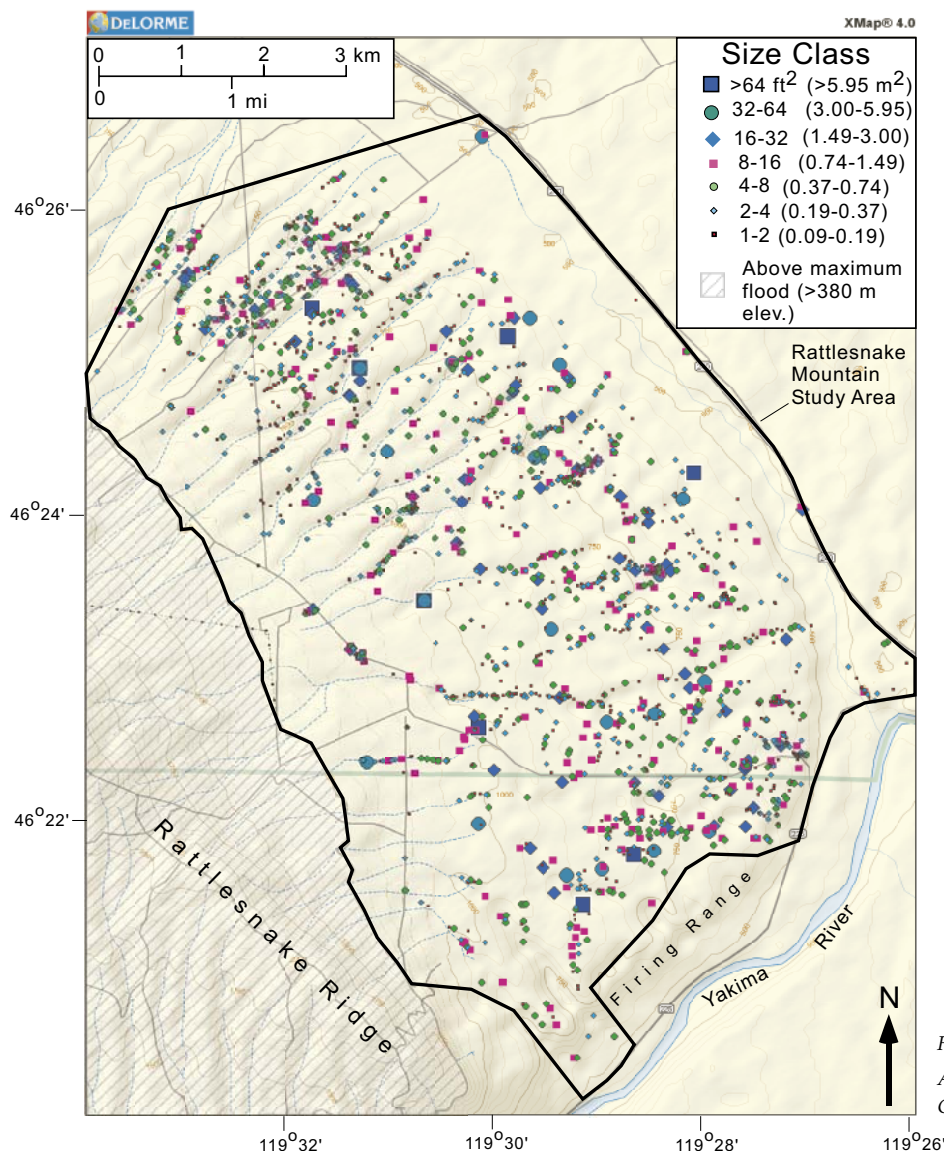


Fig. 13: Distribution of erratics by size.

Abb. 13: Verteilung von Findlingen gemäss Grösse.

Willamette Valley. Although many erratics of the Willamette Valley could be buried within or beneath a cover of Quaternary sediments so too are many of the erratics on Rattlesnake Mountain.

As illustrated in Figures 7 and 9 the highest concentrations of erratic debris lies partway up the slopes of Rattlesnake Mountain between 200- and 300 m elevation. The decrease in erratic debris at lower elevations may be explained by increased flow velocity that developed across the lower slopes as the last of the floodwaters in Lake Lewis drained from the basin. Accordingly, the faster flows were sufficient enough to prevent iceberg stagnation. Another possibility is more erratics lie buried beneath a cover of Holocene-age eolian, slopewash or fluvial sediment at lower elevations. The decrease in erratic debris above 300 m is consistent with what is known about the sizes of the many dozens or more outburst floods, only a limited number of which were extremely large (BENITO & O'CONNOR 2003). The peak discharges for many of the floods were relatively small (<1 million m³/sec) compared to the largest flood(s) estimated at 17.5 million m³/sec (O'CONNOR & BAKER 1992). Naturally, smaller outburst floods would not pond as deep behind hydraulic constrictions at Wallula Gap and the Columbia River Gorge.

Erratic debris was observed up to an elevation of 362 m on Rattlesnake Mountain. Elsewhere in the Pasco Basin the author has observed erratics up to an elevation of 366 m on Red Mountain (N46.301, W119.448). This height is close to 372 m – the base elevation of a high flood-spillover channel along the east side of Wallula Gap (N46.0327, W118.9186) that is incised to an elevation of 378 m. Thus, two independent lines of evidence appear to place the highest level for outburst flooding in the Pasco Basin and upper Wallula Gap to at least 366 m and perhaps as high as 380 m shown in Figures 2 and 3. Highest floodwater indicators downstream of Wallula Gap are only about 335 m elevation, therefore it appears that some hydraulic gradient still existed between the upper and lower ends of Wallula Gap during the largest floods, although the flow was later impeded by ponding that shifted downstream into the constricted reaches of the Columbia River Gorge (BENITO & O'CONNOR 2003). These constrictions facilitated the high water levels observed in Wallula Gap – especially during waning stages of flooding.

Size of Erratics

The size of erratics appears to be random with respect to elevation (Figure 13) (i.e., erratics of variable size exist up

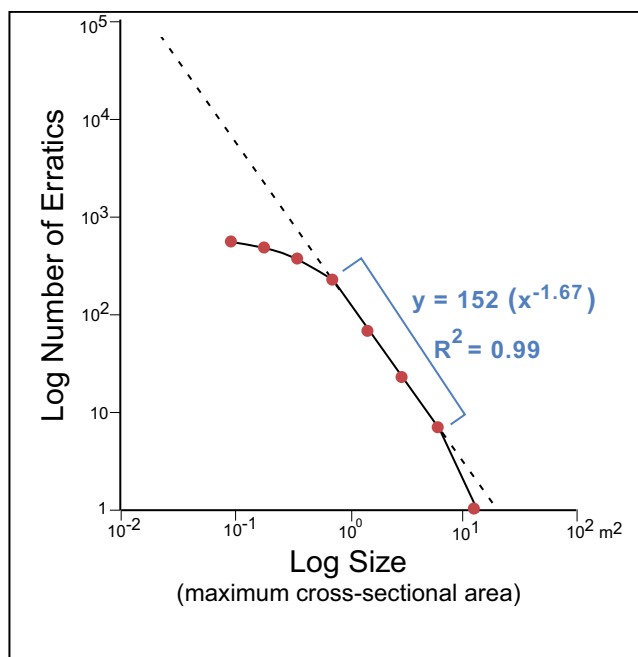


Fig. 14: Log-log plot of size vs. number of erratics measured on Rattlesnake Mountain. A linear log-log distribution may exist for erratics larger than 0.75 m² in area. The straight-line equation and correlation coefficient ($R^2 = 0.99$) are based on the number of erratics mapped between the 0.75 m² and 6 m² size classes.

Abb. 14: Log-Log plot (Anmerkung: zweidimensionale grafische Darstellung von Daten auf Logarithmenbasis): Größe vs. Anzahl ausgemessener Findlinge auf Rattlesnake Mountain. Eine lineare Log-Log Verteilung existiert möglicherweise für Findlinge größer als 0,75 m² im Untersuchungsgebiet. Die Geradengleichung und der Korrelationskoeffizient ($R^2 = 0,98$) beruhen auf der Zahl der Findlinge in den Größenklassen von 0,75 m² sowie 6 m².

to 362 m elevation). The number of erratics, distributed over seven different size classes, is shown in Figures 10A and 14. The log-log trend of smaller erratics appears to drop off around <0.75 m², which may be due to sampling bias. Erratics larger than 0.7 m² are more likely to extend above the low shrub-steppe vegetation rendering them more visible and mappable. Smaller erratics, on the other hand, are less conspicuous and more likely to be overlooked, especially if located between survey transects.

In general, considerable amounts of erratic debris smaller than 0.09 m² were observed to litter the surface, often multiple clasts per square metre, especially below 300 m elevation within the study area. Even though it was not practical to map debris of this size, it might be possible to estimate the abundance of the <0.09 m² size class based on the log-log plot illustrated in Figure 14. Accordingly, the number of larger erratics (0.7 m² to 6 m²) appear to plot along a generally straight-line, log-log function for the number of erratics versus size expressed as:

$$N = 152 A^{-1.67}$$

where N = number of erratics, and A = maximum exposed cross-sectional area in m².

Assuming erratics <0.75 m² follow a similar straight-line log function, the actual number of erratics in the 0.1–0.2 m² size range might be on the order of several thousand, which is considerably more than the 682 actually observed within the study area (Figure 10A). Accordingly, the total number of even less-visible erratic clasts down to the next smaller size class (0.01 m²) may exceed 300,000 based on the above equation. However, while it is possible the above equation

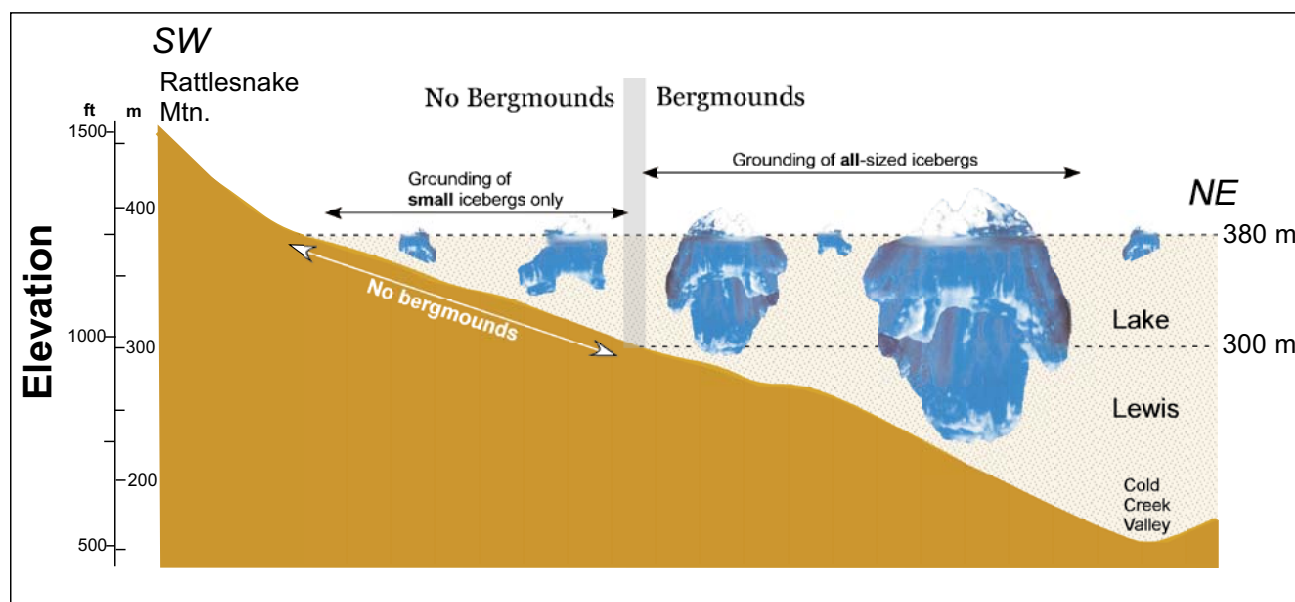


Fig. 15: A diagram to illustrate elevation-limited distribution of bergmounds on Rattlesnake Mountain. The largest icebergs were grounded on the gently sloping lake bottom, further from the ancient shores of short-lived Lake Lewis. Few well-developed bergmounds are observed above ~300 m elevation, even though floodwaters extended up to 380 m. Apparently, during the largest flood(s) all bergmound-producing icebergs became grounded below 75 vertical metres of maximum lake level because of their greater size.

Abb. 15: Diagramm zur höhenbegrenzten Verteilung von Bergmounds auf Rattlesnake Mountain. Die größten Eisberge liefen weit entfernt von der alten Uferlinie von Lake Lewis auf der sanft abfallenden Seesohle auf Grund. Obwohl der See bis auf 380 m Seehöhe anstieg wurden nur wenige gut entwickelte Bergmounds oberhalb von 300 m Seehöhe gefunden, Während der größten Fluten liefen offenbar alle Bergmound erzeugenden Eisberge aufgrund ihrer Größe unterhalb von 75 Höhenmeter unter dem maximalen Seespiegel auf Grund.



Fig. 16: Diagram showing variable flow and speed of floodwater across a portion of the study area. Where floodwaters were deeper to the right, eddy currents caused icebergs to collect and ground on upstream sides of gullies while in shallower more-restrictive waters (left side) icebergs tended to ground along the bottoms of gullies. Elevations in this standard USGS topographic map are in feet (conversion to metres = 0.3048).

Abb. 16: Diagramm zur unterschiedlichen Abflussmenge und Geschwindigkeit von Flutwassers in einem Teil des Untersuchungsgebietes. Wo die Flutwasser auf der rechten Seite tiefer waren, verursachten foucaultsche Wirbelströme („eddy currents“), so dass sich Eisberge stromaufwärts an den Seiten der Ablaufrinnen stauten und auf Grund liefen, während in flacherem, eingegrenzterem Wasser (auf der linken Seite) die Eisberge eher auf dem Grund der Ablaufrinnen strandeten. Topographische Höhe in Fuß (Umrechnung in Meter = 0.3048).

may not accurately reflect the true number of erratics for different size classes, the known bias introduced in size versus visibility suggests a log-log distribution may in fact represent a reasonable estimate.

Sources of Erratic Debris

The lithology of erratics is consistent with an origin from areas once in contact with the Cordilleran Ice Sheet. Argillite is closely associated with quartzite; both are meta-sedimentary rocks derived from the Proterozoic Belt Supergroup (ORR & ORR 1996). These rocks crop out along the eastern side of the Purcell Trench at the site of the ice dam for glacial Lake Missoula (MILLER et al. 1999; BJORNSTAD & KIVER, 2012). Other erratics include Cretaceous and Eocene plutonic rocks such as granite and granodiorite and high-grade metamorphics (i.e., gneiss, schist), which crop out along the western side of the Purcell Trench and in northern Washington (STOFFEL et al. 1991) (Figure 6).

Some basalt, while indigenous to the area, was also entrapped in glacial ice and carried along with outburst floodwaters. Although Columbia River basalt did not come in direct contact with the ice dam for Lake Missoula, there are other places downstream where the ice sheet overrode basalt: 1) Okanogan lobe where it extended across Waterville

Plateau (Figure 1), and 2) the Colville Valley (Figures 1 and 6). Thus, ice-rafted basalt may be derived from: 1) icebergs that calved off the Okanogan Lobe (and possibly the Colville Lobe) into glacial Lake Columbia, or 2) a final flood associated with the breakup of Lake Columbia several centuries after the last Missoula flood (ATWATER 1987; WAITT 1994; WAITT et al. 2009).

Surficial Weathering and Roundness of Erratics

The majority of erratics (especially granitic) on the surface show some degree of roundness and weathering (Figure 8). Yet the spatial distribution of surface weathering and rounding appeared random with respect to elevation and location within the study area. Because ice-rafted erratics undergo little or no abrasion during ice transport, the rounding must have occurred either before or after melting out of icebergs. Like erratics on Rattlesnake Mountain, many glacial erratics observed today near the former Cordilleran Ice Sheet show some degree of rounding, either inherited from their place of origin prior to ice transport, or due to in-situ weathering since the end of the Pleistocene.

Although it may be tempting to attribute weathering and roundness to in-situ weathering (via exfoliation and spalling) since iceberg emplacement, this does not appear to

be the case based on a limited number of cosmogenic-exposure ages on erratics (KESZTHELYI et al. 2009). Three dated erratics shown in Table 1 are variably weathered (two of the three strongly weathered), yet their exposure ages fall into a narrow range of time. All three erratics are associated with the last cycle (Late Wisconsin) glaciation between 16 and 17 thousand years. This amount of time appears insufficient to produce a strongly weathered erratic surface especially considering some granitic erratics on Rattlesnake Mountain show little or no weathering. Therefore, the degree of weathering and rounding of granitic erratics does not appear to be a reliable indicator of age since grounding of erratics and suggests at least some weathering and roundness of erratics is inherited from their place of origin, prior to being ice rafted to Rattlesnake Mountain.

Because of the extremely limited number of dated erratics, however, the possibility still exists that other, undated erratics could have undergone strong weathering in situ. Outburst floods from previous glacial cycles going back to Early Pleistocene (>780 ka) are known to have occurred in eastern Washington (PATTON AND BAKER 1978; BJORNSTAD et al. 2001; PLUHAR et al. 2006). Erratics associated with these earlier flood events would be expected to display strong weathering if continuously exposed at the ground surface. On the other hand, those erratics derived from pre-Wisconsin floods that lie buried beneath an aggrading mantle of post-basalt Pleistocene sediments may have escaped surficial weathering.

Origin of Bergmounds

Although it seems logical to ascribe bergmounds (Figure 4C) entirely to the buildup of ice-rafted debris from large, sediment-laden icebergs, the model for bergmound development may be more complicated. Not all topographic relief observed on bergmounds is necessarily constructional. Some of the relief may also be the result of increased erosion around the flanks of the bergmound, which occurred during or since flooding. This is suggested from trenching studies of bergmounds, which reveal at least some of the interiors are composed of mostly fine-grained sediments, which are not necessarily derived from ice-rafted debris (FECHT & TALLMAN 1978; CHAMNESS 1993, 1994). Thus, coarser ice-rafted debris may merely blanket bergmounds creating a lag-gravel cap that acts to armor and protect the underlying finer-grained sediments. Areas between bergmounds, lacking a coarse, armored cap, then were more susceptible to erosion by either: 1) receding floodwaters, or 2) post-flood eolian deflation. However, problematic is the almost perfectly symmetrical and conical shape of bergmounds that show little or no sign of streamlining, which might be expected if subsequently eroded by either moving floodwater or wind.

Distribution of Ice-Rafted Debris

BRETZ et al. (1956) noted two populations of ice-rafted debris along Rattlesnake Mountain. They reported bergmounds up to an elevation of 260 m; while single erratics or groups of erratics extended up to 335 m elevation they speculated the two populations were the result of two separate flood events. However, knowing there were at least dozens (WAITT 1980) to perhaps hundreds of separate flood events (BJORNSTAD 2006) of various magnitudes, a different explanation for

the distribution of ice-rafted debris shown in Figure 7 seems warranted.

Based on Archimedes' principle the minimum size of an iceberg needed to float a boulder of granite composition (average density ~2.70 g/cm³) is estimated at about 20 times the volume of the erratic boulder. In marine waters only the upper one-eighth of icebergs rise above water level (BRUNEAU 2004). However, due to higher sediment loads and densities of continental floodwaters, it is likely icebergs were even more buoyant and floated higher when compared to marine water containing little or no suspended sediment. Nevertheless, the bulk of floating icebergs lie underwater and protrude well below the waterline. Essentially, the larger the iceberg the deeper it extended into temporary Lake Lewis (Figure 15). It follows then that larger icebergs would become grounded well away from the ancient shores of Lake Lewis. This would explain why far fewer bergmounds lie above 300 m elevation. Furthermore, the greater concentration of ice-rafted debris at lower elevations (Figure 7) is consistent with evidence to indicate there were many more smaller floods versus large floods toward the end of the Wisconsin Glaciation (WAITT 1980; BENITO & O'CONNOR 2003).

Flow Dynamics

A notable pattern, illustrated in Figure 7, is the alignment of many erratics and erratic clusters along gullies. At lower elevations, erratics are concentrated along the south-facing (upstream) slopes of gullies. At higher elevations, erratics lie mostly along the bases of these same gullies. The pattern of concentration for erratics along gullies is attributed to eddy currents set up by floodwaters flowing at slightly different velocities across the uneven surface of Rattlesnake Mountain. It is envisioned that, as floodwater levels rapidly lowered, icebergs were trapped in eddies that circulated in the deep and slower moving water within the gullies. At higher elevations where the lake was shallower, deep-rooted icebergs migrated to the deeper water overlying the bottoms of gullies. At lower elevations icebergs were less constrained because of deeper water and therefore more likely to become grounded on the high margins to the upstream side of gullies where eddy circulation concentrated their movement (Figure 16).

Erratics are also more concentrated along roadways within the study area; this concentration suggests exhumation during road construction from fine-textured, post-basalt sediments. Furthermore, cisterns dug by early 20th century settlers in this older loose, fine sediment often produced erratic boulders. These observations indicate: 1) many erratics from older floods lie buried within the post-basalt sediment cover, and 2) there are far more erratic boulders on Rattlesnake Mountain than those currently exposed at the surface. Some of these buried boulders could be associated with older, pre-Wisconsin (>120 ka) outburst floods (PATTON & BAKER 1978; BJORNSTAD et al. 2001).

7 Conclusions

Ice-rafted debris accumulated in slackwater areas up to an elevation of ~365 m within south-central Washington during repeated Pleistocene cataclysmic floods. Erosion by floods along the basin outlet at Wallula Gap indicates maximum

flood depths approached 380 m. Floodwaters impounded behind Wallula Gap temporarily created Lake Lewis and deposited ice-rafted erratics and bergmounds along the gentle slopes of Rattlesnake Mountain. Ice-rafted debris is of three types: 1) isolated erratics, 2) erratic clusters, and 3) bergmounds. Unlike erratics and clusters, bergmounds display topographic relief in the form of scattered, low-relief, conical mounds.

The present study of ice-rafted debris was performed in a long-protected, sparsely vegetated, 60 km² area on the north-eastern flank of Rattlesnake Mountain. More than 2,100 locations of erratics >0.09 m² and bergmounds were recorded with a GPS unit. Additional information was gathered on 1) elevation, 2) lithology, 3) size, 4) roundness, 5) shape, and 6) surface characteristics of erratics. All but a few percent of erratics consist of rocks other than indigenous basalt; approximately 75% are of granitic composition. Other lithologies, in order of decreasing abundance, are quartzite, diorite, argillite, basalt, gneiss, schist, and gabbro.

³⁶Cl-exposure ages on three, widely dispersed erratics were 16–17 ka, suggesting all were emplaced during the last glacial cycle (i.e., Late Wisconsin). Most erratics are either subrounded or rounded, followed by subangular; angular clasts are least common. Greater than three quarters of erratics are moderately to strongly weathered, but a wide range of weathering and roundness was observed on the cosmogenically-dated erratics suggesting these features are likely inherited from their place of origin.

The distribution of ice-rafted debris is non-uniform; maximum concentration is around 200 m elevation. Bergmounds are mostly restricted to a band between 200- and 300 m elevation; especially large icebergs that produced bergmounds could not approach the shores of the ancient shoaling lake, which extended up to 80 m higher. As floodwaters moved across an uneven surface backwater eddies concentrated icebergs and their entrapped erratics along the upstream sides of northeast-trending gullies within the study area. This study affirms that the north slope of Rattlesnake Mountain likely contains the most prolific erratic and bergmound population of anywhere along the route for glacial-outburst floods in the northwestern US.

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Driving mechanisms of Holocene lagoon development and barrier accretion in Northern Elis, Peloponnese, inferred from the sedimentary record of the Kotychi Lagoon

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Abstract:

As part of the lagoon barrier accretions plain characterizing the NW coast of the Peloponnese, the Kotychi Lagoon is believed to have formed in the prograding delta of the Palaeo-Peneus River over 7000 years ago. Geochemical/sedimentological proxies-data (XRF, grain size, OC-, IC-, C/N-analysis) combined with Bayesian age-depth-modeling revealed that from 8500 to 8000 cal BP marine conditions were prevailing. Around 8000 cal BP, a short-lived sequence of coastline progradation and barrier accretion created lagoonal conditions. Thus, the first chronological control for the onset of lagoon formation in coastal Elis is presented. Pronounced lagoonal conditions developed approximately 6300 cal BP, simultaneously to the period of circum-Mediterranean lagoon formation. A rapidly varying sedimentary record indicates a phase of geomorphological instability between 5200 and 3500 cal BP terminating with the erosional unconformity of a river channel. This evolution reflects a two-phase development: (1) Early Holocene morphology was controlled by the postglacial sea level rise; (2) with receding of the ice sheets by mid-Holocene, the preeminent role of the eustatic signal was overwhelmed giving local and regional processes, such as human-induced soil erosion and climatic forcing an accentuated role. Thus, the evolution of the Elean coastline shows analogies to circum-Mediterranean lagoon formation.

Mechanismen Holozäner Lagunenentwicklung und Küstenwachstum in der nördlichen Elis (Peloponnes, Griechenland), abgeleitet aus dem Sedimentarchiv der Kotychi Lagune

Kurzfassung:

Die Kotychi Lagune liegt im Nordwesten der Peloponnes und hatte ihren Ursprung vor mehr als 7000 Jahren im progradierenden Delta des Paläo-Peneus. Anhand geochemischer und sedimentologischer Proxydaten (XRF, Korngröße, OC, IC, CN) kombiniert mit bayesianischer Alters-Tiefen-Modellierung wurden für den Zeitraum von 8500 bis 8000 cal BP marine Verhältnisse nachgewiesen. Eine erste Phase der Küstenprogradation um 8000 cal BP führte zu kurzlebiger Lagunenbildung, womit zum ersten Mal das Einsetzen der *lagunären* Bedingungen chronologisch erfasst wurde. Die Lagune manifestierte sich nachhaltig um 6300 cal BP, zeitgleich zu einer Phase weitverbreiteter Lagunenbildung im Mittelmeerraum. Zwischen 5200 und 3500 cal BP deutet eine rasche Abfolge variierender Sedimentationsbedingungen auf eine Phase geomorphologischer Instabilität. Abgeschlossen wird die sedimentäre Abfolge um 3500 cal BP mit der Erosionsdiskordanz eines Flusslaufes. Die rekonstruierte Entstehungsgeschichte der Kotychi Lagune lässt sich in zwei Phasen unterteilen: Die frühholozäne Küstenentwicklung wurde durch den postglazialen Meeresspiegelanstieg dominiert. Mit dem Abschmelzen der Eiskappen verlangsamte sich der Anstieg im mittleren Holozän. Dadurch schwächte die dominierende Wirkung des Meeresspiegels ab und lokale und regionale Faktoren wie klimatisch oder anthropogen induzierte Bodenerosion prägten die Küstenmorphologie durch ihren Beitrag zum Sedimentbudget. Somit zeigt die Genese der Kotychi Lagune Parallelen zu vielen Lagunen im Mittelmeerraum.

Keywords:

Greece, Holocene coastal evolution, XRF, geochemistry, lagoon, sea level change

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1 Introduction

Situated at the dynamic interface between land and sea, coasts are transient landscapes shaped by the interplay between macroscale (sea-level changes, climate, and tectonic stability) and microscale processes (sediment supply, topographical inheritance, wave energy, and tidal range (BAO et al., 2007)). Any change in the power of these controlling factors directly translates into morphological adjustment of the

coast (COOPER, 1994; BAO et al., 2007). Where rivers enter the sea, hinterland sediment supply associated with natural climate fluctuations or anthropogenic activity provides a significant contribution to the gross sediment budget and thus forms the coast. If the rate of sediments supplied by rivers exceeds the combined ability of waves and tides to carry away the sediment, a delta forms at the mouth of the river. Delivered to the delta front, sediments may be re-deposited to block the river mouth and shape the coast by forming

barriers, lagoons or marshes (McMANUS, 2002; KRAFT et al., 2005; DINIS et al., 2006). In this context, sea level has an exceptional role as it serves as the ultimate base level (PLATER & KIRBY, 2011).

During the Holocene, sea level exhibited significant fluctuations, mostly responding to the advance and retreat of the continental ice sheets (LAMBECK et al., 2010). In general, two distinct periods of sea level rise can be distinguished. An early period of rapid sea level rise spanning from 18000 to 6000 BP contrasts with a period of near stability from 6000 BP to the present. During the early phase, coastal evolution was predominantly controlled by the post-glacial sea-level, while by 6000 BP local and regional factors (isostasy, subsidence, sediment supply, anthropic activity) overwhelmed the power of attenuated sea level rise (BAO, FREITAS & ANDRADE, 1999; FREITAS et al., 2003; DINIS et al., 2006). This mid-Holocene sea level stabilization had major influence on fluvial sedimentation in coastal regions and triggering delta progradation on a global scale (STANLEY & WARNE, 1993). The formation of many Mediterranean lagoons for example on the Thessaloniki plain, Greece (FOUACHE et al., 2008), on the Iberian Peninsula (FREITAS et al., 2003; BAO et al., 2007; COSTAS et al., 2009), at the coastal plain of Marathon, Greece (PAVLOPOULOS et al., 2006) or on the northern Peloponnese, Greece (KONTOPOULOS & AVRAMIDIS, 2003) can be traced back to this period.

To understand coastal evolution and the driving mechanism, the sedimentary record of lagoons has great potential. As a net area of deposition, lagoons preserve the sedimentary signature of multiple processes acting upon the coastline and thus offer the possibility to trace spatio-temporal coastline migration and its causes.

The present study addresses the Holocene coastal evolution of the northern Elis, Northwest Peloponnese, Greece. It focuses on the coastal stripe between the limestone headlands of the Mavra Vouna (Black Mountains) in the north and Cape Kyllini in the south both connected by a crescentic beach isolating a coast parallel sequence of lagoons and marshes (Fig. 1). Out of these lagoons, the Kotychi Lagoon, serves as a geoarchive. Previous studies indicate a series of lagoons fringing the Elean coast for at least 7000 years (KRAFT et al., 2005; KONTOPOULOS & KOUTSIOS, 2010). However, there is no evidence about the onset of barrier-lagoon accretion in Elis. Research at the Elian coast extends back to RAPHAEL (1973, 1978) who combined geomorphological and archaeological evidence to identify cycles of coastal pro- and retrogradation from the late Holocene until Turkish Times. Likewise, a synthesis of geomorphology, literary sources and drill cores allowed KRAFT et al. (2005) to reconstruct mid-Holocene evolution of the Elean coastline. The authors identified a sequence of ancient lagoons and barrier islands resulting from the interplay of varying sediment fluxes and sea level rise. Recent studies at the Kotychi Lagoon focused on a local palaeogeographic reconstruction by pollen analyses of sediment cores (LAZAROVA, KOUTSIOS, & KONTOPOULOS, 2012) and the analysis of microfossil assemblage and sedimentology (KONTOPOULOS & KOUTSIOS, 2010).

The aim of this study is to decipher coastal morphogenesis of northern coastal Elis during the Holocene and to put the evolution into a general, circum Mediterranean context. Special emphasis is placed on the driving mechanisms of coastal

evolution and the role of post glacial sea level rise as one potential driver of change. In order to achieve these aims, high resolution XRF-data is combined with grain size and C/N analyses from a sediment core retrieved at the shore of the Kotychi Lagoon. As XRF-analysis is mostly applied to limnic or deep sea environments, this study is one of few using XRF-data to reconstruct coastline evolution.

2 Regional Setting

2.1 Geography and Geology

The study area comprises a coastal stripe in the northwestern part of the Peloponnese, which belongs to the northern part of the prefecture of Elis (Fig. 1).

The convergence of the African and the Eurasian plate which culminated in the Alpine orogeny is crucial for present geology and tectonic activity in the western Peloponnese because geological basement and major fault systems are related to this compressional regime. Located in the direct vicinity to the Hellenic arc, the center of subduction since Miocene times (23–7 Ma BP), the study area is part of one of the most tectonically and seismically active regions in Greece (Fountoulis et al., 2011). A detailed earthquake catalogue spanning the twentieth century is summarized by BURTONA et al. (2004). In the study area, Miocene diapirism caused by Triassic evaporates of the Alpine basement contributes to complicated tectonic deformation and high seismicity (UNDERHILL, 1988; MAROUKIAN et al., 2000). According to Fountoulis et al. (2011), major faults form neotectonic blocks which exhibit diverging trends of vertical movement. While uplift rates between 0.16 and 0.67 mm/year characterize the hinterland, the coastal strip is subsiding (Fountoulis et al., 2011).

The area is characterized by a broad coastal plain where the limestone headlands of Cape Kyllini, Cape Kunupeli and the Mavra Vouna are the only relicts of the Alpine formations. They comprise series of white to light brown limestones (geological map of Greece, sheet Amalias, sheet Nea Manolas, scale 1:50,000, IGME (1977); JACOBSHAGEN, 1986). Further remains of the alpine basement in this area are covered by Neogene to Pleistocene marine sandstones, sandy clays and clayey marls which intercalate with Holocene terrestrial and torrential clayey and sandy deposits to the east (geological map of Greece, sheet Amalias, sheet Nea Manolas, scale 1:50,000, IGME (1977); JACOBSHAGEN, 1986; MAROUKIAN et al., 2000). The latter represent the most extensive Quaternary alluvial deposits in the Peloponnese (RAPHAEL, 1978; POULOS et al., 2002). Primarily responsible for the accumulation of these deposits is the Peneus River, originating to the east in the Arcadian Mountains and currently entering the Ionian Sea south of Cape Kyllini. However, the palaeo-delta of the Peneus River is believed to have been located north of Cape Kyllini giving rise to a sequence of lagoons and marshes embedded in the prograding delta and fed by sediments from the Elean uplands. As a relic of this past river course, a system of ancient levees depicted in historic maps (cf. CURTIUS, 1851) has been detected a few kilometers south of the Kotychi Lagoon (Fig. 1) (PHILIPPSON, 1959; RAPHAEL, 1978; KRAFT et al., 2005; Fountoulis et al., 2011). An anthropogenic or natural (tectonic) diversion of the Peneus River is still subject of discussion but is believed

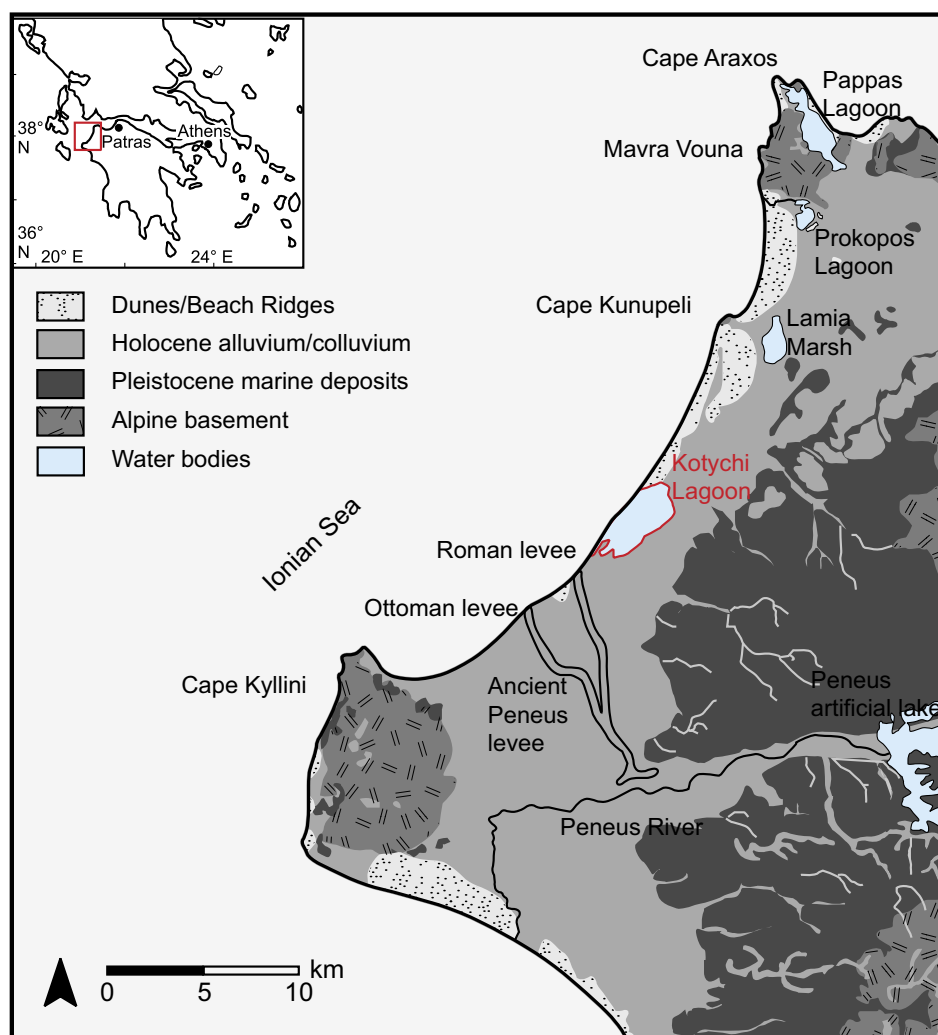


Fig. 1: The study area (modified after FOUNTOULIS et al., 2011).

Abb. 1: Das Untersuchungsgebiet (nach FOUNTOULIS et al., 2011, verändert).

to have occurred in the 17th century AD (KRAFT et al., 2005; KONTOPOULOS & KOUTSIOS, 2010; FOUNTOULIS et al., 2011). Starved of its sediment supply, erosion has dominated the north Elean ever since (KRAFT et al., 2005).

The high seismicity accompanied by gradual as well as episodically crustal displacement limits the utilization of a uniform sea level curve. The relative sea level data for the Elis coastal plain varied between 0.5 m/ka (VÖTT, 2007) and 0.95 m/ka (KONTOPOULOS & KOUTSIOS, 2010) from approximately 5000 cal BC to roughly 1000 cal BC and since then, increased to values of 0.7 m/ka (VÖTT, 2007) up to 1.2 m/ka (KONTOPOULOS & KOUTSIOS, 2010).

2.2 Lagoonal settings

As part of the coastal lagoon barrier accretions plain which characterizes the NW part of the Peloponnese, the Kotychi Lagoon is believed to have formed in the prograding delta of the Palaeo-Peneus River more than 7000 years ago (KRAFT et al., 2005; KONTOPOULOS & KOUTSIOS, 2010). The lagoon is bound to a gentle basin structure (Avramidis pers. comm.) and is separated from the wave dominated and microtidal coast of the Ionian Sea by a low relief barrier, dissected by a small channel. On the eastern, landward shore, small deltas prograde into the lagoon and intertidal/supratidal mud flats covered with halophytic vegetation are prevailing (Fig. 2) (AVRAMIDIS et al., 2008; LAZAROVA, KOUTSIOS, & KONTOPOU-

LOS, 2012). With a maximum depth of 2.5 m and an average depth of 0.5 m the Kotychi Lagoon is considerably shallow (KONTOPOULOS & KOUTSIOS, 2010). During the wet season in winter, salinity values range from 8‰ and 17‰, whereas during the dry summer month, restricted runoff and enhanced evaporation lead to salinity from 20‰ to 37‰ (KONTOPOULOS & KOUTSIOS, 2010). Six main streams and a few torrential rivers mostly originating in the foothills of the Arcadian Mountains characterize the hydrological network. Fluvial sediments supplied by the streams are the major constituent of sediments deposited in the lagoon whereas autochthonous shell fragments are of minor importance. Aeolian sands, suspended sands eroded on tidal flats, or sand washed into the lagoon by washover fans are negligible (AVRAMIDIS et al., 2008). The torrential discharge regime of the rivers results in a seasonal expansion of the water body between 710 and 850 ha (DOUKAKIS, 2003).

2.3 Climate

Meteorological data from the nearby Araxos station (38°08'N/21°23.5'E; 12 m above sea level) reflects a typical Mediterranean climate. In the period from 1965 to 2003, the mean annual temperature was 17.8 °C with a relative humidity of 69.4%. The mean annual precipitation between 1948 and 2007 was 669 mm with a pronounced winter maximum between September/October and March/April. This season-

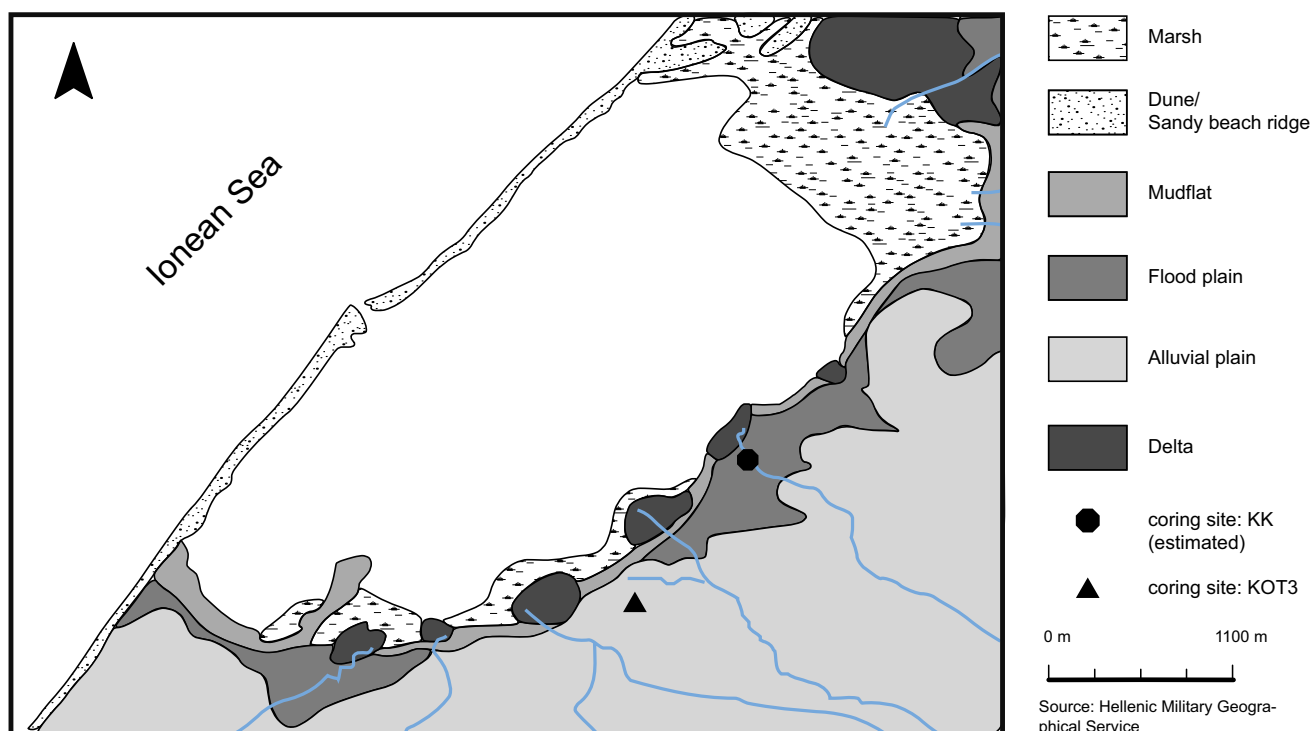


Fig. 2: Lagoonal settings (modified after AVRAMIDIS et al., 2008).

Abb. 2: Die Kotychi Lagune (nach AVRAMIDIS et al., 2008, verändert).

ality is reflected by cyclic sediment fluxes within the year. They are highest at the beginning of the wet season removing easily erodible soils barely vegetated and desiccated during the summer (POULOS et al., 2002).

3 Material and Methods

3.1 Field and Laboratory Methods

In April 2011, the two parallel sediment cores KOT3A and KOT3B (37°59,752N / 21°18,133E, Fig.2) were retrieved on the alluvial plain at the eastern shore of the Kotychi Lagoon using a high-precision rod-operated piston corer (Usinger-System) (MINGRAM et al., 2007). With a total length of 980 cm, KOT3A comprises 10 segments each approximately one meter in length.

The sediments in the uppermost centimeter of each core segment might be disturbed during the coring process, which introduces gaps to the sedimentary sequence. Further, a sand layer in KOT3A at depth from 85 to 123 cm could not be recovered. To overcome these gaps, the parallel core KOT3B comprising 940 cm split into 10 segments was retrieved in a distance of about 2 m from KOT3A.

All sediment cores were transported to Kiel University in densely packed plastic inliners, where they were opened, split lengthwise in the lab and stored in a core repository at 4°C. One half of each core was used for non-destructive scanning methods and subsequently archived. The other half of the core served for different material-consuming analyses such as grain size analysis and CN-analysis.

Prior to any analytical procedures, macroscopic components, sediment structures and sediment texture were documented on split and cleaned sediment surfaces. Sediment color was determined according to Munsell Soil Color Charts.

The archive halves of all cores were scanned at 1 cm resolution for major elements with an Avaatech X-ray Fluorescence (XRF) core scanner equipped with a Rhodium tube (RICHTER et al., 2006). The results of the scans are displayed as element counts per seconds (cps) at each measuring point. For elements with lower atomic number (Al, Si, P, S, Cl, K, Ca, Ti, Mn, Fe) generator settings of 10 kV and 1000 µA with an exposure time of 10 seconds were used. For elements with higher atomic number (Br, Rb, Sr, Zr) setting of 30 kV and 2000 µA with an exposure time of 20 second were chosen. At each measuring point, element counts were normalized to the total element counts (cf. CUVEN, FRANCUS & LAMOUREUX, 2011) to avoid possible bias caused by non-linear absorption (matrix effects) or by dilution of elements outside the measuring range of the spectrometer (closed-sum effect) (RICHTER et al., 2006; LÖWEMARK et al., 2011; CUVEN, FRANCUS & LAMOUREUX, 2011). Furthermore, ratios of selected elements were established and used as geochemical proxies (cf. chapter 6.1).

The XRF-data was used to support the visual correlation of the parallel cores KOT3A and KOT3B. By matching distinct peaks in both core profiles, the respective gaps of KOT3A could be completed with data from KOT3B to obtain the continuous, undisturbed sedimentary sequence KOT3. Due to inhomogeneous sediment surface with abundant sharp shells, XRF-scanning between 313 and 330 cm was not possible. Hence, this section was excluded from further interpretation.

To assess the degree of association between chemical elements, Pearson's correlation coefficients were calculated for the major elements and plotted in a correlation matrix (Fig. 4) using the corrgram package in the statistics software R version 2.13.2 (R DEVELOPMENT CORE TEAM, 2011; WRIGHT, 2011). A strong positive correlation between chemical ele-

ments suggests a common driving force whereas negative values imply an inverse association of elements.

All samples for destructive analyses were solely extracted from KOT3A.

The concentrations of total nitrogen (TN), total organic carbon (TOC) and inorganic carbon (IC) were determined on 86 dried, grinded and homogenized samples through combustion in an elemental analyzer (Euro EA, Elementanalyzer). To quantify the relative abundance of terrestrially and aquatically organic matter C/N (TOC/TN) ratios were calculated (MEYERS & LALLIER-VERGÈS, 1999).

The grain size distribution of the sediments <1.8 mm was determined using a laser-particle-analyzer (Mastersizer 2000, Malvern). 78 samples (~0.5 g) were taken and successively treated with Hydrogen Peroxide until all organics were removed. Prior to measurements, rinsed samples were suspended in 10 ml of distilled water and 0.5 ml 0.1 M Sodium Pyrophosphat to prevent coagulation. Final grain-size distribution/histograms is the mean of 12 runs with a measuring time of 60 seconds for each sample run.

3.2 Proxy Interpretation

Depending on the environmental conditions in a water body or its catchment, geochemical and sedimentological proxies can be interpreted in multiple ways. The grain size distribution in limnic or lagoonal sediments is an indicator for horizontal as well as vertical distance to the shoreline and hence water level fluctuations. In close proximity to the shoreline a greater proportion of coarse grains is deposited while the amount of fine sediments increase with vertical transport distance to the shoreline (BURNETT et al., 2011). However, the grain size distribution can also be a signal of catchment erosion, either reflecting shifts from mainly physical to more chemical weathering or enhanced precipitation and depositions of coarse sediments (KOINING et al., 2003).

In lagoonal systems, organic matter is a binary mixture of terrestrial and aquatic sources. The origin of organic matter can be distinguished by a characteristic C/N-ratio. Aquatic organic matter generally has C/N-ratios between 4 and 10 in contrast to cellulose-rich and protein-poor vascular land plants which have C/N-ratios higher than 20 (MEYERS & LALLIER-VERGÈS, 1999; BERTRAND et al., 2010). Understanding the temporal changes in the origin of sedimentary organic matter may help identifying periods of enhanced terrestrial input versus autochthonous algal productivity, which in turn sheds light on palaeoenvironmental conditions.

Chemical elements may play various roles within an ecosystem. In near coastal environments, Br and Cl act as geochemical markers that record marine influence. Diverging elemental profiles observed in the sedimentary record can result from a large affinity of Br to organic compounds or from a terrestrial source of Cl as a common constituent of sedimentary rocks (BOYLE, 2001; SCHOFIELD et al., 2010).

It is assumed that the amount of Ca and Sr represents the carbonate fraction which can have both an allochthonous as well as an autochthonous origin (COHEN, 2003). Autochthonous carbonate precipitation with co-precipitation of Sr-CO₃ occurs during the summer months when the water is saturated with respect to carbonate due to increased evaporation accompanied by lake level fluctuation and/or algal

photosynthesis in the epilimnion (COHEN, 2003). When the relationship between Ca and Sr is out of phase and one, or both chemical profiles follow the immobile, clastic fraction, the detrital contribution controlled by weathering in the catchment must be considered (LAST, 1994; COHEN, 2003). A likely source for Sr may be silicates, particularly plagioclase feldspars (KYLINDER et al., 2011). To assess the amount of detrital and biogenic carbonate the Ca/Sr ratio can be used. Biogenic calcite precipitated from microorganism is more enriched in Sr than inorganically precipitated calcite/dolomite. Thus, Ca/Sr is high when the contribution of detrital carbonate is elevated and the biogenic component is low (HODELL et al., 2008). To visualize layers rich in biogenic carbonate, opposed to those composed of detrital silicate minerals, the Si/Sr ratio is common tool. The presence of detrital rich layers is inferred from a high Si/Sr (HODELL et al., 2008).

The amount of clastic material in the sediment is generally inferred from the signal of Al, Si, K, Ti, Fe, Rb and Zr reflecting weathering, runoff, erosion, and/or aeolian activity within the catchment (MÜGLER et al., 2010). However, diverging chemical profiles of the clastic elements may be the response of superimposing signals. Si, for example, has a detrital component representing aluminosilicate minerals but also has an autochthon biogenic component influenced by the abundance of diatom frustules (KYLINDER et al., 2011). Some chemical proxies are strongly correlated with the grain size of clastic sediments. Ti, Rb and K are usually concentrated in clay minerals and mica, while the distribution of Zr and Si is linked to their association with the medium to coarse silt fraction (KOINING et al., 2003; KYLANDER et al., 2011). Hence, sorting during sedimentation has an influence on the chemical record and the Zr/Rb ratio is often used as a proxy for changes in grain size distribution. Lower values point to fine-grained sediments rich in Rb while higher values indicating coarse-grained material rich in Zr (DYPVİK & HARRIS, 2001).

Mn and Fe both have an exceptional position. They are associated with the lithogenic fraction but unlike the stable, immobile components, their concentration is strongly controlled by redox remobilization. Under anoxic conditions, the reduced, mobilized Fe²⁺ is less stable in the water column than Mn²⁺ and may re-precipitate as FeCO₃ or more frequently as FeS. Hence, during periods of anoxia Mn/Fe ratio is low. Mn²⁺ will primarily precipitate as MnCO₃ and hence, the reaction is strongly pH-dependent and will most likely occur under anoxic conditions with high pH (EUSTERHUES et al., 2002; KOINING et al., 2003).

When associated with Fe, S may also indicate anoxic conditions but generally, it is present in inorganic S-complexes reacting with organic matter and therefore, reflects the organic content of sediments. S-retention can also be linked to gypsum deposition when correlated with Ca (COHEN, 2003; STRIEWSKI et al., 2009).

4 Chronology

Two bulk samples and three shell samples from KOT3A were processed and radiocarbon dated at the Leibniz Laboratory, Kiel, Germany (Tab. 1). To minimize the risk of dating re-deposited shells, only intact valves in growth position were sampled. Another four ¹⁴C samples dated at Beta Analytics

were available from the core KK retrieved in the direct vicinity of KOT3 by KONTOPOULOS and KOUTSIOS (2010). The five samples from KOT3 taken in between 318 and 908 cm below sediment surface (b.s.s.) cover an age range of 4500–19900 ¹⁴C-years uncalibrated.

The four shell-samples from the core KK were taken between 355.5 and 737 cm b.s.s. providing ages in stratigraphical order between 2800 and 6400 ¹⁴C-years uncalibrated.

While the shell samples of KOT3 are in good stratigraphical agreement with the previously published ages from KK, the bulk samples (KIA45677 and KIA45678, Tab. 1) appear to be unusually old.

Most likely, the two bulk samples contain a considerable amount of re-worked, old carbon, derived from the catchment of the lagoon. Hence, the two bulk samples were treated as outliers (BRONK RAMSEY, 2009) and only the shell samples were included in the age depth model.

As neither shell samples nor macro-remains could be retrieved from KOT3 below 375 cm, the two lowermost ¹⁴C dates of the core KK (Beta-194653 and Beta-194654, Tab.1) were included in the age depth model to build a chronology for the entire sedimentary sequence. Based on the lithology, it is assumed that the lower part of both cores was deposited simultaneously under the same lagoonal conditions, while the cores show a different sedimentary evolution in the upper part, which is strongly influenced by fluvial processes such as channel formation.

All ¹⁴C ages were calibrated using Oxcal 4.1. (BRONK RAMSEY, 2001, 2008, 2009) with respect to the IntCal04 calibration curve (REIMER et al., 2004) and are reported with a 1-sigma (68.2%) probability range. Calibrated years are

denoted as “cal BP” (before AD 1950) according to MOOK & VAN DER PLICHT (1999). For all samples a marine/brackish environment was assumed and a marine reservoir correction of 390 ± 85 BP as suggested by SIANI et al. (2000) was applied. An age-depth-model based on 5 shell samples from KOT3 and KK was calculated applying the *P-Sequence model* of OxCal, which assumes random deposition (BRONK RAMSEY, 2008). The resulting age-depth polygon was directly extracted from Oxcal, returning maximal and minimal ages for every centimeter (Fig.3). A curve based on mean age values – $(age_{max} + age_{min})/2$ – was then used to plot the proxy data.

While the GPS coordinates of the two core sites are known, the exact elevation above sea level could not be determined for neither site. Thus, correlating the depths of the respective ¹⁴C samples introduces an additional uncertainty to the age-depth-model. Due to all these challenges in dating, the paper aims at a general chronological context and ages reported here are rounded to the nearest hundred.

5 Results

5.1 Lithology

Based on sediment color, grains size distribution, and distinct changes in geochemistry, the sediment core KOT3 is subdivided into 16 units. A detailed description is presented in table 2.

On average, the sedimentation rate is around 0.17 cm/a with a maximum of 0.2 cm/a and a minimum of 0.02 cm/a. As the age-dept-model and thus the sedimentation rate is result of a core synthesis, only average rates are given.

Tab. 1: List of radiocarbon samples taken from KOT3 and KK (Kotihi 2-8) (KONTOPOULOS & KOUTSIOS, 2010) from the Kotychi Lagoon.

b.s.s. = below sediment surface.

* = ages in cal BP, based on REIMER et al. (2004).

** = samples integrated in the OxCal-Model, calibrated using OxCal 4.1. Ages are in 1σ range. For all samples a marine reservoir correction of 390 ± 85 a, based on SIANI et al. (2000) was applied.

Samples from KK are published in KONTOPOULOS & KOUTSIOS (2010).

Tab. 1: Liste der Radiokarbon-Alter von KOT3 und KK (Kotihi 2-8) (KONTOPOULOS & KOUTSIOS, 2010) von der Kotychi Lagune.

b.s.s. = below sediment surface (unterhalb der Sedimentoberfläche)

* = Alter in cal BP, nach REIMER et al. (2004).

** = Radiokarbon-Proben die in das OxCal-Modell integriert wurden. Alter wurden mit OxCal 4.1 kalibriert und sind im 1σ-Bereich angegeben. Eine Reservoir-Korrektur nach SIANI et al. (2000) von 390 ± 85 Jahren wurde für alle Proben vorgenommen. Radionuklid-Alter vom Kern KK sind in KONTOPOULOS & KOUTSIOS (2010) publiziert.

sample No.	analysis No.	sample material	¹⁴ C age [BP]	1σ error [BP]	IntCal04*	depth [cm b.s.s.]
KOT3A318**	KIA48442	<i>Spisula</i> sp. valve	4625	40	4880-4790	318
KOT3A355**	KIA48443	<i>Spisula</i> sp. valve	5140	40	5130-5050	355
KOT3A375**	KIA48444	<i>Cerastoderma</i> valve	5285	40	5230-5160	375
KOT3-565	KIA45677	Bulk sample	11330	50	12940-12770	565
KOT3-908	KIA45678	Bulk sample	19930	100	23680-23300	908
Kotihi 2	Beta-194651	<i>Cerastoderma</i> valve	2830	40	2690-2500	355.5
Kotihi 4	Beta-194652	<i>Cerastoderma</i> valve	3860	40	3990-3780	397.80
Kotihi 6**	Beta-194653	<i>Cerastoderma</i> valve	5490	40	6360-6250	604.80
Kotihi 8**	Beta-194654	<i>Cerastoderma</i> valve	6410	40	7110-6960	737.00

Tab. 2: Summarized lithostratigraphy of KOT3. Fig. 3 is the graphical version of this table. UB = upper boundary.

Tab. 2: Zusammengefasste Lithostratigraphie von KOT3. Abb. 3 ist die graphische Version der Tabelle. UB = upper boundary (obere Grenze).

Unit #	Depth [cm]	Lithologic description
16	0-73	Recent soil, very dark brown [10YR 2/1] at the top to yellowish brown [10YR 5/4] at the medium sandy bottom, finer layer with charcoal between 68-73, edgy chert fragments, few snails
15	73-85	Sandy clayey silt coarsening upward, brown [10YR3/4], gravel at the top, few charcoal fragments, UB: sharp
14	85-123	Core loss
13	138-179	Clayey silt, greenish grey [Gley2 5/10G] with abundant yellowish brown [10YR 5/6] striae almost entirely brown [10YR 4/3] at the top, intercalated silt layers [cm-mm scale], charcoal-rich at the top [123-138], UB: sharp
12c	179-197	Strong clayey silt, greenish grey [Gley2 5/10G] with abundant yellowish brown [10YR 5/6] striae, few shell fragments, UB: sharp, bright, carbonate-rich layer
12b	197-215	Strong clayey silt, greenish black [Gley2 2.5/10BG], homogenous, UB: gradational
12a	215-235	Strong clayey silt, greenish grey [Gley2 5/10G] with yellowish brown [10YR 5/6] striae, homogenous. UB: gradational
11	235-307	Strong clayey silt, greenish grey [Gley2 5/10G], abundant shell fragments [up to several mm], several organic-rich layers [e.g. 239-244], UB: gradational
10	307-329	Silty clay, dark greenish grey [Gley2 3/5BG], lower boundary is marked by a shell detritus layer [329-326], several shells [up to 3 cm] and shell fragments [mm-cm] especially at the bottom, UB: gradational
9	329-354	Strong silty sand, dark greenish grey [Gley2 3/10BG], homogenous, UB: sharp [shell detritus layer]
8	354-383	Strong clayey silt, black [2.5Y 2.5/1], abundant shell fragments [mm-scale] especially at the bottom, few shells [cm] at the bottom, UB: sharp
7a	383-470	Clayey silt, greenish grey [Gley2 5/10G], few yellowish brown [10YR 5/6] striae/layers, carbonate concretions between 440 and 418, UB: undulating [root channel/bioturbation?]
7b	470-515	Sandy silt, greenish grey [Gley2 5/10G], homogenous,
7a	470-534	Clayey silt, greenish grey [Gley2 5/10G], few yellowish brown [10YR 5/6] striae/layers
6	534-554	Strong clayey silt, yellowish brown [10YR 5/6] striae increasing upward until sediment is entirely yellowish brown, homogeneous, UB: gradational
5	554-599	Clayey silt, decreasing clay content to the top, greenish grey [Gley2 5/10G], organic detritus, few shell fragments at 558, silt-dominated layers at 560-62 and 567, UB: gradational
4	599-705	Alternating, homogenous sequences [10-20 cm] of sandy silt to silty sand, greenish grey [Gley2 5/10G], UB: sharp/undulating
3	705-863	Weak clayey silt to medium clayey silt, heterogeneous lamination of dark greenish grey [Gley2 4/10G] silty and greenish grey [Gley2 5/10G] clay layers of varying thickness [mm-cm], sometimes silt or organic-rich and/or bright, light greenish grey [Gley2 7/10G] layers intersect fine lamination especially in the upper part, abundant organic fragments, silt-filled root channels, 744 - 728: sandy silt layer fining upwards, 823-818: sandy silt layer, UB: sharp
2	863-909	Silty clay, greenish grey [Gley2 5/10G], sometimes intersected by yellowish brown [10YR 5/4] and organic-rich layers up to 0.5 cm thick, UB: gradational
1c	909-920	Silty clay, abundant shell fragments, dark greenish grey [Gley2 4/10G], UB: sharp
1b	920-930	Medium sand, increasing silt content, greenish black [Gley2 2.5/10BG]
1a	930-980	Medium sand, slightly fining upward, greenish black [Gley2 2.5/10BG], gravel fragments at the bottom

5.2 Laboratory analysis

When deciphering processes within the lagoon, relative changes and altering trends of elemental concentrations throughout the profile should be interpreted rather than absolute values (KYLANDER et al., 2011). Positive correlations among elements (Fig. 4) confirm the common classification into the following three geochemical groups (cf. chapter 6.1): the marine fraction (Cl, Br), the carbonate fraction (Ca, Sr), and the detrital fraction (Al, Si, K, Ti, Rb). Mn and Fe both have an exceptional position and only partially follow the trend of the detrital fraction. S shows no continuous correlation with the other elements. It sometimes tracks the marine indicators or the organic content but is obviously involved in different lagoonal processes (Fig. 5). Zr is decoupled from the detrital fraction but is vaguely correlated ($r_{Zr-Cl}=0.41$) with Cl. It is further associated with the coarse silt/sand fraction (Fig. 5 and 7). As additional proxies, elemental ratios such as

Mn/Fe, Zr/Rb and Si/Sr were analyzed (cf. chapter 6.1).

The chemical composition of the sand-dominated sediment in unit 1 is controlled by the marine indicators (Cl, Br) and the carbonate fraction (Sr, Ca). Linked to the latter one, IC is at its maximum (3.9%). Mn/Fe, Zr/Rb and sulfur also display high values while OC and the C/N-ratio are at their minimum (Fig. 7).

With the transition to unit 1b, a gradual increase in terrestrial elements is accompanied by decreasing Zr/Rb and an enrichment of silt and clay which culminates in unit 2. Most proxies show a significant change at the transition to unit 2 (8000–7800 cal BP). It is characterized by an abrupt increase of the terrestrial fraction and Si/Sr mirroring the declining trend of the IC, the carbonate fraction, the salinity indicators, and Mn/Fe. In contrast, the OC rises of up to 2%.

Throughout units 3 and 4, terrestrial elements exhibit a gradual decreasing trend, however, not synchronously. Si is decoupled from the terrestrial fraction. Zr in fact follows the

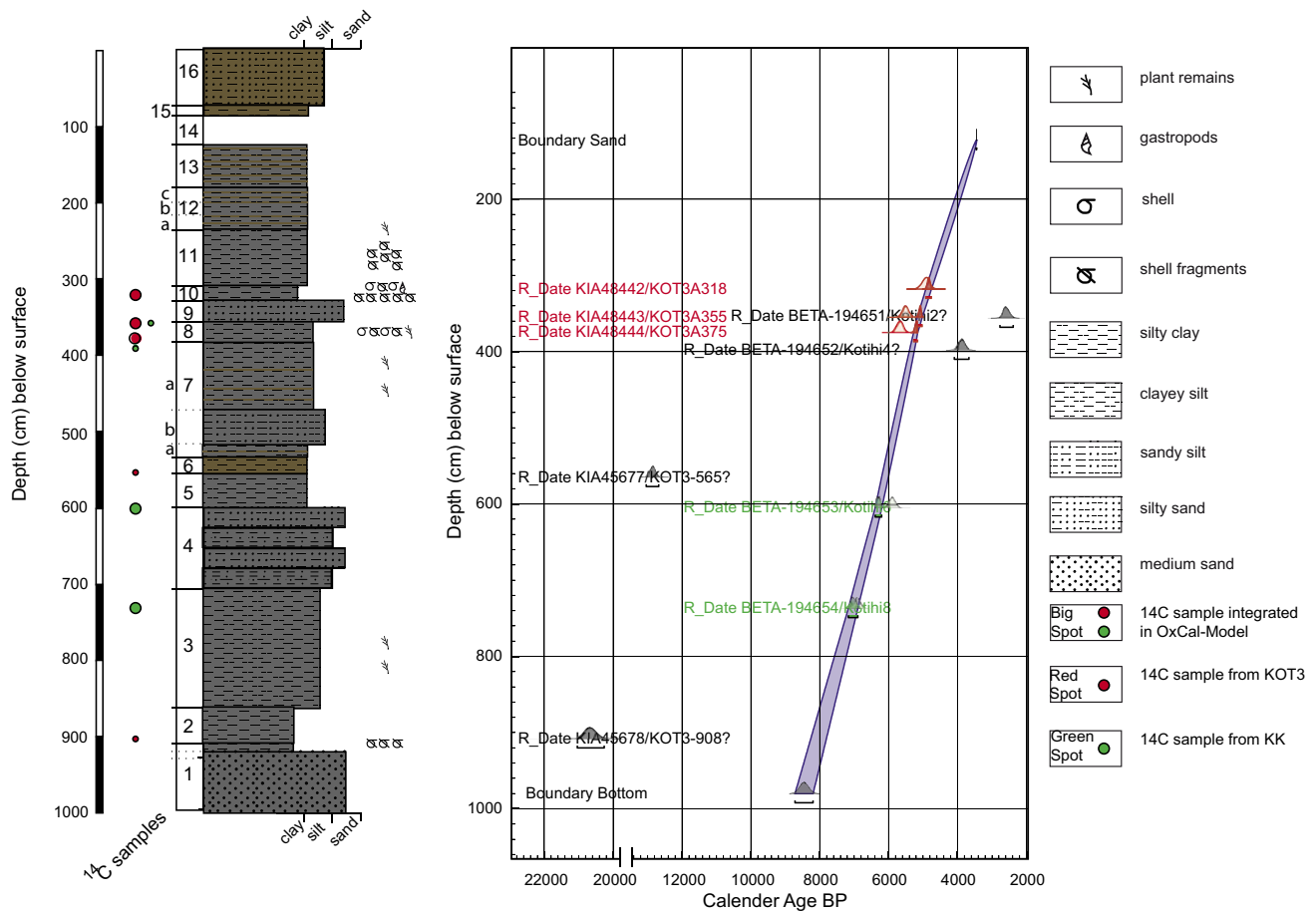


Fig. 3: Lithological units of KOT3 and radiocarbon dates of KOT3 (red) and KK (green) (KONTOPOULOS & KOUTSIOS, 2010). Radiocarbon dates used for the establishment of the chronology are marked with big spots. The resulting age-depth-model was calculated with OxCal 4.1.

Abb. 3: Lithologische Units von KOT3 und Radiokarbon-Alter von KOT3 (rot) und KK (grün) (KONTOPOULOS & KOUTSIOS, 2010). ¹⁴C-Datierungen, die in die Chronologie integriert wurden sind mit einem großen Punkt markiert. Das resultierende Alters-Tiefen-Modell wurde mit der Kalibrations Software OxCal 4.1 erstellt.

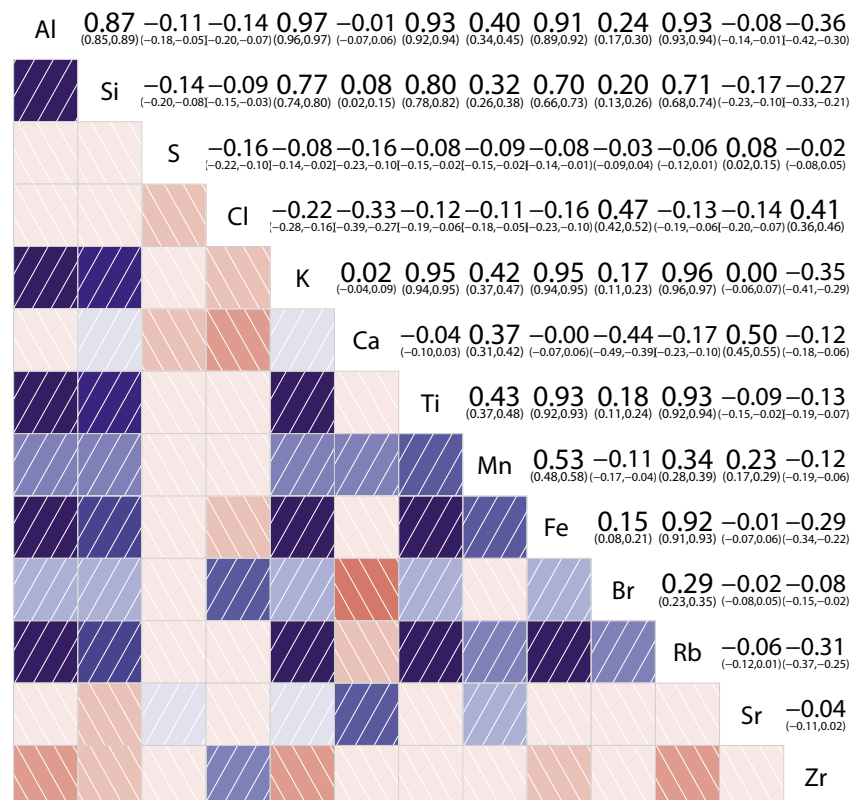


Fig. 4: Correlogram for KOT3 based on Pearson's correlation coefficients. Blue colors and rising line pattern indicate a positive correlation whereas red colors and falling line pattern indicate negative values. The strength of correlation or anticorrelation increases with color shades. A strong positive correlation between chemical elements suggests a mutual driving mechanism.

Abb. 4: Korrelationsmatrix für KOT3 basierend auf Pearson's Korrelationskoeffizienten. Blaue Kästchen und aufsteigende Linien stehen für eine positive Korrelation und rote Kästchen mit absteigenden Linien zeigen eine negative Assoziation. Der Grad der Korrelation oder Antikorrelation steigt und fällt mit Intensität der Farbe. Ein stark positiver Zusammenhang zwischen chemischen Elementen impliziert, dass die Elementhäufigkeiten vom selben Prozess gesteuert werden.

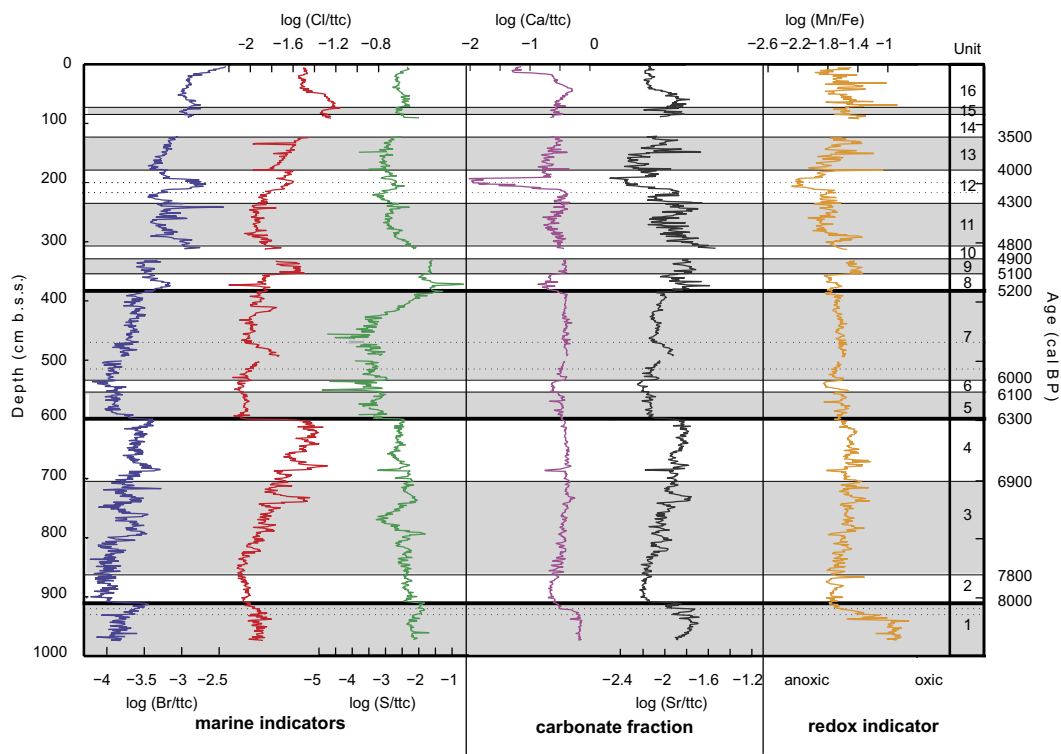


Fig. 5: Geochemical XRF-profiles of the marine fraction, the carbonate fraction, and Mn/Fe (redox-indicator) plotted against depth. Sulphur is often considered a marine indicator but in this case shows a mixed signal. Grouping of elements is based on Pearson's correlation coefficients (Fig. 3). Elemental profiles obtained in cps (counts per second) by XRF are normalized by the total number of counts. The solid black line separates different evolutionary stages.

Abb. 5: Geochemische XRF Profile (in counts per second, cps) von der marinen Fraktion, der Karbonat Fraktion und Mn/Fe (redox Indikatoren). Schwefel zählt generell zu den marinen Indikatoren, zeigt in diesem Fall aber eine Signalüberlagerung, gesteuert von verschiedenen Prozessen. Die Gruppierung der Elemente basiert auf Pearson's Korrelationskoeffizienten. Die geochemischen XRF Profile wurden gegen totale Anzahl der Element-counts normalisiert. Schwarze Linien unterteilen verschiedene Entwicklungsstufen der Lagune.

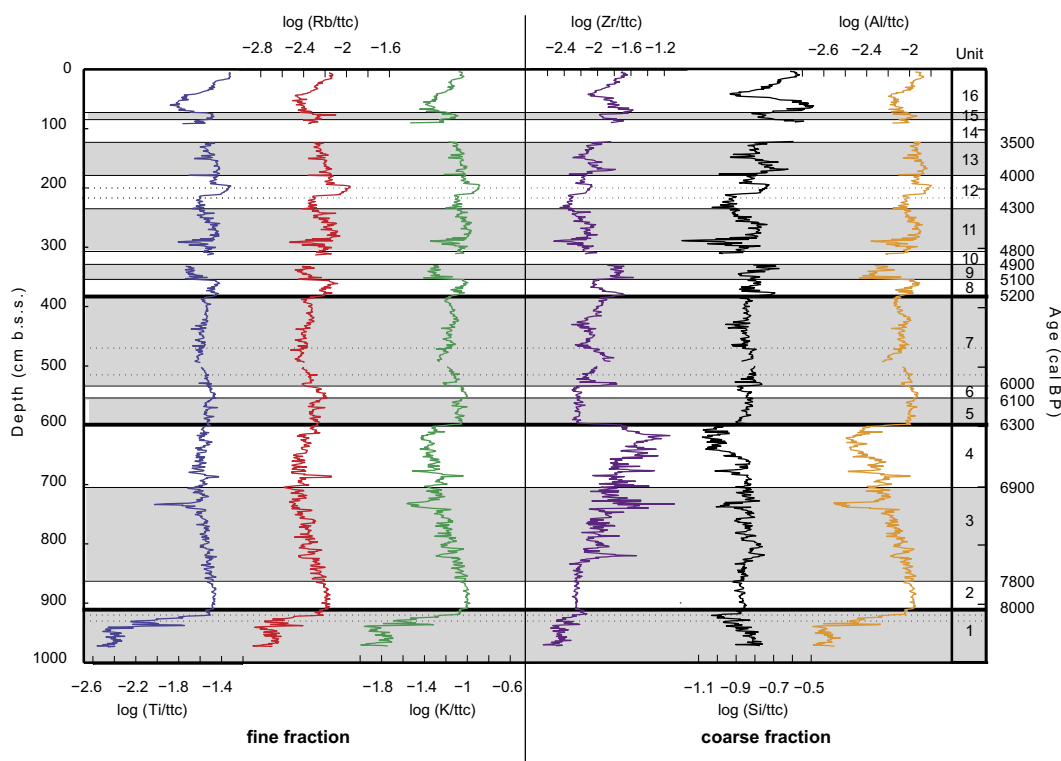


Fig. 6: Geochemical profiles of the terrestrial fraction subdivided into elements predominately associated with fine and coarse-grained minerals. Elemental profiles obtained in cps (counts per second) by XRF are normalized by the total number of counts (ttc). The solid black line separates different evolutionary stages.

Abb. 6: Geochemische XRF Profile (in counts per second, cps) der terrestrischen Fraktion unterteilt nach Elementen, die an grob- und feinkörnige Mineralien gebunden sind. Die geochemischen XRF Profile wurden gegen totale Anzahl der Element-counts normalisiert. Schwarze Linien unterteilen verschiedene Entwicklungsstufen der Lagune.

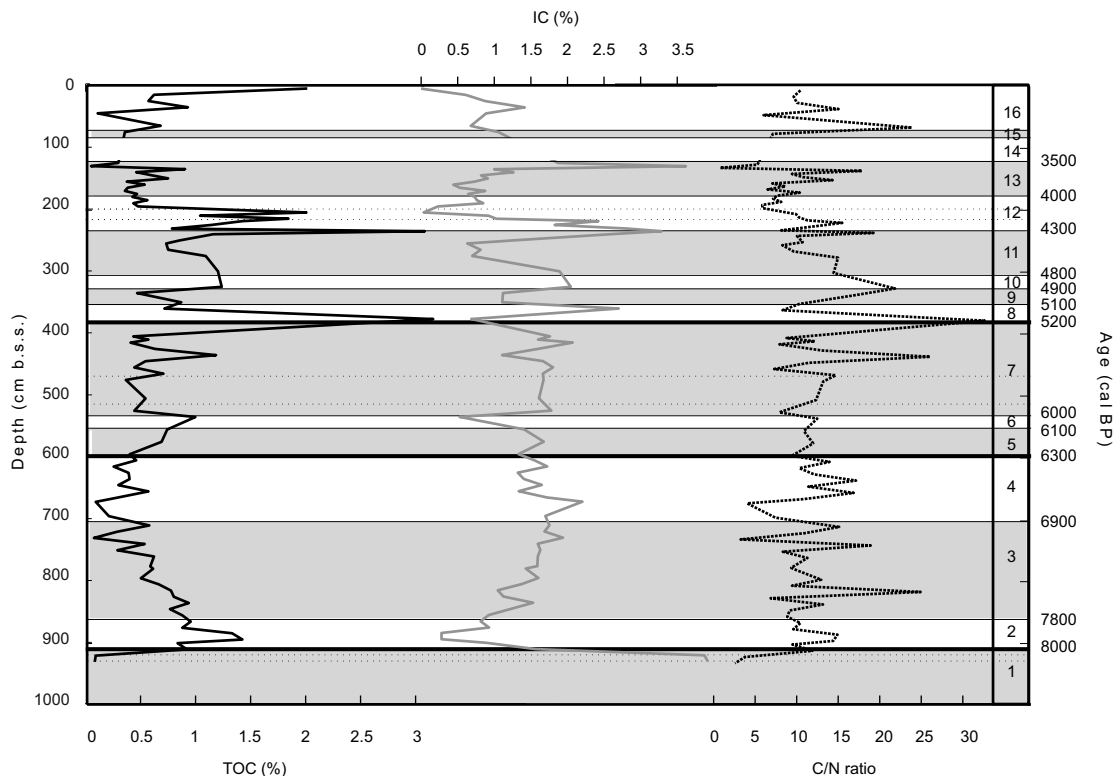


Fig. 7: Inorganic carbon (IC), organic carbon (OC) and CN-ratio of KOT3 plotted against depth. The solid black line separates different evolutionary stages.

Abb. 7: Inorganischer Kohlenstoff (IC), organischer Kohlenstoff (OC) das C/N-Verhältnis. Schwarze Linien unterteilen verschiedene Entwicklungsstufen der Lagune.

increase of the marine indicators and the carbonate fraction which is also reflected by an increased contribution of the sand and silt fraction at the expense of clay. High amplitude fluctuations of the sand and silt fraction reflect rhythmically, alternating sequences of the prevailing homogenous sediment in unit 4. The silt content varies between 21 and 71% generally mirroring to the trend of the sand fraction. Sporadically, the C/N-ratio reaches values of 15 (7900 cal BP), 13 (7600 cal BP) and 25 (7500 cal BP) but shows no pronounced trend in the entire profile.

A considerable change reflected by almost every proxy occurs at the transition to unit 5 (6300 cal BP). The geochemical salinity indicators, Sulphur, and the carbonate fraction abruptly decrease whereas terrestrial elements show elevated values, especially K, Al and Si. The changing chemical composition is accompanied by a pronounced shift towards the clay and silt fraction reflected by a decreasing Zr/Rb ratio. A change in the carbonate sedimentation is indicated by increasing Si/Sr and Ca/Sr ratios.

Until the end of unit 7 (5200 cal BP) elemental profiles remain relatively stable. The sediment is dominated by silt (up to 90%) while the contribution of the sand fraction remains negligible (< 1%) despite a peak of 13% around 5700 cal BP which is reflected by the marine proxies.

With the onset of unit 8 (5200 cal BP), proxies start to show short-lived fluctuations, well documented by an oscillating Si/Sr ratio. The marine proxies are increasing again. However, around 5200 cal BP, parallel to a maximum in S, OC (3.7%) and C/N-ratio (32), Br increases more significantly than Cl which is explained by the large affinity of Br to organic compounds. In unit 9, the elements are in phase again and exhibit high values corresponding to a peak in the sand fraction (13%) and Zr/Rb. In contrast, in unit 11 the terres-

trial proxies show slightly elevated values and the sediment is enriched in clay. High values of OC (3.6%), IC (3.2%) and a high C/N-ratio mark the transition to unit 12 where the marine influence is temporally re-established around 4100 cal BP. At the same time, a peak in OC (2.3%) corresponds to a high C/N-ratio (10), a low Mn/Fe and minimum values of IC (~0%) and the carbonate fraction. Si/Sr is high, mirroring the decline in Ca/Sr. Together with a rising C/N-ratio, the marine proxies, the carbonate fraction and Mn/Fe increase gradually in unit 13. Terrestrial elements especially those enriched in the fine grained fraction are decreasing. The sedimentary record terminates with an erosional unconformity at the upper boundary of unit 13. Parts of unit 14 where lost during coring and the uppermost meter of the core is transformed by anthropogenic activities.

In summary, in coarse sediments (high Zr/Rb) the chemical composition is generally rich in marine proxies and the carbonate fraction while fine grained sediments (low Zr/Rb) show a high concentration of terrestrial proxies.

6 Discussion

6.1 Shoreline evolution

Inferred from simultaneous changes in proxy data, the sedimentary sequence can be subdivided into four evolutionary stages: Stage 1 (8500–8000 cal BP) is characterized by coarse, marine influenced elements. Stage 2 (8000–6300 cal BP) comprises an enhanced concentration of fine grained sediments enriched in terrestrial elements which are progressively replaced by coarse marine influenced sediments. Stage 3 (6300–5200 cal BP) is controlled by silt-dominated, terrestrial sediments with only sporadic traces of marine proxies. Stage 4 (5200 cal BP–present), in contrast shows oscillating

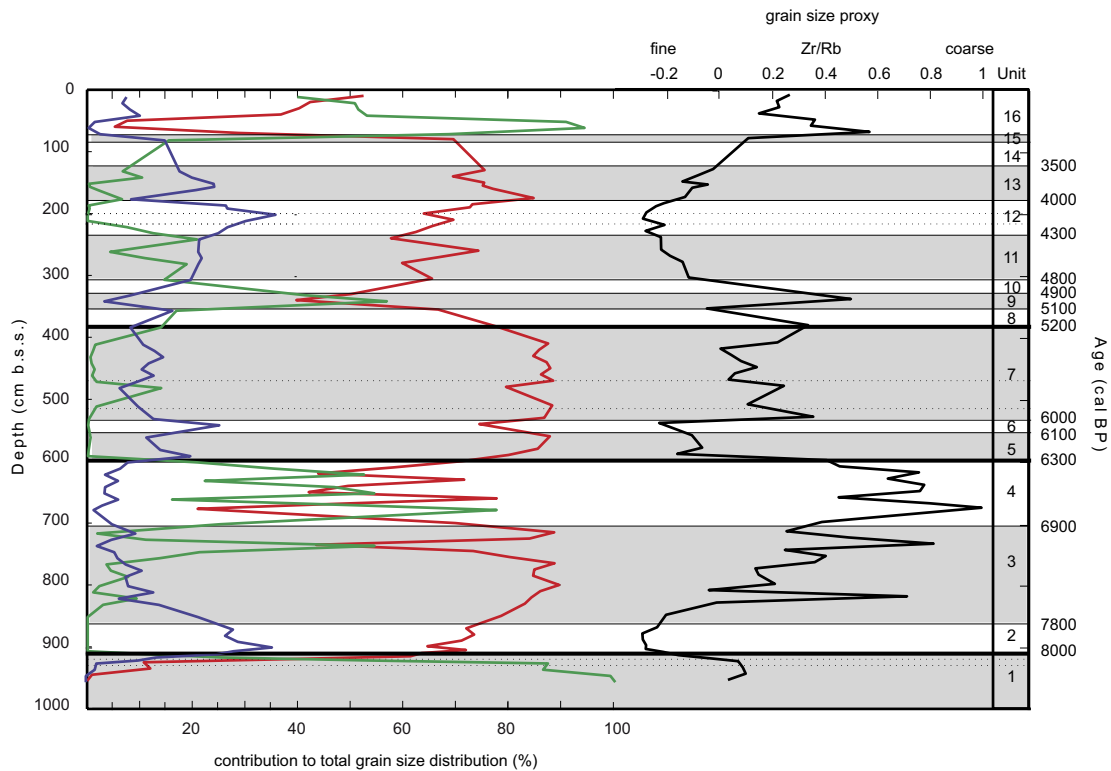


Fig. 8: Percentual contribution of clay (blue), silt (red) and sand (green) to the total grain size distribution in comparison with $\log (Zr/Rb)$ (grain size proxy). The solid black line separates different evolutionary stages.

Abb. 8: Prozentualer Anteil von Ton (blau), Schluff (rot) und Sand (grün) an der gesamten Korngrößenverteilung in Vergleich zu $\log (Zr/Rb)$ (Korngrößenproxy). Schwarze Linien unterteilen verschiedene Entwicklungsstufen der Lagune.

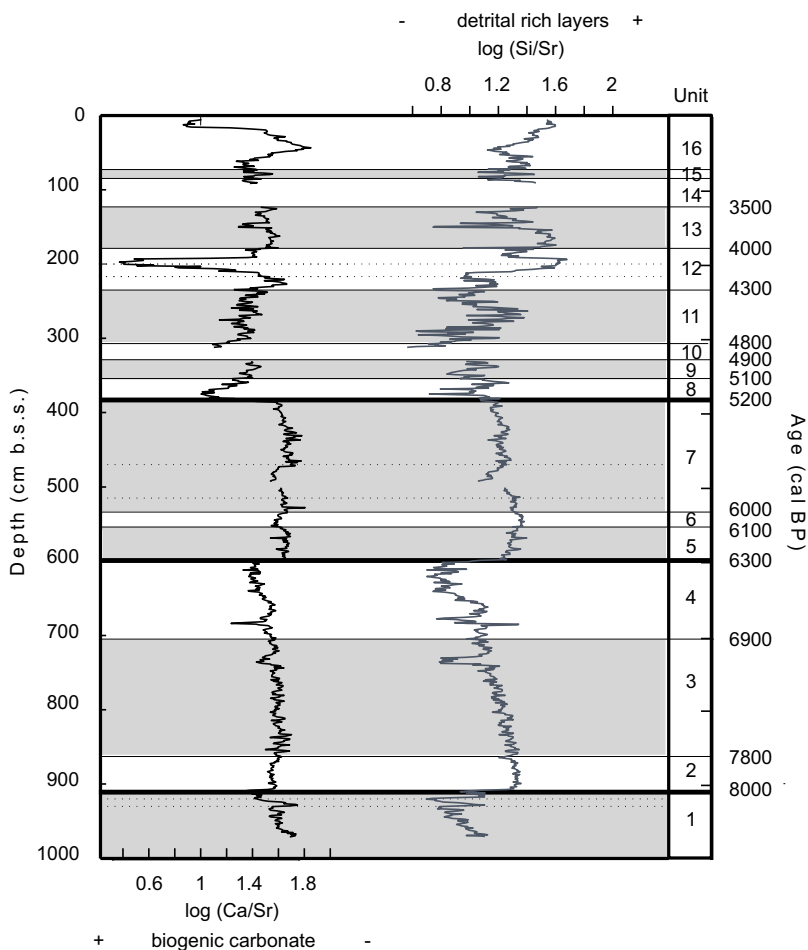


Fig. 9: Si/Sr and Ca/Sr.

Abb. 9: Si/Sr und Ca/Sr.

profiles with successive phases of both, terrestrial and marine dominance. These four evolutionary stages document the palaeoenvironmental evolution of the Kotychi Lagoon.

Stage 1, comprising only unit 1, represents the marine phase of the sedimentary succession which is characterized by coarse carbonate-rich sediments high in Cl and Br deposited in a well oxygenated environment (high Mn/Fe ratio) with low bio productivity (almost lowest OC in the profile). Based on the low C/N-ratio the contribution of terrestrial organic matter rich in cellulose and lignin is negligible.

Based on this data, it is assumed that the palaeoshoreline was located several km east of its present position, which confirms the assumptions of KRAFT et al. (2005). The slowly increasing silt content and the simultaneously increasing signal of the detrital fraction gradually starting around 8200 cal BP (Unit 1b) indicate the gradual development of a back barrier environment where fluvial sediments from the hinterland start to accumulate. Abundant shell fragments indicate the transition to ecologically favorable conditions.

With the onset of stage 2, comprising units 2, 3, and 4, lagoonal conditions have developed. A tenfold increase of the sedimentation rate supports the hypothesis of a sediment trap in the back barrier environment. Clayey sediments (low Zr/Rb) enriched in terrestrial elements point to a quiescent deposition environment where sediments derived from the hinterland accumulate. Peaks in C/N-ratio indicated a binary mixture of terrestrial and aquatic organic matter which confirms the terrestrial sediment source. It is assumed that in the stagnating water vertical mixing was restricted and in the oxygen depleted environment (low Mn/Fe) productivity was still extremely low (low OC). The greyish color of the sediments confirms that no oxygen was available for the oxidization of Fe^{2+} . The drop of the IC content reflects the absence of marine Ca-rich sand and a rising Si/Sr points to enhanced detrital silicate deposition. This indicates the transition to a restricted marine influence.

The continuing sediment supply from the hinterland throughout stage 2 lead to further accumulation of fine grained terrestrial sediments (high amount of detrital fraction and low Zr/Rb) in the back barrier environment, and promoted the growth of the barrier island. Hence, the lagoon was perfectly sheltered favoring the development and preservation of finely laminated sediments (Unit 3), characteristic for a quiescent, oxygen-poor environment with limited bioproductivity (low Mn/Fe and low OC) and no bioturbation of higher organisms which would destroy the lamination.

A decoupling of Al and Si between 822 and 660 cm (7500 and 6600 cal BP) suggests an increased deposition of biogenic silicate possibly associated with diatom productivity. However, this hypothesis needs further verification. Additionally, rising IC indicate carbonate precipitation in the lagoonal environment most likely associated with light colored layers in the laminated sediments (higher IC). They probably reflect an annual cycle. During spring/summer, enhanced biological activity of microorganism such as diatoms results in CO_2 consumption leading to autochthonous, biochemical precipitation of whitish calcite layers. In winter, surface run off and creeks draining the lagoon provide clastic sediments from the catchment and build up a layer composed of mineral grains, reworked carbonate, shell fragments and organic

detritus (BRAUER, 2004). However, the poor state of preservation of the laminae hampers detailed interpretations.

In unit 3, layers of sand for example at 731 cm (7000 cal BP) correlating with peaks in Cl and Br, with rising IC, and higher Zr/Rb imply marine intrusion into the lagoonal environment.

Marine intrusion into coastal water bodies is often ascribed to high-energy transport associated with extreme wave events such as tsunamis or storms. Several authors believe to have found evidence for the occurrence of tsunamis in coastal environments all around the world (GOFF et al., 2012). It is even hypothesized that the ancient city of Olympia approximately 50 km southeast of Kotychi was destroyed by a tsunami (VÖTT et al., 2011).

However, due to the homogenous nature of the marine sediments in KOT3 deposition associated with high-energy transport is neglected. Sediment structures like muddy intra-clasts, were not detected as well as fining upward sequences including abundant faunal remains of marine origin which are often associated with extreme wave events like tsunamis or severe storms (MORTON et al., 2007; GOFF et al., 2012).

By 6900 cal BP (unit 4) the barrier, which entrapped the lagoon, was at least partially flooded as indicated by high Cl and Br and a reduction of the terrestrial fraction. Homogenous, marine sandy silt was deposited and saltwater intrusion initiated bottom ventilation as suggested by a slightly increasing trend of Mn/Fe.

Based on this evidence, stage 2 is assumed to represent a first, short-lived episode of shoreline progradation and barrier island accretion, which isolated the shallow lagoon but successively the coastline retreated. So far, the existence of a sequence of lagoons fringing the Elean coast for at least 7000 years has only been postulated (RAPHAEL, 1973, 1978; KRAFT, 2005; KONTOPOULOS & KOUTSIOS, 2010). With an age of 8000 cal BP, the sedimentary evidence from KOT3 for the first time presents chronological control for the onset of barrier accretion and lagoon development.

A distinct change in sedimentation occurred around 6300 cal BP and marks the beginning of stage 3 encompassing units 5, 6, and 7. The marine influence (Cl, Br and S) abruptly ceases and an increase in terrestrial sediments dominated by clayey silt (low Zr/Rb) implies that riverine outflow to the sea is blocked again and a sediment trap developed behind a barrier. The shift towards a terrestrial sediment accumulation is again well documented by the elevated Si/Sr pointing to a detrital dominated sedimentation at the expense of carbonates.

Around 6000 cal BP, shell fragments and altering layers of grey and brown shades rich in clay (Tab. 2) indicate an alternation of waterlogged phases and drier conditions allowing the oxidization of ferric compounds. A high concentration of terrestrial elements (Rb, Ti, K) linked to the fine fraction opposed to low values of Cl and Br is enriched in unit 6 implying terrestrial condition at this time. The decline of Mn/Fe seems to reflect a lowering of the pH due to oxygenation of the sediments, a process often contributing to soil acidification. A minimum of IC caused by carbonate dissolution in the acid environment supports this hypothesis.

Isolated brown, mm-scale layers in unit 7 result from episodic drought but are rapidly succeeded by lagoonal deposits indicating a low water table susceptible to a changing

hydrological regime. The precipitation of carbonate concretions between 440 and 418 cm (verified by a strong response to HCl) confirms an increased trend towards aridity around 5500–5400 cal BP.

As stage 3 is interpreted to represent the terrestrial/marginal part of the lagoon, the lagoon itself and hence the shoreline must have been located further west or the lagoon has been silted up by progradation of the fluvial influences facies. However, the presence of marine layers associated with homogenous, coarse sediments high in Cl, Br and S deposited in an oxygenated (rising Mn/Fe) environment 7b (515–470 cm; 5800–5700 cal BP) suggest short-lived marine intrusion and the return to marine influenced conditions in the lagoon.

Stage 4 including units 8 to 16 is characterized by pronounced, short-lived fluctuation of many proxies. A sharp undulating contact between unit 7 and 8 (383 cm; 5200 cal BP) marks the transition to stage 4. Abundant shell fragments and shells like *Cerastoderma* sp. and *Spisula* sp., some of them in growth position, indicate the return of lagoonal conditions which is also confirmed by the marine proxies Br and Cl. The decoupling of Cl and Br coincides with a maximum in OC indicating that the opposing trend may be explained by a large affinity of Br to organic compounds. Likewise, the correlation of S and OC implies that S in this unit is also bound to organic complexes. Terrestrial elements linked to the fine-grained fraction are dominating in the strong clayey silt (low Zr/Rb) derived from the hinterland.

The opposing trends of Ca and Sr are attributed to a changing composition of the carbonate source. The decline in Ca seems to be connected to the dissolution of IC which might be attributed to the partial decomposition of enriched OC. A subsequent release of CO₂ lowers the pH to favor carbonate dissolution (DEAN, 1999). However, the carbonate available in the sediment seems to be dominated by Sr-rich, biogenic carbonate as inferred from low Ca/Sr. Keeping in mind high OC values, the undulating sharp lower contact to unit 7 may either be interpreted as a consequence of bioturbation or represents a sediment-filled root channel.

With another sharp transition towards unit 9 around 4900 cal BP, the lagoonal sediments are replaced by homogeneous, strong silty sand showing the typical marine signature characterized by coarse sediments high in Cl and Br and the Ca fraction. The apparent decrease of Br compared to unit 8 can be ascribed to a decline of the OC no longer fixing Br. The sharp contact indicates an event of erosion. However, the homogenous sediments represent an episode of sea level rise rather than an extreme wave event. In contrast, a layer of fragmented shells uncomformably overlaying the marine deposits could be attributed to an extreme wave event around 4900 cal BP. The sharp, erosional contact indicates high-energy transport and the distinct degree of fragmentation of shells results from deposition under turbulent conditions. Following the marine influenced deposition, there was a return to quiescent, lagoonal conditions characterized by the abundance of shells; *Cerastoderma* sp. reaches up to 3 cm.

The gradual decrease of shells in growth position, OC, and the geochemical proxies indication marine influence imply the termination of the fully lagoonal conditions between 4100 and 4300 cal BP (unit 11). A slight coarsening of the sediments towards clayey silt may be the result of seaward

progradation of the profundal zone of the lagoon. However, the grey color typical for reduced mineral compounds due to an oxygen-depleted environment indicates still waterlogged conditions. Numerous shell fragments indicate ecologically favorable condition and the presence of wave action required for their fragmentation and deposition. The high OC towards the end of unit 11 results from a combined contribution of aquatic and terrestrial matter (C/N-ratio: 19) and supports the hypothesis of ecologically favorable condition.

The appearance of brown shades in unit 12 indicates a lowering of the water level, which allowed bottom ventilation and subsequent oxidization of ferric compounds giving the sediment its characteristic color. The environment progressively changed from a fully lagoonal to a lagoonal/marginal environment. However, this trend is interrupted by a short-lived episode of lagoonal predominance around 4100–4000 cal BP. High values of Cl, Br, and S associated with the black color of the sediments suggest the return of water-saturated, anoxic conditions (lowest Mn/Fe). The abrupt decrease of both, Sr and Ca seems to be related to IC dissolution (low IC).

Towards the end of unit 12, around 4000 cal BP the water level has dropped and oxygenated conditions (slightly rising Mn/Fe) returned. The marine influence is declining and a light colored horizon at 179 cm (3900 cal BP) strongly responding to diluted HCl indicates carbonate precipitation possibly as a result of dry conditions.

Around 3900 cal BP, the lagoonal/marginal part has further prograded seaward and the sedimentary evidence of unit 13 (3500–3900 cal BP) documents the coexistence of stream channel and adjacent flooded areas characteristic for a floodplain. During periods of high discharge, coarse sediments are deposited in episodically activated stream channels (silt layers unit 13), which are partly cutting the floodplain. The brownish color of the sediments implies the presence of oxygen supported by rising Mn/Fe. Elevated Cl and Br values suggest periodic flooding but charcoal accumulation at the top of the unit clearly points to a terrestrial provenance of the sediments.

The coarse material of unit 14 (85–123 cm) uncomformably overlaying unit 13 could not be recovered. Judging from a small sample that could be retained in the field, the grain size, the fragmented, redeposited shells (> 2mm) and the incorporated, sharp edged gravel seem to reflect high energy deposits probably associated with a river channel crossing the floodplain and discharging into the lagoon. A temporal onset for the river activity cannot be given based on the age-depth-model as it has to be assumed that the river deeply cut in to the floodplain introducing a non quantifiable erosional gap and hence uncertainty into the age-depth-model. Accordingly, the interpretation of the sedimentary succession terminates with the lower, erosive boundary of the sand layer around 3500 cal BP. The sediments overlying this fluvial deposit are subject to soil formation and intensive agriculture and can therefore not be interpreted in a palaeoenvironmental context.

Based on proxy evidence, it can be assumed that stage four is characterized by a rapid succession of lagoonal and terrestrial deposition milieus, which points to unstable environmental conditions.

Depth (cm b.s.s.)	Stage	Proxy evidence	Palaeoenvironmental Interpretation	Driving mechanism	Age (cal BP)
200	4	Oscillating proxy profiles	Phase of geomorphological instability indicated by a rapid succession of marine, lagoonal and marginal environments	Influence of post glacial sea level rise is influenced by local and regional processes (climate, sediment supply, topography, bathymetry, wave regime.....)	3500
300					
400	3	Marine proxies: low Silty sediments, brownish, mottled	Second, pronounced period of lagoon/barrier formation, shallow lagoon in the back barrier environment, temporal termination of water-logging	Mid-Holocene cessation of sea level rise	5200
500					
600	2	Lower part: clayey sediments rich in OC and terrestrial proxies, high CN, greyish sediments partly laminated Upper part: coarse, homogeneous layers enriched in carbonate and marine proxies increase	First episode of barrier/lagoon formation and subsequent barrier breaching -> reestablishment of marine conditions	First, short-lived episode of a deceleration of sea level rise	6300
700					
800					
900					
1000	1	Coarse greyish sediments Marine/carbonate proxies: high OC: low	Marine phase	Controlled by post glacial sea level rise	8000

Fig. 10: Summary of proxy evidence and palaeoenvironmental interpretation.

Abb. 10: Zusammenfassung der Proxies und ihre umweltgeschichtliche Interpretation.

6.2 Regional Synthesis

The data presented in this study provide evidence for the existence and yet the temporal onset for the development of a Holocene coastal lagoon as proposed by KRAFT et al. (2005), RAPHAEL (1973, 1978) and KONTOPOULOS & KOUTSIOS (2010). However, RAPHAEL (1973, 1978) assumed an early phase of coastal progradation during Hellenistic Times, and KONTOPOULOS & KOUTSIOS (2010) suggested an age of 4000 BP, results from this study provide evidence for a period of pronounced, coastal progradation starting 6300 cal BP (stage 3) and an early but only short-lived phase around 8000 cal BP (stage 2). The latter is approximately in accordance with the oldest proposed barrier accretion cycles proposed by KRAFT et al. (2005) during Late Mesolithic to Neolithic Times.

Further, KRAFT et al. (2005) present evidence for major sediment surges in the Early Helladic (3000–2000 cal BC), in Early Mycenaean times (1700–1400 cal BC), and from Classical (500–323 cal BC) to modern times.

The pronounced period of coastal progradation starting 6300 cal BP (stage 3) is not in accordance with the evidence presented by KRAFT et al. (2005). However, enhanced sediment supply in the Early Helladic and Early Mycenaean Times can also be inferred from the sedimentary record of the Kotychi Lagoon (unit 11 and unit 13). A certain time lag can be attributed to a site-specific evolution or is a result of restricted dating accuracy, which cannot be neglected in coastal environments where ^{14}C dating is limited by a hardly quantifiable marine reservoir correction.

Evidence for three major sediment surges during Hellenistic (323–30 cal BC), Roman (30 cal BC – 330 cal AD) and early Medieval times (330 cal AD – 1453 cal AD) proposed by RAPHAEL (1973, 1978) could not be supported by this study because the sedimentary record of the past 3500 years is not

preserved in KOT3. The river channel producing the erosional gap further hampers a comparison to the palaeoenvironmental interpretation of KONTOPOULOS & KOUTSIOS (2010) derived from the same lagoon. KONTOPOULOS & KOUTSIOS (2010) distinguished three evolutionary stages from lagoonal (earlier than 7000–3810 cal BP) towards a terrestrial period (3810–1400 cal BP) followed by a transgressive phase (1400 cal BP–present). Whereas the basal part of the cores show a similar evolution, the mid-Holocene phase of coastal progradation corresponding to stage 3 in KOT3 is not identified by KONTOPOULOS and KOUTSIOS (2010). In contrast, in KK lagoonal bottom facies interrupted by an episode of stream channel influence in a depth of 440–470 cm are prevailing (KONTOPOULOS & KOUTSIOS, 2010). This probably introduced the erosional gap in KK much earlier than in KOT3 and is held responsible for the evolutionary discrepancy of the cores in the upper part.

6.3 Driving forces of coastal pro- and retrogradation

On a local scale, coastal morphology is highly susceptible to terrestrial sediment supply. Terrestrial environments play a profound role controlling sediment dynamics as material is eroded from the hinterland, transported by rivers, and entrained in littoral currents to shape the coast by creating barriers, lagoons, marshes or deltas. However, the sea level acts as a baseline to which a broad spectrum of processes adjusts. Hence, it controls the long profile of rivers, which influences sedimentation and erosion in the lowlands. Consequential, in the transitional zone were rivers and the sea merge barrier migration is the morphological response to the relation of sediment supply rates and the rate of sea level rise. A disequilibrium results in an adjustment of process and form (COSTAS et al., 2009; PLATER & KIRBY, 2011).

As causes for the perturbation of this disequilibrium and the balancing processes initiating shoreline migration, either changing sea level rates associated with the decay of the continental ice sheets and changes the ocean volume or varying sediment fluxes can be considered (LAMBECK & PURCELL, 2005; COSTAS et al., 2009). The latter one can be a composite of both, naturally induced soil erosion or anthropogenically enforced hinterland exploitation and soil degradation that translates into sedimentation in the lowlands (RAPHAEL 1973, 1978; KRAFT et al., 2005; DINIS et al., 2006).

Of course, local tectonic processes can influence the interplay of sea level rise and sediment supply. It controls the nature of the coastal environment either by gradual uplift or subsidence or by a sudden, impulsive vertical displacement. In this context, tectonic activity in the coastal area itself as well as in the hinterland has to be considered because it controls relief energy and hence sediment delivery. However, the quantification of tectonic control over geological time-scales is difficult (PLATER & KIRBY, 2011).

Further, short-term processes as extreme wave events can create an episodic disturbance. COSTAS & ALEJO (2007) for example report sporadic barrier breaching associated with increased wave energy striking the barrier during periods of increased storminess. After the initial disturbance, the coastal system needs a certain time to readjust and adapt to the boundary conditions.

With regard to the diverse driving factors of coastal evolution in the Kotychi area, stage 1 is influenced by the aftermath of the melt down of the continental ice sheets, which initiated ocean volume expansion and subsequently rapid sea level rise. Until 8000 cal BP the shoreline was located several km landward of its present position.

With vanishing of the ice sheets ocean volume remained constant attenuating the rate of sea level rise and allowing local signals to overcompensate the eustatic signal (FREITAS et al., 2003; BAO et al., 2007; COSTAS et al., 2009). In the Mediterranean, widespread evidence for a weakening of the eustatic signal and the development of lagoon barrier systems clusters around 5500–6500 BP (DINIS et al., 2006; BAO et al., 2007; COSTAS et al., 2009). However, stage two starting around 8000 cal BP seems to represent an early phase the mid-Holocene cessation of sea level rise which gives local factors such as sediment availability an accentuated role.

This early stage sea level deceleration reflects the hypothesis of STANLEY & WARNE (1994), who postulate worldwide delta progradation as a result of fluvial sediment input overcompensating the declining rate of sea level rise between 8500 and 6500 BP. On a regional scale the hypothesis of a stabilization of the sea level shows analogies to the results KRAFT et al. (2005) and it approximately confirms the results of AVRAMIDIS et al. (2012) who described sand barrier evolution and brackish/lagoonal back barrier conditions around 8540 cal BP in the Alykes Lagoon on Zakynthos Island 30–40 km west of Kotychi Lagoon.

Human induced soil erosion creating a surplus of sediments must also be considered. An array of studies evaluated and summarized by DUSAR et al. (2011) proposes a causal relationship between anthropogenic landscapes modification and enhanced sediment dynamics in the Mediterranean. It is common assumption that deforestation in the Mediterranean dates back to the Neolithic (KRAFT et al., 2005). Indeed, there is evidence for Early Neolithic human activity in

the study area (RAPHAEL, 1973; WILLIAMS, 2004). However, their ecological footprint is considered to have been small as indicated by pollen data (LAZAROVA, KOUTSIOS, & KONTOPOULOS, 2012) and so far no anthropogenically induced sediment surges have been reported in the region during Neolithic Times (RAPHAEL, 1973, 1978; KRAFT et al., 2005; KONTOPOULOS & KOUTSIOS, 2010). Hence, human induced soil erosion can be neglected as a source for the perturbation of the balance between sea level rise and sediment supply around 8000 cal BP. In fact, hinterland erosion and sedimentation in the Kotychi area appears to have been on a moderate level inferred from the finely laminated sediments representing a quiescent, low-energy sedimentation regime and further from an immature, semipermeable sand barrier which was flooded by a slower but after all rising sea level around 6600 cal BP.

Accordingly, temporal analogies of circum-Mediterranean lagoon formation indicate that an early stage of sea level deceleration led to a surplus of sediments in the transitional area where the Peneus River and the Ionian Sea merge which, initiated barrier formation in the prograding delta. Hence, an attenuation of the eustatic signal was the driving force of coastal progradation around 8000 cal BP. The gradual, subsequent flooding of the barrier environment is ascribed to a slower but after all rising sea level. However, an episode of subsidence, lowering the overflow threshold of the barrier could also be possible but is hard to prove.

A shift toward terrestrial marginal conditions either associated with a seaward migration of the lagoon or silting up of the back barrier environment characterize stage 3 and indicate an additional imbalance between sea level rise and sediment supply. The beginning of this stage coincides with circum Mediterranean lagoon formation and the traditionally mentioned cessation of the post-glacial sea level rise around 6500 and 5500 BP (LAMBECK & PURCELL, 2005; DINIS et al., 2006). For this period, FOUACHE et al. (2008) report a complex of coastal barriers encompassing the Thessaloniki plain, on the Iberian Peninsula COSTAS et al. (2009), BAO et al. (2007) and FREITAS et al. (2003) just to name a few, postulate lagoon and wetland formation. At the coastal plain of Marathon, Greece, PAVLOPOULOS et al. (2006) ascribe a sequence of lagoonal deposits to the stabilization of sea level rise just as KONTOPOULOS & AVRAMIDIS (2003) at the Aliko Lagoon, north Peloponnese.

However, the deceleration of sea level rise and a subsequent surplus of sediments must not be the sole reason for the cessation of the waterlogged conditions during stage 3. In the mid-Holocene, a trend towards a general aridification is inferred from $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ records of speleothems in Israel (BAR-MATTHEWS et al., 2003), or by geochemistry and pollen data from crater-lake sediments in central Turkey (Roberts et al., 2001). However, the transition towards aridity shows spatial variability throughout the Mediterranean (cf. LESPEZ, 2003; JALUT et al., 2009; DUSAR et al., 2011; FINNÉ et al., 2011). In a comprehensive synopsis of Holocene climate in the eastern Mediterranean, FINNÉ et al. (2011) propose a transition towards increased aridity gradually starting 5400 BP, which could have triggered climatically induced water table oscillation. The appearance of carbonate concretions in the Kotychi record around 5500–5400 cal BP supports the hypothesis and implies dry conditions in the Kotychi area during this time.

Anthropogenic induced soil erosion in stage 3, coinciding with the Late Neolithic is still considered to be on a moderate level and hence, appears to be a minor factor initiating enhanced sediment supply.

Consequently, prevalent temporal analogies of lagoon formation justify the hypothesis that at the beginning of stage 3 a widespread attenuation of the eustatic signal resulted in delta progradation of the Palaeo-Peneus River delta. Entrained in littoral currents the surplus sediments was redistributed and reorganized to form a lagoonal back barrier environment. The cessation of the waterlogged conditions might have been a composite of an attenuated sea level rise amplified by climate forcing which with diminishing dominance of the eustatic signal is able to exert influence on coastal morphology.

Stage 4 is characterized by a comparable rapid succession of varying deposition environments indicating reoccurring fluctuation in the sediment budget. After the pronounced period of coastal progradation (stage 3), lagoonal conditions are returning again indicating an episode of barrier breakdown around 5200 cal BP. It seems that hinterland sediment supply was restricted. With a still moderate level of human activity in the Elean uplands and climate conditions obviously not favoring enhanced sediment fluxes, a slower but nevertheless rising Ionian Sea flooded the lagoon. Apparently starved of continuous sediment supply, the barrier became progressively impermeable allowing marine intrusion (unit 9) and it was of course not mature enough to withstand extreme wave events (shell fragments unit 10).

Starting 4700 BP, lagoonal/marginal conditions are prevailing passing into progressive terrestrial conditions around 4300 BP indicating that the profundal zone must have migrated seaward or has been silted up. At least a slight increase in sediment supply must have occurred which could be attributed to gradually increasing human activity with the onset of the Early Helladic (5000 BP) (WILLIAMS, 2004). A short lived phase of lagoonal predominance between 4000 and 4100 cal BP could be the result of a proposed climate anomaly around 4200 BP creating cool and dry conditions (Mayewski et al., 2004) and hence hampering sediment dynamics and the consolidation of the barrier allowing salt water intrusion. However, the event-like character of the so-called 4.2 ka event is controversial (FINNÉ et al., 2011). The reestablishment of marginal conditions and the gradual transition to fully terrestrial conditions starts around 4000 cal BP contemporaneously to the onset of the Middle Helladic (2000 cal BC). In this period, the first peak of settlement activity was reached as verified by an increased number of sites (WILLIAMS, 2004). A palynological signal of anthropogenic landscape modification displaying peaks in indicators of agriculture and stockbreeding appears in the Kotychi area in early Helladic times and thereby supports the archaeological evidence of increased human activity. The cultivation of *Olea* assigned to 3810 cal BP is an additional indicator for progressive human interference into nature (LAZAROVA, KOUTSIOS & KONTOPOULOS, 2012). Further, abundant charcoal fragments clustering between 3500 and 3600 cal BP might indicate human induced fire activity in the hinterland but could also be the result of natural fire activity. However, apparently rising demographic pressure resulted in forest clearing and overgrazing irrevocably destroying endemic vegetation. Devoid of its natural, protective cover, the soil

was susceptible to erosion, which translated in silting up of lowlands (*cf.* DUSAR et al., 2011) and the development of a broad floodplain prograding into the lagoon (unit 13). Hence, it must be considered that the cessation of the waterlogged conditions at the end of stage 4 is attributed to human induced soil erosion providing a surplus of sediments.

The erosional force of a meandering river channel deeply incised into the floodplain eroded the sedimentary evidence deposited after 3500 cal BP. Accordingly, no statements about this period can be derived from the sedimentary sequence KOT3.

In summary, it is hypothesized that after the deceleration of sea level rise the influence on coastal evolution of variables such as anthropogenic induced hinterland erosion or climatic factors became more important. With a multiplicity of factors acting upon coastal morphology unstable environmental conditions in the coastal area prevailed.

7 Summary and conclusion

The sedimentary sequence of KOT3 provides insights into the evolution of the north Elean coastline from around 8500–3500 cal BP. Combining geochemical and sedimentological methods (XRF, grain size, OC, IC, and C/N analysis) with Bayesian age-depth-modeling, four evolutionary stages documenting the transition from a marine to a marginal lagoonal environment could be identified. With regards to the driving forces of Holocene coastal evolution, this study demonstrates that Elis shows a two-phase development as reported from lagoons throughout the Mediterranean. Temporal analogies in coastal evolution across the Mediterranean indicate that early Holocene morphology results from the global aftermath of postglacial sea level rise. With vanishing of the ice sheets, the so far preeminent role of the eustatic signal was overwhelmed giving local and regional processes an accentuated role. Geomorphological instability in the coastal area indicates that a multiplicity of factors is acting upon morphology demanding constant adjustment. Hence, in mid to late Holocene natural and/or anthropogenically controlled sediment supply which translated in rapid morphological adjustment of the coastline was the driving force of coastal evolution.

In this study the application of the fast, non-destructive, high-resolution XRF-analysis technique proved to be a powerful tool providing a general overview of the chemical composition and altering element concentrations in a sedimentary sequence. The XRF-technique is a low-cost approach to discriminate marine and terrestrial environments and is especially useful when the concentration/preservation of indicative microorganism is poor or too time-consuming (e.g. CUNDY et al., 2006). However, the semi-quantitative nature of the XRF-results needs to be kept in mind to avoid over-interpretation of the element concentrations.

Despite its potential to answer an array of palaeoenvironmental questions, deciphering sedimentary sequences in coastal environments presents certain pitfalls. Disentangling the superimposing, driving forces of coastal evolution is challenging and requires an independent chronology of involved processes to support their identification in the sedimentary sequence. Using neighboring cores as validation is restricted due to the rapid re-deposition and reorganization of unconsolidated sediments in the transitional zone where

rivers and the sea merge. Another crucial point when studying lagoonal environments is the accuracy of radiocarbon dating, which is limited by a hardly quantifiable reservoir correction and re-deposited organic material. Chronological control is additionally hindered in tectonic active regions, where crustal movement might introduce uncertainties in age depth relation of the sediments.

Hence, the interplay of tectonic, sea level changes, human and natural induced changes of sediment budgets presents a challenge when deciphering sedimentary sequences in tectonic active coastal regions. Despite all that, the present study successfully demonstrate that Holocene coastal evolution of the North Elean coast shows significant analogies to circum-Mediterranean lagoon formation during the Holocene.

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Holotype skulls, stratigraphy, bone taphonomy and excavation history in the Zoolithen Cave and new theory about Esper's "great deluge"

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Abstract:

The Zoolithen Cave, in the Wiesent River Valley of Upper Franconia, Bavaria, South Germany, has a very long excavation history. The site is of international paleobiological importance as the Type site for five Pleistocene top predators (cave bear, Ice Age hyena, lion, wolf, dhole). This large cave system has developed in three elevations and preserves three fluvial sedimentary sequences including two speleothem genesis phases representing changing ponor, dry and wet stages from the Oligocene/Miocene (Neogene), over the Pliocene/Early Pleistocene to Late Pleistocene. The cave bear *Ursus deningeri* used the cave as den during the MIS 6–9 (Holsteinian interglacial-Saalian glacial). Single P4 tooth and skull shape analyses ("= cave bear clock") date different cave bear species (*U. spelaeus eremus/spelaeus*, *U. ingressus*) within the Late Pleistocene (MIS 3–5d). Finally the bones of other Pleistocene megamammals were washed from two former cave entrances at elevations of about 455 m a.s.l. up to 30 meters deep into lower elevation cave parts, during the Last Glacial Maximum (Post-*U. deningeri* times or Postglacial), -historically believed to be the result of the "great deluge". The young "river terrace dolomite gravels" which occur as relic sediments at elevations of about 455 a.s.l. in several caves around Muggendorf cannot be explained by natural erosion/river terrace stratigraphy, and must relate to an uncertain glacial context. Finally Iron Age (La Tène) humans left secondary burials (human skulls and long bones with pottery and after-life food animal donations) only in the first deep vertical shaft (Aufzugsschacht) similar to the situation in the nearby Esper's Cave.

Holotypen-Schädel, Stratigraphie, Knochen-Taphonomie und Ausgrabungs-Historie in der Zoolithenhöhle und eine neue Theorie über Esper's „biblische Sintflut“

Kurzfassung:

Kurzfassung: Die Zoolithenhöhle liegt entlang des Wiesentals (Oberfranken, Bayern, Süd-Deutschland) und hat eine lange „Spatenforschungs“-Historie. Die Fundstelle ist von internationaler Bedeutung aufgrund ihrer fünf validen Holotypen-Eiszeittier-Schädeln des „Höhlenbären“ sowie Top-Prädatoren (Eiszeit-Löwe, -Hyäne, -Wolf und -Rotwolf). Das große Höhlensystem entwickelte sich auf drei Etagen und hat drei fluviale Haupt-Sedimentsequenzen inklusive zwei Haupt-Speleothem-Genesephase unter wechselnden Ponor-, Trocken- und Nassphasen während des Oligozän/Miozän (Neogen) über das Pliozän/Frühpleistozän bis hin zum Spät-Pleistozän. Die ersten Höhlenbären-Populationen Oberfrankens mit *Ursus deningeri* nutzen die Höhle als Horst bereits im MIS 6–8 (Holstein-Interglazial/Saale-Glazial). Isolierte P4 Zahn- sowie die Schädelmorphotypen (= „Höhlenbären-Uhr“) datieren verschiedene Höhlenbären-Arten/Unterarten (*U. spelaeus eremus/spelaeus*, *U. ingressus*) in das Spät-Pleistozän (MIS 3–5d). Ihre Knochen wurden in etlichen Fällen zuerst durch Top-Prädatoren beschädigt. Letztendlich wurden die Knochen aller pleistozänen Großsäuger während des Hochglazials (= Last Glacial Maximum, Post-*U. deningeri*-Zeit oder Postglazial) von zwei ehemaligen Eingangsbereichen in Höhenlagen um 455 m NHN bis zu 30 Meter in tiefere Höhlenbereiche besonders über die Vertikalschäfte durch Hochflutereignisse verschwenkt. Dieses wurde in historischer Zeit als „biblische Sintflut“ interpretiert. Die jüngsten „Flussterrassen-Dolomitkiese“ in Höhenlagen um 455 m NHN werden als Reliktsedimente in verschiedenen Höhlen um Muggendorf angetroffen und können in solchen extremen Höhenlagen 130 Meter über der heutigen Wiesent nicht mehr mit „natürlicher Erosion/Flussterrassenstratigraphie“ erklärt werden. Sie müssen im noch unklaren glazialen Kontext stehen. Letztendlich hinterließen Eisenzeit-La Tène-Menschen Sekundärbestattungen (Schädel, Langknochen, Keramik und Jenseits-Nahrungs-Haustier-Beigaben) nur im ersten tiefen Vertikalschacht (= Aufzugsschacht), ähnlich wie in der nahegelegenden Esperhöhle.

Keywords:

Holotype skulls, stratigraphy, bone taphonomy, excavation history of the Zoolithen Cave, new theory about Esper's "great deluge"

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1 Introduction

The Franconian Karst in Germany is one of the most important regions for megafaunal and palaeoclimatic research of the Middle/Late Pleistocene in Europe due to its multiple archaeo-biological-archives. The extremely Pleistocene bone-rich caves, which also contain important sedimentary sequences, allow landscape and erosion modeling which is potentially important for comparisons within other regions of central Europe.

Upper Franconia (Bavaria) along the Wiesent and tributary valleys has the highest density of caves in the Franconian Alb (e.g. KAULICH & SCHAAF 1993, GROISS et al. 1998) and is one of the most cave-rich regions of Europe, worthy of "speleopark" designation in the future. The caves are eroded into the massive Upper Jurassic reef and lagoon/inter-reef dolomites (MEYER & SCHMIDT-KALER 1992, GROISS et al. 1998), which additionally are famous climbing areas. Most of the caves are only small clefts or cavities, with only a few larger caves are present.

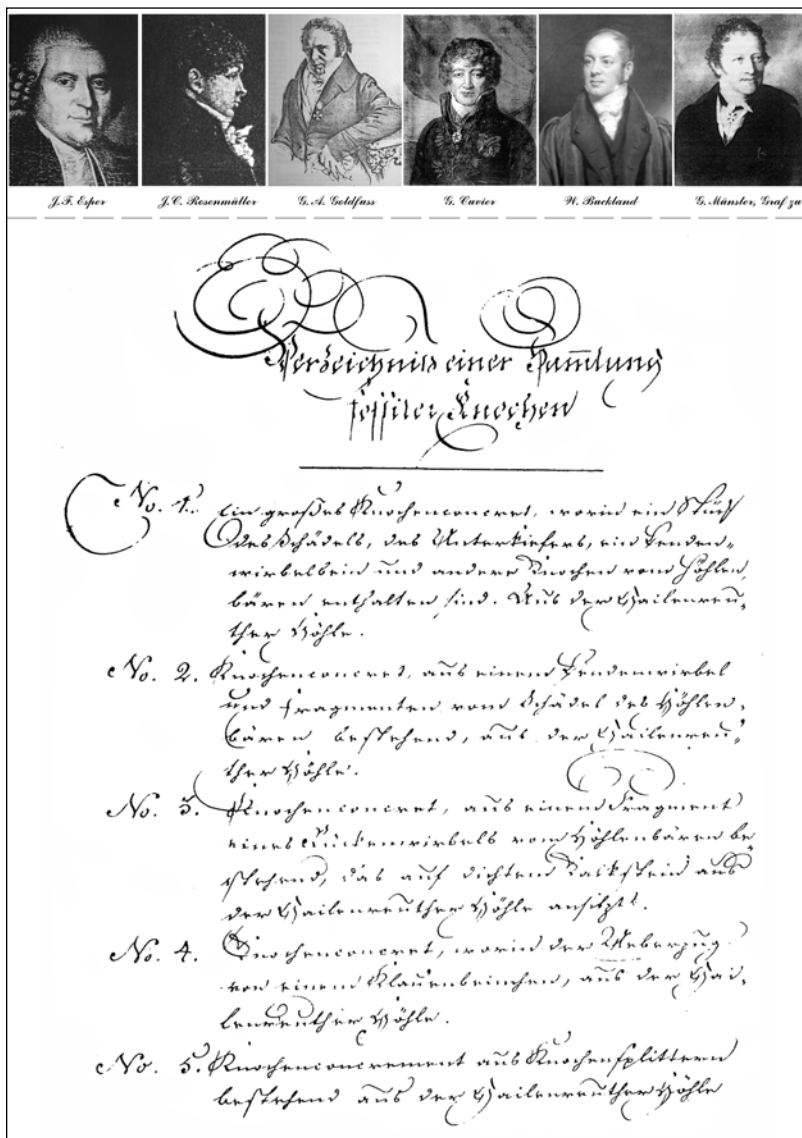


Fig. 1: The first Zoolithen Cave researchers of the „early spade research time: J.F. ESPER, J.C. ROSEN-MÜLLER, A. GOLDFUSS, G. CUVIER, W. BUCKLAND and Graf zu MÜNSTER. Below: First page of the catalogue of the selling contract of the Rosenmüller-1794 collection (former times Preußische Geologische Reichsanstalt Berlin, today Naturkundemuseum der Humboldt-Universität Berlin).

Abb. 1: Die ersten Zoolithenhöhlen-Erforscher während der „frühen Spatenforschungszeit“, J.F. ESPER, J.C. ROSENMÜLLER, A. GOLDFUSS, G. CUVIER, W. BUCKLAND und Graf zu MÜNSTER. Unten: Erste Seite der Kataloges des Kaufvertrages der Rosenmüller-1794-Sammlung (damals Preußische Geologische Reichsanstalt Berlin, heute Naturkundemuseum der Humboldt-Universität Berlin).

Those caves, and especially the Zoolithen Cave, was one of the first and most famously targeted fossil cave bear localities, where many famous German, French and English pioneering researchers excavated or studied material, including ESPER, ROSENMÜLLER, GOLDFUSS, CUVIER, BUCKLAND & Graf zu MÜNSTER (Fig. 1). The most famous and largest fossil collection assembled by ROSENMÜLLER (labeled in 1797) that contains several of the famous skulls, was thought to be lost, but has been relocated by the current author, in the Preußische Geologische Landesanstalt (see ROSENMÜLLER catalogue, Fig. 1). During the DDR socialist Republic times this collection was forgotten but is now recognized to be of international importance. Further important researchers included GOLDFUSS & BUCKLAND who provided the first illustrations of the cave and excavation areas (Fig. 2A). Sadly, in more recent times (after the Second World War), large newly discovered parts and old areas were emptied with old-style methods by GROISS (Fig. 2B).

Mainly cave bear remains have been found from the Pleistocene layers in the Zoolithen Cave as the first and most well-known and one of the richest cave bear bone site of Europe (ESPER 1774, ROSENMÜLLER 1794, GOLDFUSS 1810, 1818, 1821, 1823, BUCKLAND 1823), which herein estimated has/had

a half million of remains. This and other larger bone-rich caves in Upper Franconia (Fig. 3A) are all in higher elevations between 550 to 400 m a.s.l. (Zoolithen Cave former entrance 455 m NHN, Sophie's and Große Teufels Cave former entrances 410 m NHN) and contain mainly Late Pleistocene megafaunas in the Große Teufels Cave, Sophie's Cave, or Geisloch Cave such as the herein discussed Zoolithen Cave, which will be demonstrated to have been used by early cave bears even earlier already in the Middle Pleistocene. At minimum four larger Pleistocene cave bear dens are known today (Fig. 3A). Smaller and fewer cave bear remain containing caves are the Zahnloch Cave, Neideck Cave, Moggaster Cave, König-Ludwigs-Cave, Wunder Cave and Esper Cave (DIEDRICH 2013a, Fig. 3B).

1774–1794 – ESPER to ROSENMÜLLER – “surface collecting times”

The reports of HELLER (1972) do contain a compilation of the history, but do not reflect the real situation. After the first reports of the cave fauna by ESPER (1774), he collected, described and figured “bones of extinct animals” (= cave bears and others) from the Zoolithen Cave. These bones were not found in the Entrance Hall (455 a.s.l.), but in the area of the

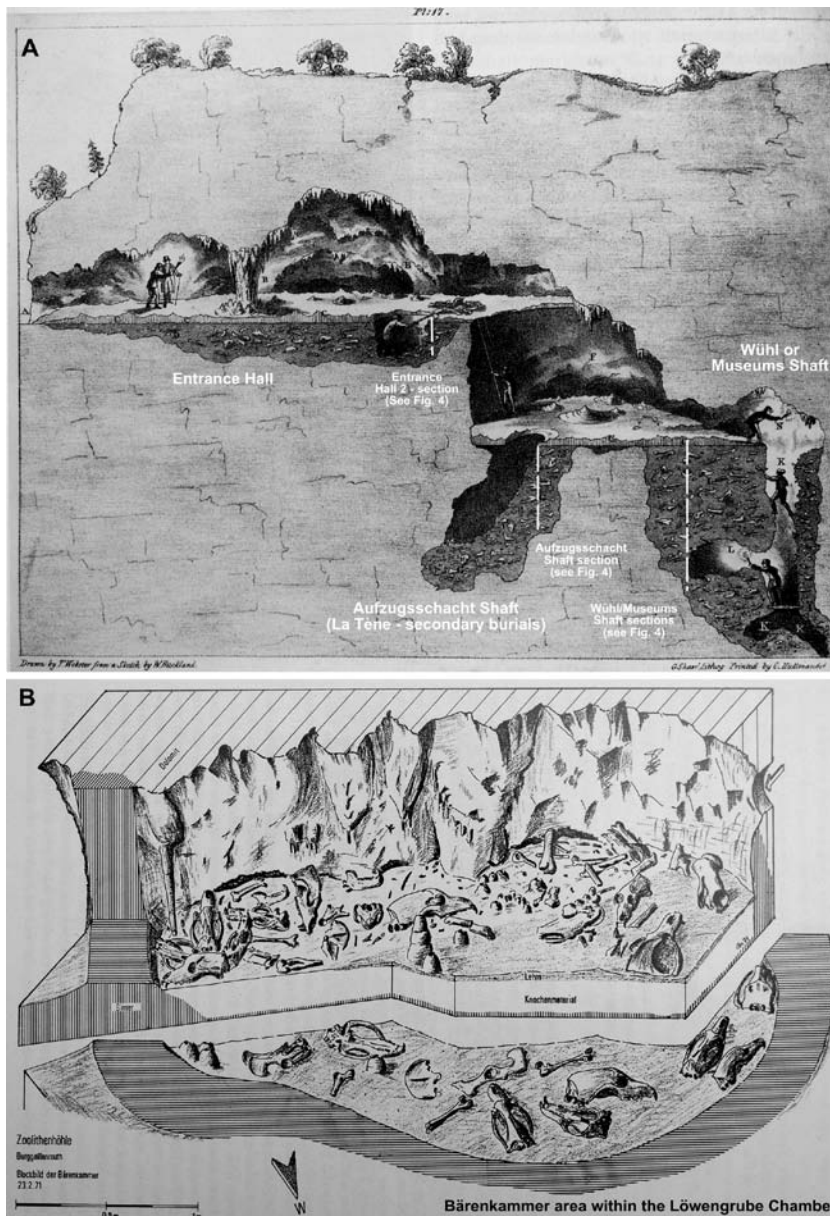


Fig. 2: A. Cross section of the historical excavation areas of W. BUCKLAND (from BUCKLAND 1823). This section *Dieser Aufriss-Querschnitt* shows the historic discoveries of the anterior cave parts and excavation areas in the Entrance Hall, Wühlschacht Shaft (= Iron Age secondary burials) and the Aufzugs-/Museumsschacht Shafts (sections cf. Fig. 4) of the "early spade research time". In the Museumsschacht-Shaft partly the herein illustrated wooden planks of the ladders are preserved until today in the Museumsschacht Shaft. B. Not time-adequate and sadly useless "finding map sketch" of the bonebed surface in the „Bärenkammer“ of the Löwengruben Hall of the "late spade research time" (from GROISS 1971). The removed material can not be relocated today anymore to the places (lacking documentation due to quick-emptying action), nor were the bones osteologically determined.

Abb. 2: A. Querschnitt der historischen Grabungsbereiche von W. Buckland (aus BUCKLAND 1823). Dieser Aufriss-Querschnitt zeigt die damaligen Entdeckungen des vorderen Höhlenbereiches und Grabungsstellen in der Eingangshalle, dem Wühlschacht (= eisenzeitlicher Sekundär-Bestattungsschacht) und dem Aufzugs-/Museumsschacht (Profil vgl. Abb. 4) in der „frühen Spatenforschungszeit“. Im Museumsschacht sind noch heute Balken der Leitern vorhanden, die hier teilweise eingezeichnet sind. B. Nicht zeitgemäße unbrauchbare „Befundplan-Skizze“ der Bonebed-Oberfläche in der „Bärenkammer“ der Löwengruben-Halle in der „späten Spatenforschungszeit“ (aus GROISS 1971). Die geborgenen Funde können heute weder zugeordnet werden (fehlende Dokumentation aufgrund einer Schnell-Entleerungs-Aktion), noch sind die Knochen osteologisch bestimmt worden.

“Aufzugsschacht”, which was a 2 m bone filled shaft, described as resulting from a “biblical flooding event” (ESPER 1774). Initially, bones were easy to collect from the surfaces, starting in the Aufzugsschacht Shaft area. The cave bear, lion and dhole holotype skulls must have been collected by ROSENMÜLLER from the surfaces in those first two larger shafts in the cave system before 1797, because the other cave parts (2/3 of the today’s known cave) behind the Aufzugsschacht had not been discovered at that time (cf. ROSENMÜLLER 1794, BUCKLAND 1823, NIGGEMEYER & SCHUBERT 1972).

1810–1823 – GOLDFUSS to BUCKLAND – “first excavation period”

The sketch of BUCKLAND (Fig. 2B) demonstrates the main bone excavation areas around 1823. GROISS (1979) believed this illustration not to be correct, but it is (see reidentifications in Fig. 2A), even the last shaft can be demonstrated to be the Museumsschacht (after illustrations with two speleothem layers, old ladder material and dump) and not the Wühlschacht. Even today preserved historical ladder material and dumps in the last shaft (Museumsschacht) prove the

deep and extensive excavations at that time, being well figured by BUCKLAND (1823). Those areas can be seen well until today in the cave. Buckland’s excavations began in the Entrance Hall, which can be seen in his illustrations (BUCKLAND 1823, Fig. 2B). The bones are there in primary positions (not water transported) in a silty medium brown and dolomite ash sediment (cf. section in Fig. 4), which minerals gave the bones a typical medium-to dark brown colour. Two preservation and bone colour types GOLDFUSS (1823) are known, the brown form and the more abundant “white yellowish” form. This old excavation area was partly reopened during the own studies (Fig. 4) and some bones (large proportions) and teeth (multiconed enamel surfaces) of *U. ingressus* were found, which are important for dating. The Entrance Hall was not – as wrongly believed – fully excavated, because the sections demonstrate autochthonous and dark-brown coloured cave bear bones, which form a loose bonebed. Furthermore larger speleothems still cover most of the hall, which is only overlain by few Iron Age period and Holocene sediments (Fig. 4). This Hall was less of interest, because in the first shafts (Wühlschacht/Aufzugsschacht/

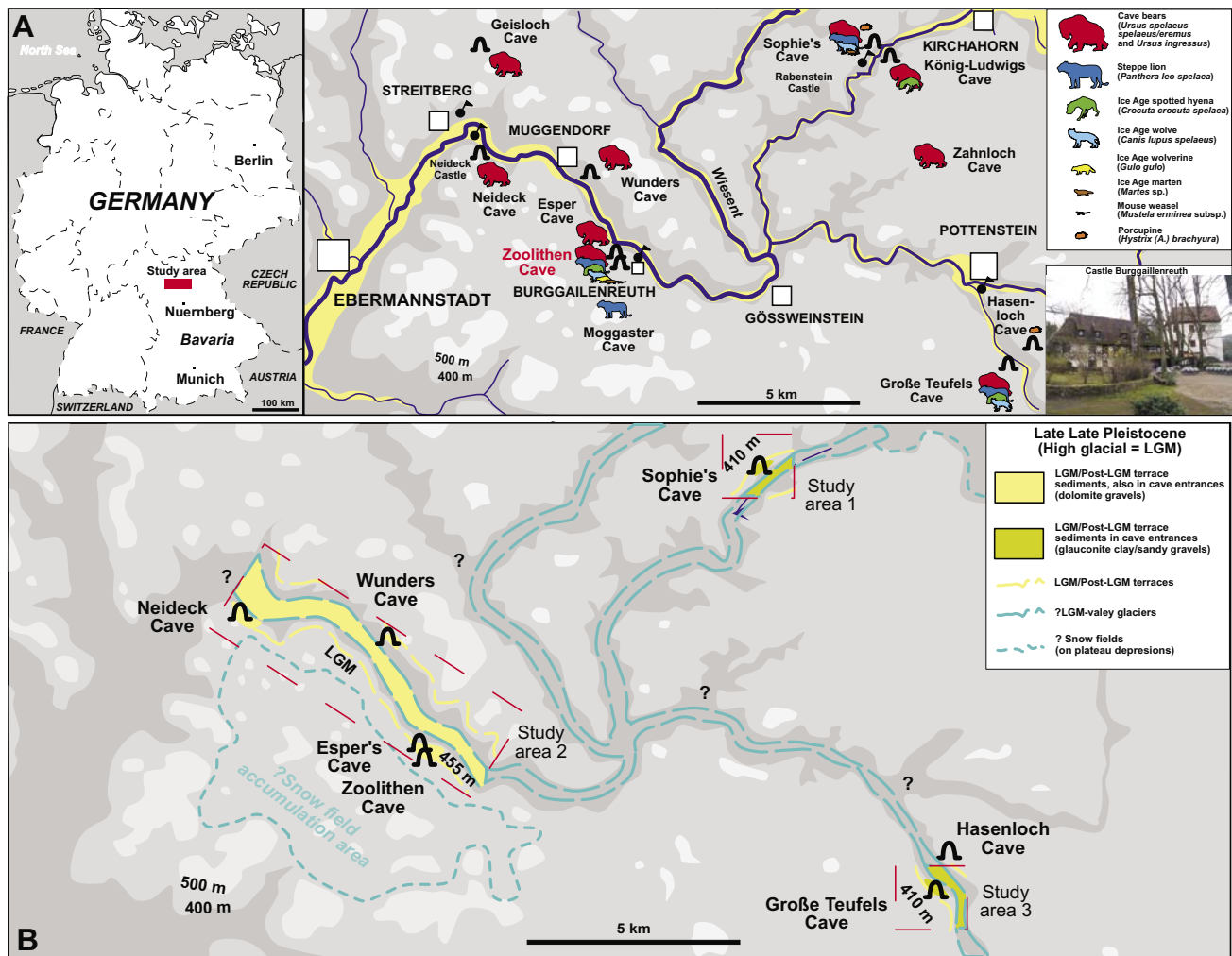


Fig. 3: A. Late Pleistocene cave bear, hyena, wolf, marten and weasel den cave sites in Upper Franconia (northern Bavaria). B. Three sedimentological studied areas in Upper Franconia (Große Teufels Cave, Sophie's Cave and Zoolithen Cave) with questionable "valley glacier situation" explaining best the high elevation of the Wiesent River terrace on 455 a.s.l. elevations and large gorges along during the late Late Pleistocene (high glacial) and terrace relict sediments (especially glauconite till sediments) explaining the possibilities of the flooding situation. The problem is the correlation of the different elevated Sophie's/Große Teufels caves (410 m a.s.l.) elevation and similar old different terrace sediments in the Muggendorf area (Neideck/Wunders/Zoolithen caves about 455 m a.s.l.).

Abb. 3: A. Spät-Pleistozäne Höhlenbären, Hyänen, Wolf, Marder und Wiesel-Horst-Höhlen in Oberfranken (Nord-Bayern). B. Drei sedimentologisch untersuchte Regionen in Oberfranken (Große Teufelschöhle, Sophienhöhle und Zoolithenhöhle) mit fraglicher „Tal-Vergletscherungs-Situation“, die am besten die hohe Wiesent-Flussterrassenlage auf 455 m ü. NHN und tiefen Schluchten während des späten Ober-Pleistozäns (Hochglazial) und Terrassen-Reliktsedimente (besonders Glaukonit-Tillsedimente) und Flutung der Zoolithenhöhle erklären könnte. Das Problem ist die Korrelation der unterschiedlich hoch gelegenen Sophien-/Großen Teufelschöhle (410 m NHN) und gleichaltrige unterschiedliche Terrassensedimente in der Muggendorf-Region (Neideck-/Wunders-/Zoolithenhöhle ca. 455 m NHN).

Museumsschacht) the bone beds were extremely dense and material was easy to obtain. The material was found in the "bonebed breccias" (= cemented dolomite sand-dolomite pebble layers cemented by speleothems) below the upper speleothem layer. Also Graf zu MÜNSTER collected there in the 19th century (WEISS 1937), with his collections now curated in the Urweltmuseum Oberfranken Bayreuth. The sediment of those excavations and further ones of the 19th century was back-filled into the Wühlschacht (pers. com. M. CONRAD). This shaft was closed again after 1823 already historically with reworked sediments. These first discovered areas of the anterior cave separated for long the deeper cave parts which were discovered in 1971 after removal of the sediments (cf. NIGGEMEYER & SCHUBERT 1972). The many hundreds of cubic meters of material protected the deeper parts of the cave and were reworked 25–20 years ago (pers. com. M. CONRAD).

1971 – "second research period"

The second chimney (Wühlschacht) was also filled up with densely packed bones and was the key to today's much larger known cave system, the middle part of which was discovered in 1971 (cf. GROISS 1971, 1979). After the reopening of the Wühlschacht about 25 years ago, a new cave part and thousands (e.g. the University Erlangen collection of 100,000 bones) of untouched bonebeds were quickly removed. Those bones were taken without adequate documentation in the Knochenschacht and partly in the Bärenkammer (cf. Fig. 2B). The Guloloch and Wolfskammer shafts produced mainly cave bear bones, and many new skulls, but also other Pleistocene carnivore faunal remains (cf. GROISS 1979). From those excavations a perfectly preserved hyena skull subsequently became a paratype skull (DIEDRICH 2011a, 2014). Some thousands of other bones have been dumped after

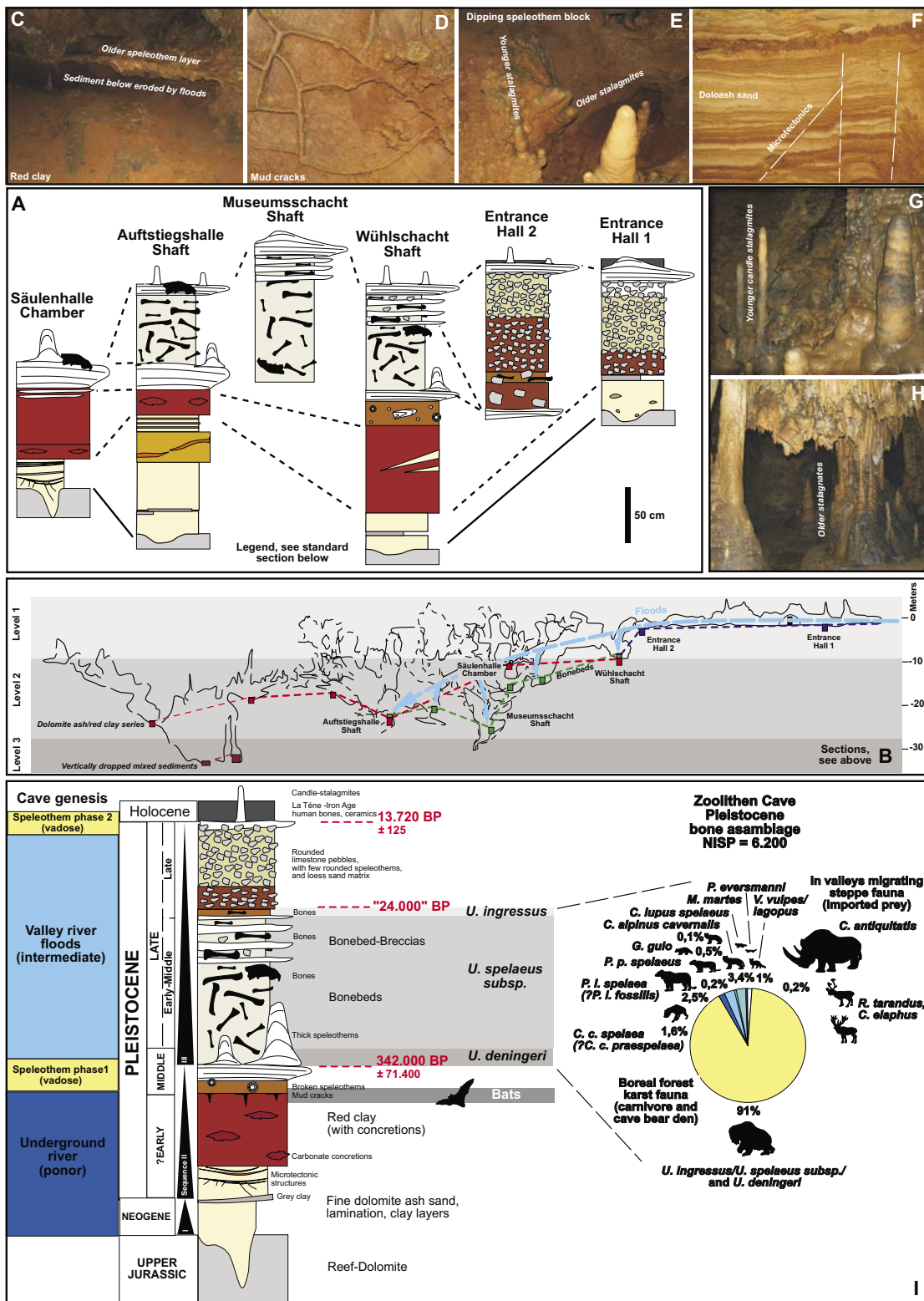


Fig. 4: Sections in the Zoolithen Cave, dating and interpretation of main sedimentary fluvial (underground river and valley river) series (absolute data from KEMPE et al. 2002, ROSENDAHL & KEMPE 2004). A. Important sections. B. Cave levels and similar elevated sedimentary sequences. C. Collapsed main speleothem layer, below eroded clay and present red clay of the sequence 2 in the Lehmhalle. D. Mud-crack negatives on the base of the main speleothem layer in the Löwengrube. E. Tipping speleothem layer in a vertical shaft, which moved downwards, and on which candle stalagmites with different angles developed. F. Dolomite ash sands/red clay layers of the sequence 2 with microtectonic horst structures. G. Candle stalagmites of the Alleröd humid phase (second speleothem generation) in the Löwengrube. H. Large stalagmites from the first speleothem phase in the Säulenhalle. I. Compiled generalized section, and faunal composition of the Late Pleistocene bonebeds.

Abb. 4: Profil in der Zoolithenhöhle, Datierung und Interpretation von fluvialen (Untergrundfluss und Talfluss) sedimentären Hauptzyklen (Absolut-Daten nach KEMPE et al. 2002, ROSENDAHL & KEMPE 2004). A. Wichtige Profile. B. Höhlen-Etagen und in gleichen Höhenlagen vorhandene Sediment-Sequenzen. C. Eingestürzte Sinterdecke, unterhalb erodierter Lehm und vorhandener roter Lehm der Sequenz 2 in der Lehmhalle. D. Trockenriss-Negative auf der Unterseite der ersten Hauptspeleothem-Lage in der Löwengrube. E. In einem Vertikalschacht verkippte Sinterdecke, die sich gravitativ nach unten bewegte und auf der sich mehrere Kerzenstalagmiten-Generationen mit unterschiedlichen Winkeln entwickelten. F. Dolomite-Aschen/rote Lehm-Lagen der Sequenz 2 mit Mikrotektonik-Horststrukturen. G. Kerzenstalagmiten der humiden Alleröd-Zeit (zweite Speleothem-Generation) in der Löwengrube. H. Große Stalagmiten der ersten Speleothem-Generation in der Säulenhalle. I. Generalisiertes Gesamtprofil und Faunenkomposition der spätpleistozänen Bonebeds.

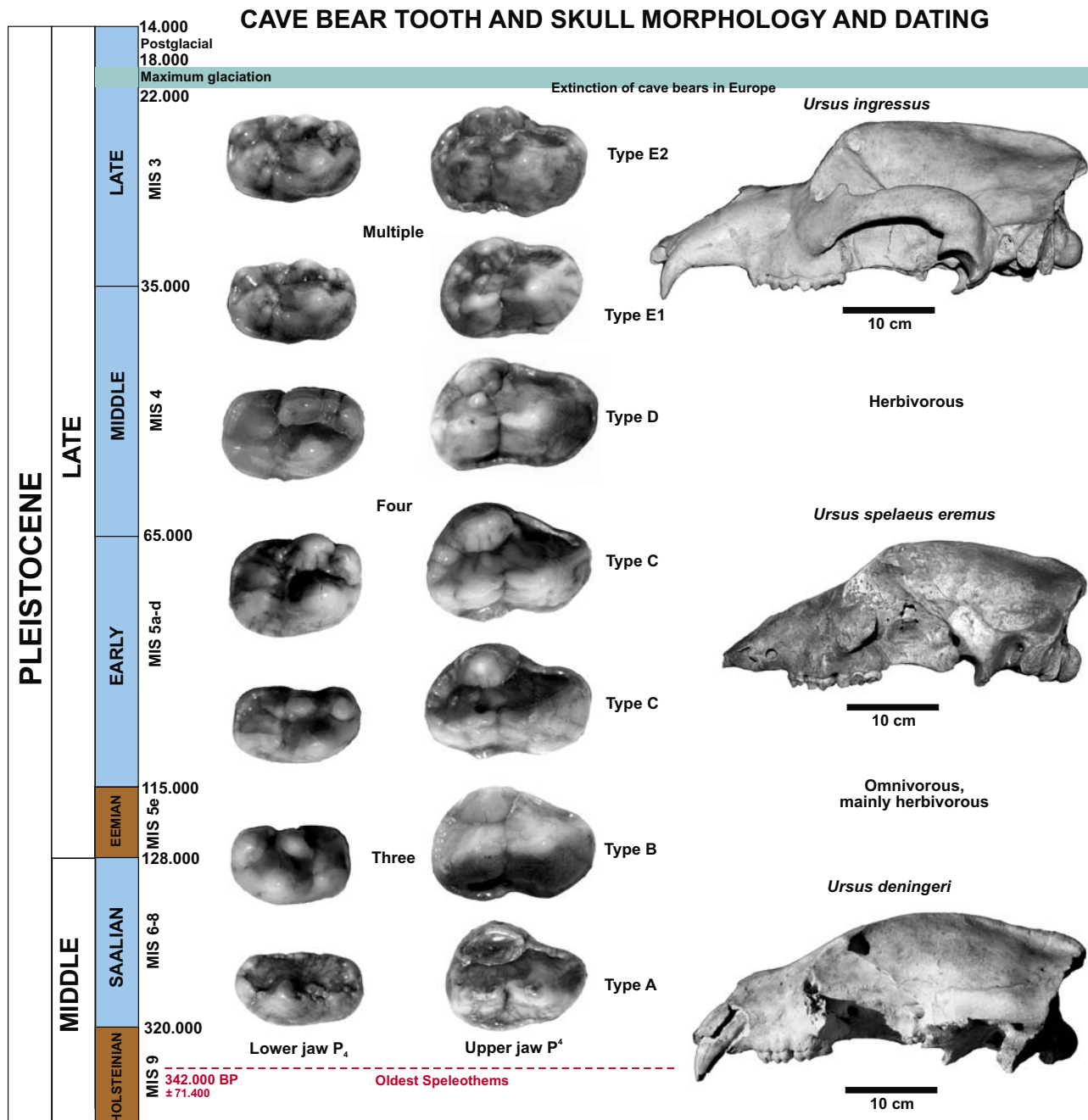


Fig. 5: "Cave bear clock" – dating the cave bears by P₄ and skull shape morphologies. The tooth material is from the dump in front of the cave but demonstrate a cross-section through all primitive deningeri to ingressus cave bear tooth morphotypes sensu RABEDER (1999). All teeth in occlusal view. *U. ingressus* skull Graf zu Münster-coll. Urweltmuseum Oberfranken Bayreuth; *U. spelaeus eremus* skull Buchhaupt-coll.; *U. deningeri* skull ROSENMÜLLER 1794-coll. MB).

Abb. 5: „Höhlenbären-Uhr“ – Datierung der Höhlenbären mit Hilfe der P₄ und Schädel-Morphotypen. Das Zahnmaterial stammt aus der Halde vor der Höhle, zeigt aber den Querschnitt durch alle primitiven deningeri bis ingressus-Höhlenbärenzahn-Morphotypen sensu RABEDER (1999). Alle Zähne in Occlusalansicht. *U. ingressus*-Schädel Graf zu Münster-Slg. Urweltmuseum Oberfranken Bayreuth; *U. spelaeus eremus*-Schädel Buchhaupt-Slg.; *U. deningeri*-Schädel ROSENMÜLLER 1794 Slg. MB).

1971 due to the continuing excavations (FHKF = Forschungsgruppe Höhle und Karst Franken e.V., Nürnberg) in the "Museum" in the cave, and still remain there. This bone material is from the approximately two meter thick bonebeds of Aufzugsschacht/Wühlschacht and the Museumsschacht vertical shafts. Most of the sediment was simply transported in front of the cave, whereas the teeth were taken by the spelunkers, who dumped the "bad bones" in the Museums area. The new "old" reworked bone material is still of high importance in the reconstruction of the exact locations of lion and hyena

remains (Fig. 14), to the understanding of the cave bear bone taphonomy, and to the compilation of the rare non-cave bear skeletons necessary to obtain complete faunal accounts.

2000–2010 – Modern Research

The descriptions of HELLER (1966) do not fit this cave or other caves, where he believed, that those sites were "emptied". After the new discoveries of untouched bonebeds which are luckily still intact in at least some areas (Fig. 15), the cave was explored further in the past decade leading to the dis-

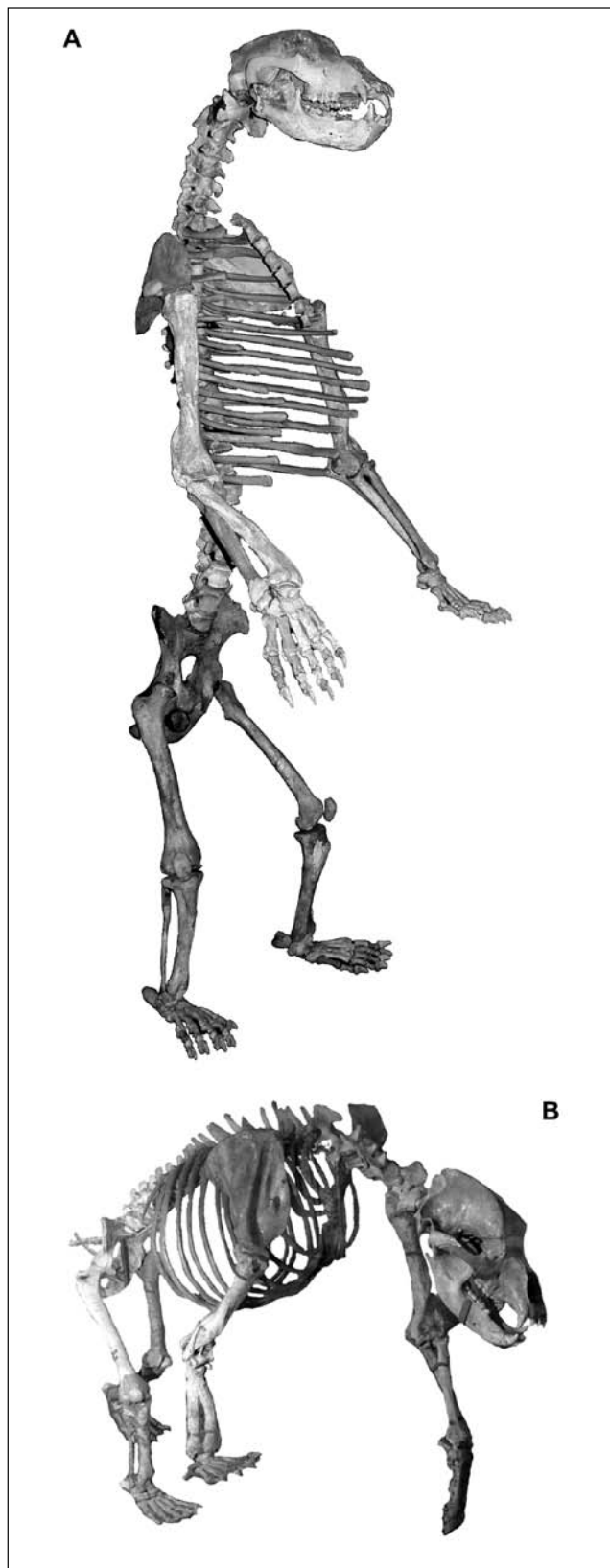


Fig. 6: A. Compiled cave bear skeleton of different individual remains and species of the Zoolithen Cave (exhibition of the Museum für Ur- und Ortsgeschichte Bottrop). B. Compiled cave bear skeleton of different individual remains of the Zoolithen Cave (skeleton exhibition of the Museum Tüchersfeld).

Abb. 6: A. Höhlenbären-Skelett-Komposit verschiedener Individuen und Arten aus der Zoolithenhöhle (Skelette-Ausstellung Museum für Ur- und Ortsgeschichte Bottrop). B. Höhlenbären-Skelett-Komposit verschiedener Individuen aus der Zoolithenhöhle (Ausstellung Museum Tüchersfeld).

covery of another extension to the south (= Dreyerhalle area, pers. com. M. CONRAD, Fig. 14). Also, a few more bones and skulls were found even deeper in the cave system, obviously in secondary (carnivore caused scattering) or even third (floods, gravity-driven transport) positions. The Museums area dump was sorted by the author in 2010. During this preliminary collections management, all non-cave bear material was extracted (together with human remains, and pottery, and Holocene fauna). The hyenas and lion material has already been published, together with the accessible historical finds (cf. DIEDRICH 2011a/b, 2014).

The international important forgotten Pleistocene holotype skull collection

Two “cave bear” skeletons were compiled of bones from different individuals and even cave bear subspecies/species as known today (Fig. 5, 6A–B). In total five Pleistocene species (Figs. 7–13) were named based on the Zoolithen Cave skulls – which make this site to the most important Pleistocene cave megafauna locality in Europe. Today only five of the six holotype skulls remain as valid Pleistocene species. The holotype of the “cave bear *Ursus spelaeus*” described by ROSENMÜLLER (1794) was identified in his collection (DIEDRICH 2009, Fig. 7). Newly identified here is the *Ursus deningeri* REICHENAU 1904 skull (Fig. 8) of the ROSENMÜLLER collection, which was historically believed to represent a “brown bear”. A larger bone collection including most of the known large lions remains and the holotype of *Panthera leo spelaea* (GOLDFUSS 1810) (Fig. 9) from the Zoolithen Cave was collected/excavated by ROSENMÜLLER himself since the end of the 18th century. Also the rediscovered hyena *Crocuta crocuta spelaea* (GOLDFUSS 1823) (Fig. 10) holotype skull (DIEDRICH 2008, 2014), revalidated wolf *Canis lupus spelaeus* (GOLDFUSS 1823) (Fig. 11), the new rediscovered dhole skull of *Cuon alpinus cavernalis* (ROSENMÜLLER 1797) (Fig. 12), and the subsequently invalidated *Gulo gulo spelaea* (GOLDFUSS 1818) (Fig. 13) must have been found in the first two vertical shafts (Aufzugs-/Wühlschacht). However, the invalid “cave tiger *Panthera tigris spelaea* (GOLDFUSS)” holotype material described by GROISS (1996) was revised to represent subadult individual remains of *P. l. spelaea* (cf. DIEDRICH 2011b).

2 Material and methods

To understand the cave bear bone taphonomy and distribution of the “bone breccias” (= bone beds), and to document two articulated skeletons (Fig. 14), a new cave survey was made in spring of 2010, in parallel with a History Channel film project. Open sections in the cave, that resulted from historical digs (cf. BUCKLAND 1823, Fig. 2B) and also the digs of the FHKF since 1971, were studied sedimentologically and stratigraphically to allow the presentation of a generalized overview section for the currently known cave system (Fig. 4). The most complete Middle to Late Pleistocene (MIS 3–9) section was found in the “Aufzugsschacht”. Some thinner sections at other places allow the reconstruction of three main fluvial sedimentary cave filling sequences and two main speleothem genetic phases in three cave levels (Fig. 4). The bone material in the “Museum” (= sorted bones from the 1971 and later reworked sediments from FHKF activities)

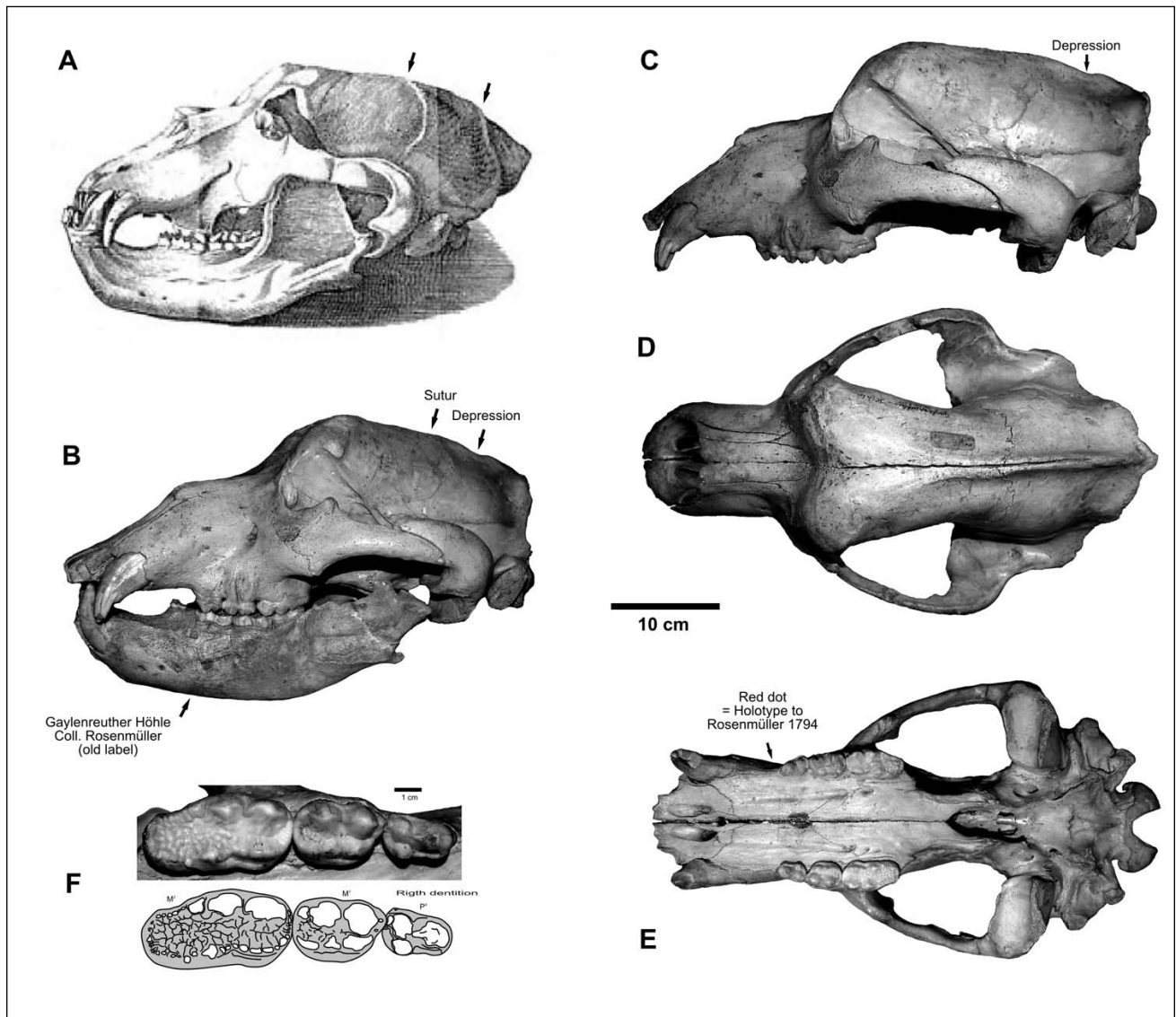


Fig. 7: *Ursus ingressus* RABEDER & HOFREITER 2004 holotype skull of “the cave bear *Ursus spelaeus*” of an early adult male. Latest Late Pleistocene (which seems to be a subadult *Ursus ingressus*; coll. MB, from DIEDRICH 2009).

Abb. 7: *Ursus ingressus* RABEDER & HOFREITER 2004 Holotyp-Schädel „des Höhlenbären *Ursus spelaeus*“ eines frühadulten männlichen Bären, späteres Spät-Pleistozän (*Ursus spelaeus spelaeus* RABEDER & HOFREITER 2004, vermutlich ein subadultler *Ursus ingressus*; Slg MB, aus DIEDRICH 2009).

was sorted and protected against further trampling damage by spelunkers. Non-cave bear bone material was extracted (Pleistocene carnivores, Holocene fauna, human bones). About 2,000 bones remain until today there behind a locked door. Furthermore the dump in front of the cave was checked for its contents with a smaller trench. From this about 1,500 finds (Pleistocene bones, pottery and human teeth) were rescued from the illegal excavation activities of private collectors.

The 1971 and later excavated and already twice redeposited bones and fragments of the non-cave bears are in the ZIEGLER-collection (former owner of the cave at that time) of the Forschungsgruppe Höhle und Karst Franken e.V., Nürnberg (= FHKF). The collection of ESPER cannot be relocated. The most famous and largest collection, that of ROSENMÜLLER, who labelled the site as “Gaylenreuther Höhle, 1797” (cf. Fig. 1B) was formerly stored in the collection of the “Preußische Geologische Landesanstalt”, and then in the “Bundesanstalt für Geowissenschaften und Rohstoffe”,

Berlin (= BGR). These collections were recently moved to the “Museum für Naturkunde der Humboldt-Universität Berlin” (= MB). The Goldfuss collection, which was taken from the cave between 1810–1823, is partially housed in the GOLDFUSS-Museum Bonn (e.g. hyena holotype skull, possibly also “lost wolf skull” there). The Graf zu Münster collection which containing well preserved hyena, lion and wolf material is in the Urweltmuseum Oberfranken Bayreuth (= U-OB). A composite skeleton was studied in the Museum Ur- und Ortsgeschichte Bottrop (= MUOB), and the Museum in Tüchersfeld (= MT). One hyena skull and several cave bear skulls are in the British Museum (Natural History), London (= BMNHL). The Buckland collection was not relocated in Oxford in the University Museum after requests, but might be hidden somewhere. The largest collection (estimated at 100,000 bones after pers. com. Ministry of Culture of Bavaria) is housed in the University Erlangen (= UE, 1971 GROISS “excavations” = cave owner property of R. ZIEGLER until today).

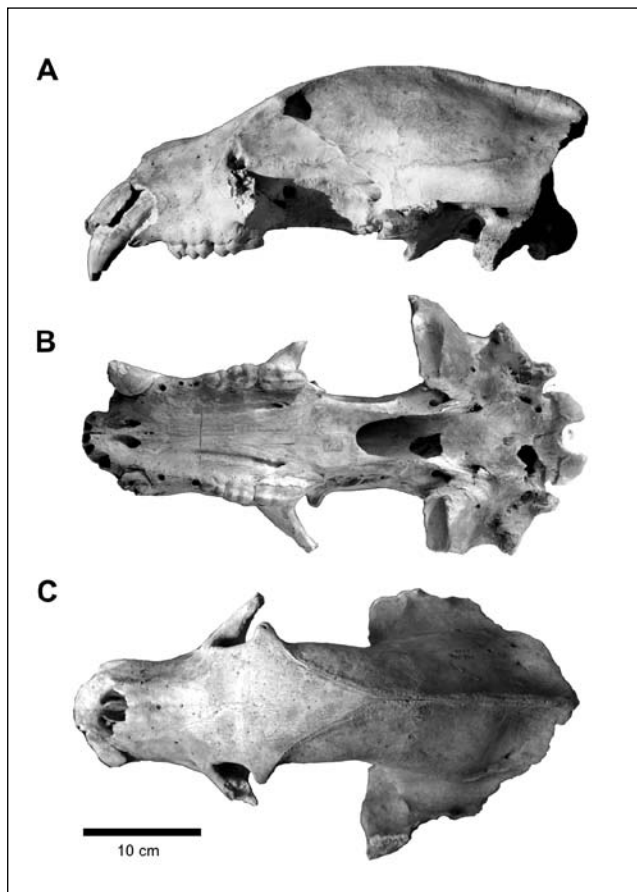


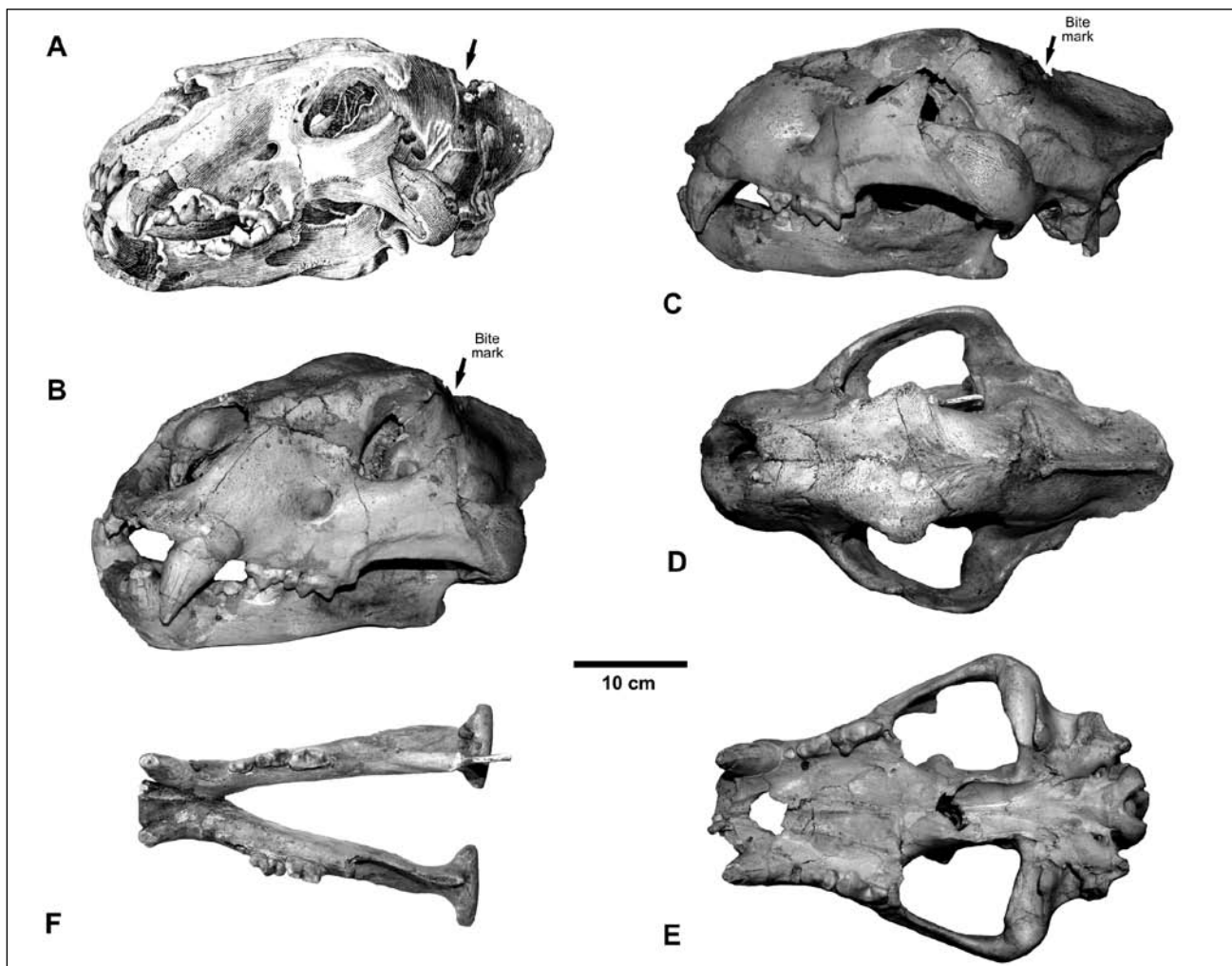
Fig. 8: A-E. Early Deninger cave bear *Ursus deningeri* REICHENAU 1904 skull, Middle Pleistocene ("Ursus arctoideus" in the ROSENMÜLLER 1794-coll. MB).

Abb. 8: A-E. Früher Deninger-Höhlenbär *Ursus deningeri* REICHENAU 1904 Schädel "Ursus arctoideus", Mittel-Pleistozän ("Ursus arctoideus" ROSEN-MÜLLER 1794-coll. MB).



Fig. 9: A-E. Steppe lion *Panthera leo spelaea* (GOLDFUSS 1810) holotype skull "Felis spelaeus", Late Pleistocene (ROSENMÜLLER 1794-coll. MB, drawing from GOLDFUSS 1810, original from DIEDRICH 2011b).

Abb. 9: A-E. Steppenlöwe *Panthera leo spelaea* (GOLDFUSS 1810) Holotyp-Schädel "Felis spelaeus", Spät-Pleistozän (ROSENMÜLLER 1794-Slg. MB, Zeichnung aus GOLDFUSS 1810, Original aus DIEDRICH 2011b).



3 Results and discussion

River terrace relicts in caves and valley genesis in Upper Franconia

NEISCHL (1904) remarked first, that sediments in caves along Upper Franconia river valleys are important for the landscape and glacial dewatering system reconstruction. The first identifications of river terraces and their possible elevations were discussed by SPÖKER (1952) for the Franconia Pegnitz valley. Problems of the valley genesis and dating including micromammal fauna containing caves (e.g. BRUNNER 1933, 1954) were reviewed (HABBE 1989). Only coarse karst evolution models were presented, especially for the earlier “Cretaceous to Tertiary” periods, but not in detail for the Pleistocene valley genesis (cf. GROISS et al. 1998). A new discussion about Plio-/Pleistocene river terraces in the valleys of Upper Franconia appeared with the new sedimentological research at Sophie’s Cave of the Ahorn Valley (DIEDRICH 2013a). At Zoolithen Cave along the Wiesent Valley, the entrance is 130 m above modern river level and must have been flooded postglacially in the Late Pleistocene, as dated by cave bear tooth morphology and stratigraphy (DIEDRICH 2011a, 2013a; Fig. 2). This presented a new idea in the understanding of the much more rapid valley genesis which is

further discussed here, but can be completed only with further studies of the many caves along the river valleys. A first model for the Wiesent Valley branching Ahorn Valley has already demonstrated the exact elevation estimates of Middle (one terrace) to Late Pleistocene (three terraces, DIEDRICH 2013a) Ailsbach River terraces, whereas the different cave bear species/subspecies are highly important for the sediment dating. Those cave sections cannot yet be correlated herein simply to the Wiesent Valley terraces, but both have similar sedimentary sequences in the Late Pleistocene. Important for the understanding of the valley genesis in Upper Franconia are the bonebeds (and different cave bear species/subspecies) and faunal remains in general with their taphonomic record, found especially in caves along the valleys.

Cave genesis, refill stages and animal den use

After a new systematic exploration of the cave in the spring of 2010, its sedimentology/morphology, cave history and geology can be reconstructed including former speleothem and micromammal age determinations, starting with its development in the Tertiary, when the cave was eroded into Upper Jurassic massive dolomites (cf. GROISS et al. 1998, MEYER & SCHMIDT-KALER 1992). The lower cave areas and the upper parts whose levels developed under phreatic underground

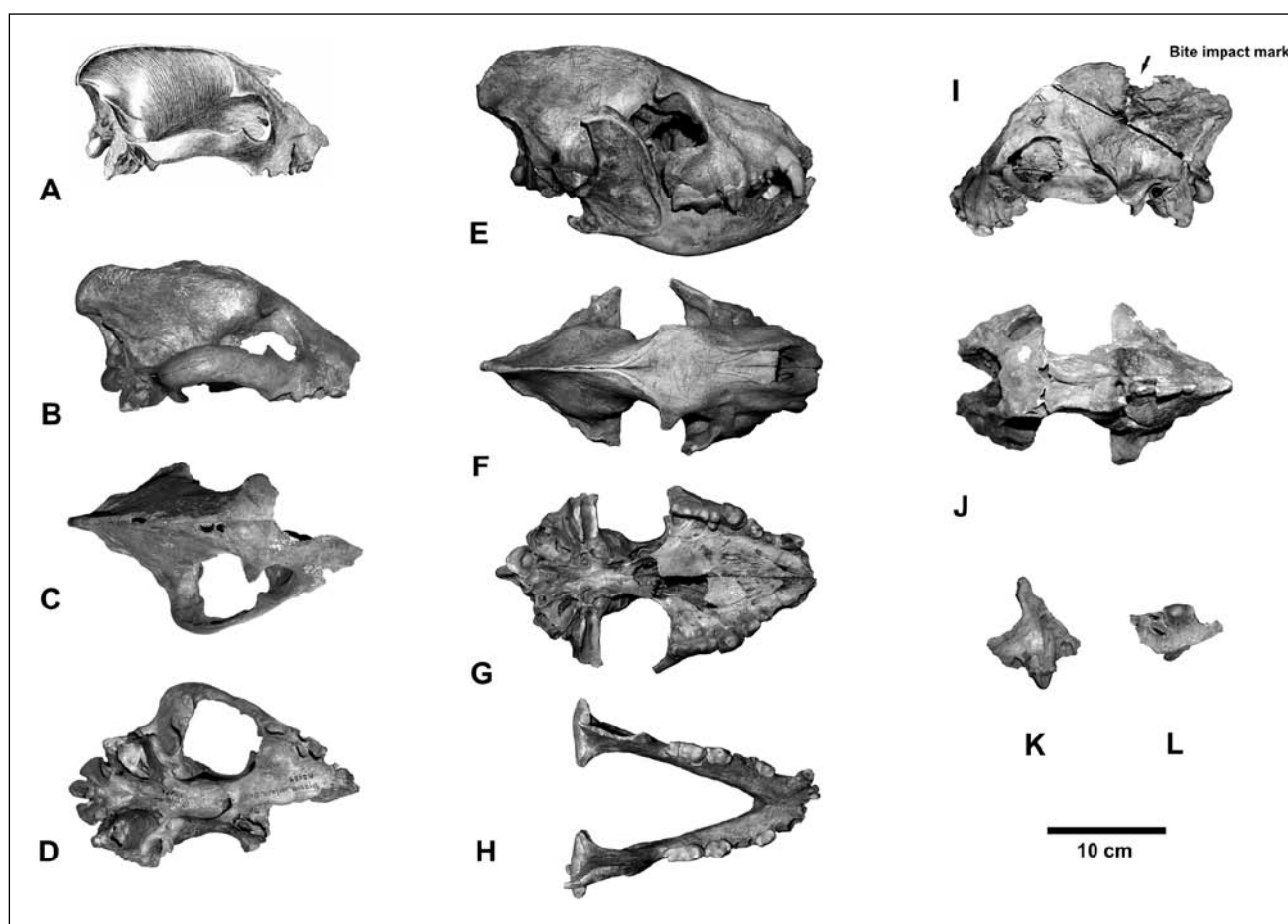


Fig. 10: A-D. Hyena *Crocuta crocuta spelaea* (GOLDFUSS 1823) holotype skull “*Hyaena spelaea*”, mirrored, Late Pleistocene (GOLDFUSS -coll. GMB, drawing from GOLDFUSS 1823), E-H. New paratype skull (ZIEGLER-coll UE, from DIEDRICH 2011a), I-J. Skull with bite damage, original of SOEMMERING, 1828 (from DIEDRICH 2011a), K-L. Maxillary, original of CUVIER, 1822 (coll. GZG).

Abb. 10: Hyäne *Crocuta crocuta spelaea* (GOLDFUSS 1823) Holotyp-Schädel “*Hyaena spelaea*”, gespiegelt, Spät-Pleistozän (GOLDFUSS-Slg. GMB, Zeichnung aus GOLDFUSS 1823), E-H. Neuer Paratyp-Schädel (ZIEGLER-Slg. UE, aus DIEDRICH 2011a), I-J. Schädel mit Bissverletzung, Original von SOEMMERING, 1828 (aus DIEDRICH 2011a), K-L. Maxillare, Original von CUVIER 1822 (coll. GZG).

river conditions seem to have been connected by vadose speleogenesis, following the cleft system (cf. POLL 1972), which explains the often small and vertical partly-branched shafts which connect the three known cave levels (Fig. 2). There must be a deeper active system, resulting in gravity movements in vertical shafts deeper than the lowermost level (about 30 m).

A. Initial ponor cave (“Oligocene/Miocene”)

The 550 a.s.l. high elevated Franconian Moggaster Cave was filled during the Early to Middle Palaeogene (= Palaeocene - Eocene, GROISS et al. 1998). The Zoolithen Cave (455 a.s.l.) is intermediate in elevation between Moggaster Cave (550 a.s.l., Palaeocene/Eocene genesis) and Sophie’s Cave (410 a.s.l., Pliocene genesis, cf. DIEDRICH 2013A) and seem to have been filled with their first sediments no earlier than the Neogene. Here an Oligocene/Miocene age for Zoolithen Cave is expected based on the elevations and dated refill history of Moggaster and Sophie’s caves (cf. GROISS et al. 1998, DIEDRICH 2013e). The first fluvial sedimentary cycle of the Zoo-

lithen Cave is the so-called “dolomite residuum/grey clay sequence” (sequence 1, Fig. 2), whereas those dolomitic sands and silts are typical products of dolomite weathering and fluvial erosion (BURGER 1989). This series was deposited in level 1 in the Entrance Hall (Fig. 2) by an underground river of a Neogene Upper Frankonian Plateau landscape origin, which seem to have been present until the Pliocene (DIEDRICH 2013a).

B. Final ponor cave (Pliocene-Early Pleistocene)

In the Zoolithen Cave, the underground river continued creating two deeper ponor cave levels reaching 25 m deep. Typical scallops (cf. JENNING 1985) on the cave walls are best visible and preserved (i.e. without vadose overprint) in then lowermost level 3 (Fig. 2) in the newest discovered last third of the cave system (branching horizontal parts around the “Siebenschläfer” chamber). The dolomite sands are overlain by up to several meters of red homogenous clay, which contains up to 20 cm large caliche-like concretions in some areas. Those clays are sometimes covered by mud cracks on the

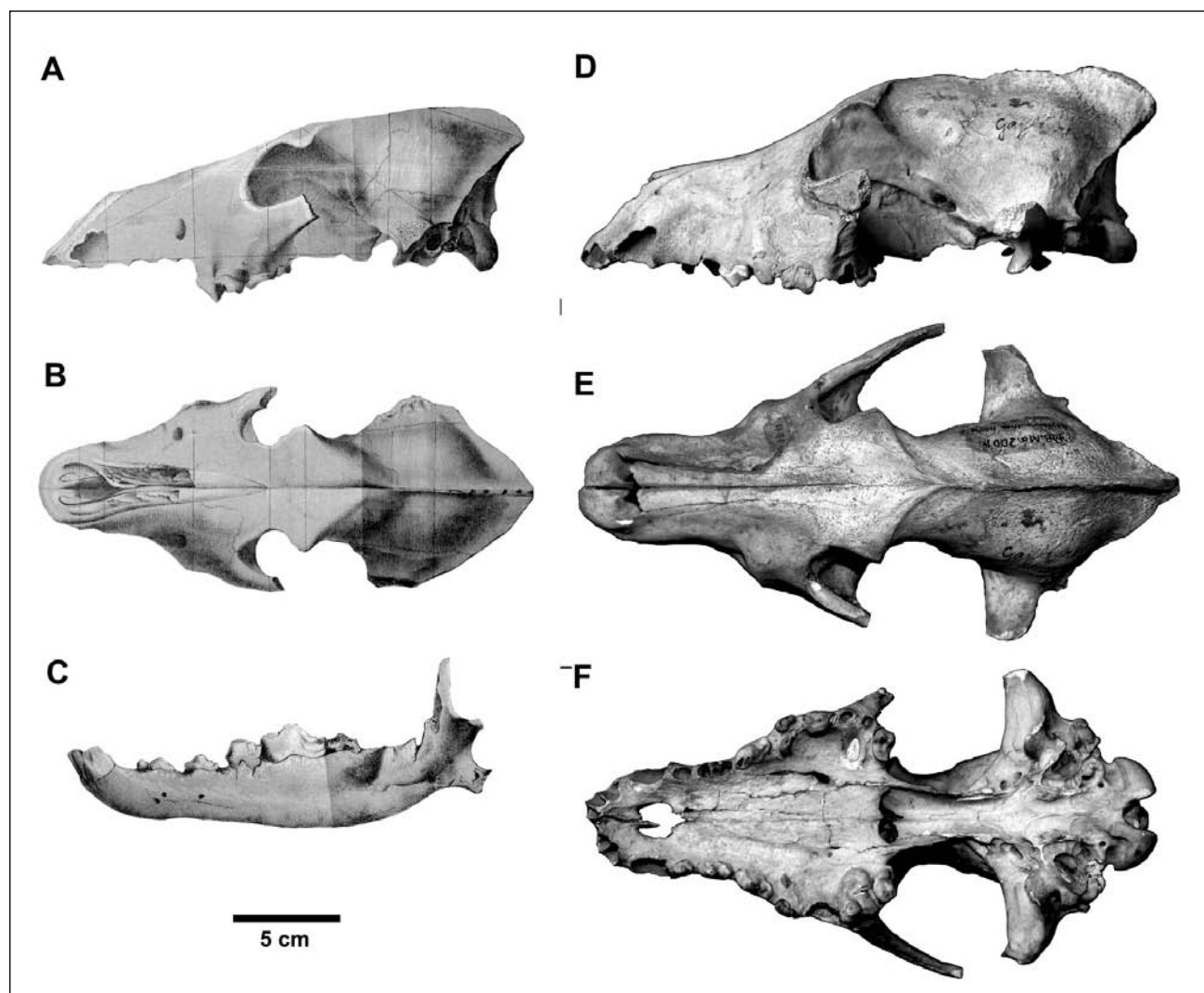


Fig. 11: A-C. Wolf *Canis lupus spelaeus* (GOLDFUSS 1823) lost holotype skull “*Canis spelaeus*” of a cub and mandible of an adult wolf (from GOLDFUSS, 1823). D-F. New lectotype skull (ROSENMÜLLER 1794-coll. MB).

Abb. 11: A-C Wolf *Canis lupus spelaeus* (GOLDFUSS 1823) Verlorener Holotyp-Schädel „*Canis spelaeus*“ eines Jungtieres und Unterkiefer eines adulten Wolfes (aus GOLDFUSS, 1823). D-F. Neuer Lectotyp-Schädel (ROSENMÜLLER 1794-Slg. MB).

uppermost surfaces (Fig. 2) which might date into Pliocene/Early Pleistocene. Absolute dating is not yet available, but a comparison to the Pliocene/Early Pleistocene similar fluvial dolomite sand/clay sequence series sediments of the Sophie's Cave (DIEDRICH 2013e) and the Bing Cave (BRAND 2006) underground river sediments permits a coarse preliminary dating. In the Zoolithen Cave those fluvial series are at higher elevations of about 445–420 a.s.l. (Fig. 2, therefore expected to be older) and in Sophie's Cave at an elevation of 410–400 a.s.l. (expected to be younger). The elevation of this sediment series is even different within the Zoolithen Cave, and sediments are found undisturbed below speleothem layers only in the “second cave level” (Fig. 2), which is some meters deeper than the sedimentary series of the Neogene (level 1 = Entrance Hall). First gravitational vertical transport below the massive speleothem layer (well seen in the Lehmgrube and Siebenschläferkammer, Fig. 2) of those sediments or even possibly earthquakes are documented by micro-horst structures (branch of Säulenhalle, Fig. 2) within the yellowish-white silt/fine sand layers. Such earthquake signs in caves are also reported for the “Middle Pleistocene” of Franconia by SPÖKER (1952) and DIEDRICH (2013a), which would indicate an uplift of Franconia even in the Ice Age, which explains fractured speleothems, or speleothem fragment layers (e.g. in Zoolithen Cave layer below first speleothem phase in the Aufzugsschacht, Fig. 2).

C. Vadose Cave (Middle Pleistocene – first speleothem deposition)

The red and mud-cracked clays demonstrate the drying of the cave, followed by a humid warm period. During this time the main (thickest) Middle Pleistocene speleothem layer developed, and also formed the larger stalagmites/stalactites in the cave. A stalagmite age determination (sample ZooH-Si2; first speleothem generation) is reported around $342,050 \pm 71,400$ years for samples of the Aufzugsschacht section (cf. KEMPE et al. 2002) which corresponds to the MIS 9 interglacial period after the Holstein Interglacial at the early Saalian of the Late Middle Pleistocene (cf. GIBBARD & COHEN 2008). After the speleothem deposition at the end of the Middle Pleistocene gravitational sediment transport happened all over the cave and more extensively later in the Late Pleistocene, whereas at many places today spaces between the clay surface and speleothem layer are between 0.5 to 2 meters. Damages like shedding on cracked and healed stalagmites (Fig. 2), seem to result from such earthquakes or gravity movements on the clay sediments, and are not to interpret as signs of an “Ice Cave”.

D. Dry cave – cave bear, hyena, wolf den (Early-Middle Late Pleistocene)

One articulated early-adult male cave bear skeleton and one skeleton of a cub, still being present in-situ, were mapped in the central but high elevation cave area (“Säulenhalle” and “Zaunikhalle”, Fig. 14). These are important to understand the cave bear bone taphonomy and possible original hibernation areas, and the flood directions. Those are in areas which are difficult to access, and isolated by vertical shafts. Their positions would fit to the theory that cave bears hibernated as deep in caves as possible to protect themselves against top predators, especially lions (DIEDRICH 2011b). The

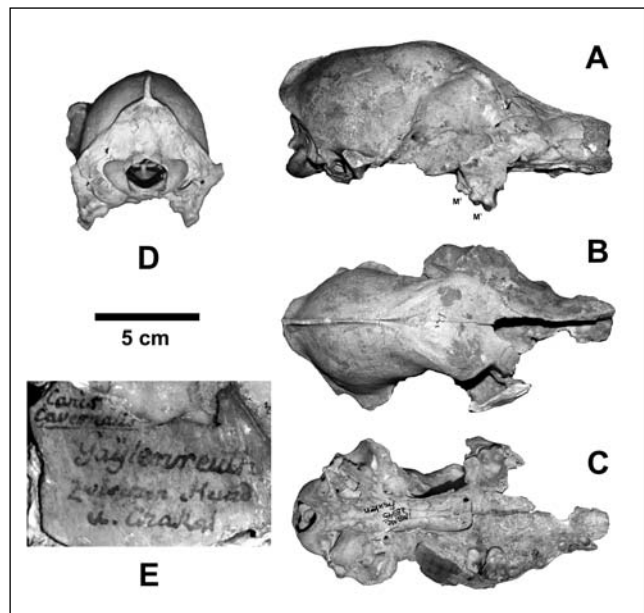


Fig. 12: Dole *Cuon alpinus cavernalis* (ROSENMÜLLER 1794). A. Holotype skull “Canis cavernalis” of an early adult, Middle or Late Pleistocene (ROSENMÜLLER 1794-coll. MB).

Abb. 12: Rotwolf *Cuon alpinus cavernalis* (ROSENMÜLLER 1794). A. Holotyp-Schädel “Canis cavernalis” eines subadulten Tieres, Mittel- oder Spät-Pleistozän (ROSENMÜLLER 1794-Slg. MB).

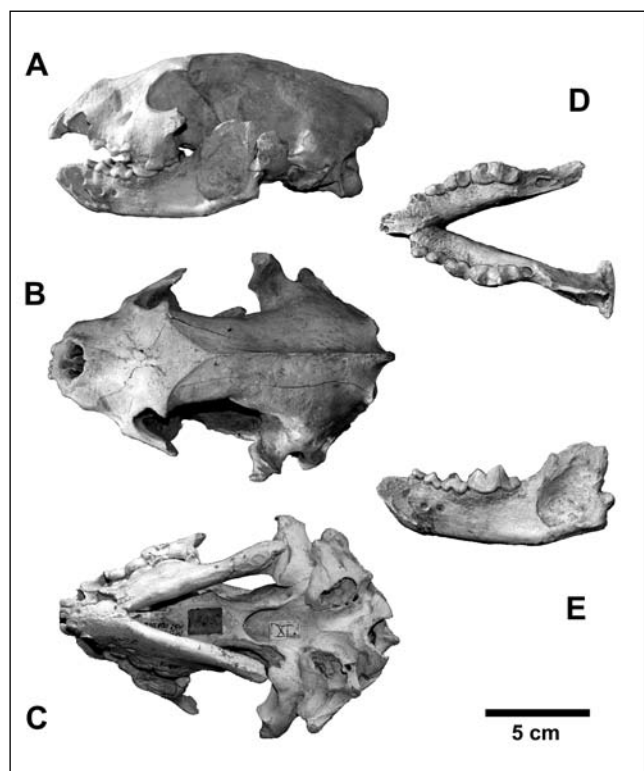


Fig. 13: Wolverine *Gulo gulo spelaea* (GOLDFUSS 1818) holotype skull (coll. MB), today attributed to *Gulo gulo* Linnaeus, 1758 (therefore no longer valid holotype).

Abb. 13: Vielfraß *Gulo gulo spelaea* (GOLDFUSS 1818) Holotyp-Schädel (Slg. MB), heute zu *Gulo gulo* Linnaeus, 1758 gestellt (daher kein valider Holotyp).

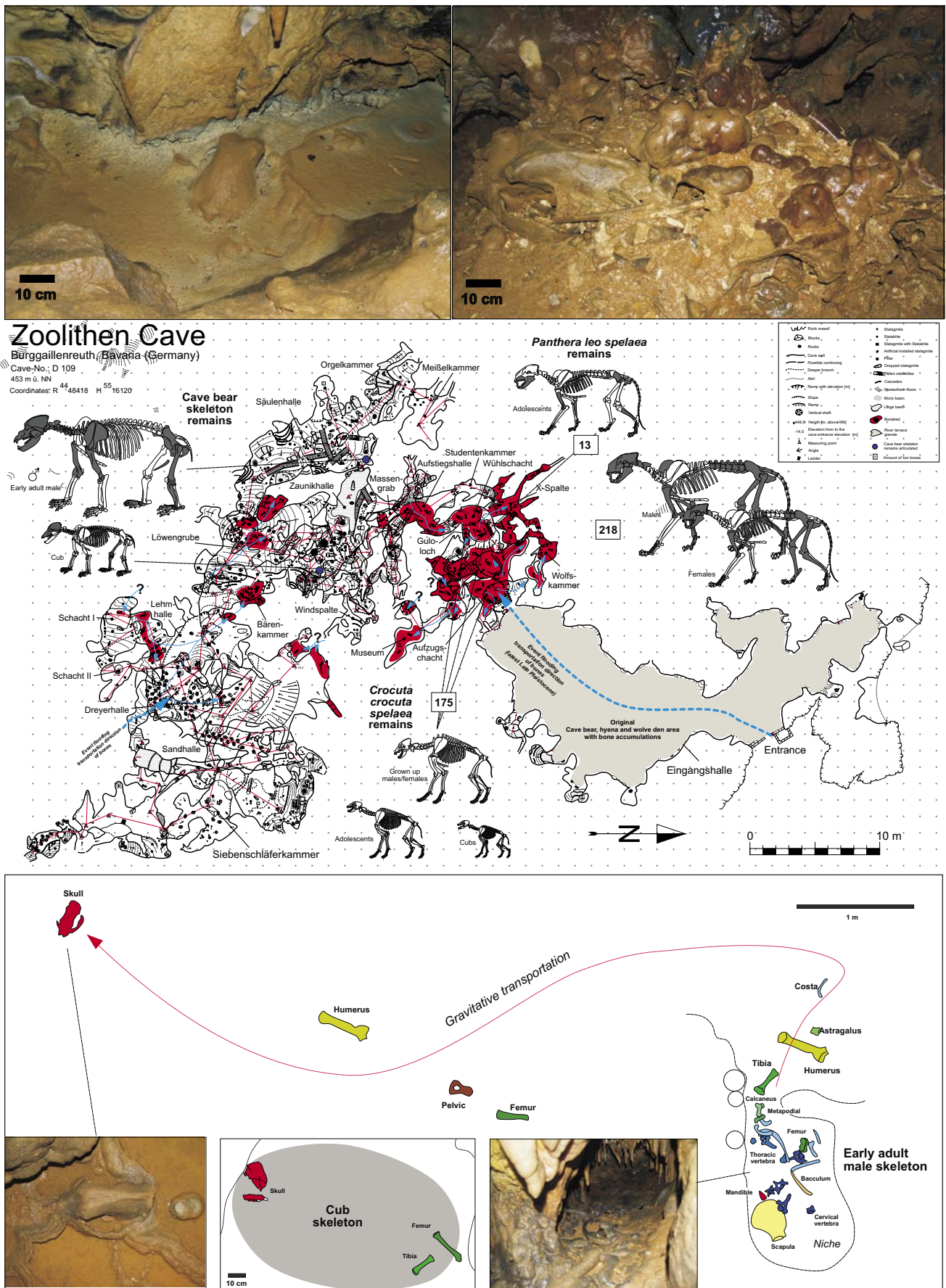


Fig. 14: Different cave bear taphonomy in Zoolithen Cave. Bones redeposited by floods accumulated in bonebeds (red) and two more or less articulated skeletons of a cub and early adult male, which skeleton parts are still on place.

Abb. 14: Unterschiedliche Höhlenbären-Taphonomie in der Zoolithenhöhle. Durch Flutung umgelagerte Knochen, akkumuliert in Bonebeds (rot), und zwei mehr oder weniger artikuliert Skelettreste eines Jungtieres und eines frühadulten Bären, die in situ liegen.

two incomplete skeletons were found on the “upper cave level”, where the Late Pleistocene floods obviously had no impact at all, so it remains unclear how those bears could have reached the “Säulenhalle” and “Zaunikhalle”. The vertical shafts surrounding those are deep and nearly impossible to climb. The early adult male skeleton is most probably of an *U. ingressus* species (skull shape and large bone proportions, covered and fixed by speleothem layer) in a niche – which was his hibernation “nesting area” (Fig. 14), and shows well the slow gravitational movement on speleothem surfaces in chambers. First the skull and larger bones such as the limb bones or the pelvic drifted downwards, through the action of dripping waters.

The hyena and lion remains in the cave bear bonebeds must have accumulated by primary deposition most probably over some thousands of years, from several hyena populations over many generations during the early to beginning of the late Late Pleistocene (DIEDRICH 2011a/b). Already in the Entrance Hall, most probably cannibalistic chewed hyena bones (Figs. 16) indicate carcass movements and damages of the material. Cave bear, hyena or lion bones themselves have sometimes well-preserved irregular chewed margins (= zigzag margins) and bite marks (Fig. 16) resulting from large carnivore activities (resulting from the breaking/scissor dentition). Similar bone bed taphonomic studies have been recently performed at Sophie’s Cave, where cave bear scavenging was convincingly proven on partly articulated vertebral columns of individual skeletons which were found with many other scattered and often bite-damaged bones or articulated body parts (DIEDRICH 2013e). Fragmentation and damage of cave bear bones must have resulted mainly from the well-known hyena scavenging activities (especially bone crushing, Fig. 16, DIEDRICH 2011a), and also by a few lion predatory activities (only joint chewing) and finally by wolf scavenging activities (cf. DIEDRICH 2013b). The Zoolithen Cave taphonomic study and model (Fig. 17) of the “historically non-collected” incomplete cave bear bones shows the same incomplete bone preservations that is being reported from many European caves (DIEDRICH 2009, 2013a/b/c), if incomplete material is included in the studies in such cave bear dens. The new interpretation of scavenging activities by hyenas, and the specialization in middle mountainous boreal forest regions of cave bears as a result of the absence or scarcity of steppe megafauna prey animals (especially mammoth, rhinoceros and steppe bison) in the Zoolithen Cave was recently proven by the large hyena population and den use as a cub raising and commuting den type (DIEDRICH 2011a), and also the very large lion population (DIEDRICH 2011b).

Skull pathologies – Neanderthal or animal conflicts?

GROISS (1978) believed the figured frontal holes and damages to adult cave bear skulls (two different species: *U. spelaeus* subsp., *U. ingressus*, Fig. 17) were the result of “Palaeolithic human hunters” (i.e. cave bear hunting and “spear attack” signs), but in the surrounding cave not one stone tool or Palaeolithic site is known. The only two small Middle Palaeolithic cave sites in the area are near Große Teufels Cave (DIEDRICH 2013a). The incomplete bear skulls are obviously in a hyena and lion predatory context and have damage to each left frontal; in one case the deep penetration is nearly healed, whereas in the other skull the hole is still open



Fig. 15: Cave bear bone taphonomy in the “Massengrab Chamber” of Zoolithen Cave with untouched and still on-place preserved non-sediment containing packed and loose cave bear bonebeds (Photos H. Schabdach, FHKF).

Abb. 15: Höhlenbären-Knochentaphonomie im „Massengrab-Raum“ der Zoolithenhöhle mit unangetasteten und noch Original ohne Sediment dicht gepackten und lose erhaltenen Höhlenbären-Knochenschichten (Photos H. Schabdach, FHKF).

and only the surrounding bone shows an early stage of the healing process (Fig. 17). The interpretation here is different in presenting a cave use model including all three large carnivores (lions, hyenas and wolves) and herbivorous cave bears (Fig. 17). There, mainly lions and possibly hyenas and wolves produced osteological damage during their attacks on the cave bears, although mostly during scavenging activities. Overlooked completely by GROISS (1971, 1978) are bite wounds on the sagittal crests of the steppe lion holotype skull from the Zoolithen Cave (DIEDRICH 2008, 2011b). The most famous hyena skull with a really deep sagittal bite wound (SOEMMERING VON 1828; DIEDRICH 2011a) is another excellent case of a skull with bite damage from Zoolithen Cave. All three specimens have canine tooth bites damage in the areas of the skull where carnivores/bears typically inflict damage during predatory or defensive attacks – the head is the main focus of attack – as in modern hyena/lion conflicts (DIEDRICH 2011c, ROTHSCILD & DIEDRICH 2012).

***Ursus ingressus* or *Ursus spelaeus spelaeus* – what is the holotype skull?**

Whereas the systematics of all the bears of the Zoolithen

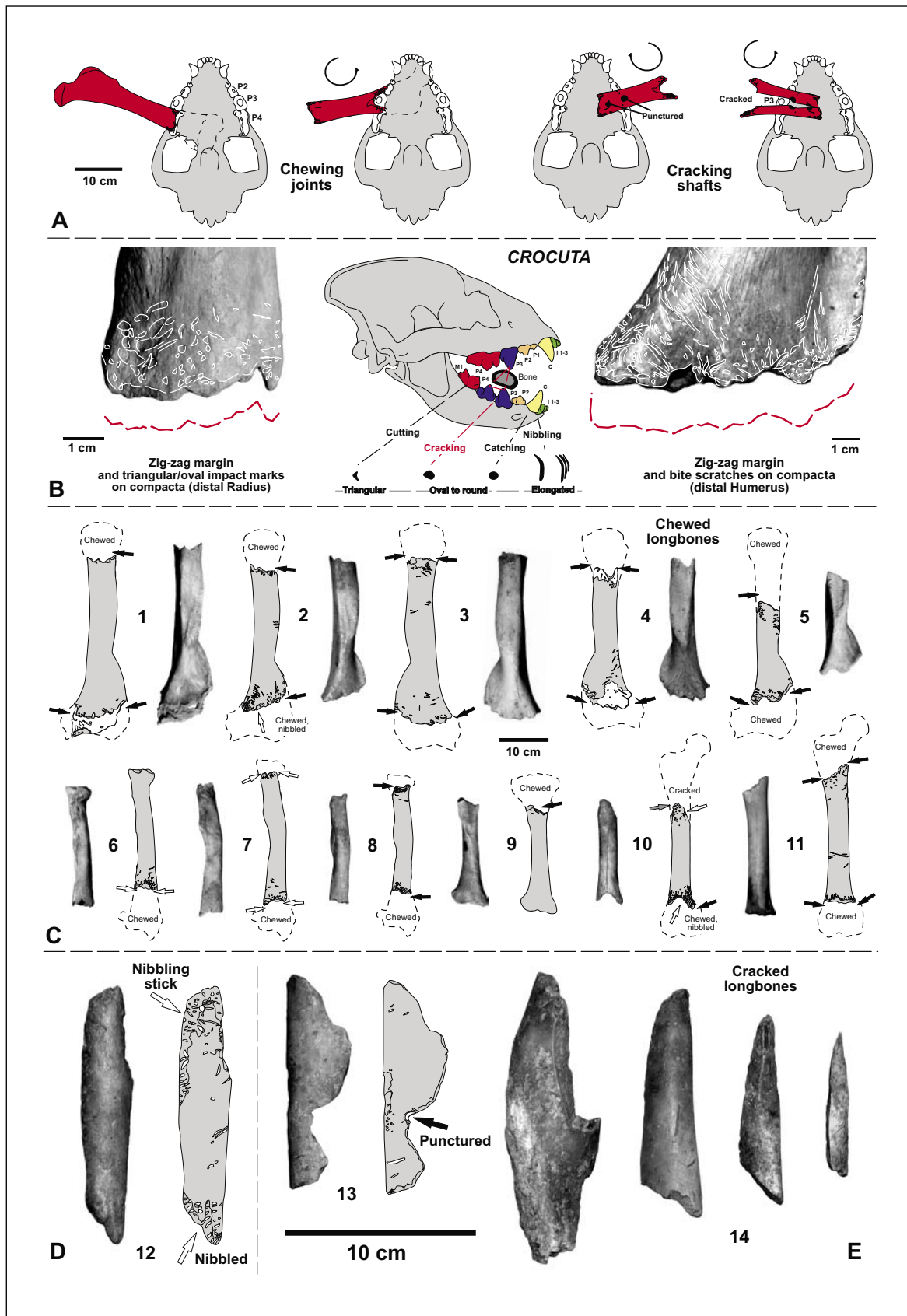


Fig. 16: Postcranial cave bear bones from the Zoolithen Cave and damage history by chew, bite and crush damages (mainly caused by Ice Age spotted hyenas). A. Damaging of a cave bear cub femur. B. Typical triangular, oval and elongated bite mars resulting of different teeth on distally chewed longbone shafts. C. Selected samples of chewed cave bear longbones. D. "Nibbling stick" – a bone fragment used mainly by hyena cubs for teething purposes. E. Fragments of cracked cave bear longbones.

Abb. 16: Postcraniale Höhlenbären-Knochen aus der Zoolithenhöhle und Zerstörungsgeschichte durch Abschneiden, Verbeißen und Zerknacken (primär verursacht durch eiszeitliche Fleckenhyänen). A. Zerstörungsgeschichte eines Jungbären-Femurs. B. Typische triangulare, ovale und längliche Bissspuren, die von unterschiedlichen Zähnen der Hyänen herrühren und an den distalen abgefrissenen Langknochenschäften zu finden sind. D. „Knabberstick“ – ein Knochenfragment, das besonders von Junghyänen zum Abzahnen verwendet wurde. E. Langknochenfragmente zerknackter Höhlenbärenknochen.

Cave has not yet been completely solved, the most recent DNA-analysis of cave bears distinguished at least two different species/subspecies, with *Ursus spelaeus spelaeus* (RABEDER & HOFREITER 2004), and *U. ingressus* (RABEDER & HOFREITER 2004) extant during the Late Pleistocene. The „cave bear“ holotype (Fig. 3C) seems to belong to the latter species/subspecies. Similar results have been published for the Sophie's Cave cave bears (DIEDRICH 2013). Finally, even *Ursus arctos* (Holocene) and another brown bear type of the Late Pleistocene seem to be represented in Zoolithen Cave.

Dating of the cave bears and species taxonomy

Some cave bear teeth in the bonebeds (*Ursus spelaeus spelaeus sensu* HOFREITER et al. 2001) were dated by radiocarbon reaching nearly the limit of this method (around 40.484 BP, HOFREITER et al. 2001) suggesting cave bears have used the cave as a den until the early Late Pleistocene. Other direct and indirect dating methods, which have to be verified with new methods, dated all bonebed material within the early to late Late Pleistocene between “72,000–24,000” BP (cf. GROISS et al. 1998). ROSENDAHL & KEMPE (2004) dated with modern methods some speleothems at the bottom of the cave (Aufzugsschacht) and suggested a mixed bonebed-fauna to range from the MIS 3–8 (late Middle Pleistocene, late Saalian to late Late Pleistocene). However, the megafauna was present already as herein now corrected by the rediscovery of a nearly complete *Ursus deningeri* skull (Fig. 8), which was misidentified as “*Ursus arctoides* REICHENAU 1904” (brown bear, after ROSENMÜLLER 1794). This skull is securely dated as MIS 6–8 (late Holsteinian interglacial or late Saalian glacial). Also the P4 tooth morphotypes (Type A) newly presented here after first studies by RABEDER (1999), that are intermediate between deningeroid and spelaeoid cave bears (see also Fig. 5) support the view that older cave bear populations were present. The Zoolithen Cave is the earliest known cave bear den in Upper Franconia, starting with cave bear denning in the cave during MIS 8–9 (Holsteinian-beginning of Saalian, Fig. 5). Using cave bear skull morphotypes (cf. the new compiled “cave bear clock” in DIEDRICH 2013b), cave bears seem to have established peak populations during the Eemian interglacial (126,000 BP). These smaller classical cave bears (most P4 are of Type C–D) of *Ursus spelaeus eremus/spelaeus* (cf. Fig. 5) occupied the cave during the early to middle Late Pleistocene (MIS 3–5d), whereas *U. ingressus* forms seem to have used the Entrance Hall during the late Late Pleistocene (32–25,000 BP), but also seem to have climbed deeper into the “Säulenhalle” (cf. Fig. 14). The cave use by carnivore and cave bear megamammals ended with the glacial peak around 25,000 BP, following climate change, flood events and cave entrance collapses.

E. River terrace floods (early post glacial peak of the Latest Pleistocene)

In the largest room of the cave, the Entrance Hall, the section starts above a speleothem layer with a medium-brown, large, well-rounded, limestone pebble layer (Fig. 4). In this ~60 cm thick layer, a few cave bear bones or teeth (also P4) have been found only in the lower part, all dark brown (iron/manganese mineral impregnation) in their color. Above those gravels a dark-brown, medium-sized, dense-packed

gravel layer (including rounded reworked speleothem gravels) is similar, but the yellowish, loess-like, one meter thick gravel sediment, is without any bone records. Those gravels are isochronous to the bone beds. The Late Pleistocene sedimentary fluvial cycle continues with the bone beds which vary in thickness as a result of their occurrence mainly in the vertical shafts and its branches (Fig. 4, 14, DIEDRICH 2011a). The sediment between the densely-packed bones (Fig. 15) is coarse grey-greenish mixed dolomite silt/sand, often consisting of very small white speleothem pieces. All the bones are in secondary positions (GROISS 1979), and must have been washed at some places up to 30 m deep (e.g. Wolfskammer). They are in most cases non-rounded and even chewing marks are well preserved (Fig. 16), indicating only short-distance and rapid redeposition within the cave. In total, on all the megafaunal and hyena bones fluvial damage or polishing of edges is rarely observed, which supports the idea of rapid transport of bones, not of “animals and carcasses as described by the “great deluge flood scenario” by ESPER (1774), but who was correct in the general idea of “floods”. The bone material was washed from two different areas and directions of the cave into its central parts (Fig. 14). Most material must have been transported from the Entrance Hall to the central vertical and diagonal shafts (= Aufzugsschacht, Wühlschacht, Guloschacht, Museumsschacht, Wolfsschacht areas, Fig. 14). In the Entrance Hall, the original bone-layers were nearly completely replaced by the river terrace gravels, which were deposited only in the upper layers also in the first vertical shaft, the Aufzugsschacht, but not deeper. The other flooding direction was estimated to have washed bones from their primary deposition site in the Dreyer Hall area into the Löwengrube Chamber and other parts of the western cave (Fig. 14), but there, the river gravels are absent. Nearly all the bones must have been transported in the final Upper Pleistocene (LGM/early Postglacial) into the middle cave part by floods due to a highly elevated Wiesent River terrace and braided Pre-Wiesent River system which must have risen enormously to an elevation of 455 a.s.l. – today's entrance level is now 130 above the today's Wiesent River valley elevation (DIEDRICH 2011a), which cannot be explained by natural river terrace stratigraphy, and only by glaciation models. The floods and river gravels being present only as “Pre-Wiesent River terrace relicts” in only the Entrance Hall area can be dated indirectly into the high to early post high glacial period with cave bear remains which were found below the river terrace gravels in the Entrance Hall (Fig. 3). Those large cave bears (bones brown in colour) are of *U. ingressus* which existed about 32,000–25,000 BP in Europe (HOFREITER 2002, PACHER & STUART 2008, STILLER et al. 2010, MÜNZEL et al. 2011). In similar elevated caves (e.g. Oswald Cave, opposite Wiesent Valley side) similar river terrace dolomite gravels are preserved again in relicts, and are absent along the steep valley margins. The final Late Pleistocene flooding events caused further gravitational vertical transport which caused speleothem collapses and sliding of those fragments into the vertical shafts. Different speleothem generations with different growth angles (well seen in the Lehmgrube, Fig. 4) on such moving speleothem slabs prove long-term mass-movement activity.

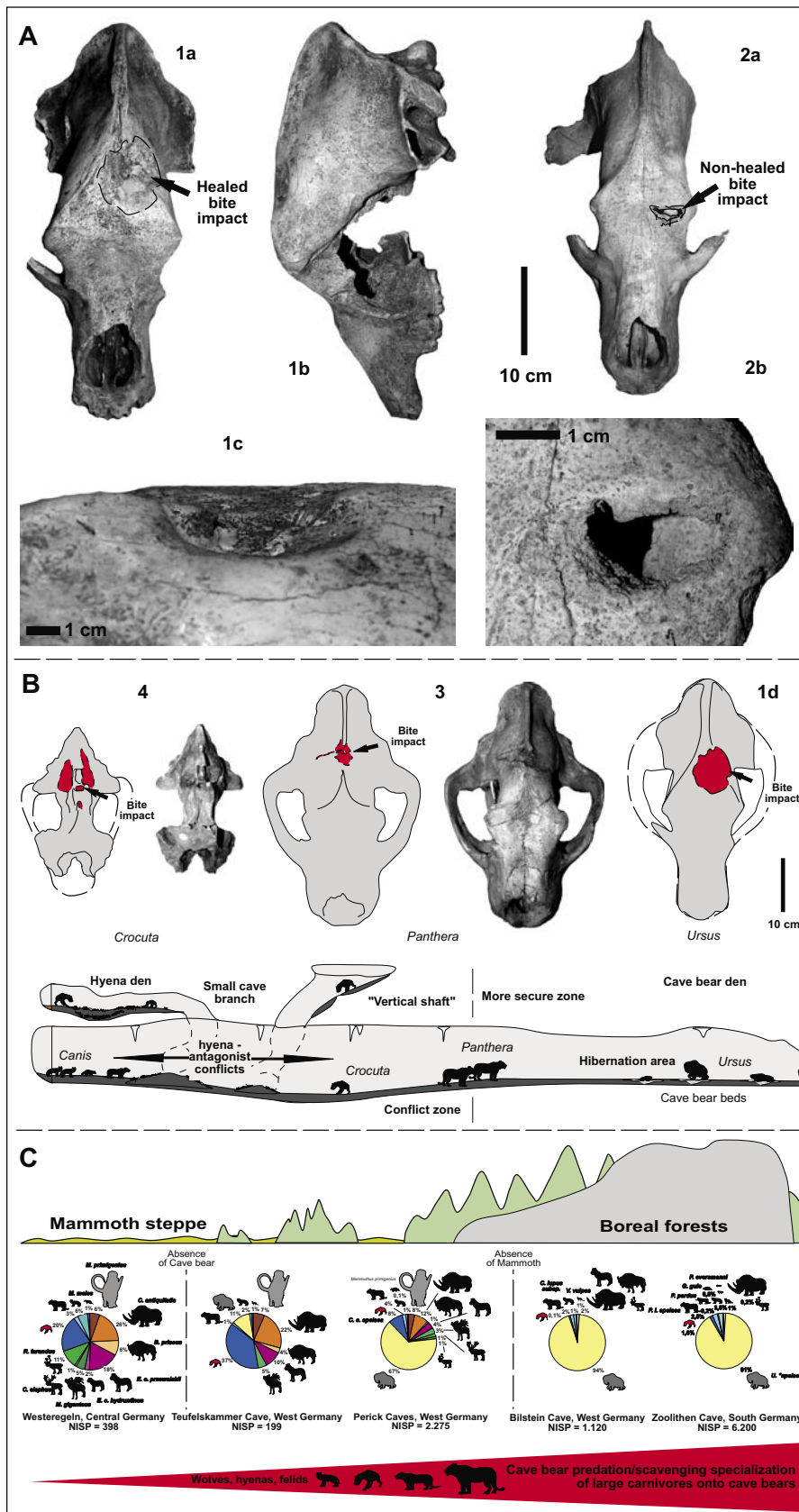


Fig. 17: Bite pathologies on skulls and conflict model for the Zoolithen Cave. A. 1. Senile male cave bear skull (*Ursus ingressus*) with bite damage in the frontal being partly healed (UE no. 142), 2. Early adult cave bear skull (*Ursus spelaeus* subsp.) with non-healed bite damage in the frontal (UE no. 132). B. 3. Lion *Panthera leo spelaea* holotype skull with bite scratch on the sagittal crest (original to GOLDFUSS 1810). 4. Hyena *Crocota crocuta spelaea* skull with massive bite damage on the sagittal crest (original to SOEMMERING 1828). Cave model with conflict zones and antagonism between all three large predators (lions, hyenas, wolves) and herbivorous cave bears hibernating as deep as possible in their den cave to protect against the carnivores. C. Cave bear hunt/scavenging specialization in boreal forest mountainous regions (modified after DIEDRICH 2011a).

Abb. 17: Bissverletzungen an Schädeln und Konfliktmodell für die Zoolithenhöhle. A. 1. Seniler männlicher Höhlenbären-Schädel (*Ursus ingressus*) mit teilverheilten Bissverletzungen im Frontalbereich (UE no. 142), 2. Frühadulter Höhlenbärenschädel (*Ursus spelaeus* subsp.) mit unverheilten Bissverletzungen im Frontalbereich (UE no. 132). B. 3. Löwen *Panthera leo spelaea* Holotyp-Schädel mit Biss auf dem Scheitelkamm (Original von GOLDFUSS 1810). 4. Hyäne *Crocota crocuta spelaea*-Schädel mit massivem Bisschaden auf dem Scheitelkamm (original von SOEMMERING 1828). Höhlen-Modell mit Konfliktzonen und Antagonismus zwischen allen drei großen Prädatoren (Löwen, Hyänen, Wölfe) und herbivoren Höhlenbären, die tief in Höhlen zum Schutz gegen die Raubtiere überwinterten. C. Spezialisierung auf Höhlenbären-Jagd/Fressen in borealen Nadelwald-Gebirgsgebieten (verändert aus DIEDRICH 2011a).

F. The Alleröd (second speleothem genesis and humid cave)

Loose limestone gravels in the Entrance Hall area on the top of the river gravels which are partly cemented by the younger speleothem generation indicate a final Late Pleistocene frost impact. Speleothem dates (uranium/thorium)

of a candle stalagmite (last speleothem generation) on the bone breccias (Aufzugsschacht) give ages on the upper and last speleothem layer of about $11,720 \pm 125$ BP (older data from 1950, see POLL 1972), which is calibrated now $13,720 \pm 125$ BP. Those final Upper Pleistocene aged thin candle-like stalagmites are typical throughout the cave system (Fig. 4).

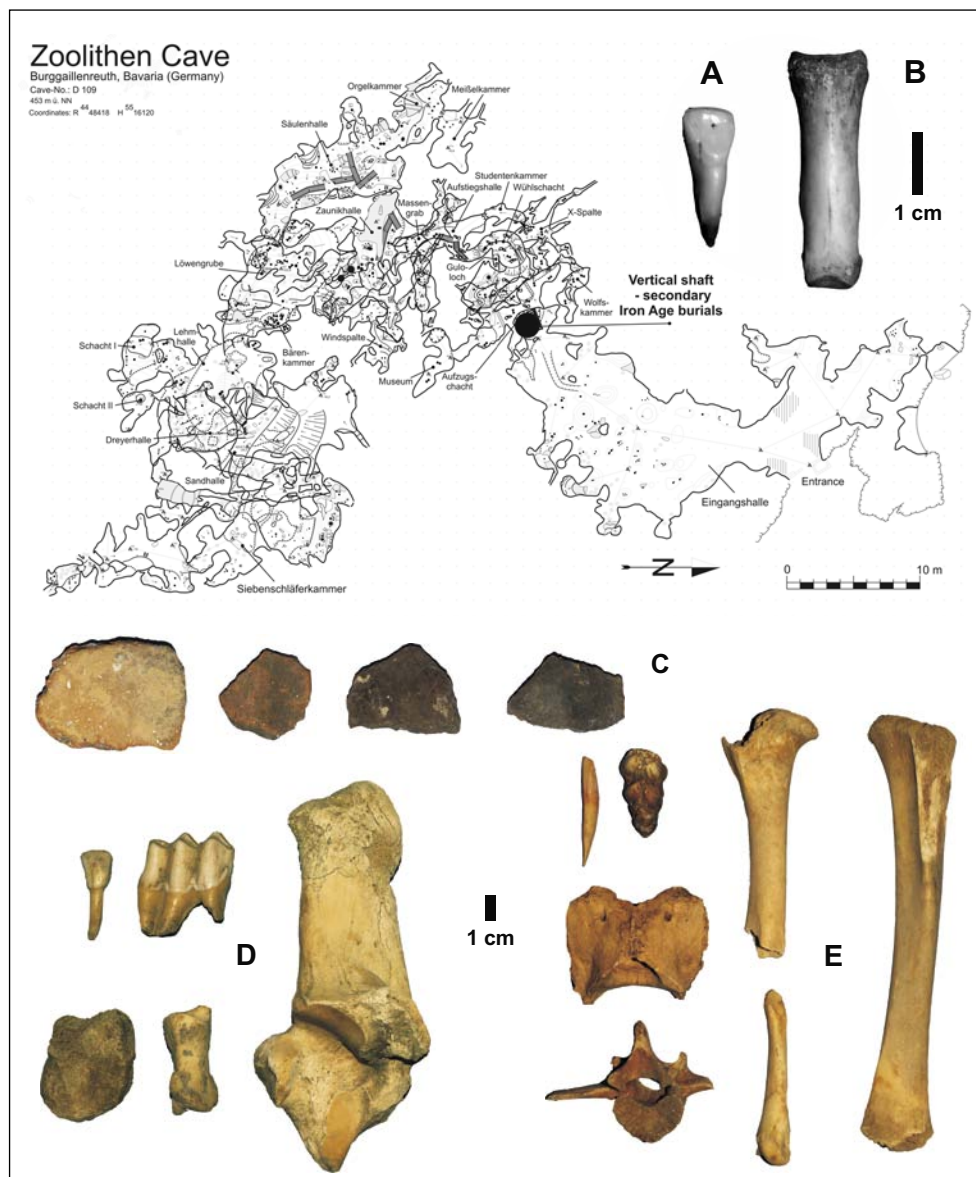


Fig. 18: Iron Age (La Tène) and Holocene bone finds – from the dump in front of the cave (originating from the Wühlloch which was secondary burial place). A. Human incisive tooth and B. Human hand phalanx I. C. Some selected pottery sherds. D. Domestic small cattle teeth and bones. E. Domestic or wild pig teeth and bones.

Abb. 18: Eisenzeit (La Tène) und Holozän-Knochenfunde – aus der Halde vor der Höhle (ursprünglich aus dem Aufzugsschacht der der Sekundärbestattungsort war). A. Menschen-Incisivus-Zahn und B. Menschen-Hand-Phalanx I. C. Einige selektierte Keramikscherven. D. Kleines Hausrind, Zähne und Knochen. E. Haus- oder Wildschwein, Zähne und Knochen.

These seem to be the result of the latest Upper Pleistocene Alleröd climatic change to a warmer period (13,500–12,700 BP; cf. KEMPE et al. 2002). At this time, higher vadose water activity resulted the last speleothem generation all over the Zoolithen Cave. This represents the end period of the massive Pre-Wiesent River terrace erosion in the Pre-Wiesent Valley, now at 130 m lower elevation (measured from Entrance Hall level). As demonstrated for Upper Franconia during the end of the Alleröd Epipalaeolithic, human reindeer hunters had already settled on rock shelters few meters above the today's river valley elevations (e.g. Ahorn and Wiesent Valleys, DIEDRICH 2013a). Possibly the bone beds also moved further downwards by gravity in some vertical shafts, and must have dropped at some places up to 30 m deep (e.g. Wolfskammer).

G. Holocene – Iron Age

The Holocene material was not well distinguished at all from the Pleistocene bones by GROISS (1971, 1979) giving an incorrect view of the “Pleistocene forest fauna” because it included domestic or Holocene wild animal remains (cf. Fig. 18). The Holocene bones were imported mainly by badgers, foxes

and martens (*Meles*, *Vulpes*, *Martes*) to their cave den with different animals (*Felis*, *Lepus*, *Cervus*, *Capreolus*, *Sus*). Also Iron Age humans deposited domestic animal bones (*Bos*, *Sus*, *Ovis/Capra*, *Canis*), which were left in the Aufzugsschacht together with human bones (juvenile and adult longbones, mainly) and pottery (Fig. 18). Also in the Entrance Hall remains of the already known La Tène (Iron Age) culture (cf. SOMMER 2006) are still present in the black sediments.

Conclusions

The sedimentological research gives a first overview of 12 accessible sections in the Zoolithen Cave along the Wiesent River Valley (Bavaria, Upper Franconia Karst, south-Germany). The cave has three main fluvial (two underground river, one valley river) sedimentary sequences, and two main speleothem phases. The first Neogene (Oligocene/Miocene) sediments are only found in the uppermost level of the Entrance Hall and consist of dolomitic sand/grey clay beds resulting from the early ponor cave stage and underground river. The second and third levels, up to 25 m deeper, formed in further underground rivers, which left facets and a second sedimen-

tary series with a yellow dolomite sand/red clay series, being most probably of Pliocene/Early Pleistocene in age. Micro-tectonic structures in dolomite silt/sand layers and clay beds document tectonic activities, such as fractured stalagmites or a speleothem fragment layer. The top of the several meters thick red clay has mud cracks on the surface, and bat remains, indicating a dry cave stage. This changed to a vadose cave in which the first massive speleothem layer developed, dated about $342,050 \pm 71,400$ in the late Middle Pleistocene warm MIS 9 zone. From this time the oldest Upper Franco-nian cave bears, of *Ursus deningeri*, used the cave during MIS 6–8 as a den site, especially during the dry late Saalian period. The cave continued to be dry and was used mainly during the Eemian to late Late Pleistocene (MIS 3–5e) by cave bears for hibernation. P4 tooth morphology and skull shapes allow the separation of three species/subspecies during the Late Pleistocene: smaller *U. spelaeus eremus/spelaeus* and large *U. ingressus*. Two incomplete skeletons are still on place in the middle of the cave, indicating hibernation as deep as possible to protect against top predator attacks. Hyena clans used the Entrance Hall periodically as a cub-raising and commuting den, and are mainly responsible for the cave bear bone damage resulting from a scavenging specialization in boreal mountain forests regions. A large steppe lion population indicates also those lions to have specialized on cave bears, but with active hunts deep in the caves. Wolves must have also sporadically used the entrance area as den site, and must have fed on cave bears, too, as has been well demonstrated at the nearby Sophie's Cave. All megafaunal bones are of a boreal forest assemblage accumulated mainly in the Entrance and Dreyer Hall areas. At the High Glacial (= LGM, about 20,000 BP), a possible valley glacier situation model is presented here using three cave sites and sedimentary sequences around Muggendorf. Pottenstein and Kirchahorn. The Pre-Wiesent River valley was filled only at Muggendorf and the terrace built at 455 a.s.l. In the other areas, glauconitic sandy clays/gravels were found only at 410 a.s.l. elevation (Sophie's Cave, Große Teufels Cave), which makes correlations difficult. The dewatering direction might also have been opposite of the today's direction but more cave sections are needed to develop a more detailed model. The dolomite gravels are found today only as relicts in the caves on both valley sides around Muggendorf. In the Zoolithen Cave, the gravels replaced at least parts of the bone beds in the Entrance Hall, when the bones were washed by flood events into the middle part of the cave and, also from another side of a today's blocked entrance close to Dreyer Hall. This bone material was transported into the Löwen-grube, Bärenkammer, and other parts surrounding the Lehmhalle and contains nearly no hyena, lion or wolf remains – those mainly used the Entrance Hall area. Stronger corrosion on speleothems (especially well below Dreyer Hall) and collapsed speleothem plateaus underline the massive flood impact at the end of the Ice Age (around LGM/Postglacial). Most of the bones were redeposited quickly by those floods (glacier melting waters and seasonally in spring time) in some cave parts and accumulated mainly in the vertical shafts and branching areas where they built up into bone beds several meters thick. With such an unexpected high river terrace position (if those gravels are river terrace layers and not side moraine till deposits flooded over glaciers

into the caves), and high ground water level at the LGM, the “great deluge theory” of ESPER becomes nearly “true” – he was right about the floods, but not with the transport of “live animals and complete carcasses” – indeed only animal bones were washed into the deeper parts of the cave.

There are no “Ice Age human” (Neanderthal or Cro-magnons: Middle to Late Palaeolithics) records from the cave. After the floods the river terrace moved rapidly down (also demonstrated at Sophie's and Große Teufels caves – 50 meters deep erosion LGM/Post LGM) which can only be explained by the presence of valley glacier. Water masses from these valley glaciers caused the steep valley morphology and 130 m of lowering within only about 10,000 years. Finally at the end of the Ice Age in the in the Alleröd humid phase, another speleothem generation with typical candle-like speleothems developed in the Zoolithen Cave around $13,720 \pm 125$ BP. During the Holocene, in the first and second vertical shafts (Aufzugsschacht), Iron Age La Tène humans left numerous secondary burials (skulls and long bones with pottery and after-life food animal donations), but those were thrown only into the first deep vertical shaft. Many new finds of human bones including children and old persons found between the old excavation bone dumps, prove the use of this cave and surrounding caves (e.g. Esper Cave), as burial places – but only in the vertical shafts, this being typical for that time.

Remarks to the future – problems of cave protection and fossil collection

In the dump in front of the cave today, visitors are able to take “souvenirs”, whereas professionals are prevented from saving remaining small bones and teeth by the “Archaeological Monument Survey of Bavaria” legislation. These remains are potentially important for the “region” (local museums project) and an international Natural Monument and potential UNESCO-world fossil heritage site designation. The cave is still in private hands and under control of the FHKF (rented the cave from recent owner). The public collections of the University Erlangen and Urweltmuseum Oberfranken or the Museum Tüchersfeld remain difficult to access, even for professionals, whereas even the Ministry of Culture and Education does not act at all positively. More positively, the Bavarian Landtag recently discussed the problems of collection accessibility in Bavarian Museums and Universities, the protection of the excavation spoil dump and the proposal of the author to make a “paleontological/archaeological” monument by purchasing the cave by the province and thereby preserving Europe's most famous Pleistocene cave fossil site (protection rejected by the Bavarian Landtag Petition no. HO.0594.16).

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