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A slope-apron succession filling a piggyback basin: the Tannheim and Losenstein Formations (Aptian – Cenomanian) of the eastern part of the Northern Calcareous Alps (Austria)

MICHAEL WAGREICH[1]

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Abstract

The Tannheim and Losenstein Formations of the northernmost tectonic units (Allgäu-Ternberg-Frankenfels nappe system) of the eastern part of the Northern Calcareous Alps comprise a coarsening upward deep-water sequence, which formed on top of the Cretaceous orogenic wedge of the Eastern Alps. Facies types include hemipelagic marlstones and shales, thin-bedded siltstone-marl intercalations, thin-bedded and thick-bedded turbidites, massive/laminated sandstones, pebbly sandstones/mudstones, deep-water conglomerates and muddy slumps. Depositional environments comprise channels, interchannel areas, debris flows and abundant muddy slumps. The coarsening-upward trend records progradation of slope-aprons into a pelagic basin due to increasing tectonic activity in the hinterland.

The Tannheim-Losenstein basin is interpreted as an early, deep-marine piggyback basin filled by deep sea clastics in front of advancing thrust sheets. The narrow basin extended over a considerable horizontal distance as suggested by similarities in timing of deposition and tectonism and clast composition. Heavy mineral assemblages comprise high amounts of zircon and tourmaline and varying amounts of chrome spinel, blue sodic amphiboles and chloritoid. Tectonic subsidence of about 460 m can be modelled by flexural loading of thrust sheets. Basin formation marks the transition from a passive continental margin to an active, transpressional margin along the northern boundary of the Austroalpine microplate.

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Eine Slope-Apron-Abfolge in einem Piggyback-Becken: die Tannheim- und Losenstein-Formationen (Aptium-Cenomanium) im Ostteil der Nördlichen Kalkalpen (Österreich)

Zusammenfassung

Die Tannheim- und Losenstein-Formationen der nördlichsten tektonischen Einheiten (Allgäu-Ternberg-Frankenfels Deckensystem) im Ostteil der Kalkalpen zeigen eine coarsening-upward Abfolge, die auf dem kretazischen Orogenkeil der Ostalpen abgelagert wurde. Folgende Faziestypen treten auf: hemipelagische Mergel und Tonsteine, dünnschichtige turbiditische Siltstein-Mergel-Wechsellagerungen, dünnbankige und dickbankige klassische Turbidite, massive und laminierte Sandsteinbänke, geröllführende Tone und Sandsteine, Tiefwasserkonglomerate und pelitdominierte Rutschmassen. Als Ablagerungsbereich lassen sich Rinnen, Zwischenrinnenbereiche, submarine Schuttströme und Rutschmassen rekonstruieren. Der coarsening-upward Trend zeigt die Progradation von linearen Hangfächern (slope apron) in ein pelagisches Becken, wahrscheinlich auf Grund erhöhter tektonischer Aktivität im Hinterland.

Das Tannheim-Losenstein-Becken kann als früher, tiefmariner, siliziklastischer Piggyback-Trog vor der Front überschiebender höherer kalkalpiner Decken interpretiert werden. Der schmale Trog zeigt eine beträchtliche Längserstreckung, wie die einheitliche Sedimentation und Materialzusammensetzung zeigt. Die Schwermineralspektren sind durch Zirkon und Turmalin sowie wechselnde Gehalte von Chromspinell, blaue Alkaliamphibole und Chloritoid gekennzeichnet. Die tektonische Subsidenz des Beckens von etwa 460 m kann mit einem flexurhaften Auflastmodell durch höhere Decken modelliert werden. Die Beckenbildung markiert den Übergang von einem passiven Kontinentalrand zu einem aktiven, transpressiven Rand am Nordrand der ostalpinen Mikroplatte.

1. Introduction

The Lower- to lower Upper Cretaceous deposits in the Northern Calcareous Alps (NCA) are characterized by a succession of fine-grained carbonates, shales, marlstones, sandstones and conglomerates. The significance of these sediments for the deformation history of the eastern NCA was already recognised by GEYER (1909). The superposition of siliciclastics above pelagic limestones is interpreted as a consequence of Cretaceous "Eo-Alpine" deformation within the Eastern Alps (e. g. GAUPP, 1982; WINKLER, 1988, 1996; FAUPL & WAGREICH, 1992a, 2000; VON EYNATTEN, 1996). Lower- to lower Upper Cretaceous deposits of the NCA were strongly influenced by folding and thrusting (e. g. GAUPP, 1982; LEISS, 1992; MAY & EISBACHER, 1999). The lithostratigraphy and the biostratigraphy of these sediments were clarified in the western NCA by GAUPP (1980, 1982, 1983) and WEIDICH (1984, 1990) and in the eastern NCA by KOLLMANN (1968) and LÖCSEI (1974). A detailed tectono-sedimentary basin model is still missing.

This paper describes the sedimentology of Aptian to Lower Cenomanian formations in the eastern part of the NCA, including the type area of the Losenstein Formation in Upper Austria. Based on these data, a depositional model and the evolution of this sedimentary basin are discussed within the geodynamic framework of the Eastern Alps and its continuation into the Western Carapthians (WAGREICH, 2001).

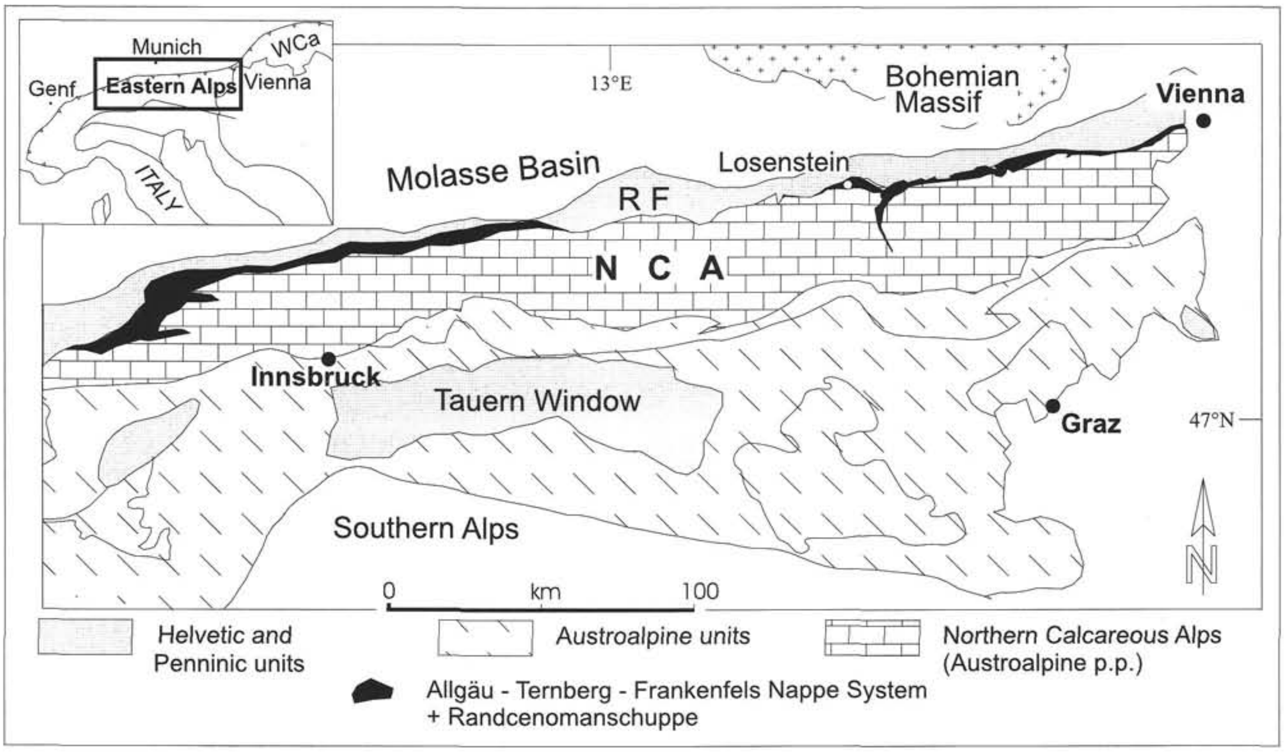

Fig. 1
Sketch of the Eastern Alps including the location of the Allgäu-Ternberg-Frankenfels Nappe System and the Randcenomanschuppe as parts of the Northern Calcareous Alps (NCA). RF – Rhenodanubian Flysch Zone; WCa – Western Carpathians.

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2. Geological Setting

The thrust stack of the NCA, as a part of the Austroalpine tectonic unit (Fig. 1), evolved during Late Jurassic to Cretaceous times due to northwestward thrusting (EISBACHER et al., 1990; LINZER et al., 1995, 1997; PERESSON & DECKER, 1997). Compressional deformation in mid-Cretaceous (Aptian to Cenomanian) times is commonly attributed to the onset of oblique convergence between the Penninic Ocean to the north and the Austroalpine microplate to the south, at the northern leading edge of the Adriatic microplate (FAUPL & WAGREICH, 1992a, 2000; VON EYNATTEN, 1996; WINKLER, 1996). During this time the NCA were situated along the northern margin of the Austroalpine microplate, and became part of the Cretaceous to Early Tertiary "Eo-Alpine" orogenic wedge (WAGREICH & FAUPL, 1994). Subsequent polyphase Tertiary tectonism (e. g. LINZER et al., 1995, 1997; DECKER & PERESSON, 1996; PERESSON & DECKER, 1997) strongly deformed this continental margin and largely destroyed the geometries of Cretaceous basins.

On top of the structurally deepest thrust sheets of the NCA, the Tannheim Formation and the Losenstein Formation filled an elongated synorogenic trough during the mid-Cretaceous. In the eastern part of the NCA, deposits of this "Tannheim-Losenstein basin" can be found in the "Randcenomanschuppe", the Frankenfels-Ternberg nappe system and the northern margin of the Lunz-Reichraming Nappe System (Fig. 1 and 2). Due to Tertiary deformation, outcrops of these deposits occur within faulted and partly overturned, narrow synclines (see e. g. EGGER, 1988; ZIMMER & WESSELY, 1996). The burial history of these formations was characterized by temperatures below 200 °C (GAUPP & BATTEN, 1985; WAGREICH & SACHSENHOFER, 1999).

Within these units, Lower Cretaceous pelagic limestones ("Aptychus limestones") and marl-limestone beds of the Schrambach Formation are overlain by marlstones and calcareous shales of the Tannheim Formation, followed by sandstones and conglomerates of the Losenstein Formation (KOLLMANN, 1968; Fig. 2). Several composite sections (S1-S3) in the eastern part of the NCA were investigated within the northernmost thrust sheets of the NCA. The most complete section in the Losenstein-Stiedelsbach area (Fig. 3) was described as the type section of the Losenstein Formation by KOLLMANN (1968) from two tributaries of the Stiedelsbach, starting at 550 m (Austrian map sheet 69, BMN coordinates re585677 / ho310063 to re535736 / ho310091; re535714 / ho310020 to re535852 / ho309560) within the Losenstein syncline of the Ternberg nappe (EGGER, 1988). The syncline continues into the Höllleitengraben/ Pechgraben area (EGGER et al., 2000) further to the east (Fig. 3). Additional sections were investigated in the Frankenfels area of the Frankenfels nappe (LÖCSEI, 1974; Hausstein and Kleinbernreith, BMN re674356 / ho316213 to re673915 / ho316 708) and in the Dachsgraben within the Rettenbach syncline of the Frankenfels nappe to the east of Großraming (see also KENNEDY & KOLLMANN, 1979; BMN re545242 / ho307528 to re545006 / ho307659). Data are also included from other localities, e. g. the Weyer Arc (Randcenomanschuppe and Frankenfels Nappe near Brunnbach, e. g. Hagauergraben, Hanslgraben, Draxlgraben and Leerensackgraben; LÖCSEI, 1974; EPPEL, 1990; FAUPL & WAGREICH, 1992b), the Bodinggraben near Frankenfels (DUMEIRY, 1990) and from the area of Kaumberg/ Lower Austria (PLÖCHINGER & SALAJ, 1991).

3. Lithostratigraphy, biostratigraphy and chronostratigraphic correlations

The lithostratigraphic subdivision of synorogenic sediments from the Aptian to Cenomanian of the NCA is based on ZACHER (1966), KOLLMANN (1968), TOLLMANN (1976), GAUPP (1980, 1982) and WEIDICH (1984). Biostratigraphic

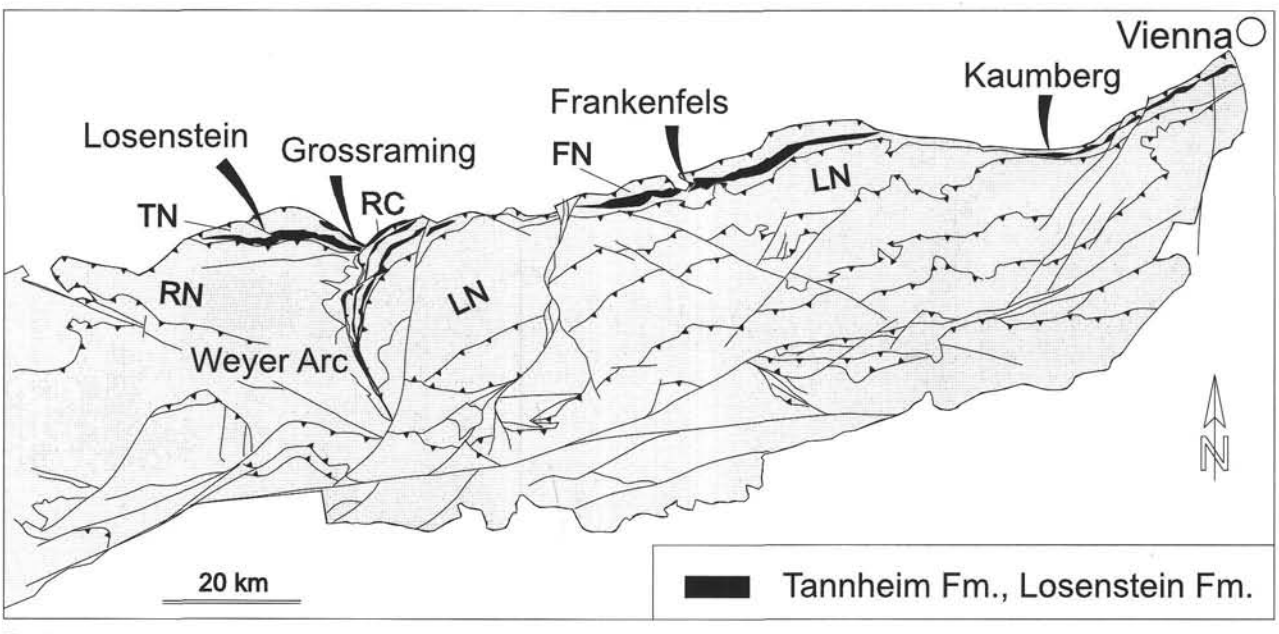

Fig. 2
Simplified structural map of the eastern part of the Northern Calcareous Alps (modified from PERESSON & DECKER, 1997). Major outcrop areas of the Tannheim and Losenstein Formations are indicated in black. TN – Ternberg Nappe, RN – Reichraming Nappe, RC – Randcenomanschuppe, LN – Lunz Nappe, FN – Frankenfels Nappe.

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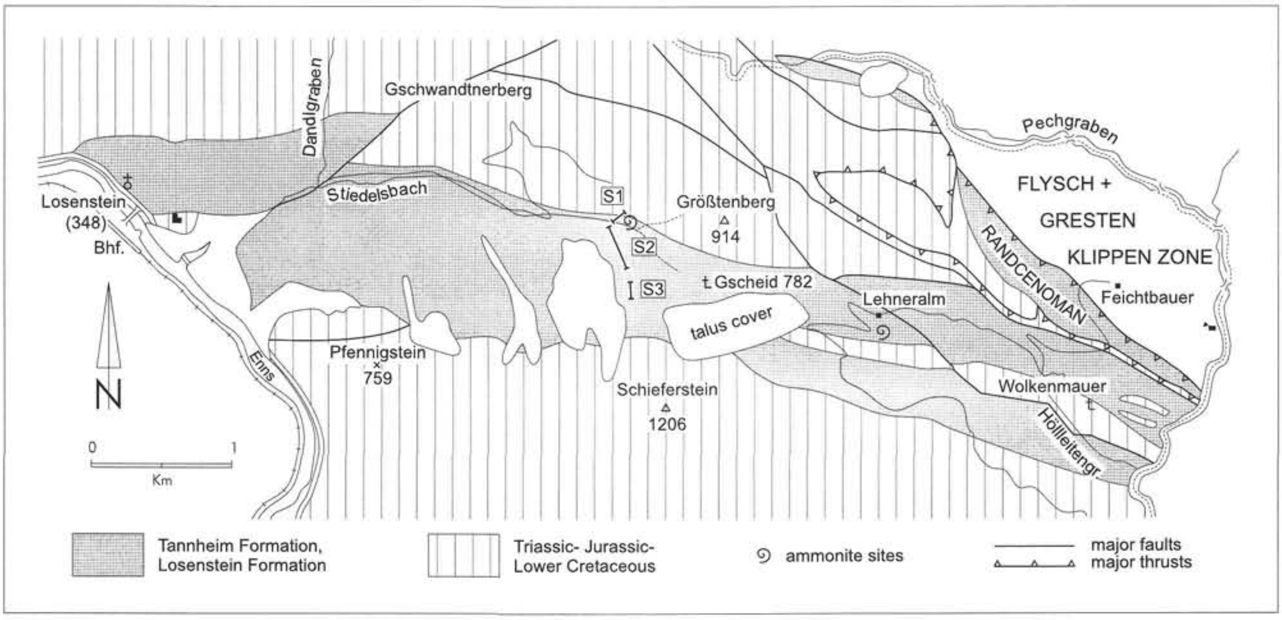

Fig. 3
Sketch map of the Losenstein syncline between Losenstein and Pechgraben, including positions of measured sections, including the type section for the Losenstein Formation (S2, S3), and ammonite sites in the Tannheim Formation at section S1 in the Stiedelsbachgraben (KENNEDY & KOLLMANN, 1979) and near Lehneralm/Höllleitengraben (EGGER et al., 2000).

data and chronostratigraphic correlations can be found in KOLLMANN (1968), RISCH (1971), KENNEDY & KOLLMANN (1979), IMMEL (1987), KAISER-WEIDICH & SCHAIRER (1990), WEIDICH (1990), and KENNEDY et al. (2000).

The lithostratigraphic term Tannheim Formation ("Tannheimer Schichten") was defined in the Lechtal/Arlberg area in the western NCA by ZACHER (1966), comprising grey, black and dark red marls and shales without conglomerates, conformably overlying limestones and marly limestones of the Schrambach Formation ("Neokom-Aptychenschichten" of ZACHER, 1966). The Tannheim Formation commonly is 10 to 30 meters in thickness, although maximum thicknesses of several tens of meters are reported from the western NCA (ZACHER, 1966).

The age of the Tannheim Formation in the eastern NCA ranges from Late Aptian to Middle/Late? Albian (WEIDICH, 1990). For the sections of the Losenstein and Großraming area, ammonites indicate an Early Albian age for a black shale interval (*Leymeriella tardefurcata* Zone; KENNEDY & KOLLMANN, 1979; IMMEL, 1987; KENNEDY et al., 2000). Based on planktonic foraminifera, WEIDICH (1990) proved a Late Aptian – Early Albian age (*Hedbergella planispira*-Zone) for the Tannheim Formation in the Losenstein syncline (Fig. 4). In the Weyer Arc area (Hanslgraben SW Brunnbach), the ammonite *Hamites* cf. *simplex* D'ORBIGNY (det. H. SUMMESBERGER) was found within the Tannheim Formation, indicating a Middle?/Late Albian age for the upper part of the Tannheim Formation. The upper boundary of the Tannheim Formation is diachronous (GAUPP, 1980, 1982; TOLLMANN, 1976), probably as a consequence of diachronous onset of turbidite sedimentation and variable incision depth of coarse channel-fills of the Losenstein Formation into the relatively soft shales.

KOLLMANN (1968) defined the Losenstein Formation ("Losensteiner Schichten") in the Stiedelsbachgraben E of Losenstein. The lower boundary of the Losenstein Formation is defined by the first occurrence of sandstones and conglomerates above the shales of the Tannheim Formation. GAUPP (1980) suggested the first occurrence of >5 cm thick sandstone beds as the definition for the base. In the eastern part of the NCA no younger, overlying strata are known from the synclines. The upper boundary of the Losenstein Formation is represented in most cases by a fault or thrust plane. In contrast, in the western NCA, breccias of the Cenomanian Branderfleck Formation are present conformably above the Losenstein Formation (GAUPP, 1980).

The Losenstein Formation has a stratigraphical range from Middle/Late Albian to early Early Cenomanian according to foraminiferal assemblages reported by KOLLMANN (1968), LÖCSEI (1974) and WEIDICH (1990). WEIDICH (1990) indicated the presence of the following foraminiferal zones within the Losenstein Formation of the Losenstein area (Fig. 4): the *primula*-Zone (Early to Middle Albian), the *ticinensis*-Zone (Middle to Late Albian; including the *raynaudi/breggiensis*-Zone and the *subticienensis/ticinensis*-Zones of WEIDICH, 1990: 74-75), the *appenninica*-Zone (Late Albian), and the *brotzeni*-Zone (Late Albian to Early Cenomanian). Most of the samples contain assemblages of the *ticinensis*- and *appenninica*-Zone, indicating a Late Albian age for the greater part of the succession, based on the chronostratigraphic correlations of CARON (1985) and BRALOWER et al. (1995). Based on the common occurrence of *Rotalipora brotzeni* (SIGAL) in a few samples of the Stiedelsbach area an early Early Cenomanian age for the uppermost part of the Losenstein Formation was inferred by WEIDICH (1990: 57). The base of the Cenomanian as proposed by the "2nd International Symposium on Cretaceous Stage Boundaries" has been placed at the first occurrence of *Rotalipora globotruncanoides* SIGAL, immediately below the first occurrence of the ammonite *Mantelliceras mantelli* (TRÖGER & KENNEDY, 1996). *R. globotruncanoides* is regarded as a synonym of *Rotalipora greenhornensis* (MORROW) by CARON (1985). *R. greenhornensis* was reported by KOLLMANN (1968) and LÖSCEI (1974) from the Losenstein

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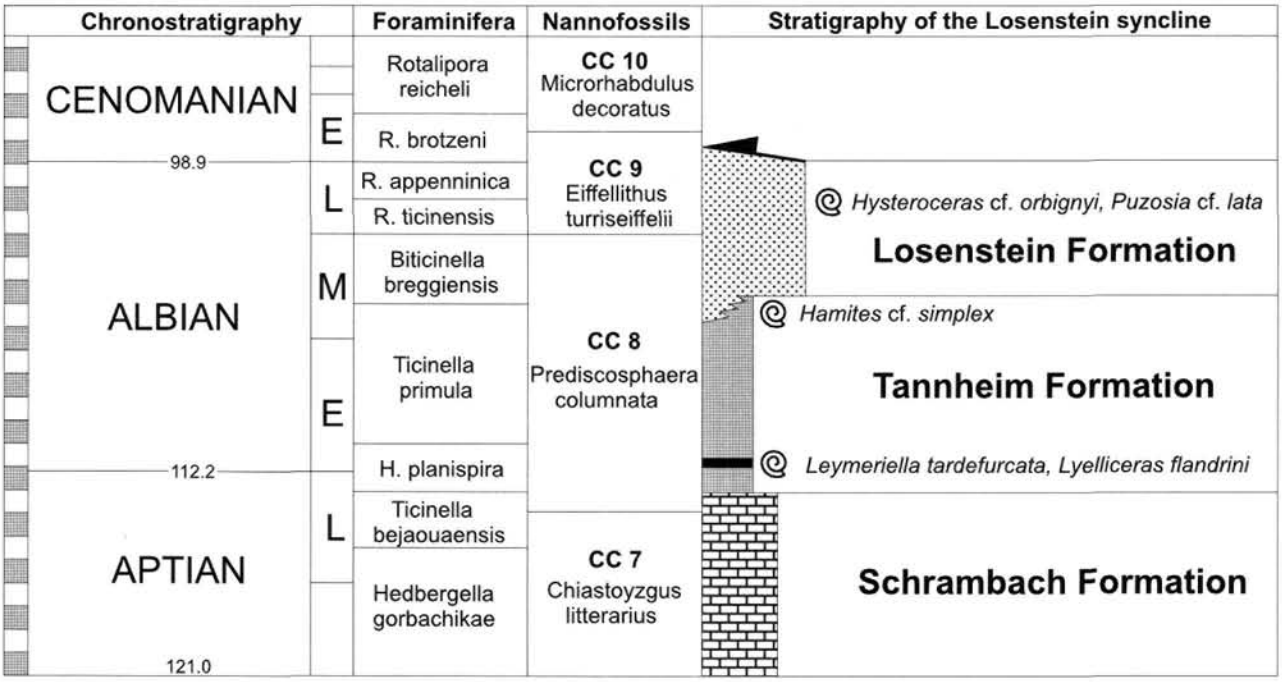

Fig. 4
Stratigraphy of the Aptian to Cenomanian of the Losenstein syncline. Chronostratigraphic correlations based on GRADSTEIN et al. (1995); planktonic foraminiferal and nannofossil standard zones based on PERCH-NIELSEN (1985) and BRALOWER et al. (1995). Positions of ammonite faunas are indicated. The base of the Albian is placed at the base of the *Leymeriella tardefurcata* ammonite Zone according to KENNEDY et al. (2000).

area, and probably included by WEIDICH (1990) into his species concept of *Rotalipora brotzeni*. Therefore, an Early Cenomanian age for the uppermost part of the Losenstein Formation is probable according to the presence of *R. globotruncanoides*, although *R. globotruncanoides* was not reported from the sections by KOLLMANN (1968) and WEIDICH (1990).

Ammonites and nannofossils provide additional biostratigraphic data. Calcareous nannofossil samples from the Losenstein Formation indicate standard zones CC8-CC9 (Fig. 4) in the standard zonation of PERCH-NIELSEN (1985), although nannofossil assemblages of the Losenstein Formation are generally poor and show significant reworking. The first occurrence of *Eiffellithus turriseiffelli*, marking the base of CC9 (Late Albian), was found within the lower part of the Losenstein Formation in the Stiedelsbach section. Fragments of two ammonites, *Hysteroceras* cf. *orbignyi* (SPATH) and *Puzosia* cf. *lata* SEITZ from the Höllleitengraben indicate the early part of the Late Albian (*Mortoniceras inflatum*-Zone, *orbignyi*-Subzone, COOPER et al., 1977). In addition, KOLLMANN (1976, 1978, 1979, 1982) described a rich Albian gastropod fauna, bivalves and a few corals from the Losenstein Formation.

Based on the chronostratigraphic scheme of GRADSTEIN et al. (1995) the time interval for sedimentation of 12 to 30 m of the Tannheim Formation (Late Aptian-Middle Albian) in the eastern NCA is about 7-11 Ma (maximum range 115-104 Ma). This indicates low sedimentation rates of about 1-4 mm/1000 a. The overlying Losenstein Formation with an age of Middle Albian to Early Cenomanian accounts for a time span of about 5-9 Ma (maximum range 106-97 Ma). Given a composite thickness of about 250 to 350 meters mean sedimentation rates of about 25-70 mm/1000 a can be reconstructed.

4. Facies analysis

Facies associations were classified according to the classical turbidite facies models of WALKER & MUTTI (1973), MUTTI & RICCI LUCCHI (1975) and WALKER (1978), with modifications according to PICKERING et al. (1989). Local subfacies are expressed e. g. as G and G'. The basic criteria for classification are grain size, matrix content, bed thickness, and sedimentary structures. The nomenclature of facies and subfacies follows WALKER & MUTTI (1973) and MUTTI & RICCI LUCCHI (1975), with the exception of facies D2.1 and D2.3, for which only the more detailed classification of PICKERING et al. (1989) was applicable. Keeping in mind the complexity of submarine fan models (e. g. SHANMUGAM & MOIOLA, 1988; READING & RICHARDS, 1994, SHANMUGAM, 2000; STOW & MAYALL, 2000), the number of distinguished facies associations was kept low to facilitate interpretations. In general, the investigated successions display a coarsening-upward cycle, starting with shales of the Tannheim Formation and ending in coarse conglomerates and slumped beds of the Losenstein Formation (Fig. 5 and 6). Facies are described from fine to coarse, more or less in the order of their appearance within the coarsening-upward cycle.

4.1 Marlstones, silty marlstones and calcareous shales (facies G, G')

Grey to olive-green marlstones to limy shales (facies G) characterize the lower part of the successions (Fig. 5 and Fig. 7a), especially within the Tannheim Formation. Organic carbon-rich, dark grey to black calcareous shales and dark red to brownish marlstones are minor in abundance. The marlstones and calcareous shales can be classified as foraminiferal mudstones to wackestones, commonly with

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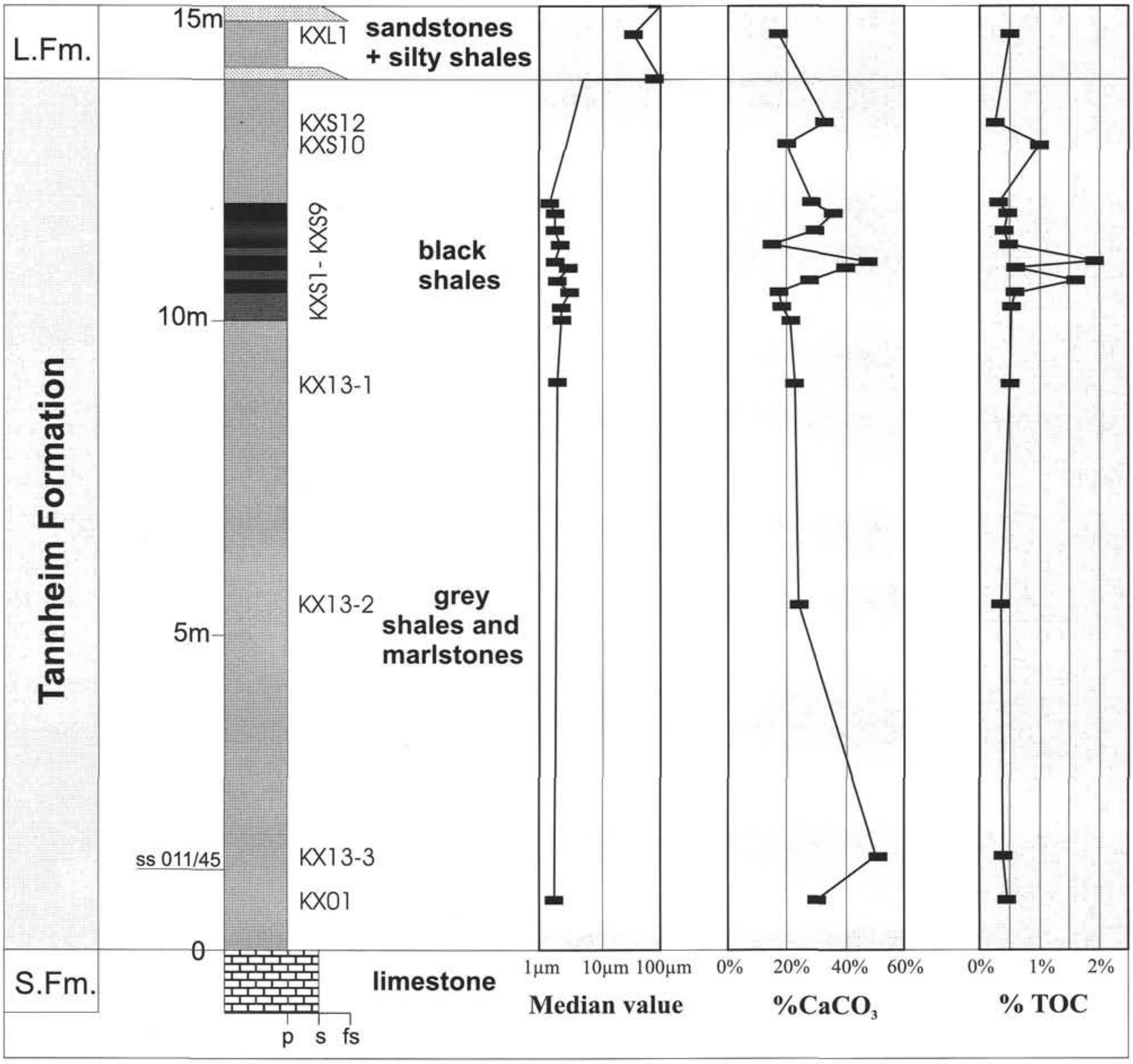

Fig. 5
Section of Tannheim Formation including black shale interval in the Stiedelsbach section S1 (see Fig. 3 for location), grain size median values (measured with Sedigraph ET5100), percentages of carbonate and total organic carbon (measured with Leco carbon analyzer). L.Fm. – Losenstein Formation; S.Fm. – Schrambach Formation; KX – samples; p, s, fs see Fig. 6. For details of black shale interval see WAGREICH & SACHSENHOFER (1999).

small *Chondrites*-type burrows and rare *Planolites*-tubes. Intervals of black calcareous shales show preserved lamination without bioturbation and ca. 5 mm thick laminae with mass occurrences of ammonites (KENNEDY & KOLLMANN, 1979). The calcium carbonate contents varies between 13 and 66% (12 samples, Fig. 5). Grain size analyses performed on the siliciclastic fraction of the marlstones indicate predominance of the fine silt- to clay size fraction (median values between 1.7 and 3.5 μm). The 10% percentile remains constantly below 10 μm. The average TOC-content (total organic carbon) of the marlstones of the Tannheim Formation is about 0.5%, but dark grey or black intervals show TOC-values up to 2% (WAGREICH & SACHSENHOFER, 1999).

The gradation from the Tannheim into the Losenstein Formation and shaly intervals in the Losenstein Formation (Fig. 6) comprise dark to medium grey clayey fine siltstones and sandy clayey marlstones (facies G'). The calcium carbonate content varies between 22 and 48% (5 samples). The marlstones have a crudely bedded appearance, but no laminae are preserved, and bioturbation is common. Median grain size values are around 20 μm (medium silt), with clay contents between 5 to 22%. Sand percentages are below 5%. More massive intervals may be due to bioturbation. No obvious grading or bed thickness cyclicity was recognized in the shaly intervals.

Interpretation

Bedded and laminated pelitic deposits were classified into facies G according to WALKER & MUTTI (1973). G' is

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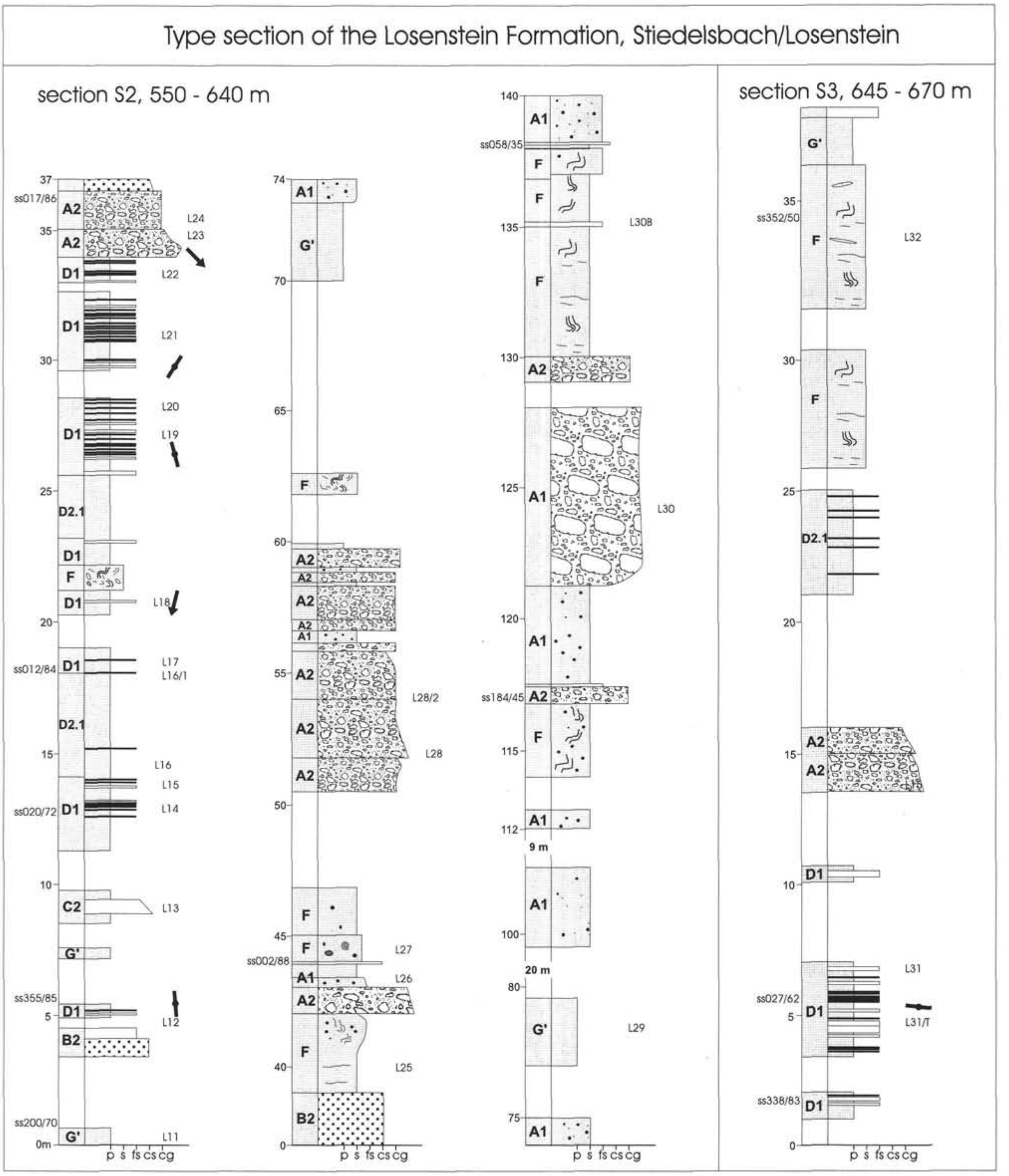

Fig. 6
Type section of the Losenstein Formation (KOLLMANN, 1968), section 2 and 3, southern tributary of the Stiedelsbachgraben east of Losenstein. Arrows indicate unipolar paleocurrent directions (mainly flute casts), lines indicate bipolar paleocurrent data. White areas comprise covered intervals. Grain size scale: p – pelites, s – pelitic slumps, fs – fine sandstone, cs – coarse sandstone, cg – conglomerates. L-numbers indicate sample locations.

recognized as a local subfacies of facies G. In the detailed classification scheme of PICKERING et al. (1989) this facies corresponds to facies group G1 (biogenic oozes and muddy oozes), although transitions into facies class E2 (organized muds) exist.

The marlstones and calcareous shales of facies G and G' are interpreted to represent "normal" marine deep-water deposits, i.e. hemipelagites (STOW & PIPER, 1984), that consist of a mixture of a biogenic pelagic carbonate (mainly calcareous nannoplankton and planktonic foraminifera),

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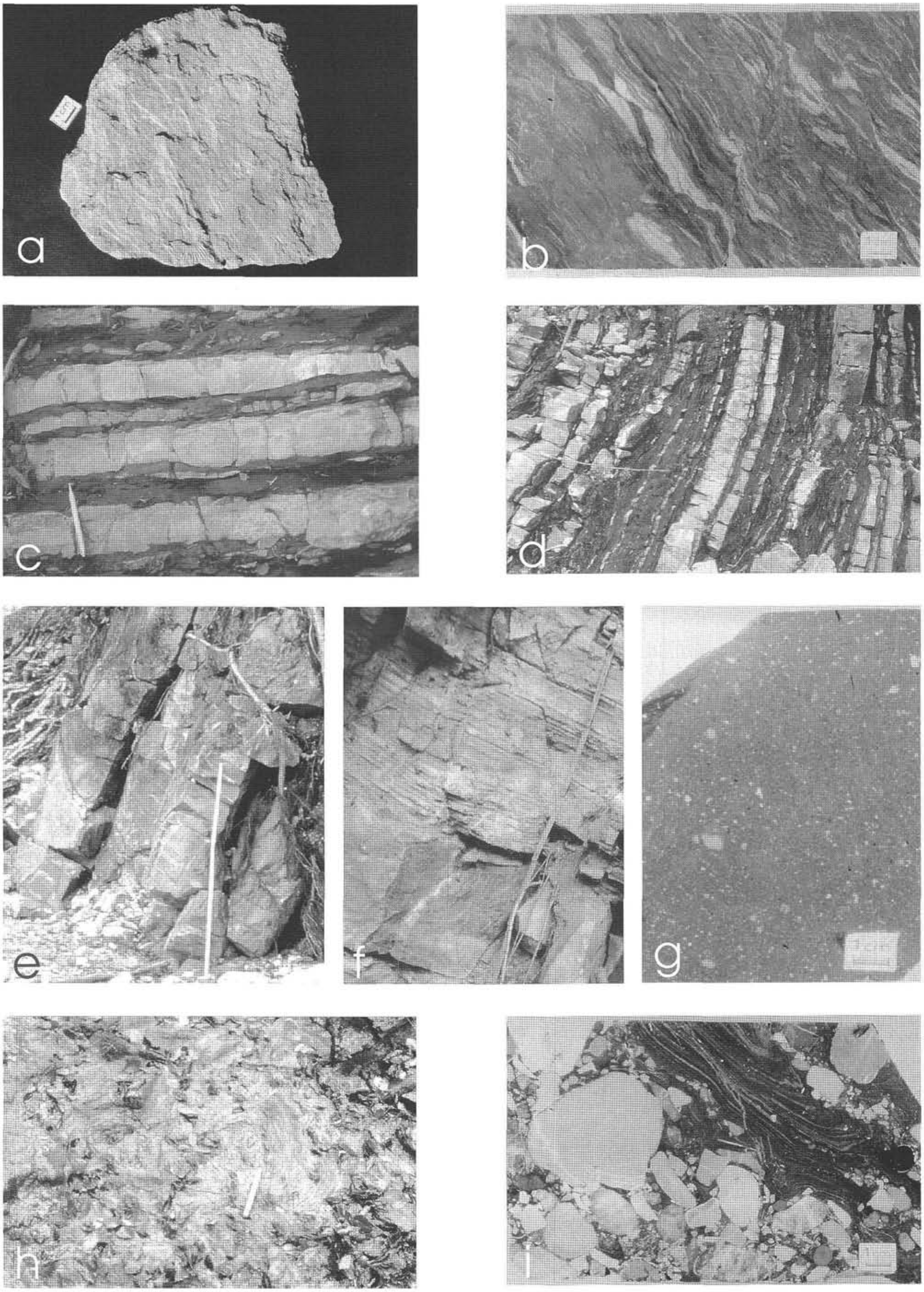

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with siliciclastic fine silt and clay, and organic carbon. Depositional processes include settling out of the water column and gravity-driven dilute bottom water currents. Based on the high content in planktonic foraminifera, bathyal depositional depths of at least a few hundreds of meters were estimated by WEIDICH (1990). The low sedimentation rates of about 1-4 mm/1000 within the Tannheim Formation are in good accordance with a pelagic-hemipelagic mode of deposition (e. g. STOW & PIPER, 1984). Facies G' is interpreted as a coarser type of hemipelagics in a more proximal slope setting, but deposition of some beds by muddy turbidity currents cannot be ruled out.

Dark grey to black shale intervals represent deposition under dysaerobic conditions based on the presence of significant bioturbation and foraminiferal assemblages containing a few benthic foraminifera (WAGREICH & SACHSENHOFER, 1999). The organic matter, dominated by vitrinite and inertinite, was mainly derived from terrestrial plants. A cyclically varying marine production resulted in organic carbon-peaks, characterized by high amounts of lamalginite. The positive correlation of organic carbon and carbonate reported by WAGREICH & SACHSENHOFER (1999) suggests that the production and preservation of organic matter was coupled to the production of planktonic carbonate ("production cycles", WAGREICH & SACHSENHOFER, 1999). According to the presence of *Leymeriella tardefurcata* and the organic facies, the black shales of the Tannheim Formation can be correlated to the oceanic anoxic event OAE 1b ("niveau Paquier", e. g. KENNEDY et al., 2000).

4.2 Thin-bedded siltstone-marlstone intercalations (facies D2.1, D2.3)

Continuous and discontinuous siltstone beds average 0.5 cm thick (Fig. 7b). They display parallel and small-scale ripple laminae, and graded bedding. Small-scale load casts are present at the base of some beds. Upper contacts are often wavy. Some siltstone beds may be classified as Tce partial Bouma sequences, others include ripple-tops, scoured load-cast bases, thin lenticular silt laminae, discontinuous and wispy convolute silt laminae. Siltstone beds form stratal packages up to 1,5 m thick. No obvious cyclicity can be observed within these packages, but they are gradational into thin-bedded sandstones.

Interpretation

These beds can be classified into facies D2.1 and D2.3 of PICKERING et al. (1989) as graded-stratified silt and thin regular silt and mud laminae. They resemble the thin-bedded silt and mud turbidite facies of STOW & PIPER (1984). Although no complete fine-grained T0 to T8 turbidite sequences sensu STOW & SHANMUGAM (1980) were found, many of the characteristic layers described by these authors are represented in this facies, e. g. silt laminae with ripple-tops (T0), thin lenticular silt laminae (T2), discontinuous silt laminae (T4), and wispy convolute silt laminae (T5).

The grading of most siltstone beds, their erosive lower contacts, and the partial Bouma sequences are consistent with deposition from turbidity currents. Thin siltstone beds record deposition from suspension of fine-grained, dilute turbidity currents and possible traction transport (PICKERING et al., 1989). The repetitive interlayering of thin turbidites without visible bioturbation suggests high-frequent turbidite events (STOW & PIPER, 1984), because time intervals between successive turbidity currents were to short for significant bioturbation. According to turbidite facies models, such silty facies types may be deposited in inter-channel areas, levees or overbank areas of submarine fans of various sizes or in a distal, basinal environment (STOW & PIPER, 1984). Fine-grained turbidites may occur also in a slope setting between channelized coarse facies. Large, silty to muddy turbidity currents may have been derived from slumps on the upper slope (PICKERING et al., 1989).

4.3 Thin-bedded turbidites (facies D1)

Thin-bedded sandstone-mudstone couplets comprise packages up to 20 m thick of sandstones with marl intercalations (Fig. 6, 7c, d and Fig. 8). Individual sandstone beds are less than 0,5 cm to 30 cm thick. Most sandstone beds are normally graded, starting with medium-grade sandstones at the base. They display planar lamination and ripple lamination; convolute bedding is present, although rarely. Incomplete, base-missing Bouma sequences comprising Tbcde, Tcde and Tde predominate. Load casts, filled bioturbation structures and flute casts are present at the sandstone bases. Few paleocurrent data indicate both N-to-S and W-to-E transport directions. Packages of sandstones display commonly no thinning or thickening trends into under- or overlying conglomerates. In rare cases, an upward-thickening trend has been recognized. Sand: mud ratios are about 1:1 or less than 1. The lateral geometry of the beds is hardly to assess because of the lack of continuous outcrops. Some beds show a pinching within a few meters, indicating a lensoid rather than a sheet-like geometry of the beds.

Interpretation

This facies can be grouped into facies D1 (base-missing classical turbidites) of MUTTI & RICCI LUCCHI (1975) and

← Fig. 7
Photographs representing different facies types of the Tannheim Formation and the Losenstein Formation of the eastern part of the Northern Calcareous Alps.

a) Facies G: Black hemipelagic shales of the Tannheim Formation with *Leymeriella tardefurcata* mass occurrence (Höllleitengraben SW Lehneralm; for location see Fig. 3).
b) Facies D2.1: Thin-bedded siltstone-marl intercalations, strongly deformed (polished drilling core; Losenstein, tunnel of Ennstal-Bundesstraße).
c) Facies D1: Thin-bedded classical turbidites; note load casts at the base of lowermost sandstone bed and horizontal lamination and small-scale cross stratification in Tbcd-sandstone beds (Weyer Arc, Leerensackgraben SE Brunnbach).
d) Facies D1 and D2: Classical turbidites (Weyer Arc, Leerensackgraben SE Brunnbach; scale is 1 m).
e) Facies C2: Thick-bedded Tabcd-turbiditic sandstones (Weyer Arc, Leerensackgraben SE Brunnbach; scale is 1 m).
f) Facies B2: Massive and shear-laminated coarse sandstones (Großraming, Maria Neustift-Graben near Geyerlehen).
g) Facies A1: Pebbly mudstone (polished drilling core, Losenstein, tunnel of Ennstal-Bundesstraße).
h) Facies F: Pelitic slump with slump fold (Stiedelsbachgraben, section 2, 130 m; scale is 22 cm).
i) Facies A2: Clast supported conglomerate with intraclast of siltstone-marl facies (arrow; polished drilling core; Losenstein, tunnel of Ennstal-Bundesstraße).

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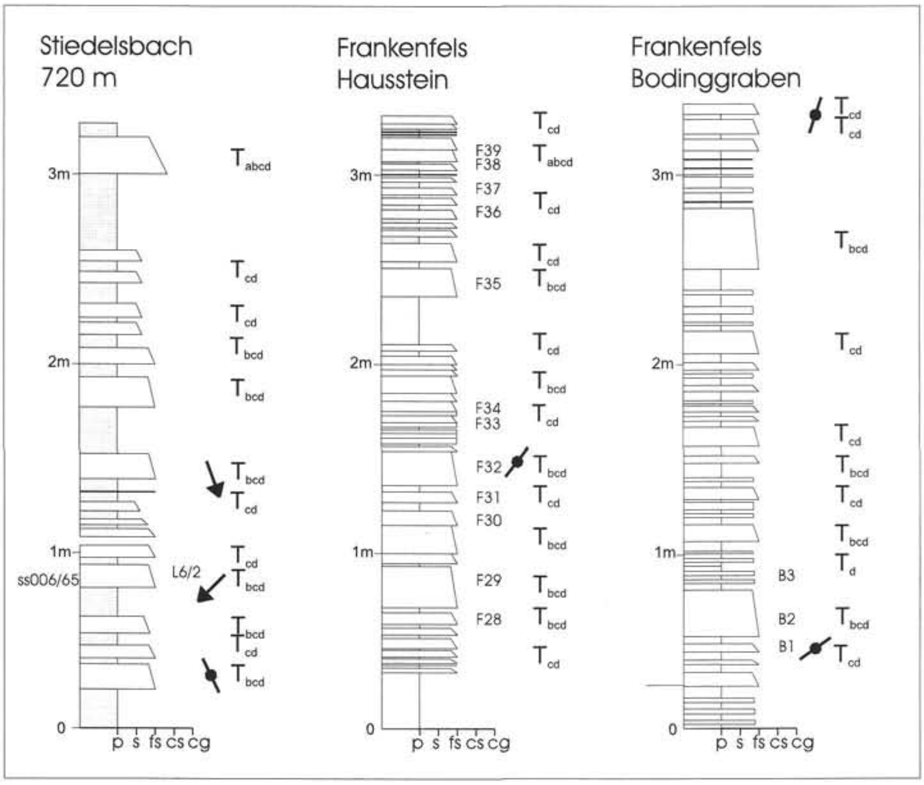

Fig. 8
Sections of thin-bedded turbidite facies D1 from the Stiedelsbach area and Frankenfels. For explanations see Fig. 6.

WALKER (1978). According to PICKERING et al. (1989) these thin-bedded sandstones can be assigned to facies C2 (organized sand-mud couplets-classic turbidites).

The occurrence of Bouma sequences suggests a turbidity current mode of transportation and deposition. The thin-bedded, fine-grained sandstones record deposition from relatively dilute turbidity currents by grain-to-grain settling out of suspension followed by tractional transport as bed load and accumulation of fine sediments from the tail of the current. The geometry of the beds, the common association with conglomerates and the scarcity of distinct thickening-upward cycles argues against an interpretation as a basin plain – sandstone lobe association of large submarine fans. Facies D1 may be also a part of channel – interchannel associations according to MUTTI & RICCI LUCCHI (1975) and the levees of large-sized upper fan channels (WALKER, 1978).

4.4 Thick-bedded turbidites (facies C1, D1)

Thick-bedded sandstones displaying complete Bouma sequences are scarce as compared to the thin-bedded facies. Tabcd and Tbcd-turbidites are present in about equal amounts in this facies. Average sandstone bed thickness is in the range of 15 to 60 cm (Fig. 7e). Sandstone bases display flute casts, load casts and bioturbation structures. Graded (Ta) and parallel laminated (Tb) Bouma divisions make up the lower three quarters of the beds, starting with coarse- to medium-grade sandstones. Convolute lamination is locally present in the upper part of the beds. Plant remnants are also concentrated in horizontal Td-layers near the top of beds. Sand: mud ratios are about 3:1 to 10:1, and amalgamation of beds is common. Thick-bedded sandstones may occur in packages a few meters thick, intercalated within thin-bedded sandstone facies without distinct thickening or thinning trends.

Interpretation

According to MUTTI & RICCI LUCCHI (1975) these sandstone beds can be classified into facies C1 (classical turbidites with Bouma Ta) and D1 (base-cut-out classical turbidites). PICKERING et al. (1989) divided their classical turbidite facies C2 (organized sand-mud couplets) according to bed thicknesses, comprising C2.1 (>30 cm) and C2.2. (10-30 cm).

The occurrence of thick-bedded classical turbidites with Bouma sequences suggests a turbidity current mode of transportation and deposition. Deposition from high concentration turbidity currents by grain-to-grain settling out of suspension followed by tractional transport as bed load and accumulation of fine sediments from the tail of the current is deduced (PICKERING et al., 1989). These beds represent proximal turbidites in the sense of MUTTI & RICCI LUCCHI (1975), which were assigned to channel fills and lobes of lower to mid-fan areas by WALKER (1978) and SHANMUGAM & MOIOLA (1988). Based on the lack of coarsening- or fining-upward cycles, deposition as large-sized channel fills and lobes of large submarine fans is unlikely for the thick-bedded facies of the Losenstein Formation. Isolated beds of facies C1 and D1 may represent parts of levee – interchannel associations (WALKER, 1978).

4.5 Massive to massive/laminated sandstones (facies B1, B2)

This facies comprises medium to coarse-grained sandstones with variable sorting. Individual beds are 30 to 100 cm in thickness. Beds may be amalgamated in packages up to 3.50 m thick. Both amalgamated massive beds and massive to laminated beds can be found (Fig. 6, 7f). A single strongly weathered bed (Maria Neustiftgraben NW Großraming) displayed faint, concave-up dish structures.

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Scours and large flute casts may be present at the base of massive beds. Grading may be present as coarse-tail grading with small pebbles concentrated at the base of otherwise structureless beds. Laminated to stratified medium-grained sandstones are observed between massive sands. Stratification is defined by 2-5 cm thick laminae, which show inverse to normal grading.

Interpretation

MUTTI & RICCI LUCCHI (1975) and WALKER (1978) classified massive sands into facies B1/B2 (massive sands with or without dish structures). They appear to be similar to divisions S2 and S3 of LOWE (1982). According to PICKERING et al. (1989) these sandstones can be assigned to facies B1.1 (thick-medium bedded, disorganized sands) and B2.1 (parallel-stratified sands). STOW & JOHANSSON (2000) included such sandstones into their DWMS-facies (deep-water massive sands).

These massive sandstones are interpreted to represent either deposits of grain flows, high-density turbidites (LOWE, 1982), or sandy debris flows (SHANMUGAM, 2000). Depositional processes include rapid mass deposition due to intergranular friction in a concentrated dispersion or freezing of a sandy debris flow. Dish structures are interpreted as fluid-escape features after deposition of beds (e. g. STOW & JOHANSSON, 2000). Laminated to stratified intervals probably represent deposits of grain flows and high-density turbidity currents with internal shear lamination or freezing of traction carpets at the base of flows (PICKERING et al., 1989). Initiation of sandy flows can be ascribed to liquefaction of unstable sandy material on submarine slopes, e. g. coarse-grained deltas. Water entrapment may have led to transformation of grain flows into sandy debris flows, which comprise highly concentrated, viscous sediment dispersions with plastic flow behaviour (STOW & JOHANSSON, 2000, and references therein).

4.6 Pebbly mudstones and pebbly sandstones (facies A1, A3)

Matrix-supported conglomerates with a sandstone matrix grade in composition into marly sandstones with dispersed pebbles. Pebbly mudstones are up to 4 m in thickness, whereas pebbly sandstones are less than 60 cm in thickness. Clasts may be concentrated in weakly defined, sub-horizontal lenses or stringers. No grading or imbrication was observed, and bimodal grain size distributions predominate, comprising a sandy matrix and a pebble population larger than 1 cm. Pebbly mudstones comprise a well-known facies of the Losenstein Formation ("Rosinenmergel", e. g. TOLLMANN, 1976; comp. Fig. 6). They are characterized by a low concentration (<10%) of average 1 to 3 cm pebbles in a muddy-silty matrix displaying a disorganized structure (Fig. 7g). Sometimes pebbly mudstones have gradational contacts to slump masses or coarse conglomerates with a pelitic matrix.

Interpretation

According to WALKER & MUTTI (1973) pebbly mudstones can be classified into facies A1 – disorganized conglomerates, including transitions into facies F (chaotic deposits). Pebbly sandstones fall into facies A3 (disorganized pebbly sandstones). Facies A1.3 (disorganized gravely mud) and facies A1.4 (disorganized pebbly sand) can be distinguished using the facies classes of PICKERING et al. (1989).

Pebbly mudstones are interpreted as deposits of cohesive submarine mud flows with abundant pelitic matrix. Mixing of a clast population and the pelitic matrix may be due to the incorporation of thin gravel beds in slides. Deposition of these muddy debris flows occurs mainly by cohesive freezing (PICKERING et al., 1989). Pebbly sandstones may represent deposits of sandy debris flows (SHANMUGAM, 2000) with transitions to massive sandstones deposited by high-density turbidity currents (WALKER, 1978; LOWE, 1982)

4.7 Conglomerates (facies A2, A1)

Polymict conglomerates constitute a characteristic facies type of the Losenstein Formation (Fig. 6, 7i). Conglomerates with "exotic" pebbles from vanished sources outside the NCA have been reported for more than 100 years (e. g. ROTHPLETZ, 1886; AMPFERER & OHNESORGE, 1909). They occur mainly in the upper part of the Losenstein Formation as stacked packages of conglomerate beds, pebbly mudstones and massive sandstones within finer-grained strata.

The thickness of individual conglomerate beds attains 4 m, amalgamated beds display thicknesses up to 10 m. Geometries of conglomerate beds are slightly lenticular, although larger outcrops are rare. Erosive contacts at the sharp base are present, although no deep incision into underlying finer-grained beds was observed. Rarely, flute casts or chutes can be seen at the base. Top of beds are planar or eroded from overlying coarse-grained beds. A lenticular, planar-convex geometry of some beds was already observed by GAUPP (1980) in the western NCA.

Internal structures of conglomerates include weakly defined horizontal bedding and normal grading; inverse grading is extremely rare. Only thinner beds up to 40 cm show grading through the whole bed. Thick massive beds display sometimes grading of the topmost 10 to 40 cm, whereas the rest of the bed shows no grading. Distinct horizontal stratification was only observed in fine-grained conglomerates or conglomerate-sandstone intervals. Most of the conglomerates of this facies are clast-supported, but frequent gradations into pebbly sandstones with matrix-support occur. Sorting is poor and both polymodal and bimodal grain-size distributions can be found. Within bimodal types component sizes above 1 cm within a matrix of medium-grained sandstone is prevailing. The maximum clast sizes observed in the Losenstein-Stiedelsbach section is 70 cm. Sub-horizontal preferred orientation of platy clasts is common in clast-supported conglomerates, whereas distinct imbrications are rarely observed, mainly in basal or upper portions of conglomerate beds. According to GAUPP (1980:99) long axes of prolate clasts are oriented parallel to paleocurrent directions. The rounding of clasts is generally good. Only intraformational clasts and carbonate clasts are subangular.

Interpretation

According to WALKER & MUTTI (1973) these conglomerates can be classified into facies A2 (organized conglomerates) with gradations into facies A1 (disorganized conglomerates) and F (chaotic deposits). Using the facies classes of PICKERING et al. (1989) these deposits are predominated by facies A2.1 (stratified gravel), A2.3 (normally graded grav-

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el), A1.1 (disorganized gravel) and transitions into facies A1.3 (disorganized gravely mud). Minor facies types include A2.7 (normally graded pebbly sand) and A2.8 (graded-stratified pebbly sand).

GAUPP (1980:92, 1982) already interpreted the conglomerates of the Losenstein Formation as deep-water conglomerates of channel fills based on their lenticular geometries. The transport mechanisms for these conglomerates are gravity-induced mass-flows. According to LOWE (1982), PICKERING et al. (1989) and SHANMUGAM (2000) these deposits can be interpreted to represent a continuum of deep-water processes such as debris flows with high concentrated dispersions including clay matrix strength, sand-rich debris flows, grain flows and highly concentrated turbidites. Facies A1 may have been deposited by frictional freezing due to cohesion of the matrix, whereas facies types A2 represent deposition by grain-to-grain deposition from suspension and bed-load processes.

4.8 Slumps (facies F)

Beds with slump folds predominate in the uppermost part of the Stiedelsbach section, where they comprise beds up to more than 5 m thick (Fig. 6 and 7h). Typical slump beds display predominantly disrupted and folded, originally laminated pelitic strata without or with very rare silt – to fine sandstone beds. Folded beds are transitional to pelitic beds with chaotic structure and a massive appearance. Sandstone beds within slumps are sheared and disrupted into lenses or phacoids. Gradations to pebbly mudstones and block-bearing mud flows can be found. Upper and lower boundaries of beds appear to be planar, although the lateral control on these surfaces is poor due to poor outcrop conditions. Locally, marlstone packages with lenses rich in well-preserved macrofossils (mainly gastropods and bivalves) are also incorporated into the slumps (e. g. Stiedelsbach, section 2, 44 m). Mixing of several shallow-water mollusc associations is reported by KOLLMANN (1976, 1978).

Interpretation

In the classification of MUTTI & RICCI LUCCHI (1975) these deposits are grouped into facies F. According to PICKERING et al. (1989) facies F2.1 (coherent folded and contorted strata) predominates, with transitions into facies F2.2 (brecciated and balled strata), A1.3 (disorganized gravely mud) and A1.2 (disorganized muddy gravel).

Transport processes can be interpreted as slides and rotational slumps on a slope, probably due to shocks by earthquakes or tsunamis (PICKERING et al., 1989). Deposition occurred due to decreasing slope gradients when gravity forces probably no longer exceeded basal and internal friction of the moving sediment mass. Transitions to cohesive debris flows and mud flows are commonly observed. Consequently, this facies type is interpreted to represent deposits of large submarine slides of mainly pelitic composition. The laminated appearance and planktonic-rich foraminiferal assemblages of the pelites point to an upper slope origin for the greater part of the slumped material. Some slumps also incorporated shallow-water material such as gastropods, which indicates that these slumps originated in shelf depths and breached the shelf break. The thickness of slumped intervals and transitions into coarse debris flows argues against an interpretation as slumped channel levee facies association (e. g. MUTTI & RICCI LUCCHI, 1975) and suggests mobilization and deposition of individual slumps on a slope. Muddy slumps probably triggered mud flows or high-density turbidites further downslope (STOW & MAYALL, 2000; SHANMUGAM, 2000) and transitions to debris flows are common.

5. Depositional Systems

The deep-water character of the Losenstein Formation was disputed up to the 1970s (e. g. ZEIL, 1955; MÜLLER, 1973), when LÖCSEI (1974), FAUPL (1978) and GAUPP (1980) proved a turbiditic deep-water depositional environment. GAUPP (1980) interpreted the coarsening-upward succession of the Tannheim Formation to the Losenstein Formation as the result of submarine fan progradation sensu MUTTI & RICCI LUCCHI (1975), with a predominance of inner/upper fan conglomerates, based on the fan model of WALKER (1978). Conglomerate packages, rarely displaying a fining-upward trend and lensoid geometry, were assigned to the fill of inner fan channels (GAUPP, 1980, 1982). In the western NCA, north-to-south transport directions were deduced from the southward decrease of clast sizes from the Randcenomanschuppe to the Allgäu Nappe (e. g. ZEIL, 1955; ZACHER, 1966; GAUPP, 1980). Few paleocurrent data in conglomerates and sandstones of the eastern NCA (Fig. 6, 8) support the idea of north-to-south paleotransport directions, although a strong variance is recognized and a deflection of turbidites in east – west directions along the basin axis can be inferred.

The facies association of coarse channel fills, fine-grained, thin-bedded intercalated packages and significant amounts of debris flow deposits and slumped beds points to a deep-water slope environment with a considerable gradient. The non-cyclic arrangement of the facies types and the non-organized distribution of channels and slumps argue against a classical submarine fan model with well-developed inner-, mid- and outer-fan areas in the sense of MUTTI & RICCI LUCCHI (1975), WALKER (1978) and SHANMUGAM & MOIOLA (1988). The uniformity of facies and composition along the entire length of the basin point to multiple sources or a linear source along the northern margin of the Tannheim-Losenstein basin, and a linear clastic wedge filling the basin from the north. Such depositional systems are either classified as submarine ramps or slope aprons (READING & RICHARDS, 1994; STOW et al., 1996). Submarine ramps are defined as turbidite systems characterized by the presence of multiple feeders, by indistinct channel and overbank deposits and the absence of lobe facies (HELLER & DICKINSON, 1985; READING & RICHARDS, 1994). Slope aprons comprise a linear depositional wedge fed from a continuous source with multiple feeders. They build narrow clastic slope systems elongated parallel to basin margins (READING & RICHARDS, 1994; STOW & MAYALL, 2000). Coarse-grained ramps and slope aprons are hardly distinguishable; therefore the interpretation of the depositional system is based on the widely used slope apron model (PICKERING et al., 1989; READING & RICHARDS, 1994; STOW et al., 1996; STOW & MAYALL, 2000).

Slope aprons are fed from multiple feeders along an essentially continuous linear source. Slump scars, slump and debris-flow masses and short-lived shallow channels

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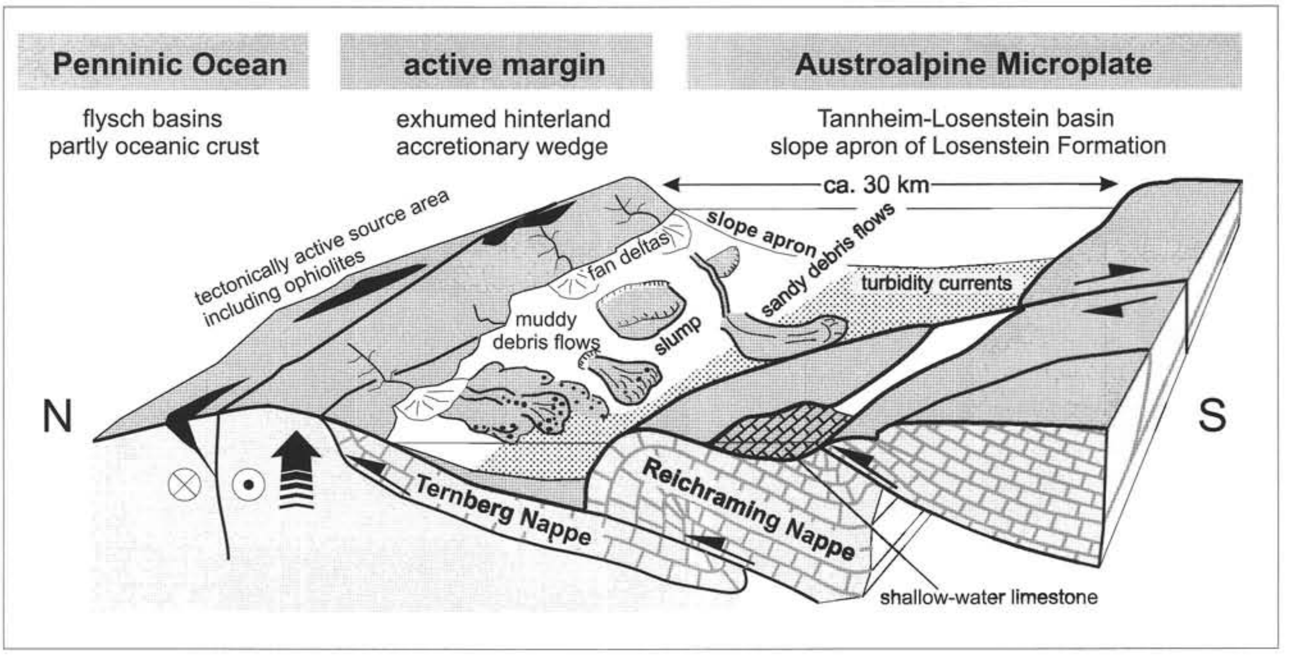

Fig. 9
Slope apron sedimentary systems and piggyback basin model of the Tannheim-Losenstein basin. No east-west differences implied.

commonly form part of a series of small fans and lobes that overlap along the length of the slope and build out a wedge-like single depositional system (STOW et al., 1996). Compared to large submarine fans, slope aprons are characterized by the absence of a large river-submarine canyon system and the absence of a master canyon. According to READING & RICHARDS (1994) slope aprons occur on very steep slopes and are characterized by small channels, multiple chutes and gullies, frequent mass-flow slumps, and river-generated turbidity currents.

Slope apron deposits are poorly organized, with rare, thin and isolated thinning- and fining-upward sequences due to infill of a single channel (PICKERING et al., 1989: 94-95). Vertical stacking of facies is poorly ordered or chaotic. Progradation of the slope forms a succession very similar to the coarsening-upward cycle of the Tannheim-Losenstein Formations, starting with basinal pelagics, followed by lower slope deposits with gravely channels and fine-grained silty/sandy interchannel areas to upper slope deposits with abundant slumps and debris flows.

Gravel-rich submarine ramps and sand-to-gravel-rich slope aprons typically tend to form across an active synsedimentary fault margin between the shelf and the basin floor. Slope aprons generally characterize areas of high relief and active tectonism (STOW & MAYALL, 2000). Activation of faults may generate sediment gravity flows and slumps on the slope. Gravel-dominated slope aprons comprise deep-water systems off tectonically active continental margins and islands, e. g. the continental borderland of California (GORSLINE & EMERY, 1959), the Paleogene North Sea (READING & RICHARDS, 1994), and New Zealand (STOW et al., 1996). The hilly hinterland of slope aprons may consist of a line-sourced coalescing braid plain or alluvial fan-fan-delta systems, which transport coarse-grained material over a narrow shelf into a high gradient slope.

Deposits of the Losenstein Formation fit quite well into the slope apron model (Fig. 9). The dominance of coarse debris flow deposits (facies A1, A2) and pelitic slumps (facies F) and the non-cyclic stacking of turbidite facies (C1, D1, D2 of WALKER & MUTTI, 1973) is very similar to described slope aprons (READING & RICHARDS, 1994; STOW et al., 1996; STOW & MAYALL, 2000). The coarsening-upward succession with abundant slumps in the upper part can be interpreted as progradation of the slope apron into a (hemi-)pelagic basin. High-frequency fine-grained turbidites (facies D1, D2.1) may characterize interchannel areas or areas of the lower slope to basinal trough distal from debris flows and slumps. Massive deep-water sands within a coarse succession are regarded also typical for coarse-grained slope-apron systems. These sands occur in small-sized chutes, channels, debris flow tongues and lobe sheets, which are fed by fan deltas (STOW & JOHANSSON, 2000). The elongated geometry of the clastic wedge of the Losenstein Formation and the small lateral extent of conglomerates transversal into the basin are similar to reported slope aprons. E. g., the 1000 m thick Miocene Blanca turbidite system off California extends 50 km parallel to its source, but conglomerates only reach about 10 km into the basin (READING & RICHARDS, 1994).

Based on the depositional system, the existence of a steep slope and a synsedimentary active basin margin is very likely along the northern margin of the Tannheim-Losenstein basin. The main controlling factor on deposition and geometry of the slope apron is tectonism in the hinterland and within the basin, whereas eustatic sea-level changes have only a minor effect on such depositional systems.

6. Source area

The occurrence of "exotic" clasts derived from source areas outside the NCA in the Losenstein Formation has been known for a long time (e. g. AMPFERER & OHNESORGE,

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1909). Both the sandstones and the conglomerates of the Losenstein Formation show a uniform composition along the entire length of the basin. Analysis of the framework grain composition of sandstones by GAUPP (1980, 1982), VON EYNATTEN (1996), VON EYNATTEN & GAUPP (1999) indicates predominantly immature litharenites composed mainly of extrabasinal grains. Source rocks include metamorphics, carbonates, and a significant serpentinized ultramafic source. The sandstones fall within the recycled orogen fields in the provenance diagrams of DICKINSON (1985). Very low degrees of weathering point to a high-gradient, mountainous source area (VON EYNATTEN & GAUPP, 1999).

6.1 Pebble composition

The "exotic" pebble composition of the Losenstein Formation and possible source areas were debated for a long time (e. g. TOLLMANN, 1976; GAUPP, 1983). Compositions of conglomerates show only minor variations from the western to the eastern NCA. The main pebble types are remarkably uniform along the entire length of the basin, including conspicuous "exotic" pebbles unknown from the NCA such as quartz porphyries, basic volcanics, pebbly quartzites, metasandstones, Lower Cretaceous neritic "Urgonian" limestones (LÖCSEI, 1974; GAUPP, 1980, 1983; SCHLAGINTWEIT, 1991). Rare components are locally present such as different types of granites (e. g. GAUPP, 1983) or metamorphic clasts. The good rounding of "exotic" pebbles consisting of resistant lithologies, e. g. quartz porphyries or quartzites, points to resedimentation processes of already well-rounded pebbles (MÜLLER, 1973; GAUPP, 1980). By contrast, some of the carbonate clasts from the NCA succession are angular to subangular and are more typical for the depositional setting within a coarse-grained slope apron. Rare carbonate-rich beds are interpreted to have been derived from the southern margin of the basin.

6.2 Heavy mineral analysis

Data on heavy minerals of Cretaceous formations of the Eastern Alps have been published by WOLETZ (1963), MÜLLER (1973), GAUPP (1980), WILDI (1985), POBER & FAUPL (1988), WINKLER (1988), FAUPL & WAGREICH (1992b), VON EYNATTEN (1996), and VON EYNATTEN & GAUPP (1999). Detailed chemical analysis of certain minerals were given by VON EYNATTEN (1996) and VON EYNATTEN & GAUPP (1999), including also a few samples from the type locality of Losenstein. In this paper data from the eastern NCA are compared to the published heavy mineral data (Tab. 1, Fig. 10). Heavy mineral samples were taken from the following localities: the type section of the Losenstein Formation in the Losenstein-Stiedelsbach area, the Höllleitengraben, the Dachsgraben and Innbachgraben near Grossraming, the Hagauergraben, Hanslgraben, Draxlgraben and Leerensackgraben south of Grossraming in the Weyer Arc area, the Hausstein, Kleinbernreith and Bodinggraben sections near Frankenfels, the Höfnerbachgraben near Kaumberg. Samples from the Western Carpathians were taken from the Malé Karpaty Mountains north of Bratislava and the Považský Inovec Mountains further to the northwest (WAGREICH, 2001).

In general, heavy mineral assemblages of the Losenstein Formation show a uniform composition from the western to the eastern NCA. Sandstones of the eastern NCA (sample localities see Fig. 1: Losenstein Syncline, Frankenfels Nappe and Randcenomanschuppe of the Weyer Arc, Großraming, Frankenfels, Kaumberg) display high amounts of chrome spinel (5-44%) and the stable minerals zircon (8-38%) and tourmaline (9-40%), besides strongly varying amounts of chloritoid (0,4-46%) and apatite (2-43%). Garnet (0-10%) and blue sodic amphibole (up to 17%) comprise a significant component of the assemblages. Blue sodic amphiboles were analysed in detail by VON EYNATTEN (1996) and VON EYNATTEN & GAUPP (1999): They reported mainly glaucophane and ferro-glaucophane, which may be derived from Variscian HP terrains (VON EYNATTEN et al., 1996). Some differences to these uniform heavy mineral assemblages are displayed by samples from the Randccenomanschuppe of the Weyer arc area. These samples show increasing amounts of metamorphic minerals such as chloritoid (up to 59%) and garnet (up to 61%), showing a more local influence.

Similar heavy mineral assemblages of Albian-Cenomanian sandstones occur within the Central West Carpathians in Slovakia (FAUPL et al., 1997). Although the stable minerals predominate, the assemblages display significant amounts of chrome spinel both from the Poruba Formation of the Krizna Nappe and the Klape unit of the Pieniny Klippen Belt (Tab. 1). Blue sodic amphiboles are present in low amounts. Chemical compositions of blue sodic amphiboles from these mid-Cretaceous formations are similar to those from the NCA, displaying a trend towards Fe-glaucophane (IVAN & SÝKORA, 1993; SÝKORA et al., 1997).

In the logratio diagram of Fig. 10 (Cr/META = ratio of chrome spinel to metamorphic minerals garnet, blue sodic amphiboles and chloritoid, versus Cr/ZTR = ratio of chrome spinel to stable mineral group zircon, tourmaline and rutile) the samples display a broader range than reported from samples of the western NCA (VON EYNATTEN, 1996). Values below 0 on the ln(Chr/META) axis distinguishes the Losenstein Formation from the Branderfleck Formation and the Lech Formation in the western NCA. In the eastern NCA, a tendency towards increasing amounts of chrome spinel from the Losenstein area to the east can be recognized, as samples from Frankenfels and Slovakia display higher values of ln(Chr/ZTR). The ln(Chr/META) values display minor variations from the eastern NCA into the Western Carpathians and do not reach the high values reported from the Branderfleck Formation and the Lech Formation. This points again to similar source lithologies for the Losenstein

Tab. 1 →

Heavy mineral data from the Losenstein Formation of the eastern part of the Northern Calcareous Alps and Albian sandstones of the Western Carpathians (grain size fraction 0.063 – 0.4 mm, minimum 250 grains). chrome sp. – chrome spinel, blue amp. – blue sodic amphiboles; sections: TN-LS – Ternberg Nappe, Losenstein Syncline (Stiedelsbach, Höllleitengraben), TN-ST – Ternberg Nappe, Losenstein Syncline (Steinbach), FN-WA – Frankenfels Nappe, (Weyer Arc Leerensackgraben, Hanslgraben), FN-GR – Frankenfels Nappe, Großraming area (Maria Neustiftgraben, Dachsgraben, Innbach), FN-FR – Frankenfels Nappe, Frankenfels area (Bodinggraben, Frankenfels), FN-KA – Frankenfels Nappe, Kaumberg (Höfnergraben), WC-KL – Western Carpathians, Klape Unit, WC-KN – Western Carpathians, Krizna Nappe, CR-MN – Randcenomanschuppe, Maria Neustiftgraben, CR-WA – Randcenomanschuppe, Weyer Arc (Hagauergraben).

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Sample No	Section	zircon	tourmaline	rutile	apatite	garnet	chloritoid	chrome sp	blue amp.	others
KB7-1	TN-LS	28,8	35,4	4,5	7,1	5,6	7,1	9,1	2,5	0,0
L3c	TN-LS	10,3	21,1	4,1	38,1	1,5	19,1	5,7	0,0	0,0
L5-1	TN-LS	14,5	17,0	4,5	30,3	3,6	6,1	11,5	11,8	0,9
L6-2	TN-LS	31,5	17,7	4,3	29,5	2,4	1,2	9,1	3,9	0,4
K29-2	TN-LS	12,6	33,3	5,7	10,3	6,9	5,7	19,5	5,7	0,0
L16	TN-LS	9,6	31,6	1,9	23,0	0,5	11,0	12,4	8,1	2,0
L18	TN-LS	26,8	15,6	8,7	9,1	2,6	8,7	22,9	5,2	0,4
L19	TN-LS	26,9	33,0	4,4	13,7	4,4	3,8	13,2	0,5	0,0
L20	TN-LS	16,3	38,5	4,8	12,0	1,4	8,2	17,3	1,0	0,5
L21	TN-LS	16,5	30,0	7,0	10,5	0,0	13,5	20,0	1,5	1,0
L30b	TN-LS	8,2	15,9	1,0	14,9	2,4	2,9	43,3	11,1	0,5
L31-5	TN-LS	32,0	26,7	7,3	13,1	2,9	6,8	10,7	0,5	0,0
L32-1	TN-LS	7,5	19,7	4,7	25,1	1,8	19,0	16,8	5,4	0,0
L38/2	TN-LS	21,8	22,4	5,4	7,8	1,4	7,1	24,8	9,2	0,0
HÖ4-1	TN-LS	18,0	30,0	4,3	23,6	2,1	8,6	8,6	4,7	0,0
HÖ97-3	TN-LS	10,6	13,1	6,6	10,9	3,6	8,8	36,5	9,5	0,4
STEINB2	TN-ST	23,3	15,3	7,3	18,1	4,9	9,7	17,7	3,8	0,0
LEE97-4	FN-WA	13,8	12,9	5,0	8,8	1,3	10,0	41,7	6,7	0,0
LEE97-2	FN-WA	14,8	21,4	4,7	27,2	5,8	7,0	14,0	5,1	0,0
HAN97-14	FN-WA	20,0	15,8	5,6	9,1	2,1	11,6	31,6	4,2	0,0
HAN97-16	FN-WA	23,3	19,4	2,7	14,7	3,9	7,8	27,1	0,4	0,8
HAN97-9	FN-WA	4,2	8,8	2,9	8,0	6,7	46,2	6,3	16,8	0,0
MN11-1	FN-GR	23,8	28,7	4,1	22,5	2,0	1,6	13,1	3,7	0,4
MN13	FN-GR	15,4	20,7	4,8	43,3	2,4	1,4	11,5	0,0	0,5
MN16-2	FN-GR	38,4	19,9	6,0	16,2	1,9	0,5	16,7	0,0	0,5
MN18-1	FN-GR	25,6	22,1	4,2	34,4	1,5	1,9	9,9	0,0	0,4
DA8	FN-GR	26,8	8,8	3,1	13,4	10,3	23,0	13,4	0,4	0,8
DA11	FN-GR	8,5	23,3	3,1	14,7	3,1	34,9	10,1	0,8	1,6
DA14	FN-GR	12,5	24,1	3,6	23,2	0,9	22,3	4,5	8,9	0,0
INB3	FN-GR	35,2	12,8	6,2	1,8	3,1	0,4	39,6	0,4	0,4
INB1	FN-GR	23,9	20,8	4,4	5,7	5,0	14,5	17,6	7,5	0,6
BOD3	FN-FR	33,3	16,1	5,4	21,5	2,2	5,4	16,1	0,0	0,0
BOD4	FN-FR	19,0	30,6	4,1	9,9	0,7	1,0	34,0	0,3	0,3
FRA29	FN-FR	17,1	27,0	8,1	18,9	4,5	9,0	14,4	0,9	0,0
FRA30	FN-FR	19,1	20,0	6,5	7,4	0,9	2,2	43,5	0,0	0,4
FRA32	FN-FR	30,5	8,6	6,6	9,0	1,2	1,2	43,0	0,0	0,0
FRA35	FN-FR	25,0	26,9	7,3	12,7	1,5	1,2	25,4	0,0	0,0
FRA40	FN-FR	15,4	39,5	2,6	17,1	1,8	1,3	21,9	0,0	0,8
KAUM4a	FN-KA	12,4	24,5	5,0	22,1	0,7	1,7	33,6	0,0	0,0
KAUM4b	FN-KA	18,7	22,5	8,2	15,7	0,4	0,4	33,7	0,0	0,4
CS3a	WC-KL	49,8	15,4	6,6	8,5	1,0	0,3	18,4	0,0	0,0
CS3b	WC-KL	26,8	38,7	5,6	13,2	1,3	0,0	13,2	0,0	1,3
MK00-1	WC-KN	11,8	35,7	12,6	35,7	0,4	0,0	3,8	0,0	0,0
MK00-1-2	WC-KN	53,4	13,4	18,7	4,2	0,4	0,4	9,2	0,0	0,4
PI00-1	WC-KN	53,8	13,5	17,7	3,8	1,2	0,0	10,0	0,0	0,0
MN98-1	CR-MN	7,7	12,8	3,1	15,3	61,2	0,0	0,0	0,0	0,0
HAG467	CR-WA	10,4	12,1	2,2	13,7	5,2	54,7	2,9	0,0	0,0
HAG87	CR-WA	13,8	27,3	1,8	33,7	3,4	13,8	4,1	0,0	0,8
HAG106	CR-WA	8,2	25,7	0,8	45,7	3,4	10,1	4,3	2,1	0,3
HAG472	CR-WA	7,9	14,4	3,1	14,7	5,2	45,8	9,1	0,0	0,4
HAG475	CR-WA	5,1	10,4	2,6	16,2	6,7	55,2	3,9	0,0	0,0
HAG93d	CR-WA	8,6	8,2	0,8	9,2	7,3	58,4	5,3	2,1	0,5
HAG119	CR-WA	6,8	22,4	0,8	10,2	9,9	43,3	4,9	0,9	0,5
HAG110	CR-WA	4,2	6,9	1,2	3,2	3,8	73,7	3,2	2,9	2,9

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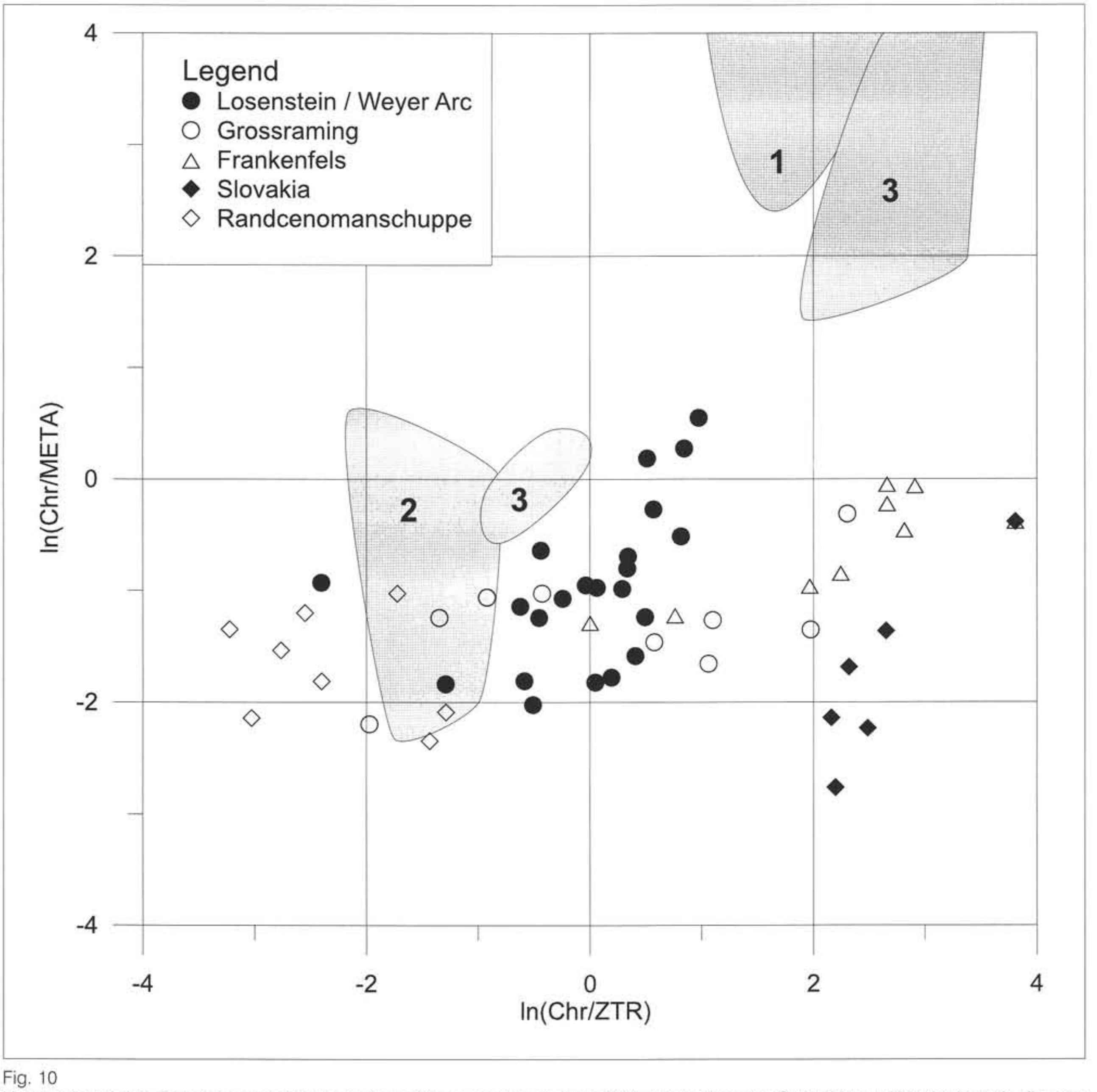

Fig. 10
Heavy mineral data from the Losenstein Formation of the eastern part of the NCA and the Western Carpathians within the logratio diagram ln(Chr/ZTR) vs. ln(Chr/META) based on VON EYNATTEN (1996). Chr – chrome spinel; ZTR – zircon, tourmaline and rutile/TiO_2-group; META – garnet, blue sodic amphibole and chloritoid. Grey fields indicate heavy mineral assemblages from the wesern NCA (VON EYNATTEN, 1996: Fig. 3.19): 1 – Lech Formation (Lechtal Formation of VON EYNATTEN, 1996); 2 – Losenstein Formation; 3 – Branderfleck Formation.

Formation and the Poruba Formation of the Western Carpathians.

The uniform composition of the detritus therefore suggests the same type of source area along the entire basin length within the NCA, and probably continuing into the western Carpathians as suggested by similar pebble compositions and heavy minerals (MIŠÍK et al., 1981). The source area consists of continental basement slices including low- to medium-grade metamorphics, granites, quartz porphyries and only rare volcanics as well as Mesozoic sedimentary rocks partly unknown from the NCA (GAUPP, 1980, 1983, VON EYNATTEN, 1996). The presence of chrome spinel and serpentine fragments indicates also ultrabasics probably from ophiolithic remnants in the source area (POBER & FAUPL, 1988; VON EYNATTEN, 1996). According to GAUPP (1980, 1983) the source area was located within a Lower Austroalpine to South Penninic position to the north of the NCA. Austroalpine basement slices, Paleozoic strata, mainly quartz porphyries, and a lithologically diverse Mesozoic succession can be compared to Lower Austroalpine rocks (GAUPP, 1983). The presence of ultrabasic rocks in the hinterland argues for ophiolitic slices.

This northern to north-western source province comprises chrome spinel, chloritoid, and blue sodic amphiboles as characteristic heavy minerals (WOLETZ, 1963; POBER & FAUPL, 1988; VON EYNATTEN, 1996; VON EYNATTEN & GAUPP, 1999). Phengites associated with (Fe-)glaucophane were dated between 339 and 357 Ma, suggesting a Variscan HP-

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metamorphic source (VON EYNATTEN et al., 1996). Zircon FT ages cluster around 170-115 Ma. The 170 Ma ages suggest a thermal event during the Early Jurassic, e. g. the rifting of the Penninic Ocean (FRISCH, 1979) and subsequent fast exhumation and erosion during the Early Cretaceous (VON EYNATTEN, 1996).

7. The Tannheim-Losenstein piggyback basin

Piggyback basins were defined as sedimentary basins that formed on active thrust sheets (ORI & FRIEND, 1984), where the entire sedimentary basins are transported during thrusting. Piggyback basins are normally the result of thrust propagation, when the frontal ramp of a thrust sheet becomes inactive and a new active ramp forms in front of the older one. In terms of the basin geometries and the depositional systems of the basin infill, two types of piggyback or thrust top basins can be distinguished (ORI & FRIEND, 1984; WAGREICH, 2001): Type 1 piggyback basins comprise deep-water basins, mainly occurring as elongated, synsedimentary deforming troughs in the footwall of advancing thrust sheets. These basins are typical for the early stages of orogenic wedges and underfilled foreland basins (e. g. ORI & FRIEND, 1984; LASH, 1990; BOIANO, 1997). Type 2 piggyback basins comprise terrestrial to marginal marine "intramontane" basins occurring predominantly in late stages of orogens (e. g. NIJMAN, 1998; CIPOLLARI et al., 1999).

The Miocene to Quaternary Po Basin Complex is reported as a typical Type 1 piggyback basin, where longitudinal deep-water basins developed between active thrust sheets (ORI & FRIEND, 1984; ZOETEMEIJER et al., 1992) and the outer thrust sheet formed an emergent mountain range during most of the deposition. Marginal fan-deltas, deep-sea fans and submarine slopes are typical depositional systems within these basins.

The interpretation of the Tannheim-Losenstein basin as a type 1 piggyback basin is based on the position at the footwall of the Reichraming-Lunz thrust sheet and the existence of a tectonically active source area (Fig. 9). The coarse deep-water slope apron facies including frequent slumps is in accordance with a tectonically active basin with a steep northern slope. However, the main sediment sources in the case of the Tannheim-Losenstein basin were to the north and not from the active thrusts to the south of the basin, were only minor carbonate breccias derived.

Although the geometry of the Tannheim-Losenstein basin is largely destroyed by the later polyphase deformation, it can be constrained based on balanced cross sections of the narrow synclines, which yield a minimum extent of the Frankenfels Nappe of about 12 km (LINZER et al., 1995: Fig. 5). Including the strongly deformed northernmost tectonic outliers of the eastern NCA (Randcenomanschuppe, e. g. EGGER, 1988) a minimum width of 20 km is suggested for this basin. As both the Losenstein syncline and the Randcenomanschuppe sections only include deep-water deposits, the shallow-water part of the northern basin margin is completely missing. Although the coarse deep-water facies argues only for a narrow, probably relatively steep shelf, a minimum of about 10 km width must be added to the total basin width. Blind thrusts are probably responsible for some of the thickness variations recorded in the Tannheim-Losenstein Formations along strike. To the south, Aptian-Albian shallow-water marly limestones with crinoidal debris and mollusc fragments along the northern margin of the Lunz Nappe near Großraming (EHRENDORFER, 1987) indicate the presence of a southern marginal facies of the Tannheim-Losenstein basin along the thrust front of the overthrusting Lunz-Reichraming sheet.

The E-W extension of the basin is given by the extend of the northernmost nappe unit of the NCA, the Frankenfels – Ternberg – Allgäu nappe system (Fig. 1). Along the whole length of this thrust system, sandstones and conglomerates of the Losenstein Formation can be found, beginning in the western Allgäu (e. g. GAUPP, 1980, 1982) to the eastern margin of the NCA and the basement of the Vienna Basin (e. g. LÖCSEI, 1974). This sums up to a minimum length of about 350 km, considering the later Tertiary E-W extension of about 40% to form the present-day elongation of the NCA (FRISCH et al., 1998). The probable continuation of the basin into the Western Carpathians as suggested by a similar coarsening-upward succession of Albian age in the Krizna Nappe (Poruba Formation, e. g. PLAŠIENKA, 1995) adds some further 100 km to the basin length, indicating an extremely elongated, narrow piggyback basin.

Within the western part of the NCA the Tannheim-Losenstein basin was probably connected with the "Kreideschiefer" basin (Lech Formation, VON EYNATTEN & GAUPP, 1999) to the south. There, both basins are clearly related to folding and thrusting in the Allgäu and Lechtal thrust sheets, respectively (LEISS, 1992; MAY & EISBACHER, 1999), and display a coarsening-upward trend, which is terminated by coarse sediments including the emplacement of olistolites in the Lech Formation during the Late Albian to earliest Cenomanian (GAUPP, 1980; WINKLER, 1988; LEISS, 1992). Whereas the Tannheim-Losenstein basin received a large quantity of non-NCA material from the northern source, the clast composition and heavy minerals of the Lech Formation point to a hinterland to the south (VON EYNATTEN & GAUPP, 1999), probably including overthrusting nappes and pre-Alpine metamorphic basement units. Large olistolites are exclusively derived from local sources, probably as a consequence of syndepositional NW-verging thrusting which triggered local slope failures along high-relief basin margins (MAY & EISBACHER, 1999).

8. Subsidence analysis and subsidence modelling

Tectonic subsidence analysis was performed for the Stiedelsbach section near Losenstein to quantify tectonic subsidence and to compare the latter with calculated results derived from forward subsidence modelling. Standard backstripping techniques (STECKLER & WATTS, 1978) were used including stepwise decompaction of the sediment column based on empirical depth-porosity relationships (BOND & KOMINZ, 1984). Local isostatic correction for sediment loading was used for simplification, because of the large error introduced in flexural models due to unknown basement parameters for the piggyback basin. Although the assumption of local isostasy is certainly an oversimplification, the resulting tectonic subsidence gives at least a best estimate for the magnitude of the maximum subsidence possible. No correction for eustatic sea level changes was

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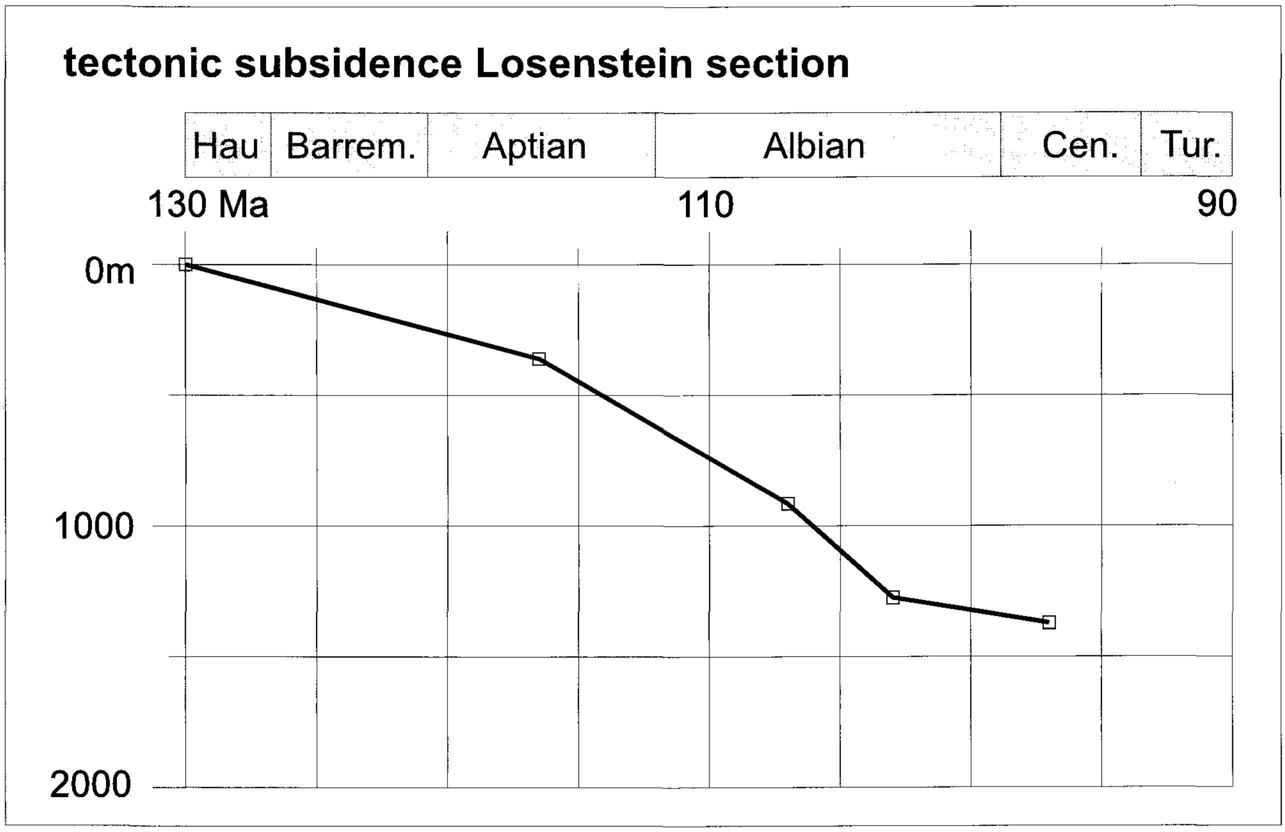

Fig. 11
Backstripped tectonic subsidence curve of the Cretaceous succession of the Losenstein syncline.

applied. Palaeo water-depths were estimated based on foraminiferal assemblages according to WEIDICH (1990). The timescale of GRADSTEIN et al. (1995) was used.

Fig. 11 displays a backstripped tectonic subsidence curve for the Stiedelsbach section from the Hauterivian to the Cenomanian. Tectonic subsidence totals up to 460 m during the time interval for the sedimentation of the Losenstein Formation (up to 9 Ma), which gives rather low tectonic subsidence rates of about 50 m/Ma for this coarse-grained succession.

The application of forward subsidence modelling for the Tannheim-Losenstein basin is strongly limited because of the large uncertainties concerning the basement type and thickness and the geometry of the basin. Therefore, a simple flexural thrust-foreland model was applied to compare observed tectonic subsidence with maximum subsidence from flexural loading of an elastic plate (e. g. JORDAN, 1981). As the geometry of the basin is largely unknown, estimates for the effective elastic thickness (Te) of the underlying crust are poorly constrained. Low Te values are assumed based on the presence of a relatively young Austroalpine lithosphere as a result of Jurassic stretching. Given the simplified 2D-equation for maximum subsidence (= maximum deflection w) adjacent to an applied load on an elastic plate (JORDAN, 1981), tectonic subsidence of 460 m would imply a load (h) about 20-30 km length and 2-4 km thickness (assuming Te values in the range of 5 to 15 km).

Maximum tectonic subsidence $w = (-h/2)\ (\rho_s/\rho_m)\ [\exp((-\lambda)(-x+s-a))\cos((\lambda)(-x+s-a)) - \exp((-\lambda)(-x+s+a))\cos((\lambda)(-x+s+a))]$

h = height of the load
ρ_s = density of sediments (2.4 g/cm^3)
ρ_m = density of the mantle (3.4 g/cm^3)
ρ_a = density of air (0.00129 g/cm^3)
x = distance of point w to origin
s = distance away from origin of load
a = half width of load
D = flexural rigidity
g = gravity acceleration (981 cm/sec^2)
λ = $[((\rho_m-\rho_a)g)/4D]^{1/4}$

9. Implications for Cretaceous deformation within the eastern part of the Northern Calcareous Alps

Cretaceous deposits of the NCA give evidence for timing of deformational events within the eastern NCA. Dating of thrust movements is constrained by the youngest sediments of syntectonic basin fills below the thrusts. A general younging of thrust movements from the south to the north within the NCA is indicated by syntectonic successions. Hauterivian – Lower Aptian deposits of the Rossfeld Formation (FAUPL & TOLLMANN, 1979; DECKER et al., 1987; SCHWEIGL & NEUBAUER, 1997) on structurally higher thrust sheets to the south are significantly older than the formation of the Tannheim-Losenstein basin below the overthrusting Lunz-Reichraming nappe system. In the area of the Weyer Arc, further time constraints are provided by the deposits of the Branderfleck Formation, which transgressed unconformably onto

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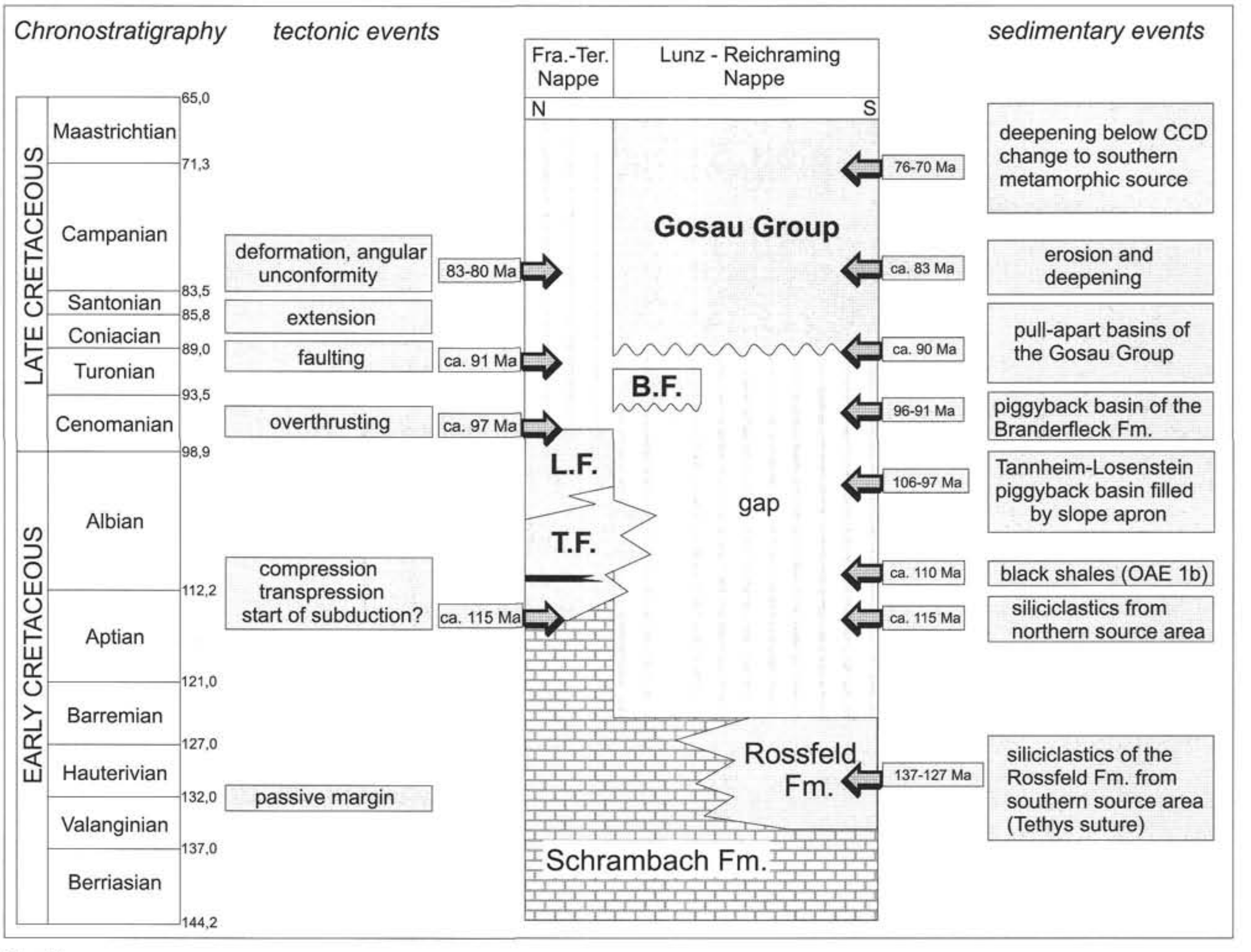

Fig. 12
Sedimentary and tectonic events in the eastern part of the Northern Calcareous Alps during the Cretaceous. Fra.-Ter. – Frankenfels-Ternberg; OAE1b – oceanic anoxic event 1b.

the northern part of the Lunz-Reichraming nappe system during the Middle/Late Cenomanian to Early/Middle Turonian (FAUPL & WAGREICH, 1992b), giving evidence for a first deformational phase starting during the Early to Middle Cenomanian. A similar timing of syntectonic strata and deformational events is reported from the western NCA (GAUPP, 1980, 1982; WEIDICH, 1984). In contrast to the western NCA, no continuation of the Branderfleck Formation into Late Turonian/Coniacian is known from the eastern part. Red alluvial conglomerates of the Gosau Group unconformably overly Middle Turonian shallow-marine marls in the Weyer Arc (FAUPL & WAGREICH, 1992b), pointing to a second phase of faulting during the Middle to early Late Turonian (Fig. 12).

The formation and the geometry of the Tannheim-Losenstein basin give further constraints for thrust geometries of the NCA. Cretaceous deformation within the NCA was interpreted to be a consequence of top-to-NW or top-to-W nappe stacking and dextral strike-slip faulting along NW-trending tear faults, followed by Paleogene top-to-N stacking (EISBACHER et al., 1990; DECKER & PERESSON, 1996; LINZER et al., 1995, 1997). Cretaceous top-to-W thrusting was reported by FROITZHEIM et al. (1994) from the western margin of the Eastern Alps. Thrusting to the west (in todays' reference system) is not compatible with the W-E extending deep-water trough of the Tannheim-Losenstein basin. Thrusting to the northwest must have been accompanied by an array of NW-SE oriented dextral transfer faults to accommodate for the continuous depositional trough to the north of the thrust front in the eastern part of the NCA. The uniformity of the basins suggests that these strike-slip faults were constricted to the higher NCA nappes (Fig. 9). Hints to the existence of such Cretaceous faults as forerunners of Tertiary strike-slip faults were also reported by fault gauge dating of KRALIK et al. (1987), which indicated a 100 Ma age for the formation of such a fault system, and by the geometry of Aptian-Albian strata south of the Losenstein basin in the western NCA (MAY & EISBACHER, 1999).

10. Geodynamics of the Cretaceous of the Eastern Alps

The paleogeographic and geodynamic evolution of the Eastern Alps regarding timing and tectonic setting of events is still a matter of controversy (e. g. WINKLER, 1988, 1996; FROITZHEIM et al., 1994, 1996; OBERHAUSER, 1995; SCHWEIGL & NEUBAUER, 1997; VON EYNATTEN & GAUPP, 1999; FAUPL & WAGREICH, 1992a, 2000; MATTERN, 1999; WILLINGSHOFER et al., 1999).

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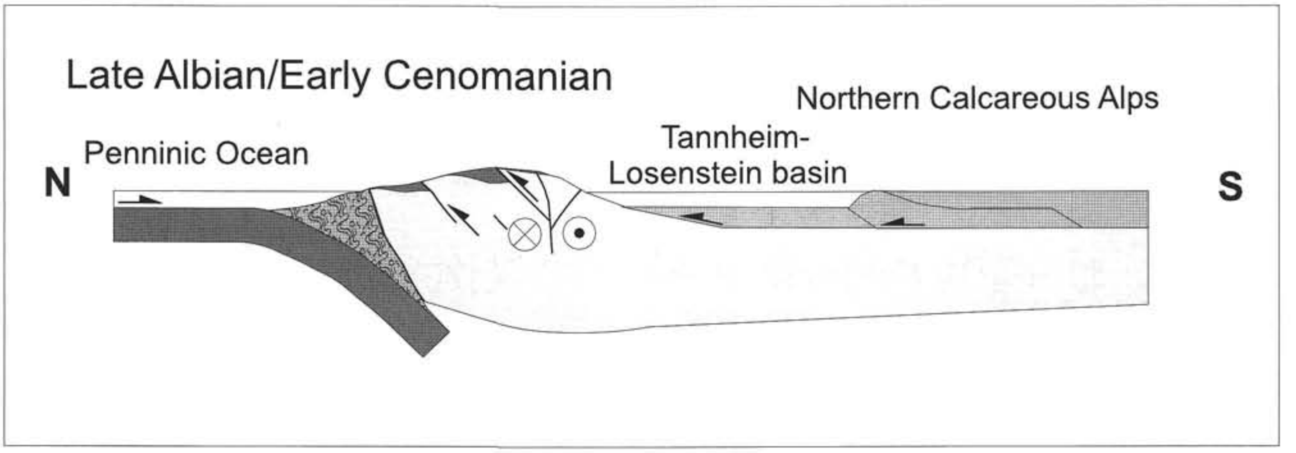

Fig. 13
Palinspastic sketch of the Austroalpine-Penninic margin during the Albian-Cenomanian. Compression along the Austroalpine-Penninic margin resulted in oblique subduction of Penninic crust below the Austroalpine microplate, the formation of a transpressional accretionary wedge, accelerated exhumation of the northern source area and thrusting in frontal parts of the Northern Calcareous Alps.

Based on the sedimentary successions of the NCA, a change from an extensional to a compressional tectonic regime is interpreted for the Early Cretaceous along the Austroalpine-Penninic margin. The formation of basins filled with deposits of the Rossfeld Formation in Hauterivian – Early Aptian time is interpreted as a consequence of suturing of an oceanic branch to the south of the NCA and subsequent thrust loading (FAUPL & TOLLMANN, 1979; SCHWEIGL & NEUBAUER, 1997). During the Early Cretaceous up to the Aptian, the northern margin of the NCA was characterized by deposition of pelagic limestones and limestone – marl beds of the Schrambach Formation, without any indications for syndepositional tectonism, and without any record of a siliciclastic source area to the north of the NCA. Farther to the north and northwest, South Penninic units such as the Arosa zone, the Walsertal zone and the Ybbsitz zone were situated.

The northern margin of the Austroalpine microplate can be classified as a passive margin succession up to the Aptian. During the Late Aptian (ca. 116-112 Ma; GRADSTEIN et al., 1995) the formation of the Tannheim-Losenstein basin and the accumulation of calcareous shales of the Tannheim Formation record a major change along the northern margin of the Austroalpine plate (Fig. 13). The Tannheim-Losenstein basin evolved as a foreland to piggyback basins in front of northwestward advancing thrust sheets. At the same time, a new source area raised at the Austroalpine-Penninic plate margin, comprising continental basement, Mesozoic sedimentary rocks and remnants of ophiolites. Uplift of this north(western) source area is documented by FT ages of detrital zircon at 120 Ma (VON EYNATTEN, 1996). A tectonically active, high-gradient source area is indicated by low weathering indices of sandstones, despite the generally humid Cretaceous climate (VON EYNATTEN & GAUPP, 1999), and coarse-grained deposition on slope aprons, which requires steep gradients and a tectonically active hinterland.

The piggyback basin formation and the rise of a northern source area during Late Aptian/Early Albian can be explained by the development of a transpressive plate margin controlled by both strike-slip faulting and thrusting (VON EYNATTEN & GAUPP, 1999). The northern source area is interpreted as a transpressive structural high (Fig. 13a). This source area formed an emergent mountain range during most of the deposition of the Tannheim-Losenstein basin. Exhumation may be the result of the formation of positive flower structures and duplexes along dextral strike-slip faults. MAY & EISBACHER (1999) also interpreted siliciclastic sedimentation of the Lech Formation in the western NCA as a consequence of the change from an extensional setting to dextral transpression. The emerging source area shed sediments also to the north into the Penninic Ocean, e. g. Aptian/Albian turbidites of the Ybbsitz zone contain also chrome spinel (HOMAYOUN & FAUPL, 1992) as well as sandstones from the Arosa zone, indicating also a derivation of the clastic material from the same mixed oceanic-continental source (WINKLER, 1988, 1996; DECKER, 1990).

Transpression along the northern margin of the Austroalpine microplate can be interpreted as a result of beginning oblique southward subduction of the (South)Penninic Ocean (Fig. 13). The conversion of this wrench zone into a subduction zone may be a consequence of plate movement reorganization and of failure of the young warm lithosphere along the Austroalpine margin due to build-up of regional compressive stresses (CLOETINGH & WORTEL, 1988). Alkaline basanitic dikes of Late Albian age (ca. 100 Ma) in the western NCA give evidence that internal parts were still underlain by a European-type basement without any indications for a large subducting slab (TROMMSDORF et al., 1990). No evidence for such dikes is known from the eastern NCA, therefore subduction may have started earlier there. A Turonian onset of subduction as suggested by WILLINGSHOFER et al. (1999) is unlikely based on the formation of piggyback basins and the rising of the source area.

The existence of ophiolitic slices along the entire northern margin of the Austroalpine microplate from at least the Albian onwards is still debated. Subduction of the Penninic Ocean below the Austroalpine microplate probably started not earlier than Aptian/Albian. WINKLER (1996) explained this contradiction by suggesting an early intraoceanic subduction within the Penninic Ocean during Hauterivian-Aptian. This led to the existence of a "north-Adriatic obduction belt" along the northern margin of the Austroalpine (WINKLER, 1996) due to ophiolite thrusting onto the Austroalpine continental margin (Fig. 13a).

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In the eastern NCA the sedimentation within the Tannheim-Losenstein basin lasted until the Early Cenomanian (ca. 97 Ma). The coarsening-upward succession records progradation of a slope apron. Subsequently, thrusting of the Lunz-Reichraming Nappe system onto the Frankenfels-Ternberg nappes ended sedimentation within this basin. Thrusting occurred simultaneously over the whole NCA and was followed by transgression of the Branderfleck Formation during the Middle/Late Cenomanian.

Within neighboring paleogeographic zones, deposits of Aptian – Albian – Cenomanian times also record significant changes during this time interval. According to FUCHS & WESSELY (1996) the sedimentary cover of the Bohemian Massif below the Molasse Basin in Upper Austria displays a Late Cenomanian to Early Turonian transgression, more or less coeval to the deposition of the Branderfleck Formation in the NCA. Major changes are reported in the Helvetic shelf, where platform drowning occurred during the Albian (e. g. FÖLLMI, 1989; OBERHAUSER, 1995), and within the Ultrahelvetic Gresten Klippen Zone, where pelagic limestones are overlain by Albian black and red shales and marls of the "Buntmergelserie" (FAUPL & WAGREICH, 2000; EGGER et al., 2000). In the Penninic Flysch Units such as the Rhenodanubian Flysch (SCHNABEL, 1992) and the Ybbsitz Zone (DECKER, 1990), no distinct changes in sedimentation are recorded during this time (HOMAYOUN & FAUPL, 1992; WORTMANN et al., 1999), although MATTERN (1999) indicated the change from a northern to a southern sediment supply during the Albian to early Cenomanian in the western Rhenodanubian flysch.

The development of the thrust-related piggyback basins of the Losenstein Formation and the Branderfleck Formation was terminated during the Late Turonian (ca. 91 Ma) by a renewed deformational event, which was followed by basin subsidence of the Gosau Group (WAGREICH, 1995). Pull-apart basin formation, followed by deepening and a change to a new southern source area was interpreted as a result of extension and tectonic erosion along the Austroalpine-Penninic margin (WAGREICH, 1995).

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