A Tethys-wide mid-Carnian (Upper Triassic) carbonate productivity crisis: Evidence for the Alpine Reingraben Event from Spiti (Indian Himalaya)?

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Abstract

A significant mid-Carnian lithological change from carbonatic to siliciclastic sedimentation is described from the Spiti Himalaya, N India. The recorded sudden increase of terrigenous influx occurs at the boundary from the Chomule to the Rama Formation; it is dated within the late Lower Carnian (uppermost aonoides Zone) and isochronous all over the Tethys Himalaya. Bio- and chemostratigraphic data point to a synchrony with changes in carbonate productivity known from on area of wide extent along the NW Tethys: The ‘Reingraben Event’ of the Northern Calcareous Alps (Austria) defines a regional decline in carbonate productivity related to both a platform demise and enhanced siliciclastic input. Laminated shales found in a distal part of the Spiti basin document the onset of anoxic conditions within the deeper shelf synchronous to the ‘Reingraben Event’ in Austria. High-resolution biostratigraphy and identical 813C_carb-curves are therefore in support of a direct link between the change in Spiti and the Reingraben Event.

Our data, combined with a comprehensive review of previously postulated scenarios regarding the ‘Reingraben Event’, suggest linking of climatic change with siliciclastic input and a carbonate productivity crisis, initiated by onset of a relatively humid, megamonsoonal late Lower Carnian “atmospheric system” that affected both the southern and northwestern Tethys and point to a supra-regional trigger of the Alpine ‘Reingraben Event’.

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1. Introduction

Abrupt lithological turnovers from highly productive Ladinian and Lower Carnian subtropical carbonate platforms to a predominantly terrigenous sedimentation is a characteristic feature of western Tethyan shelves during the late Lower Carnian. This event is widely developed; it has been introduced as the so-called “Reingraben turning point” by Schlager and Schöllnberger (1974) and marks the most prominent and severe sedimentary change within the Triassic of the Northern Calcareous Alps and the NW Tethyan passive continental margin (e.g. Hornung and Brandner, 2005). The regional demise of carbonate platforms and associated reef belts has been subject of numerous studies (e.g. Jerz, 1966; Schuler, 1968; Grottenthaler, 1978; Brandner, 1984; Riedel, 1991; Mandl, 2000; Gawlick, 2000; Hornung and Brandner, 2005; Hornung et al., 2005) and can be observed even in the Southern Alps (e.g. Rüver and Zamparelli, 1997; Flügel and Senowbary-Daryan, 2000; Keim et al., 2001; Keim and Brandner, 2001; Stefani et al., 2004; Keim et al., 2006) and NE Iran (Donofrio, 1991; Ruttner et al., 1991; Ruttner, 1993). As similar developments are recorded from Indonesia (SE Tethys; Martini et al., 2000), China (NE Tethys and Panthalassa Ocean; Lehrmann et al., 2005) and the Tethyan Himalaya, one
might assume a Tethyan-wide event. To prove this postulation, it seemed reasonable to study a region as distant as possible from Europe and the NW Tethys margin. As such the Spiti region, which has been accumulated on the former Indian shelf close to the south-eastern end of Tethys, was chosen and studied from litho-, chrono- and chemostratigraphic viewpoints.

The Spiti Valley in the Northern Indian Himalaya (Himachal Pradesh) became famous for its well-exposed and complete Triassic sedimentary succession (thickness circa 1400 m) and was studied initially by Hayden (1904) and Diener (1908, 1912). In recent years, the region has been the subject of many lithostratigraphic (e.g.; Srikantia et al., 1980; Srikantia, 1981; Fuchs, 1982; Bhargava, 1987; Bagati, 1990; Garzanti et al., 1995; Bhargava and Bassi, 1998; Steck, 2003) and biostratigraphic studies (e.g. Goel, 1977; Bhargava and Gadhoke, 1988; Bhargava et al., 2004). Two sites in the Guling (Pin Valley) and the Lalung area (earlier called “Lilang”, Lingti Valley) are known to show an abrupt change from basinal carbonates to thick argillaceous and silty marls in the mid-Carnian. This rapid turnover in sedimentary environment has been briefly described before (e.g. Bhargava and Bassi, 1998), but neither studied in detail nor from a supra-regional perspective.

2. Geological setting

Spiti is situated about 500 km north of Delhi in the northwestern part of the Indian Higher Himalaya (Fig. 1a). The Spiti Valley is about 150 km long and extends from the Kunzum La (4600 m) towards ESE to Sumdo (2965 m) near the Indian–Chinese border. Of numerous valleys draining from SW and NE into the Spiti Valley, the Pin (Guling 2 section) and the Lingti Valley (Lalung section) are the largest (Fig. 1b). At present, the two sections are located in a distance of 30 km along a line perpendicular to the original paleogeographic basin strike. Assuming a 30% tectonic shortening of the former margin (Wiesmayr and Grasemann, 2002) and a minimal basin floor inclination of less than one degree, the more northerly Lalung sequence must have been deposited more distal and considerably deeper than the one of Guling 2.

The Higher Himalaya, containing the sections described herein (Fig. 1c and d), is situated above the Main Central Thrust (MCT) atop the Lesser Himalaya and can be divided into a crystalline unit and the Tethyan sedimentary cover. Forming the northernmost tectonic parts of the Indian continental crust, the sedimentary units of the Tethys Himalaya consist mainly of low-grade metamorphic marine sandstones, shales, dolomites and limestones of Precambrian to Eocene age. These sediments represent the deformed remnants of the N Indian passive continental margin (Fig. 2) (e.g. Steck, 2003). In the region of Spiti and Kinnaur, the Tethyan sediments occur in a large syncline with an NW–SE striking fold axis (“Spiti-Zanskar Syncline”, e.g. Fuchs, 1982; Garzanti et al., 1995) which has been overthrust south-eastwards over the Precambrian “basement” (Bhargava and Bassi, 1998). The Triassic sediments are referred to the Lilang Supergroup (Garzanti et al., 1995; Bhargava et al., 2004), named after the classical 5 km thick stratigraphic succession exposed in the Lingti Valley (Precambrian to Cretaceous; e.g. Griesbach, 1891; Hayden, 1904; Garzanti et al., 1995; Bhargava et al., 2004).

Since the studies of Hayden (1904), several attempts for subdividing the Triassic sedimentary succession of Spiti (Fig. 3) have been developed: Bhargava (1987) and Bhargava and Bassi (1998), for instance, split the “Lilang Group” into eight formations which were correlated to the informal lithostratigraphy sensu Hayden (1904). This scheme was revised and amended by Bhargava et al. (2004), a classification that is followed here.

The Guling 2 section is located about 3 km NE of the Guling village at 3750 m a.s.l. at 32°03′57.63″E/78°07′00″13′N (Figs. 1b and 4). Its lower part includes the top of the Chomule Formation, its upper part the base of the Rama Formation. The section “Guling 1”, described in Bhargava et al. (2004), is the stratigraphic downward continuation of our site. The Chomule Formation, a rhythmical alternation of medium-bedded dark limestone packages (average thickness approximately 1 m) and thin- to medium-bedded grey marls (circa 0.2–1 m), differ significantly from the overlying Rama Formation with its thick silty marls (approximately 2.5 m) interbedded with limy marlstones (circa 0.1–1 m). The sequence dips moderately to SE (ss: 229/24) being a part of a large SW vergent anticlinal structure (“Chhindag Anticline”, length circa 3–4 km; Neumayer et al., 2004). The fold structure developed during the Eohimalayan deformation (Wiesmayr and Grasemann, 2002; Neumayer et al., 2004).

The Lalung section is located approximately 30 km NW of Guling at 32°08′37.07″E/78°15′05.87″N. Exposed at an altitude of about 4050–4100 m a.s.l. and located 2.5 km NNE of the Lalung village (Fig. 1b and 5), the strata dip gently towards W (ss: 265/14). The locality is poorly exposed: the topmost Chomule and basal Rama Formation (5 m thickness) had to be excavated by hand. The overlying strata of the Rama Formation were studied and sampled by Bhargava et al. (2004).

3. Methods

Limestone layers were sampled twice (bottom and top). Additional samples were taken from most of the marly beds. All limestone samples were dissolved in attenuated acetic acid, marls in hydrogen peroxide (concentration 10%). The insoluble residue was washed and fractioned by sieving (coarse: 250 μm; fine: 100 μm). All collected material is stored at the Institute of Geology and Paleontology in Innsbruck, Austria (archive Hornung: “Guling 2” and “Lalung”). To determine microfacies patterns, thin sections were prepared from every limestone bed of both sections. Slightly etched
whole rock samples aided to observe three-dimensional sedimentary and microfacies structures.

Oxygen and carbon stable isotope ratios were measured in all calcareous and marly beds. Due to the absence of articulate brachiopods, four whole-rock powder samples (0.05–0.37 mg) were prepared from every bed using a dental drill. Attention was paid in sampling only homogenous mudstones from fresh surfaces. The powders reacted in
10 ml borosilicate extainers with phosphoric acid after flushing with He. Generated CO_2 was separated from water vapour and was analysed for δ^{13}C_carb and δ^{18}O_carb on a Gas Bench II linked to a ThermoFinnigan Delta plus XL mass spectrometer (Spötl and Vennemann, 2003). The results were calibrated against VPDB (Vienna Pee Dee Belemnite). The aimed reproducibility of isotopes values is ± 0.06‰ (1σ) for δ^{13}C_carb and ± 0.08‰ (1σ) for δ^{18}O_carb.

4. Results

4.1. Fauna

Rare but important macrofossils are ammonoids (Trachyceras n. sp. 1), bivalves (Halobia fluxa) and sparsely occurring pleurotomariid gastropods. Complete faunal lists of the Chomule and Rama formations have been cited in Hayden (1904) and Diener (1912).

Both the Guling 2 and the Lalung successions are rather depauperate in micro- and macrofauna. Most of the studied layers, especially those within the Chomule Formation, are completely devoid of skeletal remains. Only ichnofaunal remains in the form of more or less intensive burrowing can be found throughout all beds.

4.1.1. Macrofauna

As has been noted by Bhargava et al. (2004), the Chomule Formation is less fossiliferous than the underlying Ladinian Kaga Formation. The most common macrofossils are pelagic halobiids, recognisable in thin sections of the limestone beds (Fig. 10f and g) and rare ammonoid fragments (Fig. 7a–c). The basal marly bed of the Rama Formation yielded a small-sized, probably dwarfed ammonoid assemblage (Fig. 6a–f and h–i), gastropods (Fig. 7d–f) and bivalves (Fig. 7i1-3). Layer GN 17, approximately 3.0 m above the Chomule Formation, contained a deformed specimen of Halobia fluxa (Mojsisovic) with faint sculpture elements (Fig. 7h). Layer GN 19, circa 4.0 m above the base of the Rama Formation, yielded a single unspeciﬁed brachiopod (Fig. 7g).
4.1.2. Microfauna

The Triassic of Spiti suffered a significant diagenetic and thermal overprint during the Neogene Himalayan orogenesis (Wiesmayr and Grasemann, 2002). Accordingly, conodonts, found in preliminary studies in the Guling region, show a conodont alteration index (CAI) of 3.0–4.0 representing an average thermal overprint of 160–245 °C (e.g. Epstein et al., 1977). Therefore, most calcite skeletal elements are thus recrystallised leading to a loss of their primary structure. Though most beds yielded skeletal grains, only few of them were classifiable: the basal Rama Formation contains radiolarians (Fig. 8a) rare brachialia of planktic roveacrinids (etching residue: Fig. 8c; thin sections: Fig. 10b). Foraminifers of the Chomule Formation in Guling 2 might be recrystallised *Tolypammina* sp. (Fig. 8b).

4.1.3. Ichnofauna

Nearly all beds are more or less intensively burrowed. Since observed only in thin sections and not in three dimensions, caution is necessary when differentiating between distinct types of bioturbation. Most abundant is mottled burrowing in cm-dimension containing dense peloidal micritic infillings. This *ichnotype 1* is present in almost all studied beds of both sections. Bedding-parallel aligned ichnofaunal remains may be deformed “Fodichnia” (Bromley, 1996) built by an unknown detritus feeder (Fig. 10c and h). Long vertical and unbranched structures may represent “Fugichnia” (Bromley, 1996), whose structures were built during rapidly increased rates of sedimentation followed by new depositional stagnation (Fig. 10c). During stages of low or no sedimentation, firmgrounds developed and were subsequently bored (Fig. 10e). *Ichnotype 2* occurs as small-sized, branched burrows showing a round to oval cross-section (circa 500 μm). Filled with dark micrite, this variation is mostly arranged in cm-sized oval-shaped clusters (Fig. 10d) and can be observed exclusively in the Guling 2 section near the Chomule-Rama formation-boundary. It is equivalent to *Chondrites* or *Phycosiphon* (“Pasichnia”) sensu Bromley (1996). *Ichnotype 3* is present rarely in both sections and consists of either lumps of pellets (Fig. 11f–h) or burrows filled with pellets (Fig. 10i). The distinct boundary between burrow and matrix indicates firmground conditions.
4.2. Biostratigraphy

This study applies the Carnian subdivision as proposed by Krystyn (1978) who uses two substages, the Julian (Lower Carnian) and the Tuvalian (Upper Carnian). Both substages are further subdivided numerically (Fig. 3) but it may be worth mentioning that the Julian 1/I corresponds to the Cordevolian substage of some authors. Biozones shown in Fig. 3 follow Gallet et al. (1994).

Rare ammonoid findings are listed in Bhargava et al. (2004). The basal marly and shaly layers of the Lalung section yielded some Trachyceras s. str., the Lower Rama Formation of the Guling 2 section rare Trachyceras n. sp. 1 sensu Gallet et al. (1994). Whereas the former record has no detailed biostratigraphical significance, Trachyceras n. sp. 1 represents a very narrow time-interval of less than 500 kyr from top of the aonoides Zone (Julian 1/Ic) to the basal austriacum Zone (Julian 2/Ia). Of further importance is the occurrence of Austrotachyceras circa 28 m above the base of the Rama Formation in Lalung (Fig. 9c) representing early Julian 2.

The Lower Carnian strata of Spiti contain generally a meagre conodont fauna. Except for its lowest part, which delivered Budurovitognathus, Metapolygnathus polygnathiformis and Gladigondolella of Julian 1/I age (Bhargava et al., 2004), the Chomule Formation is barren or extremely poor in conodonts. From the upper Chomule Formation of Muth (Upper Pin Valley), Garzanti et al. (1995) noted only M. polygnathiformis (Budurov and Stefanov) and M. auri- formis (Kovács) of indifferent early Julian age. From the very top of the Kalapani Limestone of Kumaun Himalaya, a lithostratigraphic counterpart of the Chomule Formation in Spiti, Chhabra and Kumar (1984) have described a similar conodont fauna together with Metapolygnathus carnicus (Krystyn). This species is present only in a very short time interval of a few 100 kyr at the top of the aonoides Zone (Julian 1/Ic) (Krystyn, 1983; Gallet et al., 1994). The presence of M. carnicus in the Kumaun Himalaya is seen as indicative of the carnicus time-interval in the top-part of the Chomule Formation (Fig. 3).

A few broken specimens of Trachyceras aonoides (Fig. 7a–c), found in the horizon GN 12, indicate a middle aonoides Zone age (Julian 1/Ib) for the upper Chomule Formation. Very important for biostratigraphic considerations are three specimens of Trachyceras n. sp. 1 (Fig. 6g) found at the base of the Rama Formation (GN 15a) as well as 3.50 and 7.30 m above the formation boundary (Fig. 9): hence, this part of the lithological succession can be assigned definitely to the uppermost aonoides Zone representing Julian 1/Ic. Of further importance is the presence of Halobia Xuxa (Mojsisovics) in bed GN 17, again testifying the age dating of the aonoides Zone within the basal Rama Formation. The studied part of the Rama Formation in Lalung was barren in ammonoids, and neither Guling 2 nor Lalung provided useful new conodont data.

4.3. Lithostratigraphy

As noted by Bhargava et al. (2004), the Chomule Formation always forms a cliff in the present topography between underlying Kaga and overlying Rama formations. It consists mainly of bedded dark grey limestones including some marly
and/or shaly intercalations. The limestones can be roughly classified as bioclast-bearing mudstones and bioclastic wackestones (MF 1–MF 3), which occur mostly at the base of the Chomule Formation. The Lalung succession comprises also radiolarian wackestones (MF 4). Garzanti et al. (1995) mentioned common authigenic feldspar. The Rama Formation (= “Grey Beds” sensu Hayden, 1904) was described by Bhargava et al. (2004) as thick-bedded dark shales, intercalated by bioclastic and lithoclastic wacke- to packstones as well as thin-shelled coral and oolitic pack-
stones. The authors mentioned further a local variation of mm-laminated dark-grey shales, the so-called “Paper Shales” (Fig. 9c and d). This interval is approximately 30 m thick and restricted to the basal Rama Formation of Lalung.

Both studied sections expose a distinct lithological change between the two formations (see Fig. 9): whereas the former comprises alternating well-bedded dark-grey limestones and ochre-coloured weathered, fresh grey marly
interbeds, the latter is dominated by thick silty marls (1–2.5 m) intercalated by limy marlstones (Guling 2 section), respectively dark “Paper Shales” including decimetre-bedded limy siltstones (Lalung).

### 4.3.1. Guling 2

The lower part (−9.8 to −3.0 m) of the studied Chomule Formation is built up mainly of an alternating sequence of well sorted mud- and wackestones (MF 1) and interbedded marls, the upper part (−3.0 to 0 m) mainly of graded limestone beds (calciclastic wackestones [MF 2] at the base grading into mudstones [MF 1] towards banktop). The marly interbeds contain mineral grains in silt fraction (quartz, feldspar and large amounts of light-coloured micas) next to scarce bioclasts. Above bed GN 15 (−1.0 to 0 m), the lithologic succession changes rapidly from lime-

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stones to thick-bedded marls (thickness max. 2.5 m) enriched by terrigenous mineral grains. The interbedded dm-sized limy marlstones (thickness max. 1.0 m) consist solely of burrowed elastic packstones (MF 3), dominated by planktic crinoids and sponge spicules. Towards the top of the exposed Rama Formation (approximately 10–12 m), the sily marls get thinner and limy marlstone-interbeds thicker, the latter being composed mainly of MF 3 and MF 1. In the upper Rama Formation (not studied here), the carbonatic content increases slowly (Bhargava et al., 2004).

4.4. Microfacies Types

Four microfacies types are distinguished of which MF 1 dominates the Chomule Formation and MF 3 the Rama Formation (Fig. 9). Both MF 2 and MF 4 occur sporadically in the Chomule Formation.

4.4.1. MF 1: Well sorted mud- and wackestones with sily lithoclasts (Fig. 10a–c and Fig. 11a, c and f–h)

The matrix is characterised by peloids, fused with micrite and tiny algal relics (Tubiphytes). The skeletal fragments are quite small (max. 200 μm) and comprise mainly siliceous sponge spicules, rare micritised radiolarians, foraminifers and undefined shell debris of ostracods and juvenile bivalves. All longish (bio)clasts are aligned parallel to bedding. The orientation of these spicules, filaments and clay flakes defines a very fine depositional lamination (sensu Wilson, 1969), overprinted by micropressure solution due to compaction (Fig. 10c). Nearly all beds are burrowed by ichnotype 1 (Fig. 10a, c and h), more sparsely by ichnotype 2 (Fig. 11b) and ichnotype 3 (Fig. 11h). Well sorted mineral grains of max. 50 μm (“sily lithoclasts”) are composed of calcite, quartz and angular feldspar. MF 1 is dominant particularly in Guling 2, less present in Lalung.

4.4.2. MF 2: Wackestones with sandy lithoclasts (Fig. 10c)

The densely packed and slightly laminated matrix consists of fused pelmicrites showing an increased content of mineral detritus (relative to MF 1). Quartz, rounded feldspar, less mica-shreds and rare calcite crystals (max. 200 μm: “sandy lithoclasts”) are the main mineral constituents. Framboidal pyrite was grown secondarily in pressure solution seams, tension joints and burrows. The pelmicritic texture is, relative to MF 1, somewhat coarser (up to 500 μm). Bioclasts are rare and include sponge spicules, bivalve shell fragments, filaments and small gastropods. The beds are burrowed by ichnotype 1 and 2. MF 2 occurs exclusively in the Guling 2 section.

4.4.3. MF 3: Bioturbated clastic packstones (Fig. 10g and h)

The packstones are composed of common biogenic content (mainly sponge spicules, crinoid fragments, foraminifers and ostracods), abundant calcite clasts, a coarse mixture (200–500 μm) of terrigenous clasts (quartz, feldspar and micas) and dark greyish-coloured micrite. Finely dispersed pyrite was grown secondarily and causes the intense dark greyish tint.

Densely micritisised flakes without internal structure indicate the occurrence of algal encrustations (Fig. 10j). All clasts are aligned chaotically with no preferred orientation. Relative to MF 1 and MF 2, the texture comprises a coarser mixture of significantly more abundant bioclastic detritus and terrigenous material as well as an intensified bioturbation. Ichnotype 2 is more common than ichnotype 1. As observed in the MF-types before, the texture was overprinted intensely by micropressure solution.

4.4.4. MF 4: Radiolarian wackestone (Fig. 11b, d and e)

The defining characteristic of MF 4 is its abundance of calcified spherical radiolarians, either dispersed in a dense micritic matrix (Fig. 11d and e) or enriched in distinctive bio-detritic layers (Fig. 11f). Further notable are small and thin filaments (100–200 μm) as well as bivalve shell fragments (max. 500 μm). Especially the filaments and bivalve relics show a chaotic orientation. Biotic disturbance is scarce and the matrix is much finer (ca. 5 μm) than in MF 1. Mineral grains such as quartz, feldspar and micas occur only accessorially. MF 4 is restricted to the Lalung section.

In summary, the lithological and sedimentary histories of both studied successions differ considerably from each other. Whereas the upper part of the studied Chomule Formation in Guling exhibits a bed-internal grading from MF 2 to MF 1, nearly all Chomule beds in Lalung are composed monotonously of MF 1, except for intercalated radiolarian wackestones (MF 4). Thinly laminated “Paper Shales” in the basal Rama Formation are present only in the Lalung area. Furthermore, the Lalung section comprises significantly decreased thicknesses with respect to the Guling 2 section (Fig. 9a and b) what conforms to a general thinning of the Lilang Supergroup from present southwest to northeast (Garzanti et al., 1995; Bhargava et al., 2004). The microfacies record reflects this by the more common occurrence of radiolarians, the higher content of micrite and the lack of graded horizons in Lalung, all pointing to a shallower, better oxygenated and partly current-influenced environment in Guling 2. The mm-sized and distinct bio-detritic layers (Fig. 11f) including unaligned radiolarians and filaments in MF 4 (ostracods and juvenile bivalve shells) might have been generated by weak currents transporting biogenous relics and concentrating them in small depressions.
Fig. 10. Microfacies of the Guling 2 sequence (thin sections): (a) Mottled burrow with dense micritic infilling and increased organic content relative to the surrounding matrix. Middle part of the Chomule Formation. GN 06; Ichnotype 1; MF 1. (b) Brachialia of a planktic crinoid (*Osteocrinus* sp., compare to Fig. 9c). The white arrow points to pressure solution seams. GN 07; MF 2. (c) Mottled burrows very similar to Fig. 11a. The white arrows point to possible “Fugichnia”. The dashed lines accent vertical and subhorizontal arranged depositional lamination, the black arrow pore water drainage structures. GN 13; Ichnotype 1; MF 1, MF 2. (d) *Phycosiphon* burrows. GN 18; Ichnotype 2. (e) Distinct vertical and unbranched cm-sized burrows indicating firmgrounds conditions. GN 20; Ichnotype 1. (f) Sculptured robust shallow-water bivalve contoured by pressure solution seams (white arrows). GN 07; MF 2. (g) Coarse-grained calciclastic biodetritus and halobid shell debris of the lower Rama Formation of Guling. GN 17; MF 3. (h) Intensely burrowed mudstone of the basal Rama Formation of Guling. GN 16; MF 3. (i) Corg-rich burrow filled with pellets. GN 17; MF 3. (j) Questionable thrombolite algal relics (white arrows) in the basal Rama Formation of Guling. GN 16; MF 3.
4.5. Isotope studies

The average values of isotope ratios obtained from carbonate whole rocks are given in Fig. 6. The δ¹³C_carb data range between 1.5‰ and 2.6‰ VPDB in Guling 2 and between 1.8‰ and 3.1‰ in Lalung. The average of all values plots around 2.2‰ representing a normal marine signature without freshwater input, which would displace the original marine signature to significantly lighter, almost negative values (e.g. Immenhauser et al., 1999). Several distinct shifts can be observed in the Chomule Formation of Guling 2: a relatively flat positive excursion (maximum I) in bed G 05, followed by a second one in the beds GN 08-G 10 (maximum II) and a third one just at the boundary to the overlying Rama Formation (maximum III) (Fig. 12). The limy marlstone interbeds of the latter show a nearly constant trendline at 2.0‰.

The δ¹⁸O_carb data ranging between −7.7‰ to −10.6‰ (VPDB) in Guling 2 and −4.0‰ to −7.4‰ in Lalung are the result of a strong diagenetic overprint (CAI = 3.0–4.0) which was obviously higher in Guling 2 than in the Lalung region. The values are thus not representative for paleotemperature calculation or stratigraphical correlation.

5. Discussion

5.1. Sedimentological interpretation and time steps of the Reingraben Event in Spiti

The siliciclastic event of Spiti can be divided in a pre-, syn- and post-turnover phase. The first and third steps...
imply rather “normal” sedimentary conditions that persisted in the pre- and aftermath of the event. Syn-turnover sediments, however, show conspicuous differences in material and faunistic composition that infer rapid and distinct environmental changes. Note that all bathymetric terms rely on the facies zone (FZ)-scheme given by Flügel (2004).

5.1.1. Pre-turnover (Chomule Formation)

The sparse distribution of neritic and epibenthic faunal components in the upper part of the Chomule Formation (MF 1 and MF 2), the dominance of planktic microfaunal elements (e.g., rovecrinids) and the increased content of clayish mud (MF 4) as well, point to a deposition in an environment on a distal, deep-neritic shelf below the storm wave base (FZ 2) or deeper (?). The alternation of thick limestones and thin marly interbeds can be explained by a cyclic increase and decrease of mud supply (Bhargava and Bassi, 1998). The bed-internal change of the coarse MF 2 grading into fine MF 1 (Figs. 9 and 10c) in the Chomule Formation near Guling might have originated from mud-rich calciturbidites: these “event beds” were probably caused by storms affecting the adjoining shallow shelf and shed as periplatform ooze north-eastwards into the Spiti Basin. The fused peloidal matrix of MF 1 and MF 2 including sparse algal relics (questionable Tubiphytes) infer an intact carbonate factory on the shelf; the thin-bedded argillaceous and silty marls, in turn, might represent the background sedimentation during phases of less carbonate productivity (Bhargava and Bassi, 1998). The silt-sized mineral grains occurring in MF 1, in particular angular feldspar grains, implicate an aeolian transport and, eventually, semiarid to aride climate. Towards the top of the Chomule Formation, the content of terrigenous detritus increases significantly (MF 2). The subangular and well-rounded clasts...
grains now suggest fluvial transportation from the terrigenous hinterland due to increased humidity.

5.1.2. Syn-turnover (Rama Formation)

The predominantly calcareous sedimentation trend within the Chomule Formation changes rapidly at the onset of the Rama Formation where thick-bedded siliceous and dark grey marly layers contain only a few, comparatively thin-bedded limy marlstone horizons. The significantly increased thickness of silty marls in the Rama Formation can be explained (a) by increased amounts of terrigenous input or (b) a rapidly diminished carbonate production rate. High organic bioproductivity and reduced oxidation of organic material leads to the conspicuously dark grey colour of the marls and an increased primary organic content diagenetically transferred into disperse pyrite.

The cm- and dm-bedded limy marlstone interbeds of the Rama Formation (MF 3) differ from the underlying dm- to m-bedded limestone beds of the Chomule Formation (MF 1 and MF 2) in the way that they provide (a) no evidence of shelf-derived components as peloids (as a fused peloidal matrix) and algal remains (questionable Tubiphytes) and (b) that they show a significantly increased clastic fraction in silt and sand size as well as a reduced content in carbonate. The latter infers a diminished carbonate/siliciclastic ratio, i.e. a greater clay supply (Garzanti et al., 1995), and thus a possible demise and subsequent erosion or suffocation of adjacent carbonate platforms. This, unfortunately, cannot be proved since Lower Carnian platforms or reefs are unknown from any part of the Indian shelf (Om Bhargava, pers. comm.; Kiessling et al., 2006). Nevertheless, due to their coarse clastic content (MF 3), the limy marlstones are interpreted as a mixture of reworked carbonatic material resulting from erosion of a pre-existing carbonate shelf, siliciclastic grains of terrigenous origin and the marly pelagic background sedimentation. The formation boundary thus should mark the sharp change from carbonate production to carbonate factory degradation or at least rapidly enhanced supply on terrigenous material. The interbedded packstones (MF 3) within the Lower Rama Formation of both sections seem to record short-lived “events”, such as turbiditic, bioclast-rich debris flows, shed into the basin from shallower shelf regions.

5.1.3. Post-turnover (Rongtong Formation)

The supply of terrigenous material remains high and carbonate productivity low, i.e. perturbed during the whole Rama Formation as described by Bhargava et al. (2004). The first thick limestone packages (coral packstones and peloidal wackestones) appear at the base of the overlying Rongtong-Formation, which, according to Bhargava et al. (2004), is of Upper Carnian (Tuvalian 2) age and indicate a re-established shallow-marine carbonate factory after a break of at least two million years.

5.2. Chemostratigraphic correlation of the Alpine Reingraben Event to the Spiti region

The δ13C_carb-signal, which is relatively insensitive against diagenetic alteration (e.g. Korte et al., 2005), can be retracted over long distances and help to correlate sites of quite different facies areas. Here we compare the carbon isotope trendline of the relatively complete and biostratigraphically well-controlled Lunz-Polzberg section in Lower Austria (47°49’58.79”N/15°01’06.62”E) located within the eastern sectors of the Northern Calcareous Alps (eNCA), directly to the Guling 2 section (Fig. 12). The site exposes conodont-dated Reifling Limestones (peloid-filament wacke- and packstones) of Julian 1/Iib age (auriformis I. Z.; for abbreviations see Fig. 12), followed by cm-bedded, calciturbiditic, radiolarian-rich Göstling Limestones (carnicus R. Z. Julian 1/Iic-2/Ia) and black Reingraben Shales (aus-triacum Zone; Julian 2). The conspicuous demise of the Wetterstein carbonate platform (Alpine ‘Reingraben Event’) took place at the onset of the Göstling limestones, i.e. of the carnicus R. Z. (Hornung and Brandner, 2005; Hornung et al., 2005).

Within the time frame of the auriformis I. Z. and the carnicus R. Z., both the Guling 2 and Polzberg δ13C_carb data provide similar values ranging between 1.0‰ to 3.0‰ and an equal trend including the three major positive excursions described above. Of these, the lithological boundary between Chomule and Rama respectively Reifling and Göstling Formation corresponds to maximum III (Fig. 12). Stable isotopes analyses of the Spiti sections seem not to fit into the Lower Carnian segment of the Triassic isotopic trendline proposed by Korte et al. (2005). This secular curve points to a δ13C_carb rise from 1.0‰ in the Middle Triassic to 3.5‰ (VPDB) in the Carnian including a significant δ13C_carb minimum of 1.5‰ to 2.0‰ in the mid-Carnian.
however without exact age indication. Many sites located within the NCA do confirm this rapid drop of 1.5‰ within the carnicus R. Z. (Hornung and Brandner, 2005; Hornung et al., 2006a). The Lunz section shows the negative δ^{13}C_carb shift a little delayed within the basal austriacum Zone (Fig. 12). This δ^{13}C_carb minimum could be initiated by a reactivated input of δ^{13}C_{org}+anorg into marginal marine environments. This is possible during wet climates causing rapidly enhanced weathering of continental landmasses; riverine influx might have lowered the carbon isotope composition of dissolved inorganic carbon of shallow-marine surface waters thus producing a relative enrichment on δ^{13}C (Holser et al., 1989; Baud et al., 1989; Joachimski and Buggisch, 1993; Joachimski et al., 2002). The NW Tethyan margin was closer to the Eo-Cimmerian collision belt (Brandner, 1984; Sengör, 1984; Golonka, 2002): strong uplift and elevated terrestrial relief near an active continental margin might have initiated this specific isotopic signal. The distant Spiti Basin at the southern Tethyan passive continental margin also shows siliciclastic input, which, however, might have been not as high as on the northern margin. Accordingly, the amount of siliciclastic influx have therefore not been sufficient enough to induce a drop in the δ^{13}C_carb trendline, respectively was not able to change the major ocean chemistry in this region.

5.3. A Tethyan-wide siliciclastic event in the mid-Carnian?

5.3.1. The Spiti database – implications for the Southern Tethys

Contrary to the Northern and Southern Alps, the Spiti Himalaya preserves no shallow shelf or near-shore facies belts. Facies reconstructions are thus restricted to the offshore shelf area. Accordingly, all considerations on the causes of sudden siliciclastic input are highly speculative. Dercourt et al. (1993), for instance, concluded that the increase in continental erosion and humidity was due to a latitudinal drift of the northern rim of Gondwana from high latitudes to the Southern Tropic Zone during Late Paleozoic and Lower Triassic. Garzanti (1999:810) considers “rapid up- and out-building of the continental terrace” in the Carnian, whose weathering load should have caused the observed strongly increased rate of tectonic subsidence by isostatic compensation. Parrish (1999) assumed an increasing sphere of megamonsoonal climate influence for the Southern Tethyan and the adjacent Panthalassa regions. The increase in accommodation space, however, might have also formed during extensional movements in the Southern Tethys deep shelf environment (Blechschmidt et al., 2004) due to a subsequent downward warping of oceanic crust.

Basically, two independent mechanisms may be responsible for such lithological shifts: (a) the demise of carbonate platforms, i.e. degradation of the carbonate factory due to exposure and/or suffocation beneath siliciclastic material, should favour the marly background sedimentation, or, (b) rapidly increased terrigenous input would lead to a relative dilution of carbonate influx without degradation of carbonate platforms. In the first case, we would expect a very sharp boundary from limestones made of periplatform ooze (comprising reef-related organisms and/or peloidal shoals) to marly-dominated, carbonate-depleted sediments that should contain eroded, terrestrial components and lack reefal organisms. In case (b), however, one would expect thick-bedded marls and shales intercalated with wacke- and packstones containing shallow water-derived particles thus reflecting suppressed but intact carbonate factories. Though the limy marlstones of the basal Rama Formation miss the characteristics described in (b) and show an overall low carbonate content, we call for a combination of both possibilities: wet and warm climate that stimulates enforced weathering of continental uplifts and thus the siliciclastic:calcareous ratio. Due to lack of Carnian reefs and platforms within the Tethys Himalaya belt, it is not possible to decide, whether the carbonate factory suffocated because of siliciclastic influx, died due to excess nutrients, subaerial exposure or to a global warming event. The Spiti microfacies data indicate only that, synchronously to the NW Tethys, even the Himalayan carbonate platforms suffered a major productivity crisis.

5.3.2. The Reingraben turnover – a Tethyan-wide siliciclastic event?

Late Lower Carnian terrigenous influx into epicontinental basins and shallow-marine environments are provable in many areas of the NW Tethys and can be merged under the term ‘Reingraben Event’ sensu Schlager and Schöllnberger (1974). Locations are known from the W Tethyan shelf (German Basin, Northern Calcareous Alps, Carpathians and Southern Alps, see Fig. 2), from Turkey, Iran and Afghanistan (Cimmerian Megaterrane, N Tethys, see Fig. 2) and China (NE Tethys). The Aghdarband section (NE Iran) is the most striking site to explain the drastic sedimentary turnover that affected the active continental margin of the N Tethys (Fig. 3). A conspicuous stratigraphical gap on top of the late Lower Carnian volcanoclastic Sina Formation (Donofrio, 1991; Rutten, 1993) is seen as the result of the onset of the Eo-Cimmerian orogeny (e.g. Golonka, 2002). As mentioned above, these convergent movements obviously influenced the sedimentary regime of the wider W Tethyan margin and are documented by the presence of a widespread, but short regression (Hallam, 1995).

The stepwise chronology of several terrigenous influenced stratigraphic units and abrupt declines of carbonate productivity occurring in different locations of the wider NW Tethys fit into the time slice of the uppermost Julian 1 and Julian 2. The event-progression itself, however, is still a matter of debate, but seems to be best preserved within the Austroalpine region: a rapid cessation of carbonate productivity is documented by a sharp lithological boundary in basinal sequences, whose microfaunal assemblages do not record any sign of reefal activity beyond. The aforementioned Göstling member (Fig. 3) atop the Reifling Formation consists of reworked deposits shed as calciturbidites into intraplatform-troughs. This up to 30 m thick succes-
sion is interpreted as basinal lowstand wedge that accumulated during the short late Lower Carnian regressive pulse (less than 500 kyr) causing widespread emersion and erosion of the carbonate platforms (Brandner, 1978, 1984; Henrich, 1984) before the depositional space deepened again.

The overlying Reingraben and Lunz formations (Fig. 3), present in the eNCA, are assumed to be the isochronous southerly counterpart to the Schilfsandstein Formation of the Germanic Basin (Fig. 3) (Tollmann, 1976; Krystyn, 1990; Aigner and Bachmann, 1992). The latter represents a widespread, southward directed fluvial bypass system incising up to 20-30 m deep into underlying evaporitic sediments (Gipskeuper; Aigner and Bachmann, 1992). Traceable in many sites of central Germany, southern England and eastern France, this formation defines one of the largest sedimentary Triassic unconformities within the W and NW Tethys area (Beutler and Szulc, 1999; Beutler et al., 1999). Even in the Southern Alps (Dolomites), whose paleoposition is assumed to be far away (several hundreds of kilometres) of the Austroalpine domain (Flügel, 2002; Kiessling et al., 2006), the Heiligkreuz Formation (Fig. 3) still includes several distinct sandstone layers and thus clear terrigenous-siliciclastic influence (Keim et al., 2001).

The estimated paleoposition of the NW Tethyan margin was at 23 ± 2°N and formed the northern hemisphere-counterpart of the Spiti region, whose position is estimated with 28°S (Figs. 2 and 12) (Stampfli and Borel, 2002; Kiessling et al., 2006). Both regions were thus located in narrowing paleolatitudes and equally affected by the Pangean monsoonal circulation systems (Parrish, 1999). From this point of view, the mechanisms that led to enhanced siliciclastic shedding must have been comparable. Differences between the two studied region, however, may have been persisted in that, that the wider W Tethyan margin has undergone intensified tectonics, which are recorded in the short regression–transgression couplet in mid-Carnian times (Hornung et al., 2006b). We postulate that this regressive pulse coincides with the Eo-Cimmerian collision which initiated elevation of large areas along the N and NW Tethys and forced the generally supposed monsoonal circulation system as suggested by Wilson et al. (1994); Hay et al. (1994) and Parrish (1999). Wet late Lower Carnian climate and enhanced weathering (Simms and Ruffel, 1989; Simms and Ruffel, 1990), erosion and run-off would have caused the onset of the accumulation of siliciclastics along the N and NW Tethyan continental margin leading to a widespread suppression of carbonate productivity. The Southern Tethys, however, remained relatively unaffected by Eo-Cimmerian tectonics: decided evidence for a short-lived regression–transgression couplet in the S Tethys can not be presented owing to not preserved carbonate platforms. However, its passive continental margins even had to suffer enhanced weathering and erosion influenced by enforced Lower Carnian megamonsoonal climate that, finally, was responsible for an isochronous siliciclastic event also in this region of the Tethys.

6. Conclusions

1. Integrated high-resolution biochronological dating proves that the late Lower Carnian siliciclastic turnover in Spiti is time-correlative throughout the Himalaya during a short interval (<300 kyr?) around the top of the aonoides Zone. The event is further concomitant to the onset of widespread clastic sedimentation in the NW Tethyan realm, known there as ‘Reingraben Event’, arguing for

2. A regional if not global impact and distribution of the event.

3. Facies changes reflect an enhanced siliciclastic:calcareous ratio at the boundary of Chomule und Rama formations in the basin and, in addition, support a change from carbonate production to carbonate erosion on the shelf or basin margin. This should have resulted in a scenario well-known from the NW Tethys, i.e. a wider degradation of carbonate platforms, which are assumed to have bordered even the Indian Gondwana shelf.

4. A greater scenario is envisaged, in which a mid-Carnian decline of almost all subtropical and maybe tropical Tethyan carbonate factories may have occurred. This demise probably was initiated by the interaction of global changes in plate dynamics and climate: intensive megamonsoonal climate led to enforced weathering of terrigenous uplifts and to enhanced production of siliciclastics, which shed into the former carbonate-dominated depositional environments during a susposable thermal peak in mid-Carnian time.

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References


