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Synsedimentary Deformational Structures Caused by Tectonics and Seismic Events – Examples from the Cambrian of Sweden, Permian and Cenozoic of Germany

Herbert Scholz[1], Dorothea Frieling[2] and Michaela Aehnelt[3]
[1]Chair of Engineering Geology,
Technische Universität München, München
[2]Department for Earth Sciences, Ludwig-Maximilians University of Munich,
Section Paleontology, München
[3]Institute for Earth Scienes, Friedrich Schiller Universityt Jena, Jena
Germany

1. Introduction

Most sedimentary structures occurring within siliciclastic successions are doubtless caused by a variety of depositional processes. Among thousands of these sedimentary structures, a small number can be identified as structures which are not of primary origin. Among these, many are structures generated due to a postdepositional deformation of the primary bedding planes within the sediment. Examples of such deformational structures are mostly synsedimentary folds, ball-and-pillow structures or clastic dykes. Some of these structures have been obviously generated by the liquefaction of the soft, not yet lithified sediment. The existence of unstable layers or density inversions within the succession may promote such processes. However, in most cases these deformational sedimentary structures are obviously caused by heavy earth quakes, generated at actively moving faults. Strong seismic shock waves promote a break-down of the grain structure of sensitive sand layers leading to a short termed liquefaction of these layers and causing the deformational structures mentioned above. Hence, many of these structures are lastly products of tectonic events near or at the active rims of the sedimentary basins. Examples of these tectonically triggered deformational sedimentary structures can be found within sedimentary successions throughout earth's history and are described e.g. from Proterozoic and Plaeozoic of India (Bhattacharya & Badyopathyay 1998) and Canada (Pollock & Williams), from the Cretaceous of Mexico (Blanc 1998) and Brasilia (Rosetti & Goes 2000, Rosetti & Santos 2003) or the Quaternary of Switzerland (Becker et al. 2002), Spain (Gilbert & al. 2005) or Kyrgystan (Bowman & al. 2004). Within the paper presented here examples are given from Lower Cambrian shallow marine sandstones (Vik Sandstone) in eastern Scania (Skåne, South Sweden), from terrestrial Lower Permian Rotliegend sandstone successions in Hesse (Central Germany), and from Paleogene and Neogene marine as well as terrestrial sediments within the Molasse basin in southwestern Bavaria (South Germany).

2. Ball-and-pillow structures, synsedimentary folds and clastic dykes within Oligocene and Miocene Molasse successions in the Allgäu (southwestern Bavaria)

The western part of the Bavarian Alpine foreland between River Danube in the N and the Alps in the S is made up of Oligocene and Miocene clays, sandstones and conglomerates, containing detritus from the slowly growing Alpine mountains in the S. These clastic marine and terrestrial Molasse sediments were deposited upon a steadily downsinking substrate, consisting of a thick blanket of Mesozoic strata overlying the crystalline basement of the Variscides. In the southern part of this Molasse basin the sediments attain a thickness of several kilometres, towards the N the sedimentary succession thins out and is completely eroded in places N of the Danube valley. Due to the compression caused by the prograding Alpine orogenic front the sedimentary filling of the southern part of the Molasse basin has been tectonically deformed. This Folded or Subalpine Molasse forms a 10 to 25 km wide belt in front of the Alpine nappes, characterized by widely deformed synclines and listric thrusts. Separated by a steeply inclined major fault N of the Subalpine Molasse („Südrandstörung") the Alpine foreland is made up of the flat lying, Unfolded or Autochthonous Molasse. Only at their southern margin and immediately N of the „Südrandstörung" the beds are steeply inclined (Lemcke, 1988, Scholz, 2000).

The Molasse succession exhibits two transgressive-regressive cycles. In Upper Oligocene, lower Early, Middle and Upper Miocene fluvial sediments were deposited on wide and flat lowlands with neglible relief, formig the Lower (LoFM) and the Upper Freshwater Molasse (UpFM) respectively. In times of a relative sea level rise, brackish to marine conditions have been established, evidenced by the intercalation of the Lower Marine Molasse (loMM) in Early Oligocene and the Upper Marine Molasse (UpMM) in latest Early Miocene (Lemcke, 1988).

At several sites and within sand- and silt-dominated Molasse successions deformations of the primary sedimentary structures occur, which obviously are not linked with a postsedimentary (Neoalpine) tectonic deformation of the rocks (fig. 1 / 1-6). Since overlying and underlying horizons of the deformed beds are not deformed, they clearly are of synsedimentary origin. Large ball-and-pillow structures are well developed at many sites at the contact between marls and overlying sandstone successions, mainly within the LoMM (Bausteinschichten) and LoFM. There are also some examples of small and large-scale synsedimentary folds within LoFM and UpMM. The UpMM succession exposed in the Kesselbach gorge just at the German/Austrian border NE of Bregenz contains clastic dykes, cutting through heavily bioturbated shallow marine Molasse sandstones (Scholz & Frieling, 2006).

2.1 Large ball-and-pillow structures in the Lower Marine Molasse in the Mt. Grünten area (SW Bavaria)

At the northern slope of Mt. Grünten near Sonthofen above the village of Kranzegg and below Kammerkehr-Alpe, an at least 60 m thick sequence of the LoMM is exposed in a large sandstone quarry (Grüntensteinbruch, fig. 1 / 1, northing: 52.71050, easting: 35.98500). The succession consists mainly of sandstones, belonging to the about 60 m thick Middle Oligocene „Bausteinschichten", which is the youngest member of the Lower Marine Molasse (LoMM, Schwerd, 1978). The beds dip with 17 to 20° towards ESE and are part of the southernmost syncline of the Subalpine Molasse (Murnau syncline, Schwerd et al., 1996).
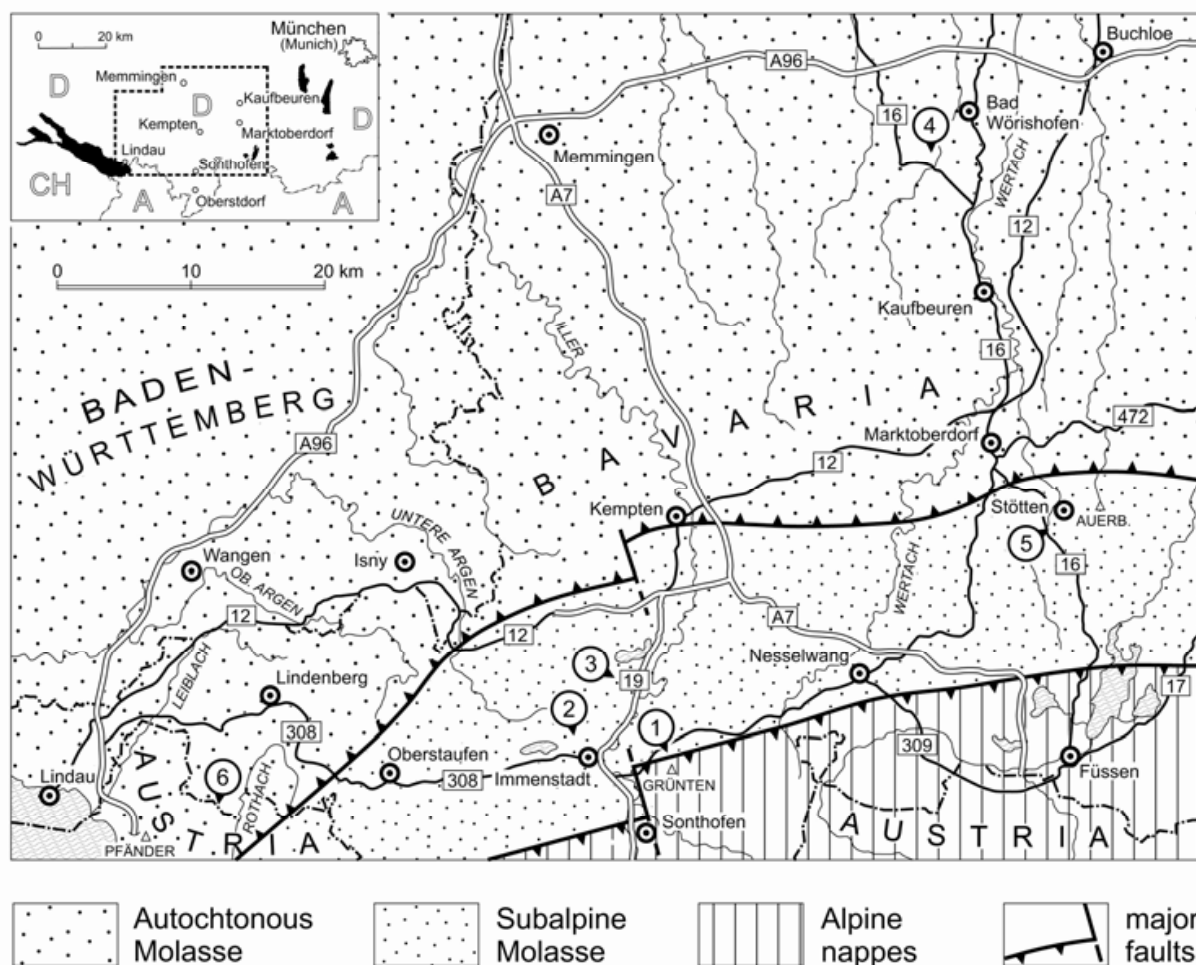
Fig. 1. Simplified geological map of the investigated area in southwestern Bavaria, Quaternary deposits are ignored. Exact locations of quarries and natural outcrops are described in the text:

1. Grüntensteinbruch (large sandstone quarry, LoMM) at the northern slope of Mt Grünten,
2. Zaumberg (old sandstone quarry, LoFM) at the blind road south of Zaumberg,
3. Niedersonthofen (old sandstone quarry, LoFM) near the road from Niedersonthofen to Seifen,
4. Hartenthal (old sand pit, UpFM) southeast of Hartenthal near Baisweil,
5. Burgleiten (old sandstone quarry, UpMM) westernmost foothills of Mt. Auerberg near Stötten,
6. Kesselbach (natural sandstone rock wall, UpFM) at Kesselbach gorge in the Mt. Pfänder area

Since Gümbel (1861: 165), this sandstone quarry is looked as the type section of the „Bausteinschichten", which have been comprehensively described by Schwerd (1978: 9, fig. 2). At the extremely steep slope down-hill of the quarry an at least some hundreds of metres thick sequence of underlying, grey silty marls with intercalated sandstone layers is exposed called „Tonmergelschichten" (LoMM). This Lower Oligocene sequence is covered with grey, middle- to coarse-grained massive sandstones, the „lower Bausteinschichten", overlain in turn by an about 5 m thick pelitic succession containing several thin sandstone beds called „middle Bausteinschichten" (Schwerd 1978: 9). This pelitic sequence is overlain by grey, glauconitic, middle- to coarse-grained, thinnly to massive bedded sandstones called „upper Bausteinschichten" (Schwerd, 1978: 9, fig. 3), towards the top grading into fine to medium-grained carbonaceous conglomerates (see Kronmüller, 1987):

Lithology and facies of the pelitic „Tonmergelschichten" and the „middle Bausteinschichten" are very similar, representing silty shelf sediments. Both pelitic sequences are overlain by sandstone-dominated successions („lower" and „upper Bausteinschichten"), that have a very similar lithology and facies as well. Their diversity of different bedding types and other sedimentary structures indicate a shallow marine environment, affected by many high energetic storm events. This facies repetition is interpreted by Reineck & Schwerd (1985: 51 ff.) as a repeated progradation of coastal sands on silty shelf sediments towards the N. According to Zweigel et al. (1998) this may be caused by a temporary reduction of sedimentary input leading to a prevailing of the subsidence rather than the result of sealevel changes. For details concerning lithology, sedimentary structures, fossil content, facies and sedimentary environment of this shallow marine sequence see Schwerd (1978: 11), Schwerd et al. (1983: 90), Reineck & Schwerd (1985: 52) and Scholz (1995: 151 f.).

Fig. 2a. Marl dominated middle Bausteinschichten and sandstone dominated Upper Bausteinschichten (LoMM), exposed in the Grünten quarry near Kranzegg (fig. 1 / 1). A large channel is visible within the middle Bausteinschichten filled with fine-sandy to silty deposits. The sediments above the interface middle / upper Bausteinschichten are affected by ball-and-pillow structures

At the interface between the pelitic „middle" and the sand-dominated „upper Baustein-schichten" erosive channels occur locally filled with fine-grained sand- and siltstones. At the basis of these more than 1 m deep channels flute casts are developed (Schwerd, 1978: 9). These channels are overlain by a 3 to 4 m thick intercalation of fine-grained sandstones and silty marls. At the interface between the pelites below and the overlying sandstone layers large load casts occur with marly flame structures (Roberts, 1989: 48) intruding the sandstones. Towards the hanging wall the penetrative mingling of marls and sandstones is increasingly intimate, grading into sphaerical bulbs, bulges and pillows with diameters of 0.1 bis zu 0.4 m (Schwerd, 1978: 9, fig. 2). In the upper part of this zone no primary bedding is preserved and replaced by a secondary ball-and-pillow structure. Within the several metres thick strata between undisturbed marls below („middle Bausteinschichten") and undisturbed sandstones above („upper Bausteinschichten") the preference for ball-and-

pillow formation obviously increases. Similar structures are very common and known to occur in many Paleozoic, Mesozoic and Cenozoic clastic successions (Richter, 1971). For more details see Scholz & Frieling (2006: 349 ff.).

The strata with ball-and-pillow structures are clearly synsedimentarily generated phenomena, that can be traced across the whole quarry over a distance of more than 500 m, although they are not developed everywhere in the quarry in the same quality and thickness and even missing in places. Nowhere else in the entire sequence similar structures occur, not even at the lithologically very similar transition zone between the „Tonmergelschichten" below and the „lower Bausteinschichten" above.
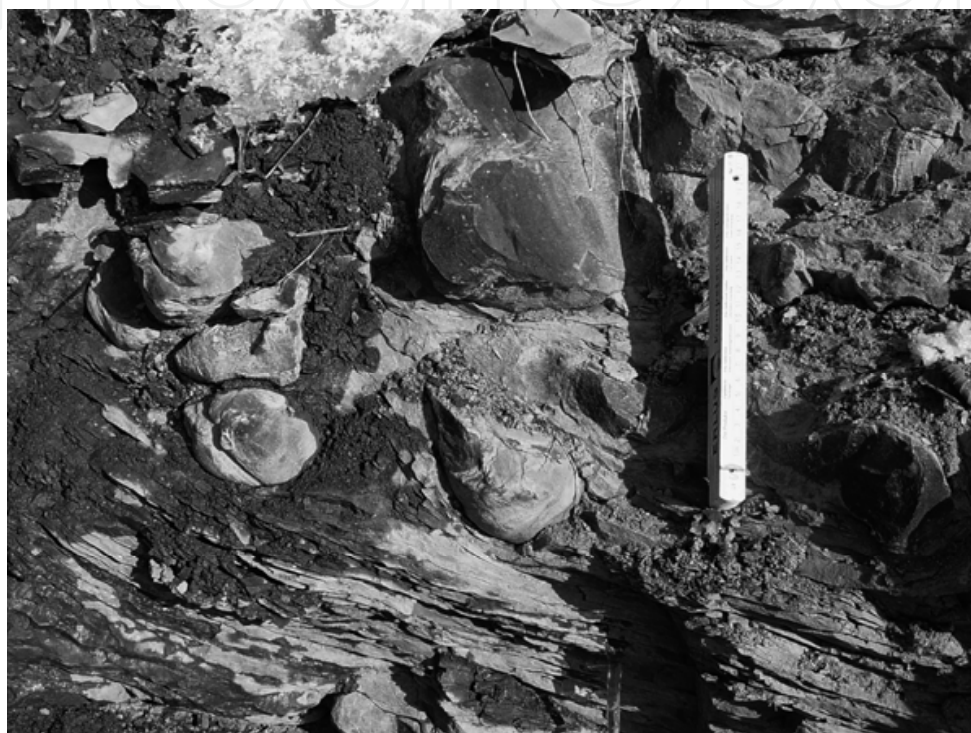

Fig. 2b. Detail from the base of the about 2 m thick deposits of the upper Bausteinschichten converted into ball-and-pillow structures

2.2 Synsedimentarily generated folds within Lower Freshwater Molasse sandstones north of Immenstadt (SW Bavaria)

The region north of Immenstadt belongs to the Subalpine Molasse. This area is dominated by an inclined interbedding of thick fluvial sandstone and marls, belonging to the Lower Freshwater Molasse (LoFM, Upper Oligocene to lowermost Miocene). Conglomeratic layers are very rare, thin and rather fine-grained. North of Immenstadt below Zaumberg, east of a hairpin bend of the old (now blind) road to Immenstadt, an only 6 to 8 m long old quarry is situated at the slope above the road (fig. 1 / 2, northing: 52.72380, easting: 35.89970, Scholz & Frieling, 2006: 357 f.). The fine- to middle-grained fluvial sandstones exposed here belong to the „Steigbachschichten" (uppermost Oligocene, Schwerd et al., 1983). They show a distinct horizontal stratification with layers a few to about 20 cm thick. These sediments belong to the northern limb of the Salmas Syncline (Jerz, 1974). Primary horizontally lying beds dip with about 21° towards SE. Some of the beds are cross bedded, indicating a sand transport towards different directions.

At two sites some layers are deformed to anticlines, a smaller one in the NE and a bigger one in the SW (fig. 3a). The bending of the layers is very distinct in the lower part of the outcrop, fading towards above. Since the layers are completely undisturbed in the upper part of the outcrop and no joints are referrable to the folding, only a synsedimentary and not a tectonic origin of the folds seem probable. They should have formed within the soft, not yet lithified sediment. For more details see Scholz & Frieling (2006: 357 f.).

Near the road from Niedersonthofen to the hamlet of Seifen in the Iller valley, a small, old, hidden and strongly vegetated sandstone quarry is situated immediately at the eastern road side (fig. 1 / 3, northing: 52.76725, easting: 35.93480, Scholz & Frieling, 2006: 359 f.). According to Jerz (1974) the sandstones belong to the LoFM („Weißachschichten", lower part of Upper Oligocene). They are part of the northern limb of the Salmas Syncline and dip 19° towards SSE (Jerz, 1974). The succession exposed here consists of sandstones partly showing planar stratification, single layers are up to 20 cm thick. Some beds are cross-stratified indicating a sedimentary transport towards varying directions. At some sites, the sandstones are made of small coss-stratified lenses indicating current ripples. At the lower part of the quarry wall a 10 to 20 cm thick marl intercalation is exposed. At the contact between the marls below and the sandstones above load casts and weakly developed ball-and-pillow structures occur (Scholz & Frieling, 2006: 357 f.).

Fig. 3a. Part of the Steigbachschichten (LoFM, Upper Oligocene) exposed at the quarry wall at Zaumberg (fig. 1 / 2) showing sandstone fold generated due to sudden dewatering. The diameter of the photo is approximately 1.6 m

Especially within an about 2 m thick bundle of layers the primary ripple-stratified sandstones are intensely deformed into 3 large and rather regular asymmetrical anticlines with amplitudes of more than 50 cm (fig. 3b). The axes of these folds strike more or less E-W.

The amplitudes of the folds fade within short distance and seem to grade into undeformed, primary planar stratified sandstones. However, 4 to 5 m further to the N, the same strata show another pronounced but weaker bending of the layers. For more details see Scholz & Frieling (2006: 359 f.). At the hanging wall the intensely deformed strata are overlayn by stratified sandstones without any deformational structures. Thus, a synsedimentary and not a tectonic origin of the folds is proved. As no joints are referrable to the folds, they should have formed within not yet lithified and still soft sediments.

Intensively folded, originally ripple-stratified sands of the upper Freshwater Molasse with a very similar shape and dimension have been found within a small and old sand pit SE of Hartenthal between Baisweil and Bad Wörishofen (Unterllgäu, fig. 1 / 4, northing: 53.16732, easting: 43.92520, see Havlik & Scholz, 2009).

Fig. 3b. Detail of the Weißachschichten (LoFM, Upper Oligocene) exposed at a quarry wall near Niedersonthofen. The ripple stratified sandstone (fig. 1 / 3) is synsedimentarily deformed and folded due to sudden dewatering

2.3 Synsedimentarily generated folds within pebbly sandstones of the Upper Marine Molasse west of Mt. Auerberg near Stötten (SW Bavaria)

The more than 1055 m high Mt. Auerberg, situated about 10 km ESE of Marktoberdorf is built up mostly by the Upper Marine Molasse (UpMM, Lower Miocene) dominated by coarse-grained conglomerates with sandstone intercalations. About 1 km SW of Stötten at the westernmost foothills of Mt. Auerberg at Burgleiten, coarse- grained pebbly sandstones are exposed within an old quarry close to the road junction, where the B 16 meets the old street from Stötten (fig. 1 / 5 northing: 52.89180, easting: 44.01030, Scholz & Frieling, 2006: 362 f.). According to Kuhnert & Rohr (1975) and Neubert (1999) the succession of the UpMM seems to be several hundred metres thick here, belonging to the northern limb of the Hauchenberg-Peißenberg syncline, which is the northernmost syncline of the Folded Molasse.
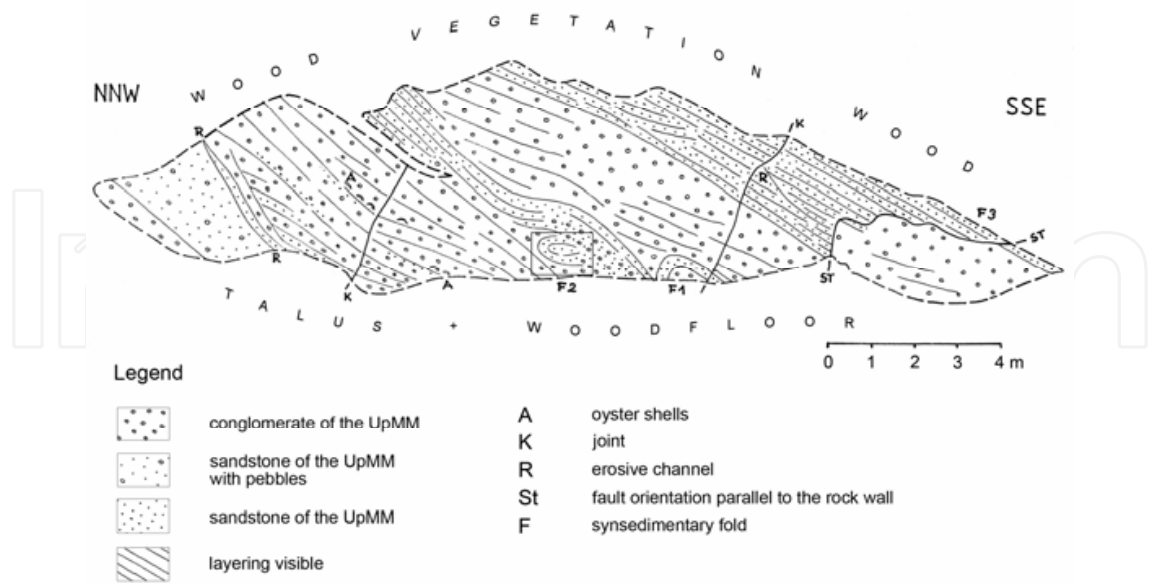
Fig. 4a. Sketch of the old quarry wall at Burgleiten near Stötten (fig. 1 / 5), where conglomerates and sandstones are exposed belonging to the UpMM (Lower Miocene)

Fig. 4b. Detail from 4 a, showing the synsedimentary fold F2 (framed in fig. 4 a) that has generated due to sudden dewatering

Lining more or less a nearly vertical normal fault with a fault throw of at least 1 m, the NNW-SSE orientated quarry wall at Burgleiten is up to 5 m high and about 25 m long. Most

of the stratified pebbly sandstones and poorly sorted conglomerates dip with 30 to 40°, at the eastern end even with 60° towards SE. The marine environment of the deposits is proven by a high content of glauconite and the occasional occurrence of marine fossils (mostly fragments of shells belonging to oisters and Pectinaceans). Additionally, coal fragments, skelletons of Bryozoans, shark teeth and carbonate pebbles with drill holes of mussles can be found. The sequence exposed here represents shallow marine near-shore sediments.

The stratification of the coarse grained deposits in the Burgleiten outcrop is indistinct to a certain degree. At some places a planar stratification or cross-bedding of the conglomerates is visible due to intercalated sandstone layers. The strata seem to be slightly bent in places, even distinctly deformed at the central part of the quarry wall. Apart from flexure-like bending of sandstone layers, two asymmetrical folds occur in the lower part of the wall with amplitudes of 50 to 100 cm (fig. 4). The axis of one of the folds with a northwestern vergency strikes 48° (fig. 4b), the axis of the other one with a northeastern vergency strikes about 120°. The overlying strata show no signs of any folding, dipping here constantly with 38° towards SE. Therefore, a tectonic generation of these folds is unlikely the deformation has been rather synsedimentary and formed within not yet lithified and still soft sediments. For more details see Scholz & Frieling (2006: 362 f.).

2.4 Sandstone dykes within Upper Marine Molasse of the Kesselbach gorge in the Pfänder area (SW Bavaria)

The Pfänder ridge situated east of Bodensee (Lake Constance) is part of the foreland dip panel, which defines the southern margin of the autochthonous Molasse. The Upper Marine Molasse (UpMM, Lower Miocene) in the Pfänder area (1064 m) is built up mostly by fine-grained glauconitic sandstones with intercalations of thick conglomerates and silty marls. They represent the shallow marine environment of a large delta, prograding from the Alps into the southern rim of the Molasse basin towards the N (Kuhlemann & Kempf, 2002, Berger et al., 2005, Frieling et al., 2009). The up to 400 m thick marine succession is completely exposed within the gorges of Wirtatobel and Kesselbach in the eastern Pfänder area, the strata dipping with about 18° towards NW. Both sections start in the youngest part of the Lower Freshwater Molasse and end in the lowermost part of the Upper Freshwater Molasse (Vollmayr & Ziegler, 1976, Kanzock, 1995, Frieling et al., 2009).

According to Hermann & Schwerd (1983) the lower quarter of the UpMM in the Kesselbach section belongs to the so called Luzern beds. They consist of a succession of platy, fine-grained sandstones showing low angle crossbedding, overlain by an intercalation of conglomerates and sandstones, reflecting a shallow marine environment with repeatedly changing energetic conditions. This part of the section is exposed on both sides of the gorge at more than 30 m high rock walls (fig. 1 / 6, northing: 52.66490, easting: 56.64720, Scholz & Frieling, 2006: 367). The lower part of the western vertical rock wall consists of grey, bedded, fine- to medium-grained, glauconitic sandstones, containing fragments of coal, mussle shells and barnacles. Some beds show a horizontal lamination or cross-bedding, are more or less bioturbated and contain long mostly vertical burrows. Due to bioturbation the primary bedding of an about 2 m thick part of the strata is obliterated or even completely destroyed. Moreover, these strongly bioturbated sandstones are discordantly cut by two subvertical sandstone dykes, about 2 m long and 10 to 30 cm thick, rooting in a subconcordant sandstone layer with a comparable thickness (fig. 5). About in the middle between the two subvertical sandstone dykes, a third sandstone dyke forks off downwards cutting through well bedded sandstones below. This sandstone dyke roots within a second subconcordant

sandstone bed exposed just at the very base of the rock wall. Here at some sites an undistinct layering can be seen, which is deformed where the dyke forks off. The subconcordant higher sandstone level and the discordant sandstone dykes are structure-less and are neither bedded nore bioturbated. The dykes are not lined by joints or faults. It seems to be a system of one source bed below, connected with a sill by a rather short dyke. The sill feeds two subvertical dykes, which thin out towards above.

The contacts between the host rock and the dykes are clear and can be traced mostly as distinct lines. The content of glauconite grains is high in the hosting sandstones and in the dykes and sills. The grain size of the sills and dykes is distinctly coarser than bioturbated host rocks. At sites where the dykes cut perpendicular through the bedded sandstones, the texture of well bedded parts of the host rock is disturbed close to the contacts. Here the lamination is dislocated along densely spaced joints or dismembered into angular or lenticular fragments. For more details see Scholz & Frieling (2006: 367 f.).
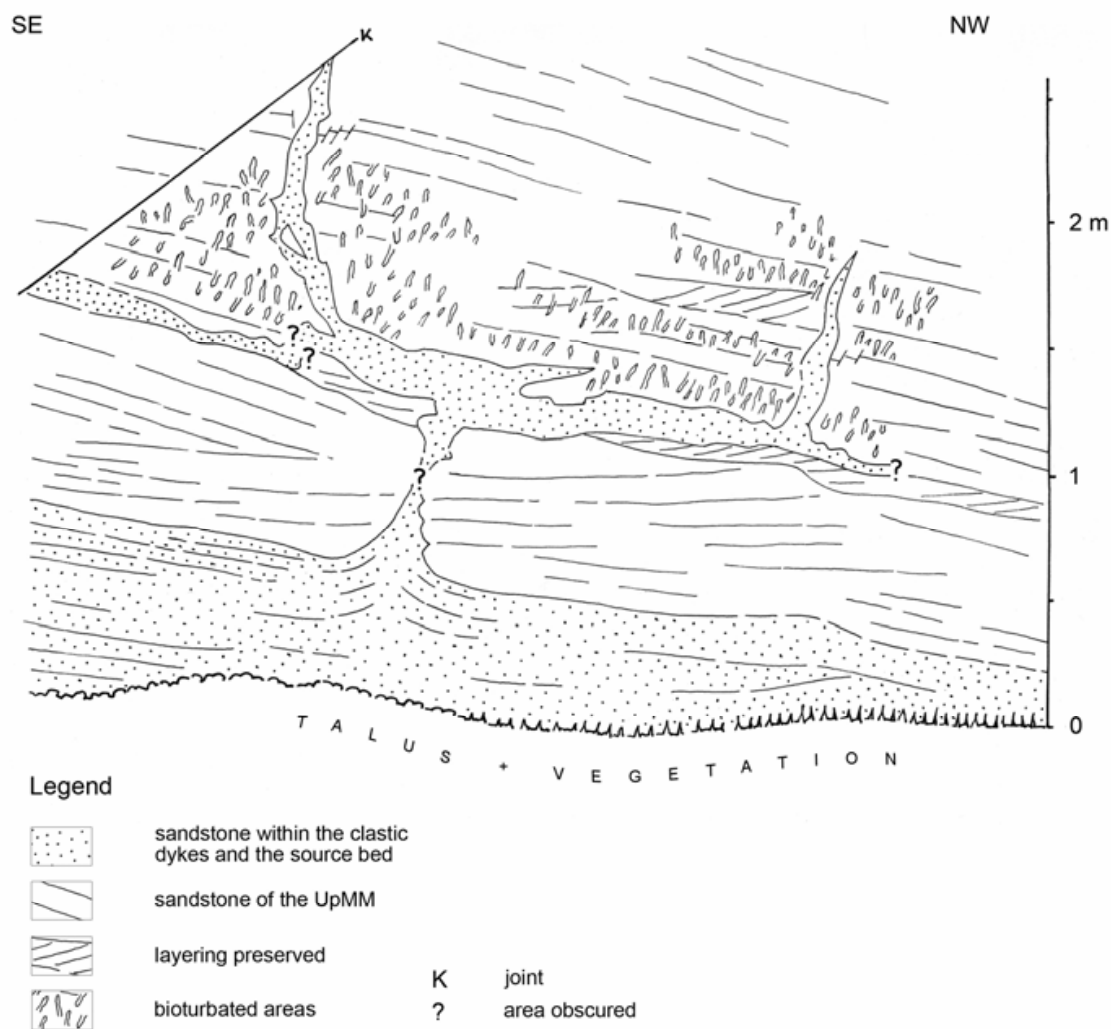

Fig. 5a. Sketch of the lowermost part of the rock wall at Kesselbach gorge (fig. 1 / 6) in the Mt. Pfänder area, showing a source bed (below), a sill and two clastic dykes within the sandstones belonging to UpMM (Lower Miocene)

Fig. 5b. Photo showing a detail of one of the clastic dykes (area left of the centre of fig. 5 near the two question marks), where it is connected with the sill

Jointing and fragmentation seem not to be of tectonic origin. Obviously, the sandy filling of the dykes was liquefied, deriving from the lower part of the strata and intruding towards above. When the sandstone dykes formed their sandy material must have been soft and not yet lithified. Thus, only their synsedimentary origin is imaginable. Discussion on their formation see below (4.3).

3. Sandstone dykes and ball-and-pillow structures within a Permian terrestrial Rotliegend succession in Hesse (Central Germany)

The Richelsdorf Mountains in Hesse (Central Germany) represent a block-like uplift of Permian strata within the Triassic Buntsandstein of the Hesse depression. It forms the westernmost part of a NW-SE trending horst structure, which includes the Thuringian Forest further to the E (fig 6a). The Richelsdorf uplift is limited to the SW by a large fault zone (Southwestern Boundary Fault), to the NE by the Sontra Graben (Sontra fault zone). Numerous faults furthermore dissect the Richelsdorf Mountain into a mosaic of individual tectonic blocks (fig 6b).

The Rotliegend was deposited above a Variscan basement with metasedimentary Paleozoic strata (greywackes, phyllites). The probably more than thousand metres thick volcanite-free Rotliegend succession consists of conglomerates and sandstones, where sandstones dominate the lower part, but conglomerates the upper part (Motzka-Nöring, 1987). Most of the succession is known exclusively from the former borehole Nentershausen 1862 (Beyrich & Moesta, 1876). Surface outcrops represent the uppermost 250 m, which consist of alluvial red beds made of an interbedding of conglomerates and sandstones together with some mudstone horizons (Motzka-Nöring et al., 1987, Aehnelt & Katzung, 2007). They are classified in age as Saxonian I (Aehnelt & Katzung, 2007). Saxonian II comprises conglomerates and pebbly sandstones, which exhibit secondary bleaching in transition to the Zechstein. Locally in the W, N and NE two different facies follow: (1) the aeolian Cornberg Sandstone s. str. and (2) a fluvial transitional facies (Oppermann, 1971, Munk et al., 1993, Gast, 1994, Aehnelt & Katzung, 2007).

Fig. 6. a) Simplified geological map of Central Germany with the location of the Permian uplift of the Richelsdorf Mountains (RM) within Mesozoic strata of the Hesse Depression. Rectangle marks detail of fig 6b; b) Simplified geological and structural map of in the Richelsdorf Mountains with location of the outcrop

In Permian times the Rotliegend deposits filled a small but deep depression within the Saar-Werra Basin, which was surrounded by NW-SE and SW-NE trending uplifts of the Variscan basement in the N, W and SW. From there material has been removed mainly as debris flows

and mud flows deposited on alluvial fans and in braided river systems on an alluvial plain. Sands blown out from the alluvial plain built up the dunes of the Cornberg Sandstone s. str. at the western to northern rim of the depression and the bordering flanks of the uplifts.

Sandstone dykes occur in the uppermost conglomeratic members of the Richelsdorf Rotliegend (Saxonian II) exposed in a road cut approx. 2 km south of Nentershausen (northing: 35.65780, easting: 56.51745; fig. 6b). The outcrop exhibits an interbedding of pebbly coarse sandstones and sandy conglomerates with internal even horizontal and cross stratification (fig 7a). Average pebble sizes are between 5 and 20 mm. The succession has been interpreted as deposition of braided river systems on an alluvial plain (Aehnelt & Katzung, 2007).

The clastic dykes are developed as straight-walled, narrow dykes, which cut perpendicular to the stratification of the host rock (fig. 7b, c). They are up to 170 cm in length and appear as filling of thin, partly wedge-shaped fissures. In parts, vertical joints follow the course of the dykes (fig. 7c). The basal part of the dykes is indistinctly defined; it derives from diffuse roots in a sand-rich bed. In parts, disturbance of primary bedding is apparent. The dykes pass through pebbly sandstone beds and terminate upwards more or less in the same stratigraphic horizon. The top is not well defined and ends indistinctly in another sand-rich bed, which is overlain by a conglomerate horizon. The filling material consists of reddish, matrix-rich sandstone with distinct, clear contact to the host rock at the dyke wall. The host rock near the dykes is greyish due to secondary bleaching, and may be enriched in gravels and imporished in sandy matrix. Especially in coarser, gravel-rich host beds the outlines of the dykes are often accompanied by streaks of granules to small pebbles or by lateral gravel lags with clast sizes of up to 30 mm (fig 7 b, c).

Beside dykes deformation structures are observed also in the host rock next to the dykes (fig. 7d). These are ball-and-pillow-like structures in the pebbly sandstone beds, which are apparent by grain size separations in (1) areas enriched in reddish, matrix-rich sandstone, (2) greyish gravel lenses and (3) areas with transitional grain sizes. The base of the upper sand-rich horizon, in which the dykes seem to terminate, shows undulation and bulbous outlines owing to deformation and smallscale diapiric intrusion of the underlying gravelly sandstone and gravel lenses respectivelly. In return, small sand dykes of a few centimeters in length finger from the sandstone horizon into the underlying bed. A similar situation is observed at the top of this sand-rich bed: bulbous outlines due to a pillow-like downsinking of gravel lenses into the sandstone bed. At some places gravelly intrusions from below and gravel bulbs from above seem to be connected forming subvertical streaks that cut dyke-like through the sandstone bed and terminate in the overlying, gravel-rich horizon.

The whole deformed complex spans the lateral extent of this road cut of about 20 m, and seems to have a thickness of nearly 2 m (fig 7a). The intensity of deformation seems to increase from the bottom to the top or the upper sand-rich bed respectively. Beds above and below the complex appear undisturbed, at least the latter ones exhibit well defined, undeformed cross beddings. Above the deformed complex massive pebbly sandstones are exposed, which overlay a conglomerate horizon with minor deformation. Sand dykes end in the sand-rich bed below.

Since the sandstone dykes start and end more or less all at the same stratigraphic levels, and can not be traced further upward, their formation must have happened in situ restricted to certain beds in times when the material was still soft. The ball-and-pillow structures as well as bulbous outlines of sand-rich horizons evidence that the host beds were also partly fluidized. Their remobilization caused the destruction of the primary bedding. Thus these

features are typical for soft-sediment deformation and prove a formation while sediment
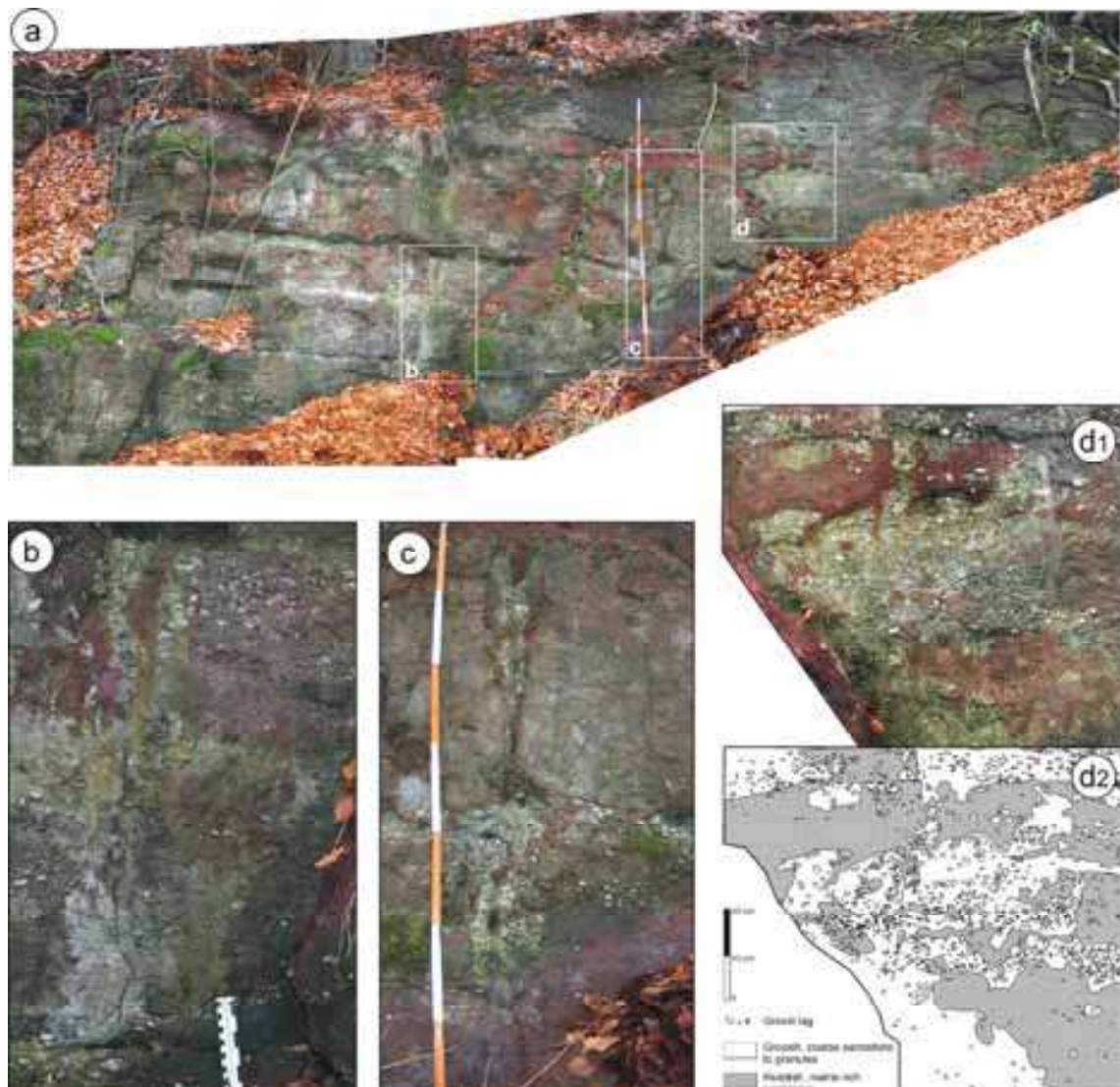was not yet lithified.

Fig. 7. a) Photograph of Rotliegend conglomerates and sandstones with synsedimentary
deformation features, exposed in a road cut near Nentershausen. Rectangle marks location
of detail photos 7b-d; b) Detail from 7a showing clastic dykes with diffuse roots in a
sandstone bed and lateral pebble lags in a conglomeratic horizon; c) Detail from 7a showing
clastic dykes with lateral pebble lags when passing through gravelly horizons. A vertical
joint follows the course of the dyke; d) Synsedimentary deformation structures in the host
rock next to the dykes generated by dewatering (d1 foto, d2 sketch)

4. Funnel structures and clastic dykes in Lower Cambrian Vik Sandstone in Scania (Sweden) and in the basement of Bornholm

Originally, marine Paleozoic sediments have been widely spread in southern Scandinavia,
primarily forming a more or less continuous blanket on top of the Precambrian crystalline
rocks of the Fennoscandian Shield. In post-Silurian times this CambroSilurian cover has
been completely removed by fluvial, coastal and glacier erosion in most places (Scholz &

Obst, 2004). The Lower Paleozoic sediments have been partly covered with thick Mesozoic sediments and down-faulted by Permosilesian to Paleogene tectonic movements within the Fennoscandian Border Zone (Lindström & Dworatzek, 1979): Remnants of these epicontinental sediments are preserved today e.g. in Scania (Skåne, southernmost Sweden). The coast of the Baltic Sea north of Simrishamn situated in eastern Scania, is made up of Lower Cambrian sediments dominated by shallow marine sandstones, resting unconformably upon Proterozoic plutonic and high-grade metamorphic rocks of the Fennoscandian Shield. The lower part of this transgressive sandstone sequence belongs to the about 110 m thick sandstone-dominated Hardeberga formation, which is rather widespread in Scania. It can be lithostratigraphically subdivided into several members (Nielsen & Schovsbo, 2006). One of these members is the light-grey, hard Vik Sandstone (Vik member) representing an about 25 m thick part in the middle of the formation (Lindström, 1972, Hamberg, 1994). The Vik sandstone is exposed e.g. near Vik north of Simrishamn (Hamberg, 1994: 38; fig. 8). The intensively bioturbated shallow marine sandstones generated near the tide-influenced Lower Cambrian coast, situated not far from the northern rim of the lower Paleozoic Tornquist Ocean.
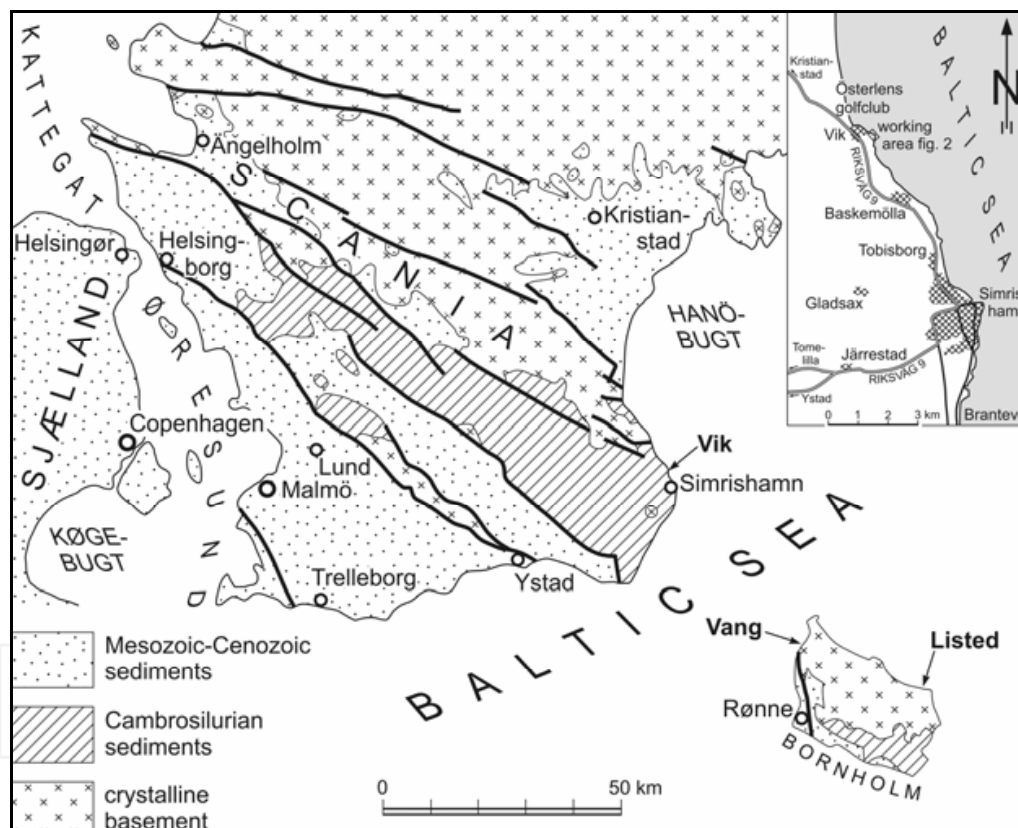

Fig. 8. Simplified geological map of Scania and Bornholm. Thick lines indicate major faults. The position of the investigated area at Vik in eastern Scania is marked with an arrow. Two additional arrows point to the positions of Vang and Listed on Bornholm, locations with sugtrusive sandstone dykes mentioned in the text

The mostly quartzitic, fine to medium grained and mature, cross bedded sandstones are nearly flat lying or gently dipping. However, they contain a couple of horizons where the lamination is completely destroyed by bioturbation. Some of them form very even bedding planes with rough and podgy surfaces, hiatus planes penetrated by densely spaced fossil

burrows, mostly to *Skolithos* and *Diplocraterion* (Häntzschel, 1975*)*. The burrows have been generated by unknown marine suspension feeders, typical colonizers of lagoons. Bromley (1996) looks this „Skolithos-Ichnofacies" as being characteristicly for medium- to high-energetic deeper intertidal to shallow subtidal marine environments near the coast. This ichnofacies is typical for wellsorted sands generated under quickly changing sedimentary conditions with fast deposition and erosion in a shallow marine environment. According to Lindström (1972) this high energetic shallow water sediments may have generated at a tide influenced coast under rather cold climatic conditions. For more details concerning the sedimentary environment, see Hamberg (1991) and Scholz et al. (2009: 357 ff.).

Irregular distributed narrow ridges rise above these podgy hiatus surfaces, forming branching bulges of several metres in length (fig. 10). Moreover these bedding planes are deformed at several sites and seem to outline conspicuous minor and large circular tub and funnel shaped structures (fig. 11).
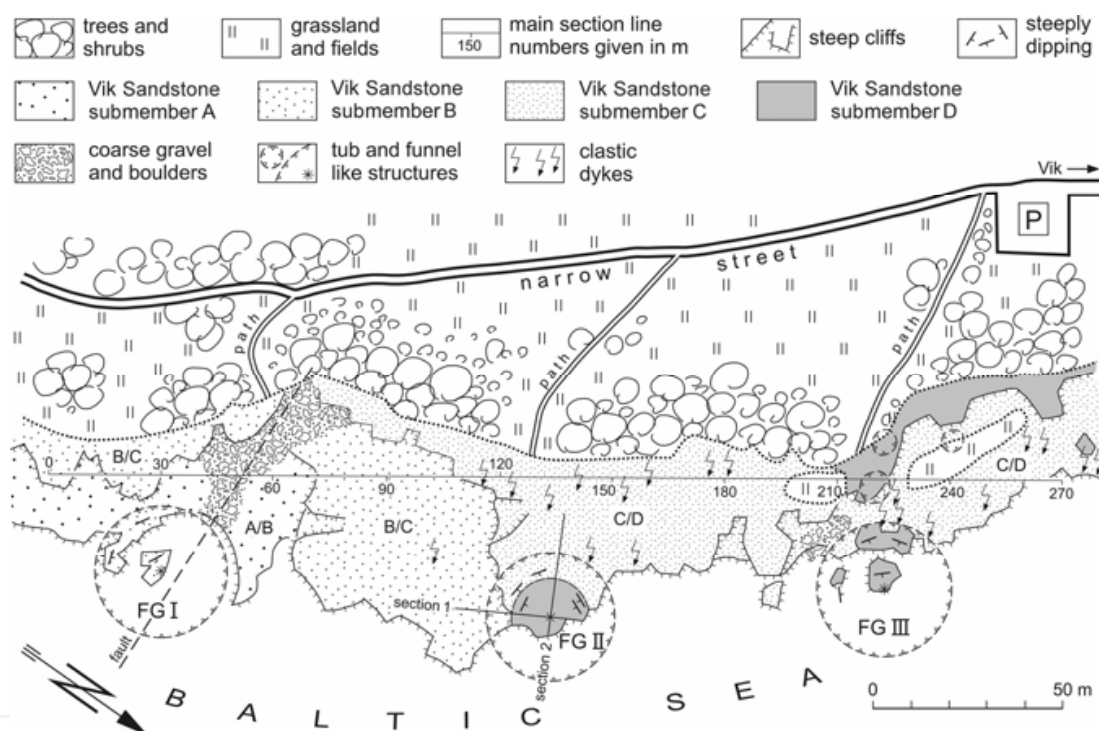

Fig. 9. Detailed geological map of the investigated area at the coast south of Vik in Scania. The location of extrusive clastic dykes can be seen, as well as the funnel structures FS I-III, forming a more or less straight chain. FS II is identical with „Prästens Badkar" mentioned in the text

4.1 Sandstone dykes within the Lower Cambrian Vik member in Scania

Several slightly curved bulges of light greyish medium grained sandstone occur on top of several hiatus surfaces south of Vik (Scholz et al. 2009: 357, 55° 36' 46.35'' N, 14° 17' 50.71'' E). Locally they rise 10 to 15 cm above the bedding planes of the Vik sandstone. A few are less than 1 m, others more than 14 m long. Some of these sandstone bulges are only some centimetres, others up to 25 cm thick. Allthough cross-cuts through these bulges are rare. It could be observed that all these bulges continue downwards, where they form dyke-like discordant bodies, which are steeply or even vertically oriented. One of these sandstone dykes is more than 15 cm wide and tracable downwards for at least 70 cm, others cut

through the sandstone layers in the hanging wall strata. Some of these sandstone dykes
have distinct contacts and differ clearly in colour and / or grain size from the wall rocks.
Thin sections of a dyke sample show strong similarities to the wall rocks in maturity and
sorting, but they have the comparably least share of matrix (Scholz et al., 2009: 365 f.).

Several strike readings suggest that there is a different favoured orientation of the bulges in
different parts of the investigated area. While most of the sandstone dykes in the SE otthe
area strike 120-150° and 50-60°, the dykes in the NW strike preferred 70-90°. There is no
parallelity between the prevailing majority of the dykes and the normal faults affecting the
sandstones. Many sandstone dykes bifurcate, some of them repeatedly at acute angles (fig.
10). Most of them branch only into one specific direction. The dykes in the SE of the mapped
area branch off only to the W, NW and NE, dependant on their basic orientation. The dykes
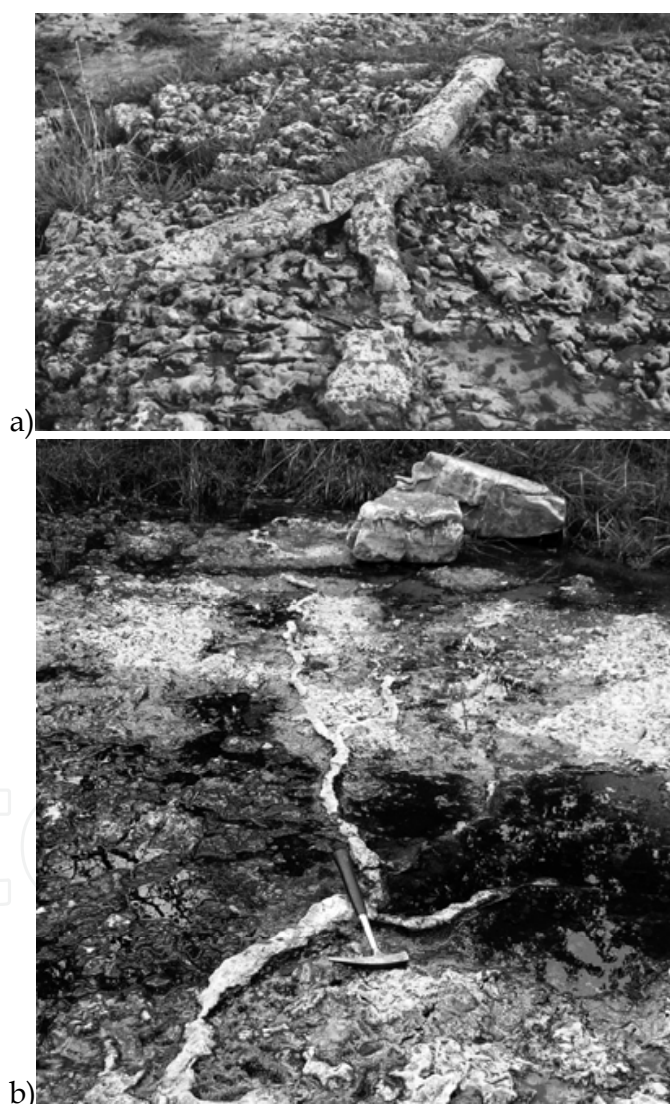in the NW branch off into western or eastern directions.

Fig. 10 a,b. Photographs of branching bulges on top of a heavily bioturbated hiatus plane
within the lower Cambrian Vik Sandstone at the coast near Vik. These bulges continue
downwards and are clastic dykes cutting through the sandstone layers below. These dykes
have generated due to the liquification of an unstable sandy layer below the visible surface
triggered by seismic shock waves

At some sites the sandy matter of dykes is weakly bioturbated differing in colour and grain size from the wall rock. All these observations point to a synsedimentary origin of the dykes. They formed presumably before the deposits had been lithified and are therefore clastic dykes. Their age relations with the funnel structures and their possible origin are discussed below. For more details see Scholz et al. (2009: 365 f.).

4.2 Tub- and funnel-shaped structures within the Lower Cambrian Vik member in Scania

Apart from these sandstone dykes and tectonic faults the nearly flat lying hiatus planes within the Cambrian Vik sandstone are intensively bent and contorted in places. There are several areas with diameters of one up to a couple of metres, where the sandstone layers are centroclinally deformed. The bedding planes within these relatively small structures are nearly horizontal at the rims, dipping with increasing angles approaching the centres and decreasing again in the very middle, forming tub or bowl like structures (Hamberg, 1994, Obst, 2004; fig. 11).

Especially at three different sites close to the water line of the Baltic Sea, three centroclinal structures are exposed with diameters up to several tens of metres (Scholz et al. 2009: 357, 55° 36′ 46.35″ N, 14° 17′ 50.71″ E; fig. 9). The hiatus planes are nearly horizontal in the periphery of these three large structures. Within the innermost parts the bedding planes show abnormal steep inclinations, showing a centroclinal strike and attaininging values of 50° to 70° in places (Scholz et al., 2009: 366 f.). Within the very center of these structures the layers flatten in a short distance until they dip steeply into the opposite direction again. Thus, the primarily horizontal bedding planes are deformed to show sterically bent funnel-like surfaces. The bedding planes within one of these three centroclinal structures are fully preserved even within the innermost core. This partly water-filled structure is well known, called „Rosenstenen" („stone rose") by the local people, „Prästens Badkar" („priest's bath-tub") on topographic and touristic maps or „Parson's Bathtub" by Lindh & Bergman (1995: 220) (fig. 11).

At the outer borders of the three large centroclinal structures (funnel structures) there are many sub-concentrically oriented, minor normal faults, displacing the bedding planes some centimetres or decimetres. The slightly bent fault planes dip steeply towards or away from the centres of the funnel structures. They are best visible within the western and southern periphery of „Prästens Badkar" (FS II) on top of the hiatus planes. The up and down of these marker horizons suggests horst- and graben-structures. Approaching the core of the structures the vertical separations along the concentrically arranged faults seem to increase, exceeding at least 1 m for each downfaulted block closer to the centre.

Due to the funnel-shaped deformation and the down-faulting within the central parts of the funnel structures, the overlying successions have sunken down and thus partly escaped erosion. The amount of down sinking can be estimated using one of the hiatus planes as a reference level. So a down-sinking of the core of this structure of at least 6 to 7 m can be estimated. For more details concerning the funnel structures of Vik see Scholz et al. (2009: 366 f.).

Because of the funnel-shaped deformation and the down-faulting of the layers within these structures, they have been named „funnel-grabens" by Lindström (1967). However, the name „graben" suggests a tectonic origin and implies the formation of a linear rather than a curved and centroclinal structure. Therefore the name „funnel structure" has been introduced by Scholz et al. (2009: 368). Their age relations compared with the clastic dykes and their possible origin are discussed below.

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Fig. 11a. Photograph of the very centre of the funnel structure FS II „Prästens Badkar" with funnel-like inclined primary bedding planes. This synsedimentary collapse structure has been synsedimentarily generated most likely above a opening fissure in the crystalline basement below

Fig. 11b. Photograph of sub-concentrically arranged faults, visible on top of a more or less flat lying hiatus plane just S of „Prästens Badkar"

4.3 Age relationship between sandstone dykes and funnel structures in Scania and the dimension of the volume loss below

Sandstone dykes: The field observations suggest that neither the clastic dykes nore the funnel structures have been generated before the Cambrian sandstone lithified and when the sediments were still soft. If this is true, they should branch off a „source bed" somewhere below the sedimentary surfaces cut by the dykes (Jolly & Lonergan, 2002: 607, Scholz & Frieling, 2006). However, their roots are not well exposed and we have no information so far, how deep the possible source beds are situated below the exposed bulges. Only at one site a sandstone horizon below is seen to grade into a dyke above, vertically cutting through at least 70 cm of the overlying sandstone succession.

In most cases the exact age of the clastic dykes in Vik remains unclear. One dyke south of funnel structure II traceable over a distance of more than 14 m on the surface of a large hiatus plane definitely does not to continue into higher stratigraphic levels. However, some dykes forming mushroom-like bulges above the hiatus planes clearly do so. They have generated after the bioturbated bedding plane has been already covered with an at least thin layer of silty sediments. On the contrary, some of the branching dykes may have more or less the same age as the hiatus planes affected. These dykes seem to have cut the former sea floor, were eroded, egalized and bioturbated subsequently in their uppermost parts. The animals causing *Chondrites* could hardly have survived an add of many sand layers and continue to bioturbate newly intruding dykes a couple of metres below the sea floor.

Thus, contrary to the opinion of Hamberg (1994: 40), not all clastic dykes have the same age and their formation covers a certain time span. They may be related to several seismic events during the lower Cambrian. Moreover, they seem to rise from different source beds belonging to different stratigraphic levels and are not rooted within the same horizon (Scholz et al., 2009: 368 f., fig. 9).

Funnel structures: Both, the small tub-like structures and the large funnel structures neither are the direct result of a tectonic controlled down sinking nore did they form after the sediments have been lithified. This is suggested by the lack of jointing related to the deformed strata, also within the tightly bent sandstone layers in the centre of „Prästens Badkar". Furthermore, the sub-concentrically orientated normal faults within the periphery of the funnel structures do not show any relationship to the suborthogonal conjugated system of regional joints. They seem to have been generated earlier, when the still unlithified sediment has been stretched towards the centres of the structures. The sediments must still have been soft at the time of their deformation, but stiff enough to tear up. Therefore, all funnel structures are obviously collapse structures of synsedimentary origin (Scholz et al., 2009: 370 f.).

The tension, deformation and downsinking of the soft sediments have been caused by a loss of volume in the underground below the Vik member. It is possible to estimate the volume of the down-sunk sand that would contain at least 1000 m³. Considering the two additional funnel structures in the vicinity which ar even larger a total material loss of more than 3200 m³ below all three funnel structures must be assumed. For details of the calculation see Scholz et al. (2009: 371 f.)

5. Origin of all these synsedimentary deformation structures

Most authors agree, that the majority of the structures presented here like load casts, synsedimentarily formed folds, sandstone dykes or funnel structures, seem to generate due to the liquefaction of water-saturated sands (e.g. Richter, 1971, Reineck, 1984, Ricchi-Lucchi,

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1995, Jolly & Lonergan, 2002). For a short time span certain beds within water-saturated sediments are converted into a very mobile suspension consisting of water and sand grains, ready to move quickly laterally or even upwards and downwards. Precondition for the liquefaction of sediments is a sudden breakdown of their unstable grain fabric (see Blanc, 1998: fig. 5). There are certain beds which seem to have a principal disposition to liquefy, mostly water-saturated well sorted sands with well rounded grains of high sphericity. Moreover, most of the coarse-grained sand layers within a clastic sedimentary succession seem to have a special predestination to fluidize (Jolly & Lonergan, 2002: 607, Ross & White, 2005). A break-down of their grain fabric causes a sudden overpressure of the pore fluid. The degree of overpressure is dependant on the position within an actively forming sedimentary succession and will increase with the thickness and consequently the lithostatic pressure of the overlying strata. According to Jolly & Lonergan, 2002: 607) this pressure can be much higher than the pore pressure within the deposits above. The fluidization of sediments situated directly at the sedimentary surface will lead to a deformation of the layers, an abrupt dewatering and a quick escape of the water at the surface. The fluidization of deposits deeper below the sedimentary surface will lead to deformations like load casts, ball-and-pillow structures or even clastic dykes, if the pore fluid overpressure is extremely high.

All these processes leading to a deformation of the sediments may happen during or soon after deposition. The principal disposition for a liquefaction of certain beds within the deposits continues to exist, until the layers loose their softness and mobility due to compaction, dewatering and cementation. This disposition should decrease with an increase of age and subsidence.

What caused the liquefaction of especially sensitive layers within the sediments leading to the generation of a variety of deformational structures? It can be principially triggered by different conditions and events. According to Richter (1971) or Gamberi (2010) some of the structures mentioned above can simply be created by rapid sedimentation, unevenly distributed load or density inversions. Osborne & Swarbrick (1997) think, that an overpressure of the porewater can also generate from disequilibrium compaction. However, high-energetic events such as heavy storm waves, tsunamis (Reineck, 1984: 15, Reineck & Singh, 1986: 59, Roberts, 1991: 54) seem to play an important role in most cases, especially seismic shock waves. However, Ricchi Lucchi (1995: 159) hesitates to interpret all these structures automatically as seismites, especially clastic dykes. He advises that every single case should be examined very carefully. As triggers for a liquefaction of sensitive layers also a sudden increase of the sedimentary load (result of mass movements or rapid input of unusual large quantities of clastic material), the migration of hydrocarbons from below (Jolly and Lonergan, 2002, Gamberi, 2010), ice tectonics or synsedimentary extensional tectonics (Demoulin, 2003) can be discussed. Additionally, dykes may also generate from the filling of pre-existing ground cracks and fissures from below, created e.g. by deccication, tectonism or slope instability, with larger fissures being more prevalent in arid climates (Holzer & Clark, 1999: 876).

5.1 Origin of ball-and-pillow structures within the Lower Marine Molasse of southwestern Bavaria

Load casts as well as ball-and-pillow structures occur e.g. within the Lower Marine Molasse (LoMM) exposed in a large sandstone quarry N of Mt. Grünten near Sonthofen (Grüntensteinbruch, fig. 1 / 1) und within the Lower Freshwater Molasse (LoFM) exposed in small abandoned sandstone quarries near Niedersonthofen (Scholz & Frieling, 2006: 349).

According to Reineck (1984: 15) or Ricchi-Lucchi (1995: 159) load casts are sedimentary structures occurring at the base of sandstone layers which have formed due to the down-sinking of water saturated sand into a soft pelitic substrate consisting of clay and silt. An essential condition for their genesis is an unevenly distributed load, which could be caused by a pinch and swell, ripples at the top or scour structures at the base of the sandy layers (Dzulinski & Walton, 1965, cit. in Reineck, 1984: 15). Density inversion and a thixotrope behavior of the underlying pelitic sediments may promote their formation, and are likewise precondition for the genesis of small flame structures, diapires and ball-and-pillow structures. According to Ricci Lucchi (1995: 163) very similar structures can be generated most easily in experiments, if sediments are strongly shaked. Therefore, these structures seem to form due to high-energetic events, like heavy storm waves, Tsunamis or seismic shock waves (Richter, 1971: 19, Reineck, 1984: 15). Also in case of the Molasse successions such high-energetic events are most probably the cause for the deformational structures described above. The increasing degree of mingling of sandy and silty sediments at the base of the „Obere Bausteinschichten" could have been caused by the originally decreasing water content towards the top of this sequence. It could also be interpreted as an indication of a synchronous dewatering of the whole strata causing an increase of the pore water pressure towards the hanging wall. Alltogether this is an indication for the generation of all these structures by a unique trigger event.

The lithology of the LoMM sequences is both, nearly identical and genetically related, as well at the base of the sandy „Untere Bausteinschichten" as at the base of the sandstone-dominated „Obere Bausteinschichten". According to Reineck & Schwerd (1985: 51) the two sequences reflect a transition from silt dominated shelf sediments into sand dominated shore face deposits, the progradation of psammitic coastal sands on pelitic shelf sediments. In spite of that, ball-and-pillow structures have generated only at the base of the younger „Obere Bausteinschichten". Below the older „Untere Bausteinschichten" only some weakly developed load casts are present (see Schwerd, 1978: fig. 3). Shocks caused by strong storms are frequently ocurring events, which should have generated identical structures at the base of the „Untere Bausteinschichten" as well. Their singular occurrence within the whole sequence supports the idea that they could have generated only by strong seismic shock waves generated by an extremely heavy earthquake. Similar structures are interpreted in the same way also by Obermeier (1996, fig. 35).

5.2 Origin of synsedimetary folds within Lower Freshwater Molasse and the Upper Marine Molasse of southwestern Bavaria

Synsedimentary folds are normally considered to be slump folds. However, it is nearly impossible to interpret the folds within Molasse successions described above as synsedimentary slump folds (LoFM in Zaumberg and near Niedersonthofen, figs 1 / 2-3, figs. 3a, b, UpMM at Burgleiten, fig. 1 / 5, fig. 4 and UpFM at Hartenthal, fig. 1 / 4). All slump folds spontaneously forming under the influence of gravity need a clearly inclined sediment surface. The inclined rims of lakes or marine basins and inclined delta fronts are environments where slump folds could develope. In three of the cases described above these synsedimentary folds generated within fluviatile sequences of the LoFM and the UpFM, within sandy, crossbedded deposits of braided rivers. Apart from a few thin intercalations of fine to medium grained conglomerates coarse-grained sediments are missing. Therefore the sedimentary environment of these Oligocene to Miocene mostly sand transporting Molasse rivers should have been mostly even flood planes with very low gradients far

below 1‰. This cannot be seen as a good precondition for slumping (Scholz & Frieling, 2006: 371 f.).

Despite of some deposits of oxbow lakes, sediments of large lakes are very rare and even absent at the central part and near the southern rim of the Molasse basin. So the assumption of lacustrine deltas with steeply inclined delta fronts can be excluded. Also marginal marine deltas can not have existed close to Immenstadt in the Upper Oligocene, as this region was situated far inland in this time. The surface of the deltaic UpMM sediments, too, could not have been originally perceptible inclined. The tectonically caused dip of the bedding within the old quarry of Burgleiten is more or less parallel to the inclination of exposed sandy bedding planes in the vicinity, that are widely covered with current and interference ripples, proving their originally more or less horizontal surface. Consequently, one essential prerequisite for a gravity-controlled sliding is missing: a palaeoslope with an inclination of at least some degrees. If the folds described here would be the product of sliding, a basal decollement plane should be visible. However, such a decollement is not visible at the examples given above from the LoFM (1.2) and UpMM (1.3). On the contrary, at Zaumberg and near Niedersonthofen the fold amplitudes seem to increase towards above whereas the fold length remains identical.

Folding of this type is rather typical for the abrupt dewatering of sediments with an unstable grain fabric (Blanc, 1998: fig. 5). The occurrence of these folds near Niedersonthofen together with ball-and-pillow structures is an additional argument for this interpretation. Similar folds are known e.g. from the Proterozoic of India (Bhattacharya & Badyopadhyay, 1998), the Upper Cretaceous of Brazili (Rosetti & Góes, 2000, Rosetti & Santos, 2003) or the Pleistocene of Spain (Gilbert et al., 2005). They too, formed most likely due to heavy seismic shocks (Scholz & Frieling, 2006: 373).

5.3 Origin of the sandstone dykes within the Upper Marine Molasse of Bavaria and the Vik member of Scania

The ocurrence of sandstone dykes within Molasse sequences is referred only from the UpMM of the Pfänder region so long (fig. 1 / 6, fig. 5, Scholz & Frieling, 2006: 365). Similar dykes have been identified by two of the authors within marine Upper Silurian siltstones of southeastern Norway (Oslo region, Steinsfjord) and within siliciclastic marine Cambrian sandstones of southern Sweden (Scania). Shape, material, geometry and contacts to the country rock of all these sandy dykes indicates, that the material has penetrated already pre-existing but still soft sediments. The original bedding of all these sediments has been partly affected by bioturbation prior to the intrusion of the dykes. Different from the intruding mobile quicksands, the sands of the overlying host rocks must have had a rather stiff consistence in this time. The material came obviously from below, using fissures which opened prior or in course of intrusion. None of these dykes has been an open fissure before the intrusion. The filling of the dykes was neither slow nor from above, the penetration occurred very rapidly and from below. This is proved especially at Kesselbach, because of the wedging out of the sandstone dykes towards the hanging wall. Therefore, all these dykes have to be classified as clastic dykes and not as neptunian dykes, open fissures which should have been filled from above slowly and grain by grain (Scholz & Frieling, 2006: 374 f.).

In some cases, the intruding material seems to attain the sedimentary surface. Sandy walls formed by protruding clastic dykes due to strong earthquakes are known to appear on dry tidal flats. Such walls generated by high energetic seismic shocks have been described and investigated in detail near Anchorage in Alaska by Reimnitz and Marshall (1965) and noted

by Reineck and Singh (1986: 59). These clastic dykes were not the only result of this earthquake in the early sixties. The tidal flat at Anchorage was uplifted a couple of metres above the water line. So these sandy walls were rather stable and survived for a couple of weeks, as the extruded material rose above a totally dry sedimentary surface.

All bulges visible on the hiatus planes of the Cambrian Vik sandstone in Sweden are obviously clastic dykes, some of them must have cut through the former sedimentary surface as well. Like tooth paste protruding from rather narrow fissures, the material initially may have formed sandy walls as well rising above the sea floor. However, the sands may quickly have flown aside as the sedimentary surface was submerged in this time. Subsequently, the extruded sands were bioturbated by animals. The clastic dykes in the Upper Marine Molasse of the Pfänder area are different, as they did obviously not attain the sedimentary surface. As the dykes wedge out towards above, they seem to have got stuck within the overlying strata.

No indications for mass movements, rapid input of unusually large quantities of clastic material, migration of hydrocarbons from below, ice tectonics or extensional tectonic have been found, which could have caused the synsedimentary clastic dykes neither in the Cambrian shallow marine deposits of Sweden nore in the Miocene shallow marine deltaic sediments of Bavaria. Extreme storm waves as triggers are unlikely at least in the Molasse sediments, as the whole Molasse basin has been a shallow and very narrow sea. Moreover, tsunamis are mostly generated at active and destructive plate boundaries that are missing in both cases (see above). Consequently, the clastic dykes are most likely the result of heavy seismic shock waves. In the case of the Vik Sandstone, this is supported by the fact, that these „extrusive clastic dykes" (see below) are not straight as the „sugtrusive clastic dykes" (see below), but rather curved or irregular, probably due to the soft substrate. Determination of the least horizontal stress direction is also difficult. Their favoured strike is 80°. This would imply a N-S oriented extension that is different from the stress regime documented by the clastic dykes in the basement indicating an extension in NE-SW direction (Katzung & Obst, 1997: 168). No data belonging strike and shape of the clastic dykes in cuts parallel to the bedding are available within the Miocene shallow marine deltaic sediments of the Pfänder area. According Jolly & Lonergan (2002: 614) the formation of clastic dykes can easily be triggered by earthquakes with magnitudes of more than 5 to 6 on the Richter scale. Suggesting that the clastic dykes of Vik are of different ages, the whole succession seems to have been repeatedly affected by heavy earthquakes. As most of the branching dykes bifurcate only into specific directions depending on their orientation (see 3.1), this polarity could perhaps be used to identify the location of the Cambrian epicentres (Scholz et al. 2009: 374).

5.4 Origin of the sandstone dykes and deformational structures within Rotliegend sandstones and fanglomerates of Hesse (Central Germany)

The deformation structures observed in the Hesse Upper Rotliegend sediments of the Richelsdorf Mountains are restricted to a certain stratigraphic level and evidence the complete in situ disintegration of primary bedding due to sediment remobilization and dyke intrusion.

Since the considered sediments have been deposited in a terrestrial environment under semiarid conditions, trigger events such as tsunamis or heavy storm waves as well as ice tectonics can be excluded. Also, no evidences are found for the assumption of a depositional slope indicating gravity sliding or slumping, or for an extremely rapid sedimentation. Pre-existing cracks created by deccication can be excluded as well when considering the associated deformation structures in the host sediments.

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The upward increasing degree of synsedimentary deformation in the host beds (segregation into sand and gravel lenses / ball-and-pillow structures) could indicate increasing pore water pressure towards the top, resulting from a synchronous dewatering of the whole strata. Similar structures have been described by Lützner (1994) from Upper Rotliegend conclomerates of the Eisenach Formation at the eastern margin of the same sedimentary basin (Saar-Werra Basin). They have been interpreted as a result of partially liquefaction of sand-rich conglomerates by seismic waves. This left a lag of pebble and coarse sand, after the sand fraction was segregated and squeezed out together with the pore water (Lützner, 1994: 1312). In the same area gravel-filled dykes have been reported in playa deposits of the Eisenach Formation (Heubeck, 2009). These have been attributed to artesian water initiating subvertical cracks and the intrusion of sand and grit breaking through semi-consolidated mudstones (Heubeck 2009: 53). Thus, the formation of hydraulic fractures could be an important factor in the movement of fluids through and out of low-permeable, semilithified sediments (Cosgrove, 2001).

Likewise these examples, the clastic dykes and deformational structures in the Richelsdorf Rotliegend are considered to be dewatering structures resulting from the generation of overpressure, probably due to a high-energetic event. In course of the break down of an unstable grain structure localized pressure build up within the sediment causing breaching of the overlying, partially sealing lithology. The clastic dykes are most likely expression of pressure-compensating connections between more permeable gravel horizons. After liquification of sediment, a segregation into a sandy and a gravelly fraction happened due to density differences and due to transport by fluid flow affecting the easier remobilizible material in the grain frame work, i.e. finer grain sizes. Uprising sand finally caused deformation by dykes.

The most probable trigger mechanism for these structures is a tectonic–palaeoseismic event caused by tectonic movements.

5.5 Origin of the funnel structures within the Vik member of Scania and their relationship to clastic dykes within the basement of Bornholm

Several opinions could be estimated in order to explain the formation of the enigmatic funnel structures occurring only within the Cambrian Vik sandstone in Scania (see 3.2). One possibility for their formation is seen in a dewatering process of sediments below the funnel structures (Hamberg, 1994, Lindh & Bergman, 1995, Katzung et al., 1995). If the water is able to separate from the liquefied sand and to escape upwards to reach the sea floor, a volume loss will occur, which could have caused the Vik funnel structures. A similar process is favoured by Hamberg (1994) who interpreted these structures as collapse structures subsequent to sand volcano eruptions, caused by arthesic water. In this case the escaping water should have left traces like vertical pipes, where bedding and other primary structures have been destroyed (Lowe, 1975, Owen, 1987, Obermeier, 1996). However, structures of this type have not been found, and even within the very centres of the funnel-shaped structures the primary sedimentary layering is fully preserved.

The swarm of clastic dykes in the vicinity of two of the funnel structures can also not be the result of the same dewatering process, favoured by Hamberg (1994: fig. 2) and Lindh & Bergman (1995: 221). In fact, most of the clastic dykes are older than the funnel structures and have been generated at different times (Scholz et al., 2009). These centroclinal structures must have sunk down after the deposition of the youngest sandstones exposed here, because they have all been affected and deformed.

Another possibility for the formation of the funnel structures is closely associated to the occurrence of clastic dykes within the crystalline basement below Vik. This has been already proposed and discussed by Katzung & Obst (1997) or Scholz et al. (2009). Sandstone-filled fissures within the crystalline basement are well known from many sites in Bornholm as well as from other regions (Baer et al., 1994). According to Jørgart & Nielsen (1995) or Katzung & Obst (1997) these nearly vertical fissures cut through Mesoproterozoic granites which are part of the Fennoscandian Shield. The basement of southeastern Scania consists of similar rocks and nearly the same age. These sandstone dykes cut down tens of metres below the supposed interface between the crystalline basement and the overlying Cambrian sediments (Katzung & Obst, 1997: 160). Sandstone-filled fissures within different granites are known e.g. at Gulehald near Listed (fig. 12) in the eastern part of Bornholm, or north of Vang at the northwestern coast of this island.

Fig. 12. Photograph of a Lower Cambrian sandstone dyke cutting through the crystalline basement (Mesoproterozoic Svaneke granite) at Gulehald near Listed at the northeastern coast of Bornholm. This sugtrusive clastic dyke has formed in the Lower Cambrian due to a tectonically opening fissure below a cover of marine sands at the northern rim of Tornquist ocean

The petrography of the sandstones within the fissures has some similarities with the marine lower Cambrian sandstones, which have originally covered the today exposed crystalline rocks of Bornholm (Katzung & Obst, 1997: 169). The sandstone dykes described are some millimetres up to few decimetres wide. In places, e.g. near Vang, they occur as clusters with a cummulative thickness of nearly 1 m. These dykes are thinning downwards but still of up to 18 cm width, even more than 50 m below the supposed Cambrian / basement interface.

Katzung (1995) proposed that the sandstone dykes of Bornholm are extensional fissures of tectonic origin. They have generated during lower Cambrian times due to the opening of the Tornquist Ocean between Baltica and Peri-Gondwana further to the south. All known sandstone dykes of Bornholm are completely filled-up with sand, even some extremely narrow fissures (only 0,1 mm), not much wider than the diameter of the sand grains filled in (Katzung & Obst, 1997: 165). Moreover, none of these dykes displays any sub-horizontal bedding, indicating normal sedimentation into a preexisting open fissure on the sea floor (Katzung & Obst, 1997: 170). Therefore, a formation as neptunian dykes (see below) in the sense of Jolly & Lonergan (2002) by the infill of sand into open, seawater filled gaps from above can be excluded.

In contrast, a sudden opening of fissures within the crystalline basement on the shelf, below the water level and below loose sediments, would cause strong underpressure within the newly created space, sucking sediments implosively downwards (Katzung & Obst, 1997: 170). The fluidization and the rapid infill of the sands will be mainly caused by a strong pressure difference: extremely low pressure within the opening gap below and a comperatively high pressure of the pore-water within the sand above. An increase of pore water pressure within the fluidized sediments, gravity and the load of the water column could have propulsed the injection. Thus a complete infill with sand grains of even very narrow fissures seems possible (see also Lindström, 1967, 1972 und Katzung & Obst, 1997). Most likely, the opening of the joints within the crystalline basement has caused and was accompanied by a seismic event. However, the seismic shock waves would not have been an inalienable precondition for the liquefaction of the sand.

Many tunneling projects are affected by severe accidents caused by comparable mechanisms. Some tunnels are mined in solid rocks, which are covered with unconsolidated and water saturated deposits, mostly Quaternary sand and gravel (e.g., Mikkola & Viitala, 2000). A small leakage within the roof of the tunnel may cause a sudden break-down of the hanging wall, a tunnel collapse accompanied by an implosion-like invasion of liquidized sediments from above. The failure generates at the tunnel roof, moves quickly upwards into the overlying deposits, finally attaining the earth surface, where a large sink hole subsides (e.g. Kontogianni & Stiros, 2006, Philipp, 1987). These sink holes are circular funnel-shaped structures, graben-like down-faulted along sub-concentrically bent normal faults. The dimension of these sink holes corresponds with the volume of the liquefied sands invading the cavity. The fluidization of the sediments, their violent ingression into the tunnel and the cavein on the earth's surface are only driven by the pressure difference between the artificial cavity below and the pore water within the water saturated sands above. Moreover, the generation and shape of sink holes (mining damages) above old galleries within active or abandoned mining areas is a very similar process (Scholz et al., 2009: 372).

Is it really imaginable that an opening vertical fissure within the crystalline basement could have generated the funnel-shaped structures in Vik? The island of Bornholm is situated within the Baltic Sea, less than 30 km away from the investigated area. Similar fissures are known to exist as well at different places near the southeastern coast of Scania and could

also be present in the underground of Vik. A sudden filling of a fissure with sand from above would cause a loss of volume and lead to a partial collapse of the overlying strata. This is followed by the formation of a sink hole at the sea floor and a funnel structure just below. An argument for a hypothetical linear structure below is the fact, that the three funnel structures at Vik form a more or less straight chain with a strike of 144-150° (fig. 9). According to Katzung & Obst (1997: 165) most of the sandstone dykes near Vang in northwestern Bornholm show a WNW to NW oriented strike (100-145°), and also the dykes at Gulehald near Listed in eastern Bornholm are running roughly parallel to the NW-SE striking bundle of major faults of the Scandinavian Border Zone (Scholz et al., 2009: 372: fig. 1).
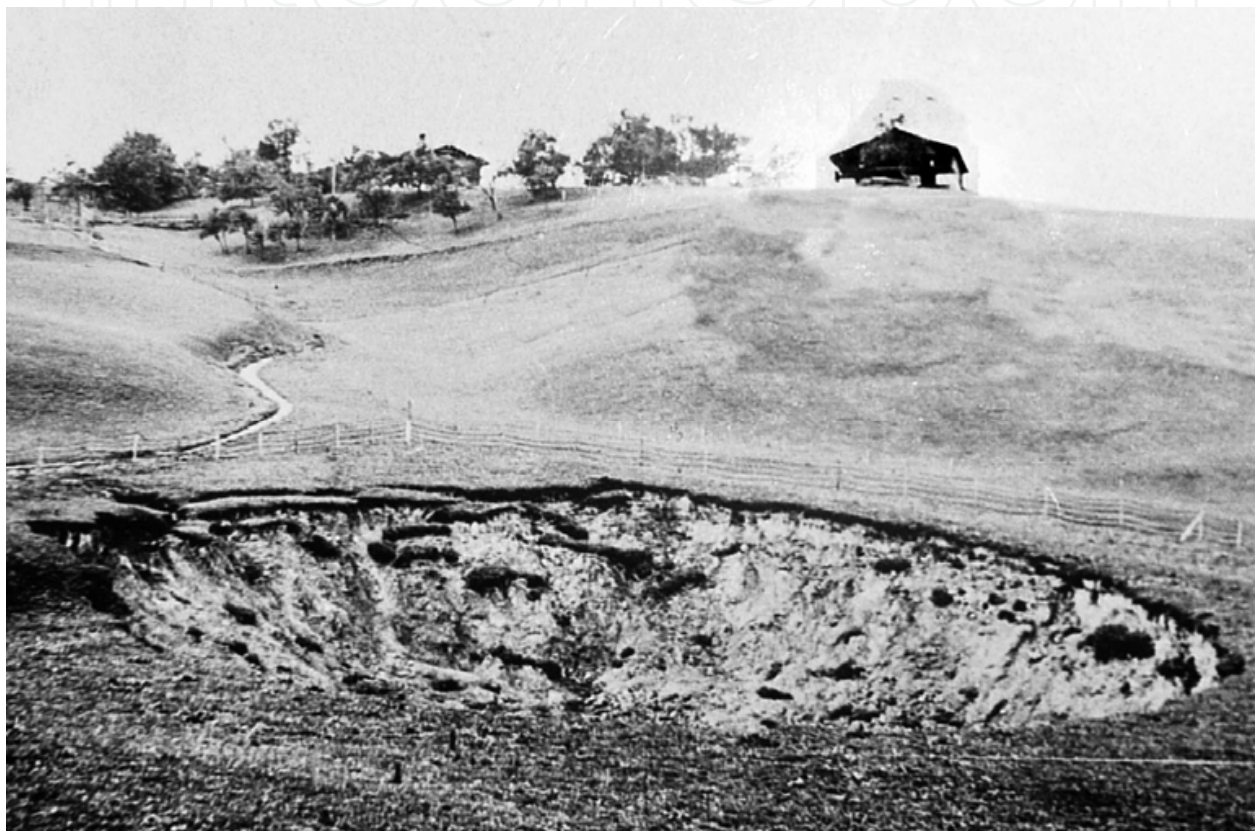

Fig. 13. Sink hole-like cave-in (diameter about 16 m) on a meadow in the Bavarian Alpine foreland near Miesbach, formed 1934 above old galleries in the underground within the Hausham area, where coal mining was active in this time. The generation of mining damages of this type is comparable with the formation of funnel structures at the bottom of the Cambrian sea above suddenly opening fissures in the basement below. Photograph taken from the Museum of coal mining in Hausham

Would the assumption be realistic that such a dyke exists in the basement rocks, regarding the volume loss of sand during formation of the funnel-shaped structures on the sea floor? The centres of the three funnel-like structures exposed south of Vik have distances of 90 to 100 m. Assuming an average distances of about 100 m, the volume missing below one of these structures would have to supply a dyke segment of about 100 m length. Thus, a simplified gap with an average width of 20 cm and a depth of 50 m could have been filled with 1000 m³ of sand, which is very close to the volume loss estimated above (for calculations see Scholz et al., 2009: 373, fig. 18).

6. The tectonic movements that could have caused the seismic shocks

The **ball-and-pillow structures** and the **synsedimentary folds** within the Oligocene and Miocene Molasse successions (fig. 1 / 1-5) have formed neither due to mass movements, nor have they been triggered by heavy storm waves, tsunamis or by the migration of hydrocarbons. A palaeoslope with inclinations sufficient for surface slides is unlikely due to the palaeogeographical situation. The same holds true for the deformational structures in the Rotliegend deposits of Hesse. We rather believe that sudden dewatering within originally soft and water saturated sediments has produced these different deformation structures, triggered by seismic shock waves in all cases (Scholz & Frieling, 2006: 371 ff., see also chapter 4.4).

The clastic dykes within the Lower Miocene Molasse succession of southern Bavaria (fig. 1 / 6), the Rotliegend of Hesse and the Lower Cambrian of southern Sweden are the results of seismic shock waves as well. This type of clastic dykes that is quickly filled with liquefied sediments, immediately after the fissure had formed by hydraulic pressure, has been called **extrusive clastic dykes** by Scholz et al. (2009: 375 ff., fig. 14). The input of sediments filling the fissures comes mostly from below. They need magnitudes of more than 5 to 6 on the Richter scale to generate (Scholz & Frieling, 2006: 377). Epicenters of such strong earthquakes have probably been located quite close. In the Molasse basin they were most likely connected with active compressive tectonics at the southern rim of or even within the Molasse Basin itself in Oligocene and Miocene times (Scholz, 2000b). These seismic shocks could have been generated at listric thrust planes of the actively prograding Alpine nappes or along blind faults of newly generating folds within the Subalpine Molasse. Within the Rotliegend succession they have been most likely connected with tectonics at deep seated faults along the Lower Permian fault-bounded sedimentary basins. The study area was surroundended by Variscan uplifts along most likely synsedimentary actice fault zones. Already in Rotliegend times the Saar-Werra Basin was dissected by NW-SE trending swells and depressions, these epirogenetic movements lasted until the Lower/Middle Zechstein (Kulick et al., 1984). Soon after deposition or during the very early stages of shallow burial of the dyke-bearing succession hydraulic fractures could have been formed by extensional tectonics initiating the segregation and uprise of a sand-water mixture as a result of tectonically induced palaeoseismic activity. The generation of the clastic dykes within the Cambrian sandstones in South Sweden was triggered by strong seismic events as well. Many of these clastic dykes branch repeatedly into different directions, most likely indicating different located epicentres as sources of the seismic shock waves at different times, but prior to the formation of the funnel structures (see below). Presumably they were generated due to tensional tectonics connected with the opening of the lower Paleozoic Tornquist Ocean in the South.

The **funnel structures** within the Cambrian sandstones in South Sweden as well are of synsedimentary origin. Different from the clastic dykes, cutting through the same sequence, however, the funnel structures were presumably not formed by seismic shock waves. They rather formed due to tensional stress associated with suddenly opening fissures in the crystalline basement immediately below the Cambrian succession, as Katzung & Obst (1997) already had suggested. Cambrian clastic dykes are known to occur in the crystalline basement of Bornholm, which is situated only 30 km off the coast of Scania in the South. The clastic dykes within the crystalline basement rocks of Bornholm and southeastern Scania are clearly different from the extrusive clastic dykes described above, regarding the mechanism of infill and the filling direction (Scholz et al., 2009: 372 f.).
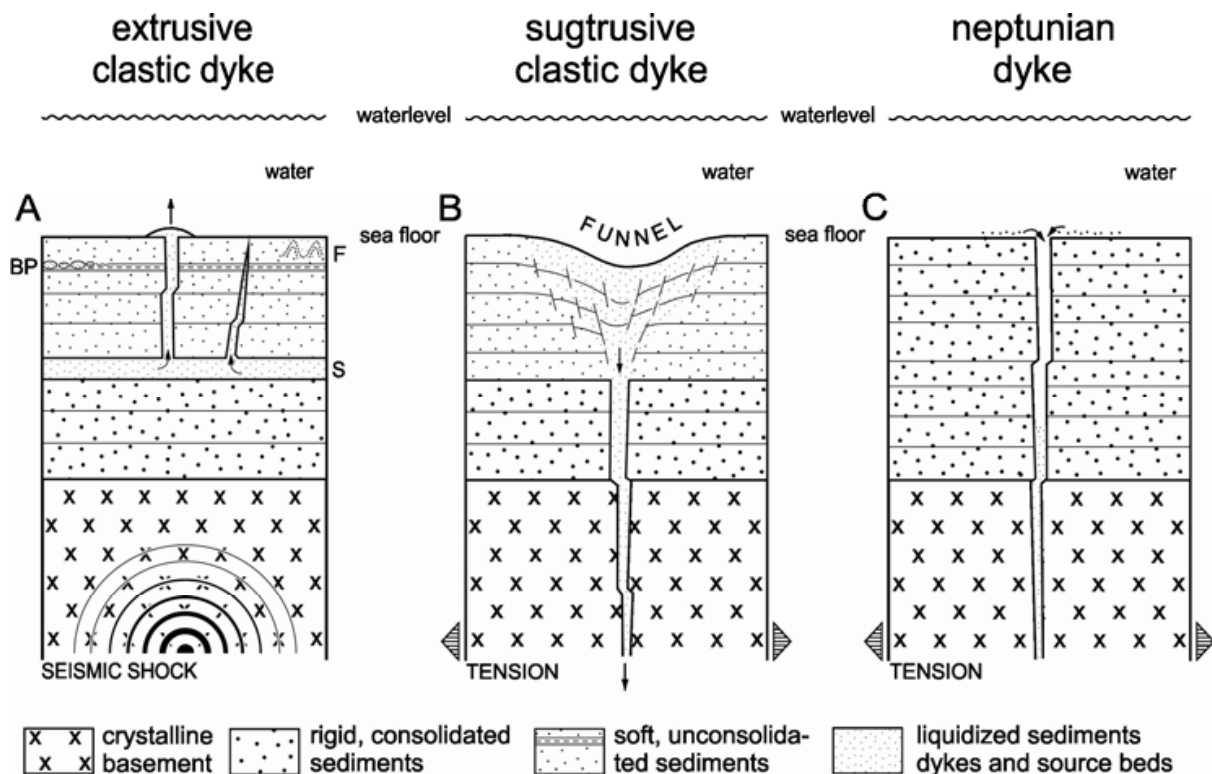
Fig. 14. Generation of ball-and-pillow structures, synsedimentary folding of bedding planes, protrusive and sugtrusive clastic dykes, neptunian dykes as well as funnel structures, demonstrated with schematic cross sections. All these structures are caused by seismic shock waves or opening fissures below the still soft and not yet lithified sediments. They are distinguishable regarding host rocks, velocity, direction and mechanism of infill.
A) Fissures within not yet lithified sediments opening due to seismic shocks are filled rapidly from below and called „**extrusive clastic dykes**", occurring in Cambrian, Permian and Neogene strata. S = source bed, BP: ball-and-pillow structure, F: synsedimentary fold.
B) Tensional fissures opening within crystalline basement rocks are filled rapidly from above and called „**sugtrusive clastic dykes**", known from the basement of Bornholm. Fissures like this cause funnel-like cave-ins at the sea floor, occurring in Cambrian strata at Scania.
C) Tensional opening crevasses at the sea floor cutting through lithified sediments. They are filled very slowly from above and are called **neptunian dykes**. Fissures of this type have been observed e.g. in Jurassic deposits of the Alps

They must have been filled anyway from above, because some of the sand-filled fissures cut down at than least 50 m deep into the crystalline rocks. Tectonic extension may have created suddenly opening fissures within the crystalline basement. Immediately opening gaps below the sedimentary succession, deposited above the crystalline basement will have caused a kind of underpressure, sucking the not yet consolidated sands implosively downwards (Katzung & Obst, 1997: 170). The rapid infill of the sands will have been mainly caused by the pressure difference between the opening gap and the pore-water within the liquefied sand. The infill may be supported by gravity and by the load of the water column. Thus even very narrow fissures have been completely filled-up with grains (Katzung & Obst, 1997). Seismic schocks will probably accompany the sudden opening of fissures within the basement rocks, however, they are not inalienable to fluidize the sands. This type of

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clastic dykes that have been quickly filled with liquefied sediments from above, immediately after the fissure had formed, has been called **sugtrusive clastic dykes** by Scholz et al. (2009: 376 f., fig. 14). The generation of this type of clastic dykes seems to be spatially as well as chronologically closely related with the formation of the funnel structures.

The dimension of such a covered, downwards filled clastic dyke below Vik can be estimated by the degree of volume loss within the funnel structures. Their generation is most likely directly connected with the tensional stress caused by the opening of the lower Paleozoic Tornquist Ocean as well (Scholz et al., 2009: 373).

In contrast, dykes which originally have been open cracks at the sea floor are called **neptunian dykes** by Jolly & Lonergan (2002) (fig. 14). These synsedimentarily generated fissures may initially have formed due to tectonic forces as well, but have been opened in many cases by gravity due to a pronounced submarine relief. Subsequently these open gaps containing sea water have been filled-up sucessively from above within a long time span, grain by grain. However, none of the sandstone dykes described above generated as a neptunian dykes. Dykes of this type are known to occur in Upper Triassic to Jurassic carbonates of the Northern Calcareous Alps (Schlager, 1969, Schöll & Wendt, 1971, Leuprecht & Moshammer, 2006). Similar neptunian dykes are also known e.g. from Devonian reef limestones of Bohemia (Chlupáč, 1967 and 1993) or from the Lower Cretaceous Schrattenkalk in the Helvetian mountains (Scholz, 1983). For details of the naming of the different types of sedimentary dykes see Scholz et al. (2009: 375 f.).

7. Conclusions

Within Oligocene and Miocene sandstone-dominated marine and terrestrial Molasse successions of southwestern Bavaria, special deformations of the primary sedimentary structures occur, which obviously do not have caused by the later tectonic deformation of the rocks. Besides of folds and ball-and-pillow structures clastic dykes occur here. Similar clastic dykes and deformation structures of the strata occur also in Lower Permian terrestrial Rotliegend successions from the Richelsdorf Mountains in the Hesse depression of Central Germany. Moreover, clastic dykes are also reported from Lower Cambrian sandstones of the Vik Member at the eastern coast of Scania in southern Sweden. As over and underlying successions are not affected by similar deformations they are clearly of synsedimentary origin. They were caused most likely by strong seismic shock waves affecting the still soft and not yet lithified sediments. In Scania the upwards filled „extrusive clastic dykes" occur together with large enigmatic centroclinal funnel-shaped deformation structures with diametres of more than 40 m, affecting this more or less flat lying shallow water sedimentary succession. Most likely, their formation is closely associated with suddenly opening tensional fissures in the crystalline basement below the Cambrian succession, sucking down soft and water saturated sands from the bottom of the Cambrian sea. A couple of such „sugtrusive clastic dykes" is known within the crystalline basement of Bornholm. Clastic dykes and other deformation structures within the Molasse succession were triggered by seismic shock waves associated with Neoalpine compressive tectonics at the southern rim of the Molasse basin. The structures in the Lower Permian successions have been formed due to seismic shocks most likely caused by tensional tectonics at major normal faults bordering the Rotliegend basins. Both, the funnel structures and the clastic dykes in Scania presumably formed due to tensional stress connected with the opening of the lower Paleozoic Tornquist Ocean.

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