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Dawn and Dusk of Late Cretaceous Basin Inversion in Central Europe

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Abstract. Central Europe was affected by a compressional tectonic event in the Late Cretaceous, caused by the convergence of Iberia and Europe. Basement uplifts, inverted graben structures and newly formed marginal troughs are the main expressions of crustal shortening. Although the maximum activity occurred in a short period between 90 and 75 Ma, the exact timing of this event is still unclear. Dating of start and end of basin inversion is very different depending on the applied method. On the basis of borehole data, facies and thickness maps, the timing of basin re-organisation was reconstructed for several basins in Central Europe. The obtained data point to a synchronous start of basin inversion already at 95 Ma (Cenomanian), 5 Million years earlier than commonly assumed. The end of the Late Cretaceous compressional event is more difficult to pinpoint, because regional uplift and salt migration disturb the signal of shifting marginal troughs. Unconformities of Late Campanian to Paleogene age on inverted structures indicate slowly declining uplift rates.

1. Introduction

During the Late Cretaceous/Earliest Paleogene, Central Europe was affected by a compression event which led to the deformation of the Central European basin. This compression induced significant shortening of the basement, accompanied by the uplift of basement anticlines within the basin, inversion of normal faults, which were appropriately oriented to the newly established stress field, and folding of sedimentary pre-inversion sequences above the thick Permian Zechstein salt (e.g. Ziegler, 1987; Baldschuhn et al., 2001; Kockel 2003, Kley and Voigt, 2008; Kley, 2018). Inverted graben fills and uplifted basement units were eroded and redeposited in newly formed flexural basins (marginal troughs), filled with late Cretaceous clastic sediments. Main deformation is focused on a 200 km wide belt running in NW-SE direction (Fig. 1) from the northern margin of the Lower Saxony Basin to the Franconian Line (southern margin of the Thüringer Wald to Bayerischer Wald). To the north and the south, the magnitude of shorting diminishes rapidly, but the expressions of deformation can be observed from southern England across the North See along the margins of the East European Craton up to the substrate of the Molasse basin in front of the Alps, beneath the Alpine nappes on the Helvetian shelf and to southern France.

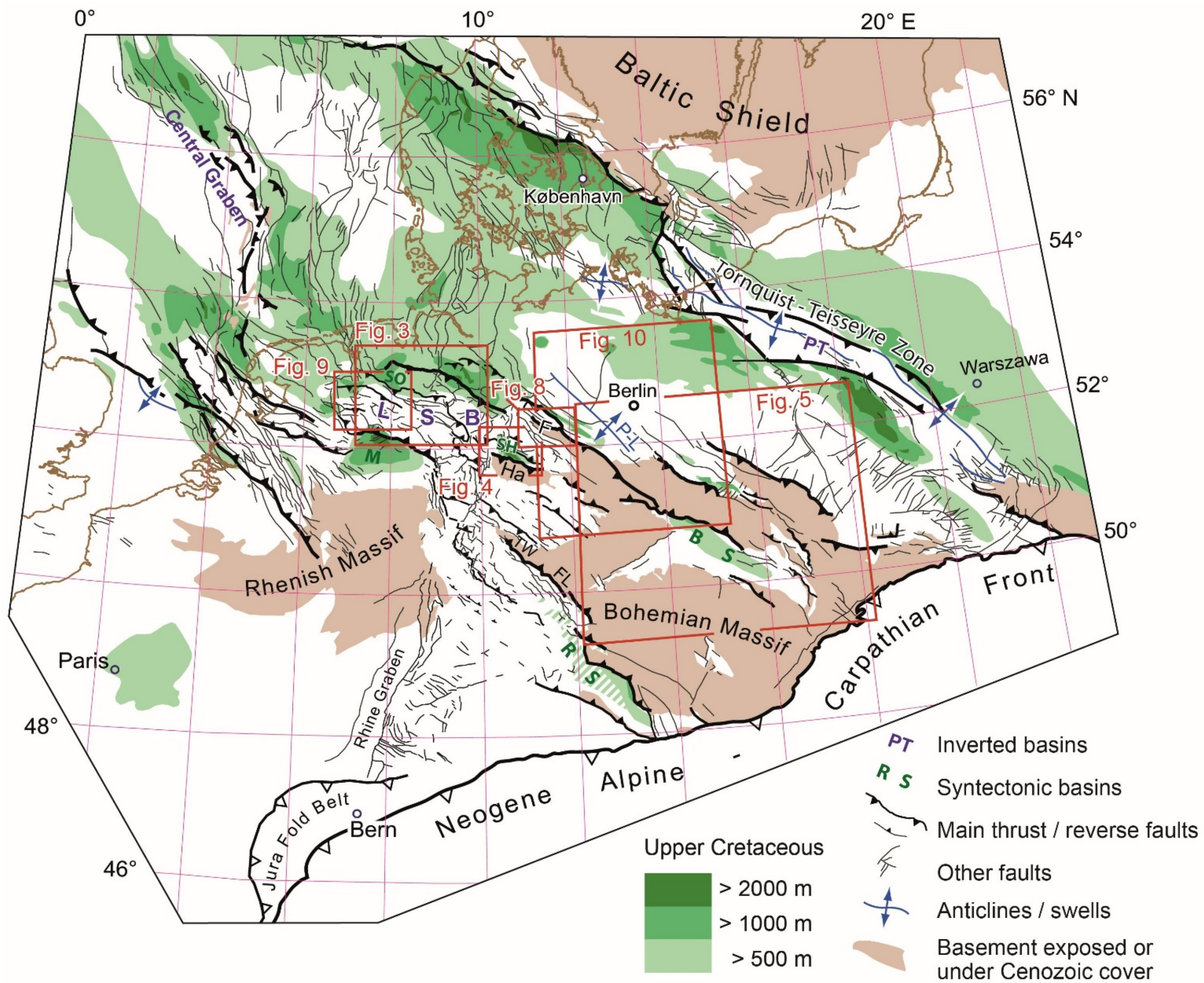

Fig. 1: Overview of Mesozoic-Cenozoic structures and Late Cretaceous basins in Central Europe and surrounding areas, modified from Kley and Voigt (2008). Cretaceous isopachs are modified from Ziegler (1990). Hachured area in RS has preserved thickness < 500 m. Red boxes show locations of Figs. 3 to 5 and 8 to 10 as indicated. Abbreviations: LSB Lower Saxony Basin, PT Polish Trough, SO South Oldenburg Basin, M Münsterland Basin, SH Subhercynian Basin, BS Bohemian-Saxonian Basin, RS Regensburg-Straubing Basin, Ha Harz, F Flechtingen High, TW Thüringer Wald, FL Franconian Line, P-L Prignitz-Lausitz uplift.

2. Late Cretaceous Central European Basin Deformation – known facts and assumptions

The strongest deformation of the European lithosphere is focused on a 200 km broad belt which runs in NW-SE direction and contains numerous basement highs uplifted by several kilometres. It comprises the inverted Lower Saxony basin, the Harz Mountains, Flechtingen High, the Thuringian Forest and the Lausitz-Krkonosze High. Milder inversion (uplift

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magnitudes of 500-2000 m) is observed along the margin of the East European platform (Tornquist-Teisseyre line; e.g. Dadlez, 1980; Krzywiec 2002, 2006; Hansen and Nielsen, 2003; van Buchem et al., 2017) and in some internal structures of the North German and Central Netherland basins (Prignitz and Roer Valley uplifts; Voigt, 2009; Malz, 2019; Zijerveld et al., 1992; Geluk et al., 1994). The amount of vertical displacement may exceed 10 km like in the case of the inverted connected systems Lower Saxony Basin – Münsterland basin (Petmecky et al., 1999; Senglaub et al., 2005, 2006), Harz – Subhercynian basin (von Eynatten et al., 2019) and the Krkonosze-Lausitz – Bohemian Cretaceous Basin (Danišík et al., 2010; Käßner et al., 2019).

Major discussions are related to the kinematics of deformation. While some authors argued for a NW-SE-directed dextral strike-slip fault system and attributed the uplifts to restraining bends (Wrede, 1988, Uličný, 2001), and related basins as transtensional basins to releasing bends, most authors agreed that frontal thrusting was the main process to develop the observed structures (Franzke et al., 2004; Kley and Voigt, 2008; Nielsen and Hansen, 2000). This was also confirmed by small-scale tectonic features at major faults related to inversion, which in many cases preserved even the extensional phase (Franzke et al., 2004; Kley, 2018; Malz et al., 2019; Coubal et al., 2014; Navabpour et al., 2017). The strike-slip model meets the problem that the principal faults should be orientated in E-W-direction to explain the subsidence anomalies at the assumed releasing Riedel shears, which were in fact never observed. The symmetric shape of the basins and their spatial relations to the inverted structures point strongly to frontal convergence as the driving force for basin formation (Voigt et al., 2009).

Navabpour et al. (2017) were able to detect an early phase of N-S compression, oblique to the main NW-SE-striking faults, between the extension and the frontal thrusting. Nevertheless, this phase is not precisely dated yet.

While former interpretations emphasized the role of collision in the Alps for compression (e.g. Ziegler, 1987; Ziegler et al., 2002) or sought the reason for basin inversion in processes of the upper mantle (Kockel et al., 2003), Kley and Voigt (2008) emphasized that synchronous deformation occurred in a broad belt from northern Africa (Morocco), and Iberia, across the North Sea and southern England to the Baltic Sea and Poland. It was connected to a change in relative motion of the European and African plates, resulting in a short-term Iberia-Europe convergence. With respect to direction or timing, this synchronous compression is not related to any deformation phase in the Alps. Instead, the opening of the southern Atlantic led to a northward drift of the African plate and led to a transfer of a compression event via the Iberian Peninsula to the European Craton and its foreland.

E. Voigt (1963) first recognized the formation of Late Cretaceous “marginal troughs” or thrust-load basins (compare Nielsen and Hansen, 2000; S. Voigt et al., 2008; Hindle and Kley, this issue) in Central Europe and detected that their development is often related to the inversion of former basin structures. Nielsen and Hansen (2000) explained the formation of primary marginal troughs through loading by thickened lithosphere of the inverted structures. Primary Cretaceous and secondary Paleogene marginal troughs, found at the margin of the inverted Danish basins, differ in structure and origin (Nielsen et al. (2005). While the first developed due to the load of thickened lithosphere and redeposited sediments on the foreland, the

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latter show a shift of the basin axis away from the inverted structure and are shallower and wider than the narrow Late Cretaceous basins. They were explained by relaxation of the lithosphere and taken as a marker for the end of inversion tectonics already in the Maastrichtian (Nielsen et al., 2005). Krzywiec and Stachowska (2016), emphasizing that such a shift is not known from the inverted Polish basin, challenged this idea. Nevertheless, the described example at the southern flank of the Polish swell shows a remarkable hiatus below an unconformity overlain by Eocene post-inversion deposits. It is therefore not clear whether a Maastrichtian/Palaeocene secondary trough never existed or was eroded during following regional uplift.

A major discussion concerns the continuous or discontinuous nature of deformation during the Late Cretaceous and Paleogene (Stille, 1924; Mortimore et al., 1998; T. Voigt et al., 2004; Kley, 2018). At the northern margin of one of the most prominent basement uplifts, the Harz Mountains, several unconformities where observed (summary in T. Voigt et al., 2004). They were used to distinguish several phases of tilting, erosion and following deposition on the newly created erosion surfaces: the Ilsede phase in the Coniacian, the Wernigeröde phase (with several sub-phases) in the Santonian/Early Campanian and the Peine phase in the Late Campanian. Stille (1924) interpreted these phases as separate (and worldwide occurring) tectonic pulses. Following the same line of reasoning, the Laramide phase was imported from Northern America to explain the major unconformity of Eocene deposits overlying Mesozoic and Palaeozoic basement and deformed Permian to Late Cretaceous deposits units through Western and Central Europe (Stille, 1924). Even younger tilting and erosional unconformities were also observed in western sub-basins of the Southern Permian Basin and related to late Eocene ("Pyrenean") and late Oligocene ("Savian") phases of inversion (de Jager, 2007).

Mortimore et al (1998) for the Cretaceous and de Jager (2007) for the Paleogene, correlated these pulses across Western and Central Europe. But mostly, the ages of these unconformities are poorly defined, because they were determined from sedimentary units covering tilted older rocks. As deposition on such unconformities needs a base-level rise, it is often related to major transgressions. The observed phases at the northern margin of the Harz Mountains therefore most likely reflect the interplay of continuous deformation and changes in base level, which led to phases of erosion and phases of deposition at the margins of continuously active structures. They represent progressive unconformities (T. Voigt et al., 2004). Late Cretaceous marginal troughs within the Polish Basin and in Denmark show continuous deformation, expressed by growth strata (Nielsen and Hansen, 2000; Krzywiec, 2002), while unconformities are limited to the margins and tops of inverted structures. There is no evidence of discrete tectonic pulses during the Late Cretaceous or changing basin configuration resulting from changing stress orientation. The Paleogene inversion in the western part of the Southern Permian Basin is also a more continuous process, spanning the late Eocene to Middle Miocene, but not related to the Cretaceous inversion event (Michon and Merle, 2000; Kley, 2018).

A crude timing of Cretaceous deformation was already established by Ewald (1862) who observed unconformities at the northern margin of one of the most prominent basement structures within the Central European basin, the Harz Mountains, and concluded a Late Cretaceous age of uplift. The following suggestions about a more precise timing were based on several methods, but came to very different conclusions. Most authors agreed that rapid inversion started about 88 Ma ago

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(Coniacian), expressed by rapidly increasing sedimentation rates and a transition from hemipelagic limestones to marly sediments (e.g. Arnold, 1964; Mortimore et al., 1998; Voigt et al., 2006). First evidence of units redeposited by submarine sliding (E. Voigt, 1962) and considerably enhanced thickness of Turonian deposits were taken as markers for the first weak phase of inversion (T. Voigt et al., 2006; Janetschke and Wilmsen, 2014). The fastest uplift of inverted structures and most pronounced subsidence of marginal troughs occurred from Coniacian to Campanian, as reflected by both cooling ages and sedimentation rates. The end of Central European basin inversion is discussed controversially: The interval proposed by different authors reaches from Late Campanian to Maastrichtian (70-65 Ma) to even Eocene or Oligocene (40 Ma). Kley (2018) summarizes the known facts of Late Cretaceous and Paleogene inversion and "Laramide" uplift in western and Central Europe and concludes that they represent three different events concerning both spatial extent and underlying causes. In this paper we will concentrate on a more precise timing of the Late Cretaceous ("Subhercynian") inversion. A second problem that we want to address is the question whether basin inversion occurred contemporaneously across the whole basin or by successive activation of different fault zones. We present sedimentological data from different marginal troughs of Germany, which pinpoint start and end of basin inversion more precisely. The database was compiled from published maps and interpretation of cores and borehole data stored at the Geological Surveys of Saxony and Saxony-Anhalt.

3. Time constraints of basin inversion

Investigations of geometrical patterns, in particular seismic stratigraphy of strata deposited during the extensional phase and in the adjacent marginal troughs (e.g. Baldschuhn et al., 1991; Krzywiec, 2006; Nielsen and Hansen, 2000; Vejbæk and Andersen, 2002), thermochronological data from uplifted basement blocks (Hejl et al., 1997; Thomson et al., 1997; Fischer et al., 2012; Lange et al., 2008; Käßner et al., 2020; von Eynatten et al., 2019) and thermal maturity of the exhumed basin fill (Petmecky et al., 1999; Senglaub et al., 2005, 2006; Luijendijk et al., 2011; Beyer et al., 2014) allowed to constrain the basin history for the majority of active structures. Additionally, sediment composition (pebbles, heavy minerals, and zircon ages) in the marginal troughs reflects the rocks that were eroded and redeposited from the uplifting structures and constrain timing and rates of inversion. This method was applied to few basins, like the Subhercynian Basin (T. Voigt et al., 2006; von Eynatten et al., 2009), the Bohemian Cretaceous Basin (T. Voigt et al., 2009; Hofmann et al., 2013, Nadaskay, 2019) and the Regensburg(-Straubing) Basin (Hofmann et al., 2014).

According to these data, most authors agree that inversion started during the Late Cretaceous and peaked during the Coniacian, Santonian and Campanian. However, the start and end of basin inversion are still debated, according to the varying sensitivity and precision of the applied methods.

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3.1 Fission track and AHe dating

Low-temperature thermochronology, in particular apatite fission track dating (AFT), has been applied to basement rocks across central Europe (e.g. Hejl et al., 1997; Ventura and Lisker, 2003; Lange et al., 2008; Thomson and Zeh, 2000; von Eynatten et al., 2019; Käßner et al., 2020). The data show in many places a rather homogenous signal of rapid uplift and associated cooling of basement rocks between 90 and 70 Ma (Santonian to Maastrichtian), in some cases continuing to 55 Ma (Palaeocene, von Eynatten et al., 2019).

Complete annealing of the apatite fission track system occurs at temperatures above 120-110° C. Partial annealing with shortening of track lengths on geologically relevant time scales occurs down to about 60° C. Fully reset samples must have moved rapidly through the ca. 60-50° C temperature window of the partial annealing zone (PAZ). Estimates of the heat flow during the Cretaceous and results of thermal modeling suggest that the PAZ was about 1.4-2.2 km thick and exhumation of this magnitude is required to cool a sample through the PAZ. Exhumation rates were estimated for the well-constrained case study of the Harz Mountains. Modeled uplift rates based on different cooling ages are in the order of >0.5 km/Ma and in good agreement with the depositional record in the adjacent basin. Earlier estimates were around 1 km/Ma. A rock residing at the base of the PAZ would take between 1.4 and 4 Ma to rise to the top of the PAZ where its age becomes fixed. The onset of deformation may thus predate the timing of cooling deduced from AFT data by a few million years. This effect is accounted for when time-temperature histories are modelled, but should be considered for ages from older studies, or when only central ages are used for comparison with other data.

Discrepancies between thermochronologic ages and stratigraphic indicators of inversion are evident for the southern basement highs, which were affected by regional uplift, leading also to the partial erosion of adjacent basins. Both in the Bohemian Cretaceous Basin and the North Sudetic Basin that are related to the uplift of the Lausitz-Krkonosze High and in the Regensburg basin, which forms the marginal trough southeast of the Franconian line, major parts of the basin fill of the marginal troughs, were removed. The remaining successions only reflect the early stages of inversion (up to the Coniacian). The main stage of basin inversion, which is known from the northern marginal troughs, is not preserved, although AFT ages point to rapid exhumation and maximum redeposition in the particular period from Santonian to Campanian as in the northern basins. In contrast to the strongly fault-controlled uplift und subsidence during basin inversion, the following regional uplift affected both the ancient source areas and the marginal troughs to regional exhumation and erosion.

3.2 Growth strata and progressive unconformities

The evolution of marginal troughs related to basin inversion is caused by thickened crust which loads and depresses the foreland (e.g. Hansen and Nielsen, 2000). As long as uplifting structures in the inverted basin remain below the erosion level in the early stages of tectonic activity, the thickness of a particular unit is increased in the marginal trough and reduced on top of the uplifting structure. However, if swells and basins remain below the influence of storms and surface currents,

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sedimentation derives only from "planktonic rain" of coccoliths and foraminifers, which forms a carpet of uniform thickness and thus obliterates the growing structure to some extent (Hancock, 1989). Nevertheless, Lykke-Anderson and Surlyk (2004, 2007) have shown by interpreting seismic profiles that inversion-controlled changes in sea-floor bathymetry may have generated both erosional features and current-induced redeposition in pelagic chalk successions below storm wave base, resulting from different speeds of bottom current flows. Continuing growth of a swell to above the erosion level leads to redeposition from the swell into the basin, leading to growth strata at the margin of the uplifting structure. As the uplift of most structures has proceeded beyond the erosion level, this early stage is rarely preserved and only the thickened basin fill reflects, probably with some delay due to the early position below the erosion level, the tectonic event. Thickening and growth strata can be detected in seismic sections, provided the thickness difference is high enough (Evans and Hopson, 2000; Lykke-Anderson and Surlyk, 2004, Surlyk and Lykke-Anderson, 2007; van Buchem et al., 2017). The resolution depends on the variability of lithology and seismic impedance in the basin fill.

Ideal conditions for the formation and preservation of growth strata are the following: 1) The fault creates a fault propagation fold and does not evolve as an emergent thrust from the beginning. 2) The front of the active structure remains fixed in one position and does not propagate into the basin. 3) Eventually, the thrust fault cuts through the steep limb of the anticline-syncline pair, leaving the syncline in its footwall. 4) The rotated sedimentary pile containing the progressive unconformities is not eroded in the course of regional uplift after the end of inversion.

Growth structures and unconformities were observed in seismic sections of the Upper Cretaceous across Europe (Krzywiec, 2006; Mortimore et al., 1998; Vejbæk and Andersen, 2002; Nielsen and Hansen, 2000; van Buchem et al., 2017) and used for dating of inversion. Nevertheless, if the basin fill is very homogenous as in the case of the up to 2500 m thick marlstones of the Subhercynian and Münsterland basins, detection of growth strata is nearly impossible. Krzywiec and Stachowska (2016) argued for a structure in the Polish trough that the observed higher total thickness of Upper Cretaceous strata only reflects folding at the margin of the inverted structure and erosional truncation, not increased subsidence in a marginal trough. The distinction between these two cases is only possible if thickness trends of single units are detectable. If the basin margin is involved in the uplift, unconformities can develop. These structures are significant markers of basin deformation, but they occur only in a few places. Only in the Subhercynian Basin, at the northern margin of the Harz anticline, progressive unconformities related to basin inversion are exposed at the surface. Precise dating of progressive unconformities is critical, because in most cases a time gap between the youngest deformed and the oldest covering units is observed. In the Subhercynian Basin, the first inversion-related unconformity occurs at the base of the Middle Santonian (~85 Ma), resting on Late Triassic deposits. Three following unconformities are developed in the Upper Santonian and in the Lower Campanian. Nevertheless, an earlier, Middle Coniacian unconformity is only exposed at the northern margin of the basin and composition and thickness of the basin fill shows clearly that inversion started much earlier (>90 Ma, Turonian). This time gap at the main structure is caused by the progressive rotation of the basin margin and accompanying erosion of older deposits. Older unconformities, which may have been present, were eroded during the main inversion phase (Voigt et al., 2004).

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Van Buchem et al. (2017) described a similar situation from the subsurface. They show an unconformity on top of the inverted Danish Central Graben at the base of the Upper Campanian to Maastrichtian chalk and argue that inversion was limited to the early Campanian. The seismic sections show nevertheless two well-expressed marginal troughs (Turonian to Lower Campanian) on both sides of the inverted structure, thus indicating that compression started earlier and was masked by the high sea level during the Late Cretaceous.

3.3 Facies and provenance

Facies changes may occur even in very early stages of basin inversion as facies is mainly controlled by water depth and source areas. They are observed in Turonian and Coniacian hemipelagic deposits of northern Germany (e.g. Mortimore et al., 1998), characterized by changes in fossil diversity and abundance, colour, and occurrence of hardgrounds or condensed sections. These features are mainly caused by shallowing, believed to be the result of uplift, but are hard to distinguish from processes related to climate variation, active salt diapirism or sea-level changes. The best marker of real inversion tectonics is represented by material shed from uplifting structures. Marginal troughs close to the southern margin of the Central European Basin contain sands, mostly derived from older Triassic to Lower Cretaceous clastic deposits. Inversion-related sandy to conglomeratic deposits allow provenance studies on the basis of pebble and grain composition, heavy mineral analysis and zircon ages. The unroofing sequence was reconstructed for the Subhercynian Cretaceous basin (von Eynatten et al., 2006) and the Bohemian-Saxonian Cretaceous basin (Voigt, 2009; Hofmann et al., 2013, 2014; Nadaskay et al., 2019), with the main result that adjacent basement uplifts had been covered by Late Paleozoic to Mesozoic sedimentary sequences. Nevertheless, most marginal troughs of the Central European Basin were filled with re-deposited fine-grained coccolithic limestones (chalk). They mostly preserve no particular provenance signal of the eroded succession. The provenance signal of uplifting basement structures is also often obscured, because Permian to Mesozoic sediments mostly covered them. Both in the Münsterland Basin and in the Subhercynian Basin, the main coarse clastic input during the early stages of inversion was apparently delivered laterally from other uplifting structures and not from the main evolving highs related to the evolution of the marginal trough. Coniacian sands in the Subhercynian Basin were probably redeposited from Lower Cretaceous sandstones covering the uplifting Gardelegen High and its southeastern prolongation (Voigt et al., 2006; von Eynatten et al., 2008), whilst the basin margin in front of the uplifting Harz Mountains shows only a very thick marlstone succession in this time. This facies probably results from the removal of thick Upper Triassic to Jurassic claystones, which covered the Harz Mountains. The same is observed in the Münsterland basin where Santonian sands were shed from the inverting Roer valley graben, while the inverting Lower Saxony Basin with its thick fine-grained Jurassic to Lower Cretaceous succession delivered the thick marly succession to the axis of the marginal trough.
Although the sedimentary record allows a precise reconstruction of uplift rates and exhumation of uplifting structures in a few cases, recognition of early inversion in the sedimentary record is often ambiguous.

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3.4 Slumps, slides and debris flow deposits

E. Voigt (1962, 1977) deduced a significantly earlier onset of deformation than previously known from unconformities by the observation of slumped and brecciated Turonian marly deposits at the faulted margin of the Münsterland basin (Osning Thrust). The oldest deposits affected are of Middle Turonian age and the slumps were initiated during the late Turonian or early Coniacian. Similar slumps and sedimentation anomalies occur frequently in the chalk of western Europe. They were described from the North Sea basin, the Danish Basin and the Anglo-Paris basin (Hardman, 1982; Bromley and Ekdale, 1987; van Buchem et al., 2017; Lykke-Andersen and Surlyk, 2004, 2007; Arfai et al 2016). Resedimentation is particularly common in Coniacian and Campanian deposits (Kennedy, 1987; Mortimore and Pomerol, 1997; Mortimore et al., 1998; Mortimore, 2011). Oldest occurrences of slumps and slides were reported already in deposits of late Turonian age (Bromley and Ekdale, 1987; Arfai et al., 2016). Surface outcrops in the Weald anticline (Sussex, Dorset) of the Anglo-Paris basin show additionally indications of tectonically induced resedimentation in the chalk (Mortimore and Pomerol, 1991) starting with the Middle Cenomanian.

Submarine slumps develop on slopes of about 3-4° inclination in marly sediments if shear strength is exceeded by gravitational forces (e.g. Embley, 1982; Hance, 2003). Especially unconsolidated, water-saturated mud is prone to such deformational processes. If additional loading of sediments produces pore water overpressure, or cohesion is low, a few degrees of steepening are sufficient to trigger a mass flow. Mass flows depend on the shear strength of unconsolidated deposits and require higher angles in pure chalk and hemipelagic limestones than in cohesive clay-rich sediments, but again the steepening has to be above 3° (Hance, 2003). Unconsolidated sandy deposits form sediment avalanches, resulting in turbidites, if the angle of repose is exceeded. Mass flows are therefore especially abundant in marly and clay-rich hemipelagic deposits (e.g. Hance, 2003). As the observed mass flows (slumps and debris flows) at the active northern margin of the Münsterland Basin are also developed in partly cemented hemipelagic limestones (isolated angular clasts of varying size in marly breccias, associated with slumps of stratified Upper Turonian deposits), the inclination of the basin floor should have been already in the order of several degrees and therefore post-dates the onset of inversion.

3.5 Changes in sediment thickness as evidence for basin formation

Flexure and subsidence of the crust under a tectonic load immediately creates new accommodation space (Nielsen and Hansen, 2000; Hindle and Kley, this issue). If this space is completely filled by deposition, enhanced sediment thickness directly reflects the onset of loading and thus basin inversion. If sedimentation rates are low and the basin deepens without compensation of sediment accumulation, only subdued facies and thickness changes may show the onset of basin inversion. In the case of mild inversion, syntectonic deposition may persist on the tops of uplifting structures that may be revealed by a reduced thickness in comparison to the neighboring marginal troughs.

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Several studies show that thickness variations of clastic deposits in inversion-related basins of central Europe become evident before the onset of inversion determined from thermochronologic ages and provenance studies. This is the case for the Subhercynian Basin (T. Voigt et al., 2006; von Eynatten et al., 2008), the Münsterland basin (Arnold, 1964), and the basin flanking the inverted Mid Polish trough (Krzywiec, 2006; Krzywiec and Stachowska, 2016). In comparison to other features taken as markers for the timing of basin inversion, the differentiation of sediment thickness probably is best suited to pinpoint the onset and end of inversion in central Europe. Therefore, we use in the following the sediment thickness of the marginal troughs to fix the onset of deformation during Cretaceous Basin inversion in Central Europe more precisely. We will use several case studies and data from the Bohemian-Saxonian Cretaceous Basin, the Subhercynian Cretaceous basin, the Altmark Basin and the basins bordering the inverted Lower Saxony Basin: The Münsterland Cretaceous basin and the hitherto unnamed Late Cretaceous basin north of the Rheder Moor-Oythe thrust system in the subsurface of Lower Saxony. Additionally, we will address the question whether all prominent basement anticlines developed at the same time or in a particular pattern.

4 Start of inversion on the basis of thickness differentiation

4.1 Münsterland basin

The Münsterland basin represents the southern marginal trough of the inverted Lower Saxony basin (Fig. 2). The monotonous Albian to Campanian basin fill reaches the highest thickness (>2000 m) close to the Osning Thrust (Arnold, 1964). Thickness of Coniacian to Campanian (marly "Emscher Facies") increases towards the thrust significantly. Slides and slumps prove Turonian to Coniacian uplift and synsedimentary deformation close to the thrust (E. Voigt, 1962). Thickness of the syn-inversion deposits increases towards the thrust although most of the sediments derived from areas in the west (Cretaceous inversion of the Roer valley Graben, Gras and Geluk, 1999) and the south (southern margin of the Cretaceous sea). As the northern margin of the Münsterland basin was tilted and even partly overturned by the activity of the Osning thrust, the increasing thickness towards the central segment of the Osning thrust can be proven even in surface outcrops (Lehmann, 1999, S. Voigt et al., 2008). Thickness differentiation occurred already in the Cenomanian and Turonian with the same depocenters as in the Coniacian to Campanian (Arnold, 1964). Sedimentation rates, nevertheless, are much lower than in the Coniacian and Santonian (Lehmann, 1999; S. Voigt et al., 2008) and in the order of 20 m / My. As the contour of the Cenomanian basin is identical to the structure of the inversion-related Coniacian to Campanian basin, a Cenomanian start of inversion is indicated by increasing sediment accumulation, although no evidence for redeposition from the uprising swell of the inverted is observed at this time. The sedimentation rates start to increase around 93 Ma (S. Voigt et al., 2008).

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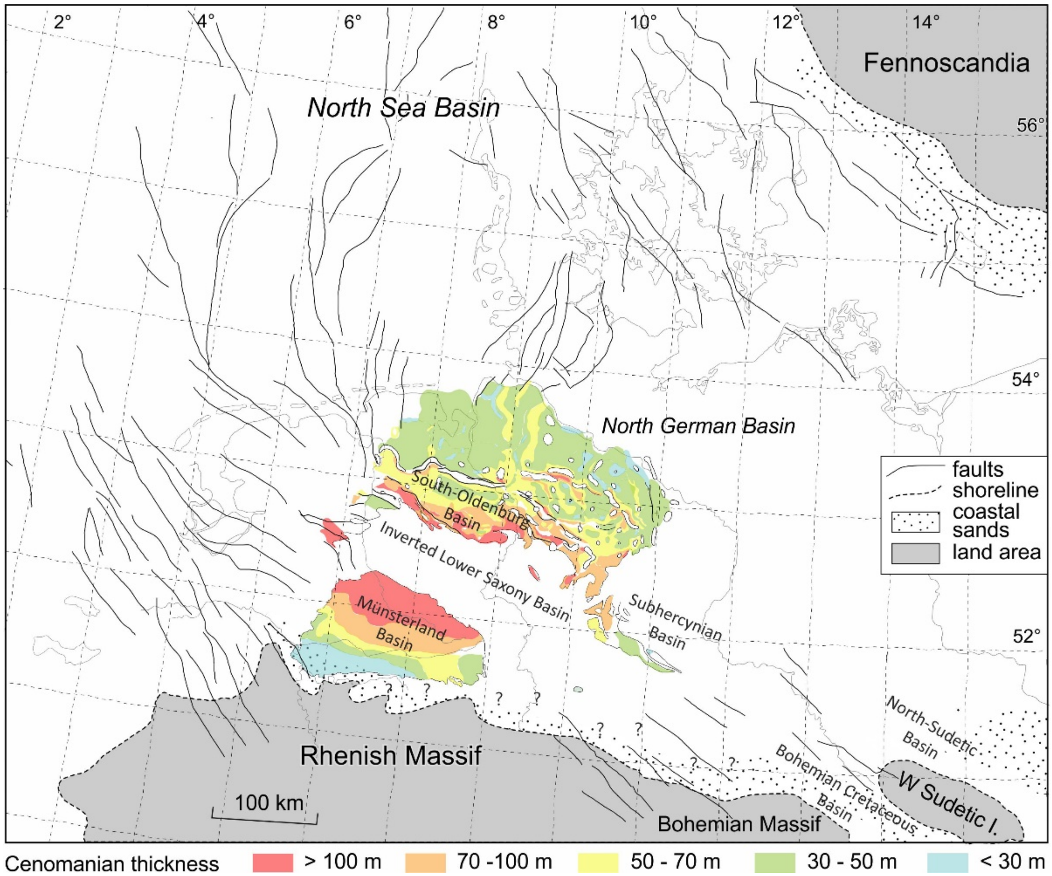

Fig. 2: The southern and the northern margin of the inverted Lower Saxony Basin show enhanced thickness of Cenomanian deposits, indicating stronger subsidence. In these fault-bounded symmetric marginal troughs, thickness of the complete late Cretaceous succession exceeds 2000 m. Enhanced thickness occurs also in the peripheral sinks of salt diapirs in the North German Basin north of the South-Oldenburg Basin (modified from S. Voigt et al., 2008; compiled from Baldschuhn et al., 2001; Arnold et al., 1964 and Frieg et al., 1990)

4.2 South Oldenburg Basin

The South Oldenburg Basin is part of the North German basin (Fig.2) and evolved as a depocentre during the Late Cretaceous north of the inverting Lower Saxony Basin. The graben fill of the Lower Saxony Basin was uplifted several kilometres during the Late Cretaceous (Senglaub et al., 2005). The thickness of Coniacian and Santonian deposits in the adjacent marginal trough (South Oldenburg Basin) increases towards the inverted normal faults of the graben structure of the Lower Saxony Basin, but different from the marginal troughs of the Münsterland basin and the Subhercynian basin it does not attain more than twice the thickness of the background sedimentation (Fig. 3). The northern margin of the inverted Jurassic to Lower Cretaceous Lower Saxony basin is marked by a system of thrust faults forming the Rheder Moor-Oythe structural trend (Fig. 3). These thrusts developed from the reverse reactivation of a swath of normal faults accompanying the

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northern margin of the Lower Saxony basin, a large Jurassic to Lower Cretaceous graben system (Baldschuhn et al., 1991; Kockel, 2003). The northern foreland (South Oldenburg Block) is characterised by a strong influence of salt diapirs on deposition, starting no later than the Jurassic and probably already during the Triassic (Kockel et al., 2003; Warsitzka et al., 2019). Rising salt domes and subsiding peripheral sinks around those diapirs also influenced the general pattern of Late Cretaceous thicknesses. The facies of northern Germany is dominated by chalk, well-investigated with respect to typical log-patterns and biostratigraphy in boreholes (Baldschuhn and Jaritz, 1977; Koch, 1977).

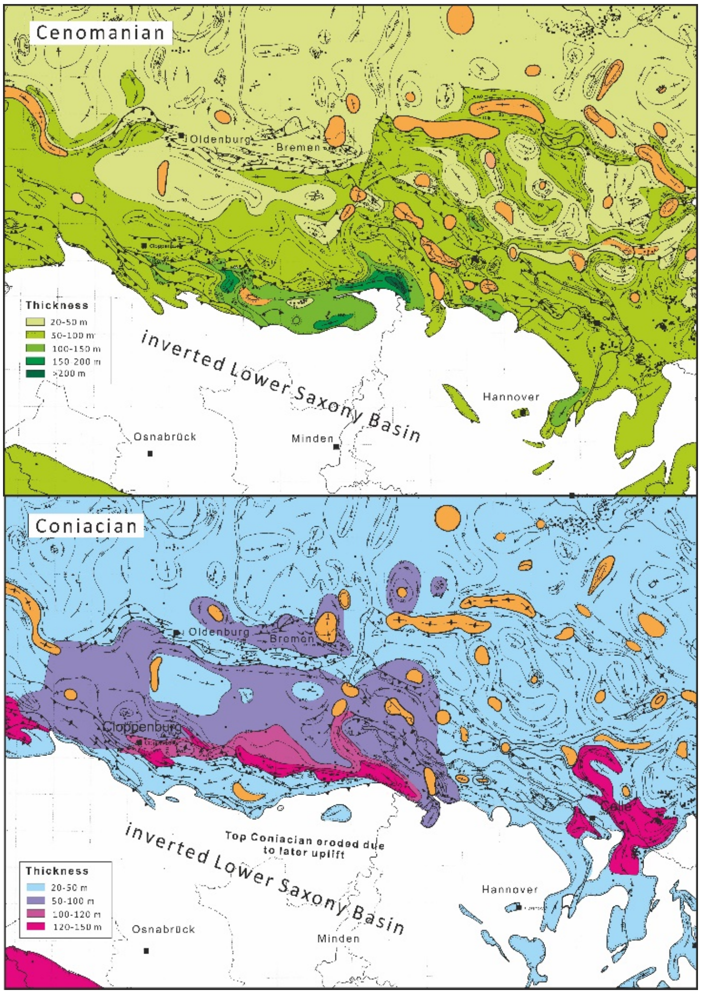

Fig. 3: Detailed thickness maps of the northern margin of the inverted Lower Saxony Basin show that the syn-inversion Coniacian thickness distribution of the marginal trough is already developed during the Cenomanian (modified from a thickness map of Baldschuhn et al., 2001). The shift of the basin axis of the marginal trough to the north can be explained by the propagation of thrusting towards the basin. Additionally, salt migration in the surroundings of salt diapirs created local highs and related local depocentres.

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Pelagic deposits characterize the facies of the paired marginal troughs on both sides of the inverted Lower Saxony Basin. Chalk with upward increasing marl content prevails, while coarser-grained deposits are missing. Most of the marls probably derived from re-deposited Jurassic and Lower Cretaceous, as the basin fill of the Lower Saxony basin was primarily composed of limestones and claystones. A key observation is that already Turonian and Cenomanian deposits reflect the same basin centres as the Coniacian to Campanian succession although sedimentation rates remain low (Fig. 3).

4.3 Subhercynian basin

The Subhercynian Cretaceous Basin contains a more than 2000-2500 m thick succession of Late Cretaceous sediments, which form a symmetric trough in front of the overthrust northern margin of the Harz basement anticline (T. Voigt et al 2006). Thickness of deposits is highest close to the thrust front. Sedimentation starts above a regional unconformity, which was formed during the global Cenomanian sea-level rise. High sedimentation rates occur during the Coniacian to Santonian (T. Voigt et al., 2006), but first enhancement of thickness in the marginal trough in front of the Harz mountains is observed already in the Middle Turonian (Karpe, 1973, T. Voigt et al., 2006, Fig. 4). No boreholes reached the base of the Cenomanian in the central marginal trough; therefore, the isopach map only displays a decreasing thickness trend to the southeast, not influenced by the Harz Mountains (Fig. 4).

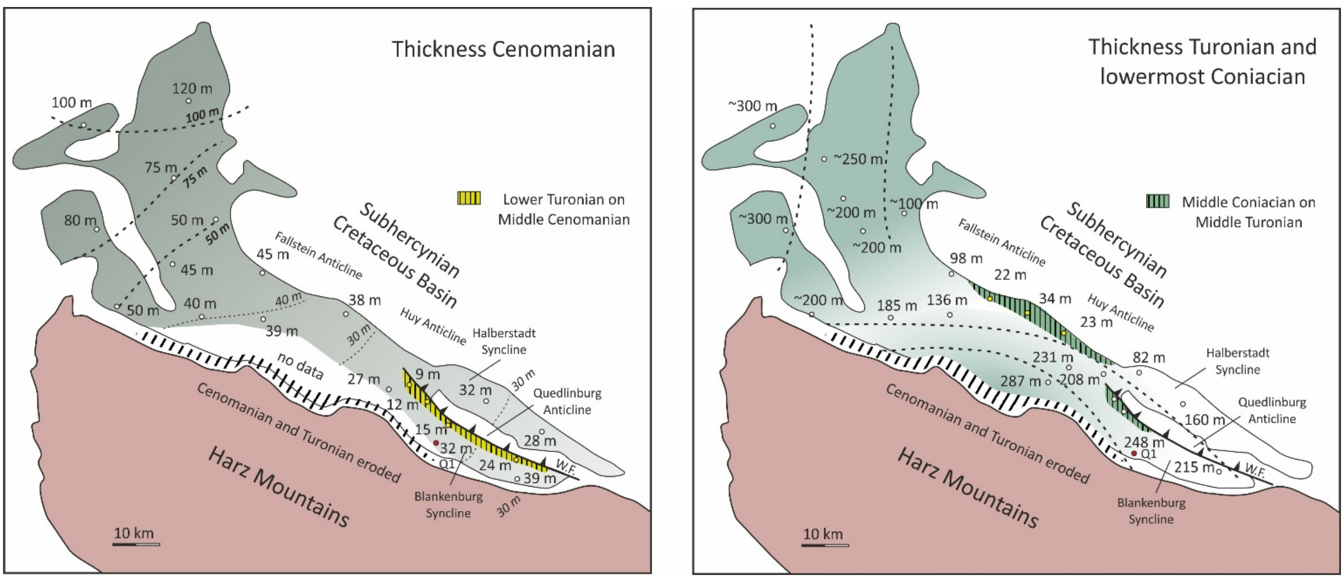

Fig. 4: First evidence of tectonic activity in the Subhercynian Cretaceous basin is given by strongly reduced thickness of Cenomanian deposits along the southern margin of the Quedlinburg anticline, the Westerhausen thrust (W.F.), which represents the master fault of an inverted early Cretaceous halfgraben. The Turonian reflects the formation of a symmetric marginal trough in front of the Harz uplift and the further uplift of the northern basin margin, accompanied by reduction of Turonian thickness and erosion of a major part of the succession before the Mid-Coniacian transgression.

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The Quedlinburg 1 (Q1) borehole, which is situated close to the thrust front but at the southeastern edge of the marginal trough, shows not an increased thickness of the Cenomanian (32 m) in comparison to the overall trend (Fig. 4).

The most striking evidence for an early (Cenomanian) start of compression in the Subhercynian Basin comes from an intra-basinal structure. The Quedlinburg anticline represents a former half graben, which formed during the Early Cretaceous. The about 40 km long master fault of the graben became re-activated as a thrust/reverse fault during late Cretaceous inversion. Along the fault, at the margin of the adjacent syncline, Lower Turonian limestones cover Middle Cenomanian marly deposits (Karpe, 1973). While the thickness of the Lower and Middle Cenomanian is similar to adjacent sections, the Upper Cenomanian is missing or condensed. This points to a late Cenomanian activity of the thrust fault. During the Turonian and early Coniacian, the structure remained active, but later erosion removed the evidence of tectonic activity close to the thrust. Nevertheless, at the western tip of the Quedlinburg anticline, the complete Turonian is preserved; hardgrounds and reduced thickness prove further activity of the thrust. Simultaneously, the Fallstein and Huy anticlines at the northern basin margin started to grow (Fig. 4, expressed in a significant unconformity of Middle Coniacian on Middle Turonian sediments and strongly reduced thickness in the Coniacian (Kölbel, 1944; T. Voigt et al., 2004).

4.4 Bohemian-Saxonian Cretaceous basin

The Bohemian-Saxonian Basin is bordered by a significant post-depositional thrust (Lausitz Thrust) underlying the basement uplift of the Lausitz-Krkonosze Massif. This thrust cuts through both coastal and hemipelagic deposits and clearly developed after the Coniacian (Fig. 5). Apatite fission track data, facies distribution and thickness data show, nevertheless, that the central segment of the fault already influenced sedimentation during Cenomanian and Turonian times (Seifert, 1955; T. Voigt, 2009; Lange et al., 2006; Danišík et al., 2010), probably by creating a fault-propagation fold (Voigt, 2009). Detritus derived from the exhumed Permian to Jurassic cover of the Neoproterozoic to early Paleozoic basement of the Lausitz High indicates the inversion of a Mesozoic graben structure (Voigt, 2009; Nadaskay, 2019). The preserved part of the basin fill ends in the Coniacian, with the exception of some deeply subsided remnants of Santonian sediments in the Ohře Graben, a segment of the European Cenozoic Rift System which was active in the Oligocene to Miocene. Fission track data point to a maximum uplift and exhumation between 85 and 75 Ma (Santonian, Campanian: Lange et al., 2006; Käßner et al., 2019), indicating that only parts of the basin fill are preserved. A following regional uplift, which ended about 40 Ma ago is shown by a regional unconformity at the base of the late Eocene? and early Oligocene deposits of the Eger graben, which cuts across both the Lausitz uplift and its marginal trough (e.g. Standke, 2008; Migon and Danišík, 2012).

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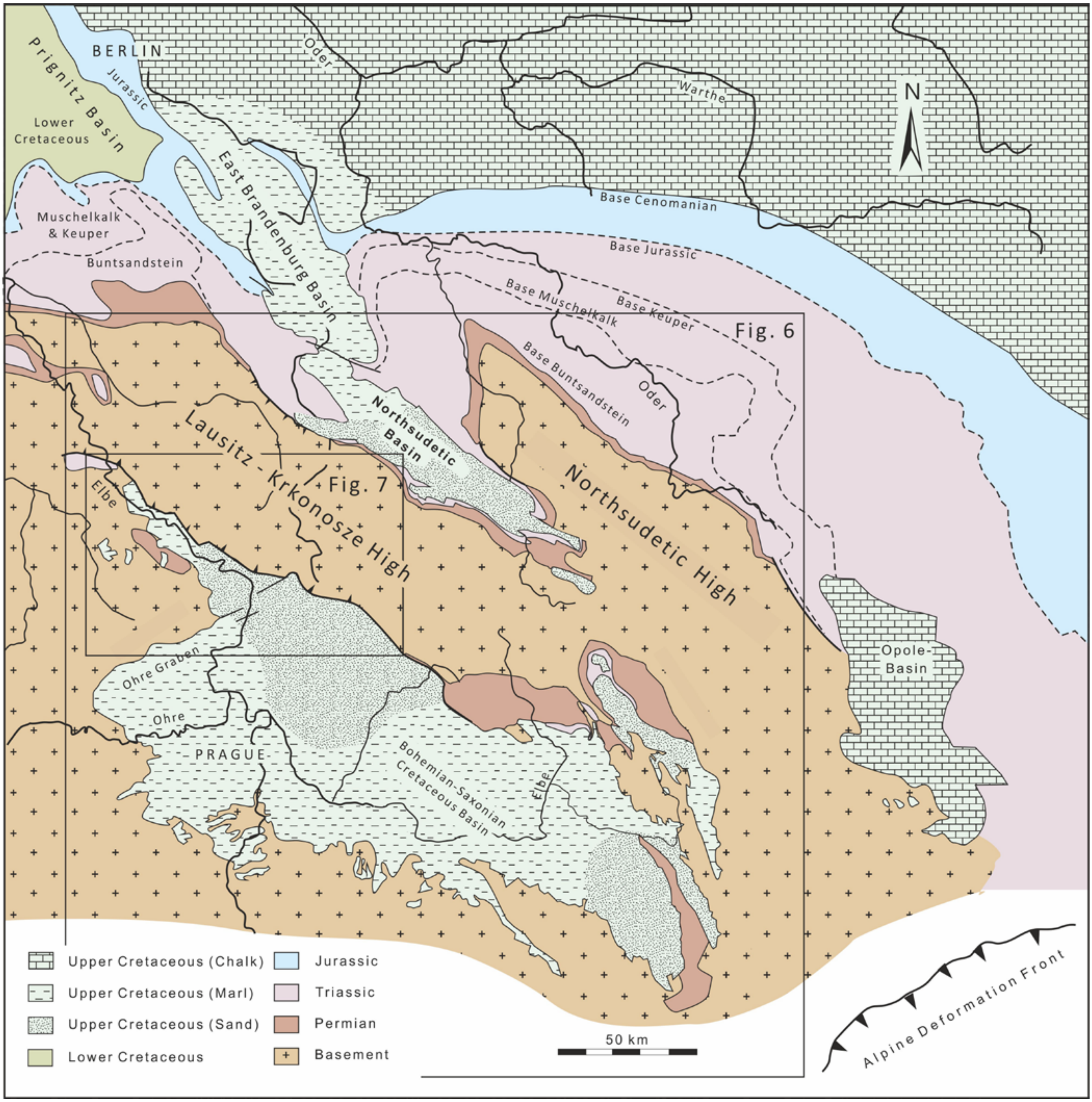

385 **Fig. 5: Late Cretaceous basins surrounding the Lausitz-Krkonosze High show a strong confinement of clastic deposits to the margins of the Lausitz-Krkonosze High. Regional Cenozoic uplift and denudation removed 1-2 km of sediments from both the Lausitz-Krkonosze Massif and the related Cretaceous basins. Therefore only remains of the primary basin fills were preserved, comprising Cenomanian to early Santonian deposits (modified from T. Voigt, 2009).**

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Sedimentation within this marginal trough started in the Cenomanian, simultaneously with the global sea-level rise. Deeply incised river valleys reflect a structured morphology with about 50 m of relief before the transgression (T. Voigt, 1996; Tonndorf, 2000; Uličný et al., 2009). The valley-fills were preserved by the successively rising sea-level from the Late Cenomanian to Lower Turonian (over 3 Ma). The pattern and evolution of these large paleodrainage systems were investigated by Uličný et al. (2009). Additionally, uranium exploration in the German part of the basin provided detailed data for the configuration of the paleovalleys in the northern part. More than 1000 uranium exploration wells determined the valley borders of the Niederschöna river, the Pirna river and the Hermsdorf river very precisely (Tonndorf, 2000, Fig. 6). Considering the whole basin, a central water shed divides a northern paleodrainage system, which was directed to the Boreal from a second system draining to the south, toward the Tethys (Uličný et al., 2009; Fig. 6). The most striking feature of the valleys in the northern paleodrainage system is their orientation, because they reflect an inclination of the valley floors to the north. The Lausitz Thrust cuts at least four large river valleys and three minor rivers discharging to the North. Uličný et al. (2009) assume a hypothetic trunk system running on the later exhumed Lausitz-Krkonosze High parallel to the Lausitz Thrust and thus collecting all the tributaries from the south. There is no evidence for a river mouth in Lower Cenomanian deposits in the northern part of the basin, where Lower and Middle Cenomanian is preserved so that also a direct connection to the Northsudetic basin can be assumed (Fig. 6).

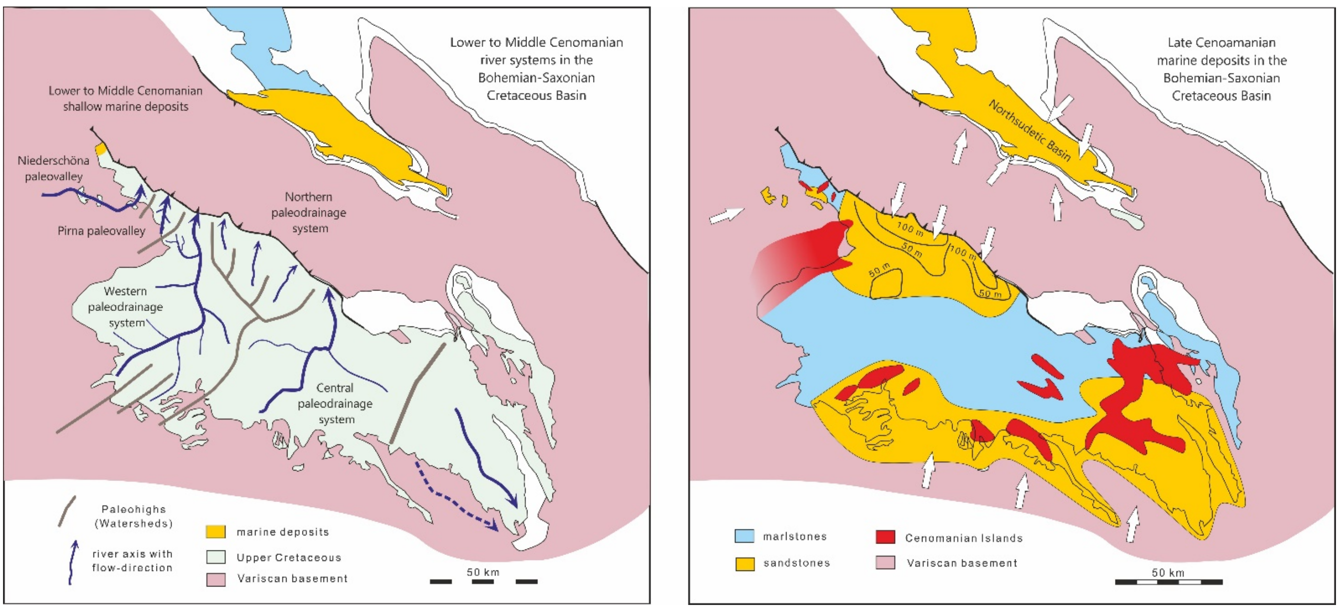

Fig. 6: Deposition within the Bohemian Cretaceous Basin started in the Lower to Middle Cenomanian with the filling of fluvial river valleys. Marine deposits were preserved at the northernmost edge of the basin. River orientation was directed to the North, towards the same area, which acted as a source area during late Cenomanian. The Late Cenomanian basin configuration reflects the onset of uplift of the Lausitz-Krkonosze High: the evolving marginal trough collects about 100 m of late Cenomanian sandstones compared to less than 30 m on the flooded shelf of the Bohemian platform. (paleodrainage pattern and Cenomanian facies and thickness after Uličný et al. 2009).

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The thickness of late Cenomanian deposits still partly reflects the morphology of the pre-transgressive landscape, because the river valleys were filled step by step by clastic deposits from the surrounding highs, while additionally an NW-SE elongated depositional centre developed outside the ancient valleys. There, marine late Cenomanian deposits reach a thickness of up to 110 m (Fig. 7). This thickness increase is observed even on the former drainage divide between the paleovalleys of the central and northern paleodrainage system (Fig. 7). In comparison to Turonian sedimentation rates (50 m/Ma), they are slightly lower (about 30 m/Ma), indicating a gradual start of basin subsidence). The hemipelagic facies on the northwestern edge of the basin shows also increased thickness compared to the Cenomanian of the Bohemian platform outside the marginal trough. The Upper Cenomanian of the Gröbern borehole reaches sedimentation rates in the order of 35-40 m/Ma (S. Voigt et al., 2006) compared to 5-15 m away from the basin axis on both the Bohemian platform and in the western Saxonian part of the basin and thus indicates the extension of the marginal trough further to the northwest.

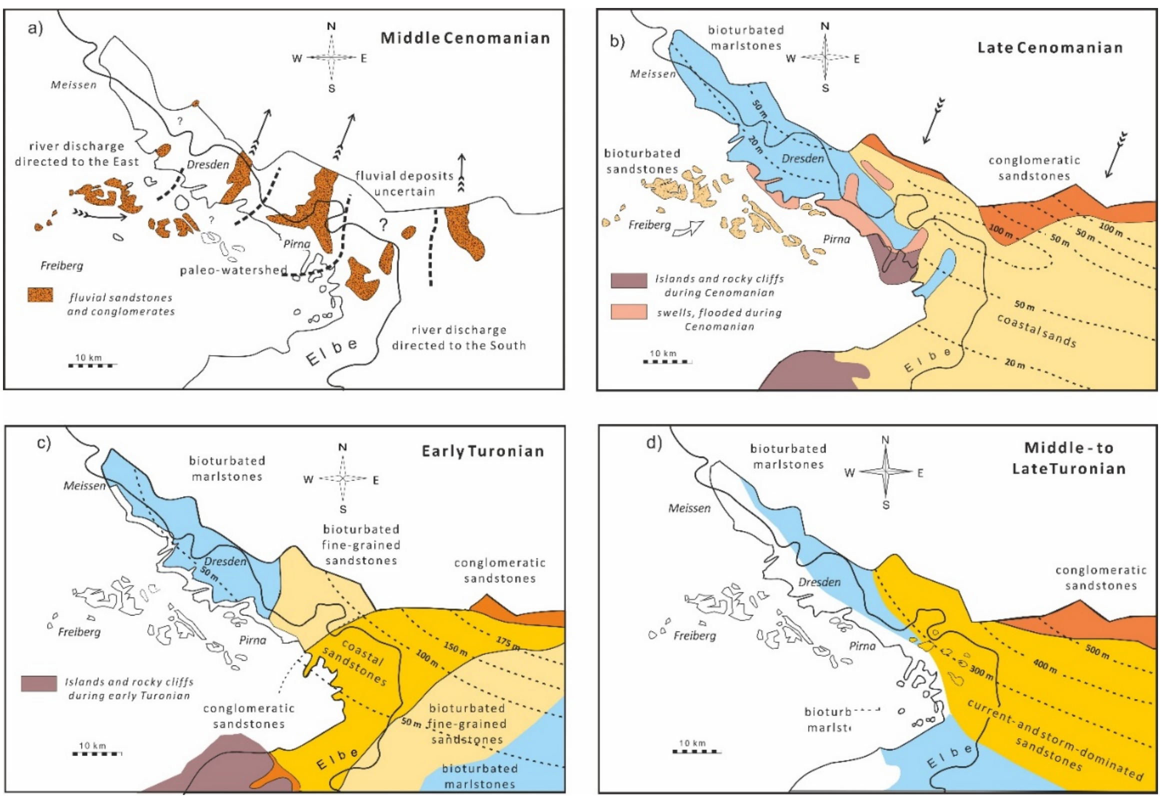

Fig. 7: Detailed facies maps of the Saxonian part of the Bohemian Cretaceous Basin. Early to Middle Cenomanian rivers discharge to the north. Distribution of sandstones in the Cenomanian and Early Turonian reflects a marginal trough in front of the uprising high, and thus the complete re-organisation of the basin configuration. Coastal sandstones of the Middle and Upper Turonian mark the northwestern edge of the Lausitz-Krkonosze High. The late Cretaceous to Paleogene Lausitz Thrust cuts through the basin margin and distal deposits. The data of the isopach maps are derived from numerous boreholes and surface mapping.

The facies distribution clearly indicates a major source area in the northeast, reflected by sandstones and conglomerates close to the northeastern basin margin (Fig. 7) and thus in opposite direction than the drainage direction during Lower and Middle Cenomanian. This basin axis is nearly identical with the marginal through which started to develop slowly during the Turonian and Lower Coniacian (Fig. 7) but consisted of several subbasins (Uličný, 2001). A possible explanation is a separate evolution of several small uplifts which later unified to one source area and the integration of the separated depocenters into one marginal trough. Alternatively, the oblique convergence phase which was observed by Navabpour et al. (2017) in small-scale structures and which predates the frontal thrusting could have induced the subsidence of oblique en échelon subbasins.

Together with the significant change of the basin floor morphology, this change during the Late Cenomanian indicates a complete reorganisation not only of the depositional system but also of the stress field within the basin. Regardless of whether the hypothetic trunk system of Uličný et al. (2009) existed, the appearance of a large source area in a direction downstream of the former drainage proves the uplift of a previous topographic low.

5 The dusk of Late Cretaceous basin inversion

The end of basin inversion/basin uplift in Central Europe is even more difficult to define than the beginning, because the region was affected by large scale regional uplift which continued until the Paleogene, and even longer south of the inverted Lower Saxony basin and uplifted Harz mountains (von Eynatten et al., this issue). Due to this event, some thermochronological data show a continuation of uplift up to 60 or even 50 Ma (e.g., summarized in von Eynatten et al., 2019). The end of the activity of a single structure can be shown if the structural configuration of highs and lows changes; new depositional centres evolve or formerly active structures and folds are covered by younger sedimentary units. If no deposition occurred during reconfiguration, because the whole region was above base level or the area was affected by later uplift, the recognition of a new stress field and differentiation between regional uplift and inversion remains ambiguous. Only in the subsurface of the deeply subsided Central European Basin, a complete succession of syn- and post-inversion deposits is preserved in the inverted Danish Trough. At the border of the inverted Danish basin, a sudden shift of the basin axis and therefore the end of basin inversion occurs still in the Maastrichtian. The Polish part shows a differentiated evolution: Deformation seems to continue until the Paleocene on the northern side of the Mid-Polish swell (Krzywiec, 2006), while the southern side experienced regional uplift, expressed by a marked unconformity across the marginal trough and the swell below an Eocene succession.

A similar situation is observed in central Germany, where Eocene deposits cover large areas of the structures resulting from Late Cretaceous inversion; including most of the basement uplifts. Only a few places allow the recognition of the change of basin configuration. In general, the youngest deposits preserved within the marginal troughs are of Lower Campanian age

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(Münsterland Basin, Subhercynian Basin, Northsudetic Basin, South Oldenburg basin). In all these basins, thermochronological data, erosional unconformities and composition of the basin fill prove a younger uplift, which involved both the source area and the adjacent marginal trough. This unconformity is partly still visible in the recent morphology. In the Krkonosze mountains, in the Erzgebirge and in the Lausitz, peneplains of Late Cretaceous to Paleogene age are still preserved (Migon and Danisik, 2012).

The time gap to younger deposits above those unconformities spans mostly more than 30 Ma, due to the absence of Late Campanian to Paleocene deposits. This is partly caused by a significant sea-level fall, which occurred during this period (Haq, 2014), but is mainly generated by regional uplift of those structures. Maastrichtian and Paleocene deposits are therefore rarely preserved in Central Europe and occur only at the margins of active diapirs and in a few narrow basins which do not reveal the configuration of former, late Cretaceous marginal troughs. These remains reflect nevertheless an extended depositional system reflecting the facies belts of a shallow shelf from continental to shallow marine environments, which grade into the hemipelagic and pelagic chalk environments of the central basins (Diener, 1968; Voigt, 2008). To better constrain the timing of the formation of this significant unconformity, we consider the examples of the Altmark basin and the inverted Lower Saxony basin with the unconformably overlying Late Campanian deposits of the Damme syncline, and the Prignitz-Lausitz High, which represents the less inverted prolongation of the Lausitz-Krkonosze High (Fig. 1). The Subhercynian basin and the Harz Mountains are taken as an example of an inversion structure with poorly constrained end of contraction and is therefore only briefly discussed.

5.1 Dusk of Cretaceous and dawn of Paleogene inversion in the Altmark basin

The Altmark basin is an elongated, about 60 km long and only 15 km wide marginal trough (Fig. 8), which formed north of the uplifted Gardelegen Block above a salt detachment linked to the Gardelegen basement fault (Schulze 1964, Kossow 2001, Malz et al. 2019). AFT ages from the Permian sandstones of the Flechtingen High, a part of the exhumed basement of the Gardelegen Block, suggest rapid cooling around 70 Ma (Fischer et al. 2012), confirming the overall pattern of Late Cretaceous syntectonic basin formation in Central Europe. The thermochronological age nevertheless is not in good agreement with the accompanying marginal trough north of the Gardelegen thrust, which preserves a syncline filled by a more than 700 m thick succession of syn-inversion deposits very similar to those of the Subhercynian Basin, indicating main inversion between 85 and 75 Ma. A late anticline divides the basin into two parts. Increased subsidence in comparison to neighbouring basins begins slowly in the Turonian and reaches its maximum during the Coniacian to Lower Campanian. The youngest preserved deposits are of early Campanian age in the central marginal trough and reach at least 450 m thickness. Close to the Gardelegen fault, Santonian sediments contain conglomerates and sands derived from the exhumed Mesozoic cover of the Gardelegen Block (Schulze, 1964). This indicates that the uplift of the Flechtingen High, which is the central part of the Gardelegen Block and was thrust onto Mesozoic deposits along the Haldensleben reverse fault, postdates the exhumation of the greater structure which demonstrably acted as a source area in the Santonian (85-82 Ma). To obtain a clear, well-constrained exhumation age, the uplift of the Flechtingen High relative to the Gardelegen Block must be in the order of an additional 2-4 km, because the partial annealing zone of the preceding uplift is not preserved.

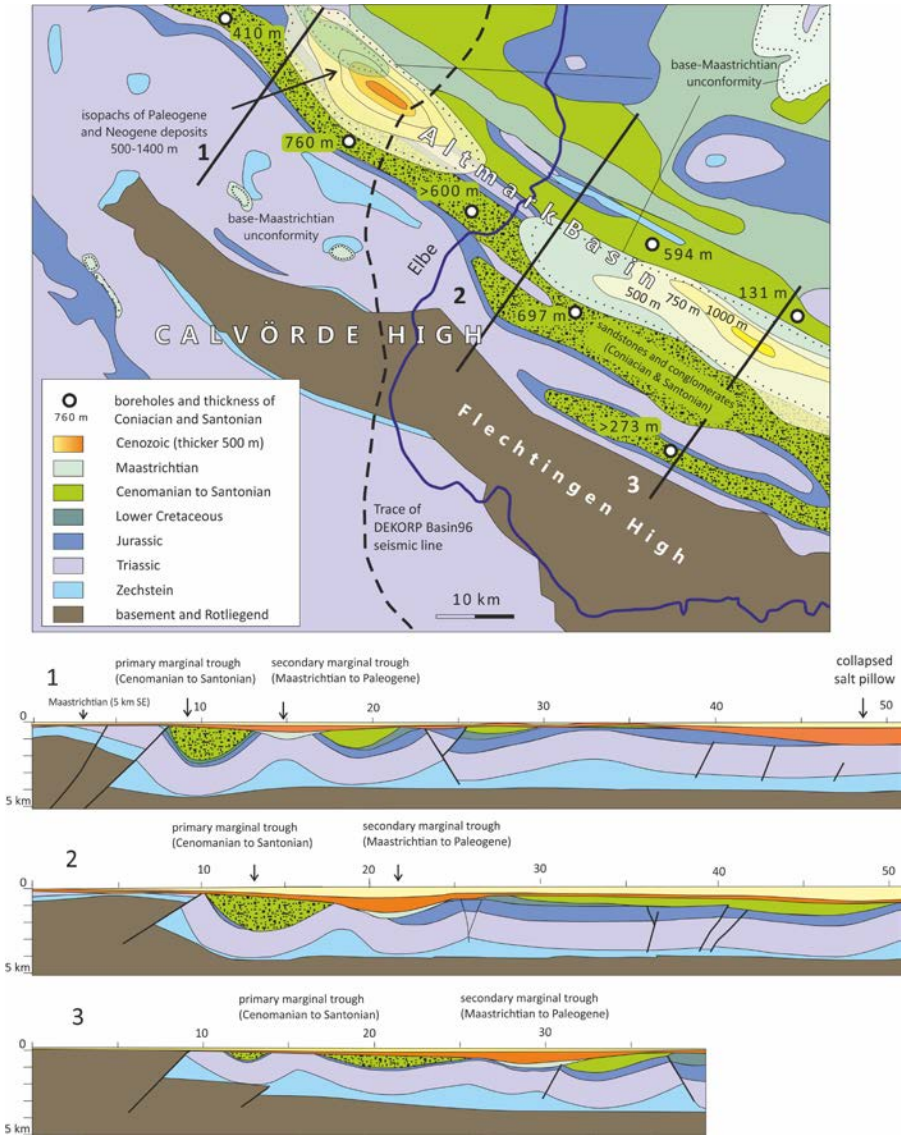

Fig. 8: The Altmark basin represents a narrow marginal trough north of the uplifted Flechtingen High. Deposition within the basin, which is dissected by a salt-intruded anticline, ended in the Lower Campanian. A shallower basin developed north of the Altmark basin above an unconformity cutting across the highs and basins since the Maastrichtian. Note that Paleogene deposits reflect the same depocentres as the Maastrichtian. The two shallow basins are therefore considered as secondary marginal troughs. The map was constructed on the base of Malz et al. (2019), Schulze (1964) and interpretation of borehole data. Location in Fig. 1.

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Zircons and volcanic quartz grains, resulting from the erosion of the Permian volcanic basement of the Flechtingen High appear late in the Maastrichtian sands of Waalbeck, south of the uplifted structure (Götze and Lewis, 1997). The provenance signal confirms the modelled AFT ages very precisely. These Maastrichtian shallow marine sands rest with an unconformity on Triassic deposits, again indicating the covering of an inverted structure, which was eroded and starts to subside again. The total uplift of the region since the mentioned 70 Ma (Maastrichtian) is less than 2 km, indicating that the situation after inversion tectonics is more or less preserved, supported by the fact that the whole area is nearly completely covered by Cenozoic deposits of Early Oligocene age (Blumenstengel and Krutzsch, 2008). Especially the base of the Rupelian transgression may be counted as a good marker horizon of the base-level. Elevation changes of this marker horizon indicate post-Rupelian tectonic movements, salt flow or both.

The regional seismic section DEKORP Basin 96 (e.g. DEKORP Basin Group, 1996; Kossow, 2001) and boreholes drilled for gas exploration allow to reconstruct the structural pattern. The succession of Cenomanian to Santonian deposits in the marginal trough, which is bounded by the Gardelegen thrust fault and in the north by a thin-skinned contractional salt anticline, which developed after deposition of the basin fill (Malz et al., 2019). Borehole stratigraphy and reflection patterns in the seismic section indicate a varying proportion of preserved strata (Schulze, 1964; Musstow, 1975). The shortened marginal trough was uplifted and eroded without further deformation. The flat erosion surface was tilted and can be traced beyond the extent of Cretaceous deposits onto the Gardelegen block (Malz et al. 2019). It is inclined to the north and forms the flank of a new depocenter, which developed north of the Cretaceous one and covers the partly eroded salt anticline. The sedimentary succession above this erosion surface shows a progressive onlap, starting with continental to shallow marine Maastrichtian sands (200-330 m), followed by a Paleocene succession (about 200 m) in the deepest parts. The Eocene reaches a maximum thickness of 400 m. Eocene marine deposits transgress even onto the Flechtingen High, demonstrating the transition from uplift to subsidence of the Gardelegen High.

This situation matches the evolution of primary and secondary marginal troughs described by Hansen and Nielsen (2000) and Nielsen et al (2002) from the inverted Danish basin. Even the timing and the described sudden shift of the basin axis may be similar, but the erosion of the Cretaceous marginal trough removed the youngest, Late Campanian part of the basin fill, so that the timing is more poorly constrained. Numerous drillings prove a saucer-shaped, symmetric structure (Fig. 8) of the secondary marginal trough. In comparison to the Late Cretaceous one, it is wider and shallower than the primary marginal trough.

A differing interpretation of the secondary marginal trough could be that collapse of the salt-cored anticline with extrusion and marine dissolution of salt caused the newly created depocenter. However, the extent, the smoothness and the undisturbed succession above the suggested solution surface disagree with this interpretation, because sediment deposition would stop further dissolution by sealing. Regardless of this interpretation, the base Maastrichtian unconformity is a prominent feature at many structures in the North German Basin, e.g. at the western Allertal fault zone (Lohr et al., 2007).

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5.2 Dusk of Cretaceous inversion in the Subhercynian Basin

The preserved sediment column of the Subhercynian Basin ends in the Lower Campanian although fission track data suggest continuous erosion of the Harz Mountains during the entire Campanian and even into the Paleogene (von Eynatten et al., 2019). Those younger deposits were eroded before Eocene times, because deposits of this age are preserved close to the front of the Harz and at the borders of some anticlines at the northern margin of the marginal trough. As the central Harz was covered by deposits of Oligocene age (König et al. 2011), inversion was in any case finished in the Eocene and only mild regional uplift affected the region afterwards (König et al 2011, von Eynatten et al 2019, Paul 2019).

Uplift continued at least until the Eocene but involved both the basement uplift and the surrounding basins. The time gap between the last preserved Lower Campanian inversion-related deposits (~82 Ma) and the Eocene/Oligocene deposits (~34 Ma) within the Subhercynian basin is in the order of 40 Ma and a definite timing of the basin configuration change therefore impossible. However, both the Harz Mountains and the foreland of the Harz show a significant peneplanation cutting across all lithologies both of the uplifted block and the basin (König et al. 2011), which formed between Lower Campanian and Oligocene. In the Harz Mountains, remains of Oligocene (Rupelium) deposits are preserved in the karst caves below such an extensive peneplain within Devonian limestones of the Elbingerode complex (Blumenstengel and Krutzsch 2008, König et al. 2011). They are about 140 m above the level of this transgressive surface in comparison to the same stratigraphic horizon south and east of the Harz mountains and prove therefore mild uplift which was not accompanied by major erosion since then.

While König et al. (2011) interpreted this elevation difference as an effect of renewed motion on the Harznordrand thrust, Paul (2019) argues that the observed offset is the result of foreland subsidence due to salt dissolution at depth.

5.3 The Damme syncline: The end of inversion in the Lower Saxony basin?

The Damme syncline is a remnant of (latest) Lower Campanian and Upper Campanian to Maastrichtian sediments of about 300 m thickness resting on the inverted Lower Saxony basin (Fig. 9). Inversion of the Lower Saxony basin was asymmetric, leading to the uplift of Triassic deposits and some small basement uplifts (Ibbenbüren High, Piesberg) to the surface in the south. In the north, decreasing uplift is observed, resulting in the preservation of parts of the Jurassic to Lower Cretaceous basin fill (Baldschuhn et al. 1991, Senglaub et al. 2005).

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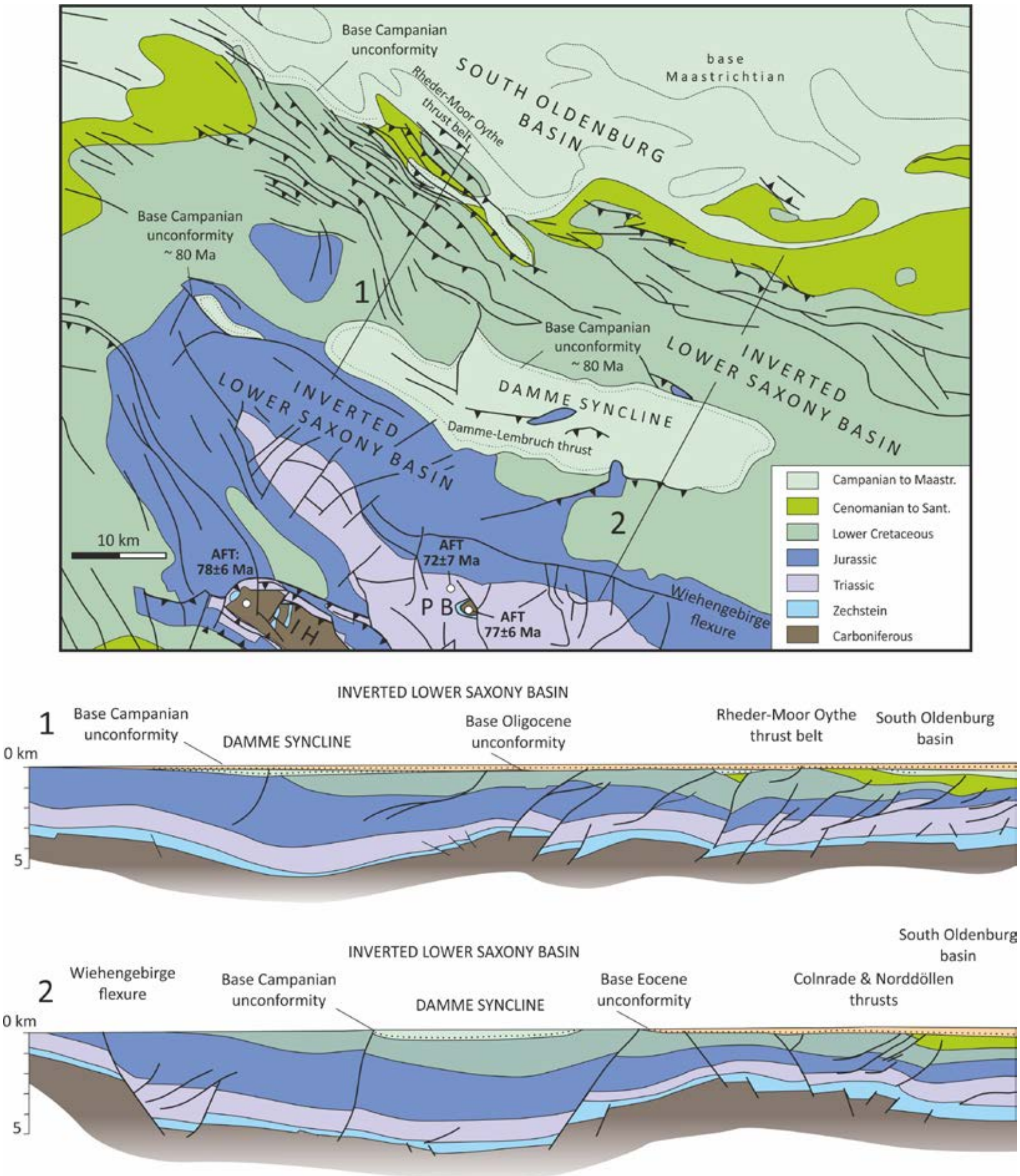

Fig. 9: The first deposits on top of the inverted Lower Saxony basin are represented by Middle to Late Campanian bioclastic limestones resting transgressively on a Jurassic and Lower Cretaceous peneplain. They predate the uplift of the strongly inverted southern part of the Lower Saxony basin and suggest that compression of the lithosphere ceased slowly. Deformation of the Damme syncline and thrusting within Campanian deposits in the north suggest progressive deformation. IH, PB: Ibbenbüren High and Piesberg basement uplifts.

Regional uplift prevented deposition of Paleogene sediments older than Oligocene both on the High and the adjacent Münsterland and south Oldenburg basins. Map and cross-sections are based on Kockel 1996; AFT-data are from Senglaub et al. 2005.

The syncline is gently folded and affected by a thrust (Damme-Lembruch thrust) of about 200 m displacement, indicating post-depositional compression (fig 9). The marine Campanian sediments unconformably covering deformed Jurassic and Lower Cretaceous strata largely postdate the marginal troughs flanking both sides of the inverted basin, which contain syntectonic basin fills of Cenomanian to Lower Campanian age. The Pompeckij Block on the north side preserved deposits of that age but with a chalk facies differing from the succession on top of the inverted basin. The transgressive succession of the Damme syncline consists of bioclastic nearshore limestones and reworked ironstones at the base (Mortimore et al. 1998) which are followed by sandy marls. In contrast, the nearest preserved Late Campanian units exhibit typical deep marine chalk facies, assumed to be deposited in water depths below 200 m and above 500 m (e.g. Hancock 1999). AFT cooling ages range between 72 +-7 to 78+-6 Ma in the hanging wall of the adjacent Wiehengebirgs flexure zone (Senglaub et al. 2005), in general spanning the same period as the sediments above the unconformity (Fig. 9). The southern Lower Saxony basin acted as a source area for the sediments of the Damme Syncline.

Eocene to Oligocene deposits cover both the inverted Lower Saxony basin and the South Oldenburg basin above a second unconformity and show that no major uplift affected this part of the inverted Lower Saxony basin since the Late Campanian. The weak folding of the first unconformity and the incipient Damme-Lembruch thrust prove deposition in the same tectonic regime as during deformation of the underlying inverted Lower Saxony basin. The preservation of these deformed late-inversion sediments indicate that no major erosion occurred since the Late Cretaceous.

5.4 End of inversion tectonics at the Lausitz-Krkonosze High

The inverted Lusatian block with the marginal troughs of the Bohemian Cretaceous Basin and the Northsudetic Basin show a similar picture. Investigations of the Late Cretaceous to Paleogene evolution by different authors are mainly based on thermochronology and thermal maturity of the hanging wall and the footwall block (Käßner et al. 2019, Lange et al. 2006, Danišík et al. 2010) and time constraints derived from geometrical relationships of strata, magmatic dykes and faults (Tietz et al. 2015, Coubal et al 2014). The sedimentary record of basin inversion ends with remains of Lower Santonian deposits, preserved in the central Ohře Graben or with Lower Campanian deposits, drilled in a syncline in the central North-Sudetic basin. A late, Paleogene, activity of a segment of the Lausitz Thrust (Pillnitz Thrust) was inferred by Käßner et al 2019. AFT cooling ages (84-70) show an accelerated uplift during the latest Cretaceous, but in the area of the Krkonosze and the Jizera mountains, even younger ages occur (40 Ma), providing evidence of Paleogene uplift. Renewed sedimentation started in the Oligocene with the formation of the Ohře Graben, which crosscuts the Lausitz Thrust and covers both the basin and the inverted Lusatian Massif (Coubal et al. 2015, Špičáková et al 2000). A later, mild reactivation of the Lausitz Thrust was reconstructed on the basis of offset Oligocene tuffs (30-27 Ma) by Tietz and Büchner (2015), but the orientation of the fault displacing the tuffs is oblique to the thrust. Under these preconditions, the end of the inversion of the Lausitz massif is poorly constrained to a period of at least 40 Ma (between Santonian/Lower Campanian to Oligocene), with the limitation that post-Cretaceous exhumation did not exceed 2 km (Kaeßner et al. 1999).

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Better timing constraints are found in the northwestern prolongation of the Krkonosze-Lausitz High, which is named the Prignitz-Lausitz High (Fig. 10). In this part, the Jurassic to Lower Cretaceous Basin shows only mild inversion of less than 1000 m. During Coniacian to Campanian uplift, the Prignitz High delivered clastic material into shallow marginal troughs north and south of the uplifting High. As the uplift rates were low, marginal parts were flooded by the sea during transgressions as is shown by the preservation of marine sediments in peripheral sinks of major diapirs on the swell (Haller 1965, Musstow 1976). To the southeast, the broad uplift of the Prignitz-Lausitz High is bounded by the deeply subsided Altmark basin, which became predominantly filled with sands and marls from the narrow uplift of Flechtingen High. The facies distribution changed significantly at the transition Campanian to Maastrichtian.

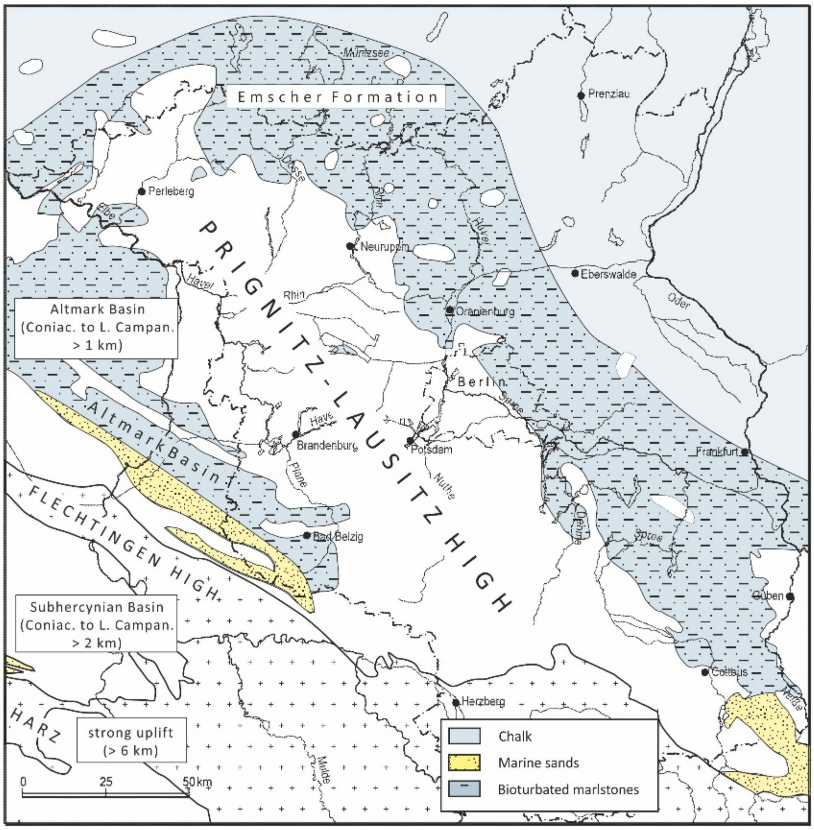

Fig. 10: The Prignitz-Lausitz High represents a gentle late Cretaceous inversion structure and represents the north-western prolongation of the Lausitz-Krkonosze High. During the main inversion phase, it was surrounded by a belt of marlstone, derived from the erosion of Jurassic to Lower Cretaceous claystones. Marine sands were restricted to the margins of the prominent inversion structures of the Flechtingen High, the Harz Mountains and the Lausitz-Krkonosze High, which brought Permian to Lower Cretaceous sandstones into the erosion level. Slightly modified from Voigt (2015).

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During the Maastrichtian, the Prignitz High was flooded and completely covered with sands of the Nennhausen Formation (Fig. 11). The facies belts show a pattern of extended shallow marine sands which give way to marlstones of the open shelf. They occur on top of the Prignitz-Lausitz High at the same elevation as on the former marginal troughs. Glauconitic sands of the Nennhausen Formation reach their highest thickness of 600-1000 m in the peripheral sinks of salt diapirs. Together with the thinner deposits they reflect one extended facies belt (Ahrens et al 1965, Voigt 2008). They are followed conformably by Paleocene marine deposits. At least in the area of the western Prignitz-Lausitz High, the end of inversion can be dated to have occurred in Late Campanian to early Maastrichtian time.

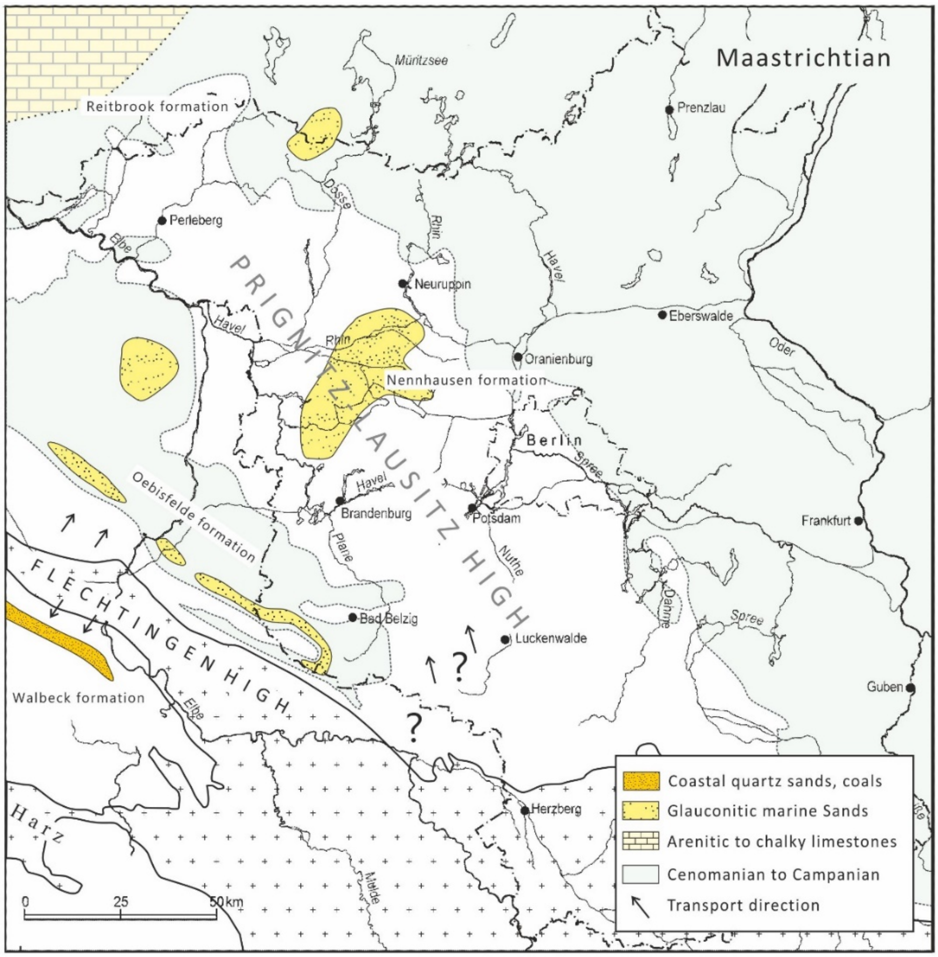

Fig. 11: The end of inversion of the Prignitz-Lausitz High is shown by its cover of Maastrichtian marine sands (Nennhausen Formation). Similar deposits extend across the Altmark basin (Oebisfelde Formation) and the northern Subhercynian Basin (Walbeck Formation), where they unconformably overlie the post-inversion structures. These deposits are mainly restricted to peripheral sinks of salt diapirs and the narrow secondary trough of the Flechtingen High, where they were protected from later erosion. The similar marine facies points to complete flooding of both former highs and basins. Slightly modified from Voigt (2015).

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6 Discussion

Basin re-organisation and formation of new depocentres that spatially coincide with those of the Coniacian and Santonian, occurred in the majority of the investigated basins already in the Late Cenomanian (96 Ma). This is about five Ma earlier than previously deduced from AFT ages (Fig. 12). In particular, basins clearly related to the compressive reactivation of normal faults seem to be affected, like the Osning thrust and the Rheder Moor-Oythe thrust bounding the Lower Saxony Basin; and the central part of the Lausitz thrust, where slices of Jurassic deposits prove the reactivation of a major normal fault (Voigt, 2009). The same is true for the re-activated normal fault of the Quedlinburg Graben within the Subhercynian Cretaceous Basin (Westerhausen thrust), where the activity during the Late Cenomanian is proven by an erosional unconformity.

The oldest AFT data point to a fast passing of the partial annealing zone (PAZ) around 89-90 Ma (boundary Turonian/Coniacian) of the basement rocks presently exposed at the surface. Exactly when these rocks entered the PAZ is probably not resolvable by thermochronology, whereas even small increments (tens of meters) of surface uplift can modify patterns of deposition and erosion. The most precise timing constraints come from the Bohemian Cretaceous basin, where the fluvial depositional system of the Lower to Middle Cenomanian is replaced by a marine environment in coincidence with a complete reversal of the main sediment input direction. North- and southeast-directed transport directions of the old northward inclining relief still prevail during the Cenomanian but a source area in the northeast provides the majority of clastic material. The sudden appearance of a northern source area which coincides with the uplift that controls the inversion during the whole Late Cretaceous (e.g. Tröger, 1965; Skoček and Valečka, 1983; Uličný et al., 2009; Voigt, 2009), is a strong hint at inversion starting already at the Middle to Late Cenomanian transition.

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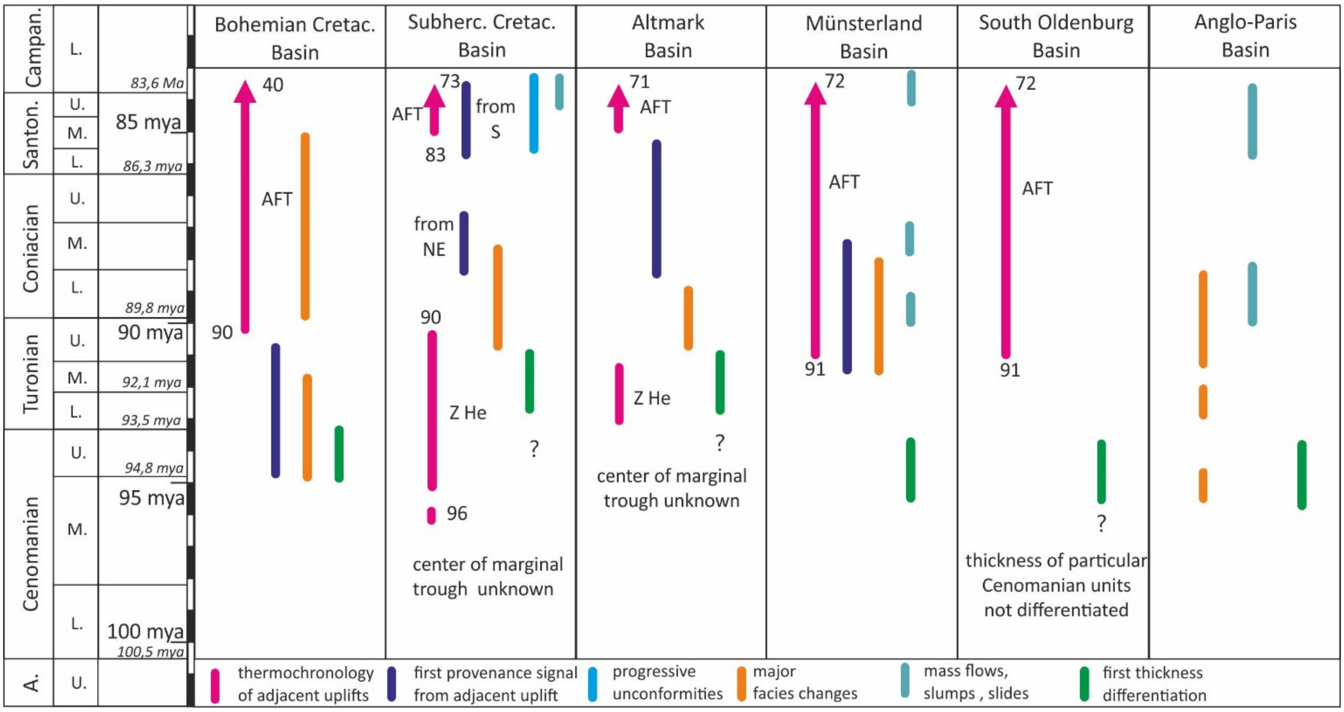

Fig. 12: Compilation of time constrains fixing the start of inversion. Varying methods, basing on thermochronology, sediment redistribution, provenance, progressive unconformities and thickness differentiation show a significant disparity in timing between Cenomanian and Coniacian/Santonian. First evidence for changes in basin configuration occurred already during Middle to Late Cenomanian across western and central Europe.

The positions of the marginal troughs and the principal regions of maximum thickness do not change from Cenomanian to Turonian (Malkovsky, 1987). Uličný et al (2009) observed the migration of the sandy margin facies to the southeast from Turonian to Coniacian. They interpreted this pattern to reflect a lateral shift of the source area and concluded that the basin evolved in a NW-SE-striking transtensional strike-slip system. However, the northwestern margin of shoreface sands in the Saxonian part of the Cretaceous basin migrates about 15 km in the opposite, northwest direction during the same time. This pattern better matches an increase in size of the uplift than unidirectional displacement of the source area. The increased sediment input which starts in middle- to late Turonian times seems to cause a general extension of the sandy facies belt.

The increasing sedimentation rates from about 50 m/Ma in the late Cenomanian to 75-110 m/Ma in the early to middle Turonian and >300 m/Ma in the late Turonian and lower Coniacian point to accelerated subsidence of the marginal trough accompanied by clastic input which was mostly accommodated by the subsidence close to the uplifting margin. The absence of early to middle Cenomanian fluvial deposits in the region of highest thickness of Upper Cretaceous sediments and their independent thickness trends prove basin re-organisation and the post-middle Cenomanian formation of the marginal trough.

The marginal troughs of the inverted Lower Saxony Basin show only a thickening signal in the Cenomanian close to the uplifting margins but no facies change, because the uplifted area did not rise above sea level at this time. As in the Bohemian Cretaceous basin, inversion starts slowly and accelerates in particular during late Turonian and Coniacian. It is still unclear whether this trend characterizes the entire Cenomanian or only the upper parts of the section, as documented in the well-investigated Bohemian Cretaceous basin.

The Subhercynian Basin clearly shows the influence on sedimentation of a basin-internal structure in the Cenomanian. Cenomanian thickness on the southern flank of the Quedlinburg anticline is strongly reduced and even an erosional unconformity developed above Middle Cenomanian and below Lower Turonian deposits. The area of this anomaly coincides with the northeast-dipping bounding fault of a half graben filled with Hauterivian and Barremian coastal sandstones. We interpret the reduced thickness to indicate the start of inversion. Instead, the uplift could be related to salt tectonics, but the northern flank of the narrow anticline shows no evidence for similar reduction of Cenomanian thickness. Surprisingly, no clear evidence for Cenomanian tectonic activity has yet been observed close to the Harz uplift as the prominent structure. The widespread erosion below the Middle Santonian unconformity at the northern margin of the Harz erased all documentation of earlier fault activity. A possible thickening of Cenomanian deposits in the axis of the marginal trough is not proven, because no borehole reached the Cenomanian in the deeply subsided basin part close to the Harz uplift margin.

In contrast to these examples of early activities in basin evolution, the deep marginal troughs of the Altmark Basin show no significant variations in Cenomanian thickness. Neither the Harznordrand Thrust nor the Gardelegen thrust are proven to have formed from inherited normal faults (e.g. Voigt et al., 2009; Malz et al., 2020). They may have nucleated later in stronger lithosphere in comparison to the easily re-activated normal faults of the Lower Saxony Basin and the Quedlinburg anticline in the Subhercynian basin. Nevertheless, ZHe-data from the Harz (von Eynatten et al., 2019) show a clear cooling signal between 96 and 90 Ma, much earlier than previously recorded by AFT ages and in agreement with a Cenomanian - early Turonian start of uplift. No drilling or high-resolution seismic data exist for the deeply subsided centers of the Subhercynian and Altmark basins so that the question of an early start of inversion-related subsidence in these basins remains open.

The early start of inversion tectonics in the Cenomanian is observed in a variety of basins of central Europe, but were not reported from the inverted Polish and Danish basins until recent. Here, a detailed areal evaluation of Cenomanian thicknesses is necessary to date the start of inversion tectonics more precisely or to conclude that only the southern basin exhibits the early start because deformation propagated northward. As the Anglo-Paris Basin shows clear evidence of Cenomanian contractional fault activity (Mortimore and Pomerol 1991), we assume a synchronous onset of inversion in Central and Western Europe.

The incomplete preservation of marginal troughs in Central Europe allows only in a few cases to constrain the end of inversion more tightly than Late Cretaceous to Paleogene (fig 13). The regional uplift after basin inversion, which affected both the highs and the adjacent syn-inversion basins, obscured the signal of the change in basin configuration, which occurs

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with the end of compression. A major erosion surface bevelled both highs and basins. This unconformity appears very flat outside the areas of salt migration, but marine deposits of varying age, reaching from the Maastrichtian to Late Eocene - early Oligocene, cover it and indicate a long-lasting evolution of a low-relief paleosurface. This surface is very gently inclined towards the northwest and the deposits covering the unconformity become continuously younger to the south.

Primary and secondary marginal troughs are therefore only preserved if an area remained below base-level. The Harz Mountains with the Subhercynian Basin and the Lausitz-Krkonosze High with its paired marginal troughs show a time gap of 30-40 Ma between the youngest, Santonian/Lower Campanian sediments of the marginal trough and the Eocene post-inversion unconformity (Fig. 13).

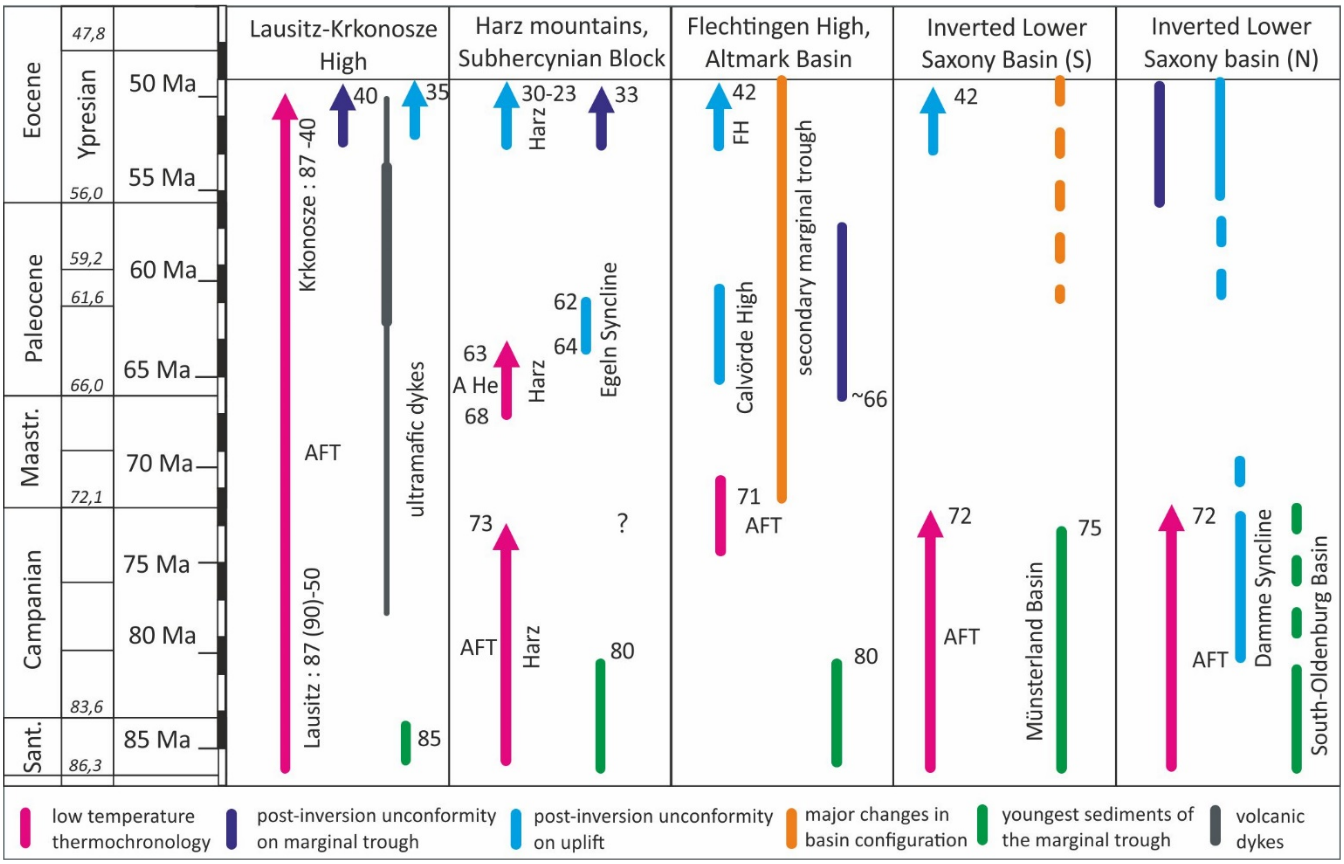

Fig. 13: Timing of the end of inversion tectonics in Central Europe is difficult to fix due to a post-inversion uplift of most structures, involving both highs and related marginal troughs. While the covering of the Lower Saxony basin with post-inversion deposits occurred already in the Campanian, a significant shift in the evolution of the marginal troughs occurred either in the Maastrichtian or later.

The marginal troughs situated more to the north preserved better evidence of the end of inversion due to the local preservation of Maastrichtian and Paleocene deposits. These indicate a shift of the basin axis. The ages of the overlying

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sequences differ between the Prignitz-Lausitz High (Lower Maastrichtian) and the inverted Lower Saxony Basin (uppermost Lower Campanian) by several million years (7 Ma to 9 Ma). In both cases, marine deposits were preserved. Although Maastrichtian and Paleocene units are often only conserved in isolated marginal troughs of salt diapirs, marine deposits prove that also the surrounding highs had been close to sea-level and were conquered by the advancing sea. In the case of the Damme syncline on top of the inverted Lower Saxony Basin, the influence of a salt diapir can be excluded.

The diachronity between the basal deposits overlying the unconformity on the Prignitz High, in the Damme syncline and in the Altmark basin probably suggest slowly waning uplift and transition to subsidence, in contrast to the sudden shift of the axis of marginal troughs in the Danish basin at the Cretaceous–Paleogene boundary. This can probably be explained by continuing salt migration: In addition to tectonics, the preservation of sediments is controlled by the re-distribution and subsolution of thick Permian (Zechstein) salt, which is or was present in the majority of inversion-related basins in Central Europe, except the Bohemian Cretaceous Basin and the Regensburg-Straubing Basin.

In comparison to the well-investigated primary and secondary marginal troughs of the Danish basin, the timing remains poorly constrained. The proposed secondary marginal trough of the Altmark basin is closer to the previously active thrust and deeper than in the Danish example. The start of subsidence in the Maastrichtian which lasted until the late Eocene could also be the result of salt migration into the adjacent pillow north of the basin due to doming of the southern region. Despite this possible alternative interpretation of the secondary marginal trough, the unconformity cuts both the Altmark Basin and the uplifts formed between the Lower Campanian and Maastrichtian and thus marks the end of subsidence in the marginal trough.

The onset of basin inversion in Central Europe coincides with major changes in relative plate motion between Africa, Iberia and Eurasia (Kley and Voigt, 2008; Rosenbaum et al., 2002; Seton et al., 2012) which in turn coincide with a mid-Cretaceous global plate reorganization event (Scotese et al., 1988; Veevers, 2000; Matthews et al., 2012). The exact age of this event is difficult to pinpoint because it occurred in the magnetic quiet period or Cretaceous Normal Superchron (CNS) between chrons M0 and 34 (120.4 – 83.5 Ma). Matthews et al. (2012) used interpolation between magnetic anomalies to suggest that the reorganization took place between 105-100 Ma, in Albian to Cenomanian time. The start of deformation in Central Europe is contemporaneous with the global Late Cenomanian sea-level rise, a consequence of the rapid formation of new oceanic crust. This supports a causal link between global plate reorganization and intraplate deformation. The termination of inversion was either due to a Paleocene drop in Africa-Iberia-Europe plate convergence (found by Rosenbaum et al. (2002), but not Vissers and Meijer (2012)) or to mechanical weakening of the Iberia-Europe plate boundary caused by incorporation of continental crust (Dielforder et al., 2019).

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7 Conclusions

The start of inversion in many Late Cretaceous basins of central Europe can be dated about 5 Ma earlier than hitherto assumed based on detailed analysis of new depocenters forming close to the inverting structures. The first signals of inversion are weak, because the sedimentation rates are only in the order of 20% of the maximum sedimentation rates attained during the Coniacian and Santonian. This slow increase in sedimentation rate probably also points to a slow development of the tectonic loads inducing the subsidence. There is no evidence for a different compression direction during the Cenomanian in the Central European basin, except in the Saxonian part of the Bohemian Cretaceous basin where a secondary oblique axis of one depocenter develops and southeast-trending highs appear, not obviously correlated to the previous paleodrainage system, which reflects in general a dissected peneplain dipping to the north.

A regional unconformity between the inversion structures (highs and basins) and a covering sequence developed in the period between the Lower Campanian and Maastrichtian. A similar surface was formed again through landscape evolution during the Maastrichtian and Paleocene with neither deposition nor major erosion. Only in the surroundings of active diapirs and in a few secondary marginal troughs, relict sediments of coastal plains and shallow marine environments witness the existence of an extended marine cover on this surface. The covering of inverted structures by deposits varying in age from Campanian to Maastrichtian indicates rather a gradual deceleration than a sudden end of compression and uplift. Large-scale salt migration is probably the main reason for preservation of a complete marine succession from Maastrichtian to Paleocene on top of the Prignitz-Lausitz High which gives strong evidence that the inversion in this part of the basin was finished before the late Maastrichtian. The end of basin inversion should be better constrained in the areas deep in the subsurface (northern slope of the inverted Mid-Polish trough, Grimmen swell, Danish trough), because the conservation of the unconformity is much better there than in the south, where several transgressions wore down the original surface.

The start of deformation in Central Europe was probably caused by a global plate reorganization event. This event induced both the changes in plate kinematics and the coeval global Late Cenomanian sea-level rise due to a peak in the production of new ocean floor. The reorganization event dated to Albian to Cenomanian time between 105-100 Ma would be the earliest possible age for the onset of contraction and basin inversion.

Author contributions. TV: general idea, conceptualization, borehole data acquisition, thickness maps and interpretations, drawing of Figs. 3-13 , writing of the original draft, funding acquisition (EU, DFG). JK: conceptualization, drawing of Fig. 1, discussion, additional and critical information, validation, critical review in the pre-publication stage, funding acquisition (DFG). SV: ideas on the general problem of Late Cretaceous basin inversion in Central Europe, data acquisition of Cenomanian thicknesses in the north, compilation of thickness data, drawing of Fig. 2., discussion and validation, revision, critical review in the pre-publication stage.

Competing interests. The authors declare that they have no conflict of interest.

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