Organic matter and palaeoenvironmental signals during the Early Triassic biotic recovery: the Salt Range and Surghar Range records

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The base and the top of the Early Triassic are marked by second order sequence boundaries (SRT1, SRT8). Within the Early Triassic two third order sequence boundaries could be delineated by means of palynofacies analysis and sedimentology, one near the Dienerian-Smithian (SRT2) and the second one near the Smithian-Spathian boundary (SRT5). The extinction event at the Smithian-Spathian boundary seems to be closely associated to the latter globally recorded sea-level low stand. Five additional sequences of undetermined order (SRT3, SRT 4, SRT5/1, SRT6, and SRT7) are reflected in the sedimentological record of the studied sections.

The observed changes in the composition of the particulate organic matter (POM) indicate a general shallowing upward trend, which is modulated by smaller transgressive-regressive cycles supporting the sedimentologically defined sequences. The POM is mostly dominated by terrestrial phytoclasts and sporomorphs. The strongest marine signal is reflected by increased abundance of amorphous organic matter (AOM) in the lower part of the Ceratite Marls at Nammal (late Dienerian) and Chhidru (earliest Smithian) and the Lower Ceratite Limestone at Chitta-Landu (late Dienerian). AOM of marine origin is characteristic for deeper, distal basinal settings and is preferentially preserved under dysoxic and anoxic conditions, indicating reduced oxygen conditions during these intervals. Up-section transgressive events are reflected by increased numbers of acritarchs, reaching up to 50% of the POM. Well oxygenated conditions and low total organic carbon contents (TOC) continue up to the top of the Early Triassic (Mianwali Formation). The most pronounced terrestrial influx is expressed in the Middle Triassic.

Organic carbon isotope data parallel the carbonate carbon isotope records from the Tethyan realm; therefore, they reflect real global changes in the carbon cycle independent of the OM composition. The biomarker study of the apolar hydrocarbons of three samples from the Nammal section indicates an enhanced bacterial productivity, especially in the Smithian and Spathian, reflected in high relative abundances of hopanes. POM, TOC data and redox sensitive biomarkers together with high resolution biostratigraphy demonstrate that well-oxygenated environmental conditions prevailed in the Early Triassic with the exception of the Dienerian to earliest Smithian interval. The POM assemblages of Late Permian to late Griesbachian age indicate well oxygenated conditions during this time interval. There is no evidence in support of an anoxic event in the late Griesbachian in these sections.
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Elke Hermanna,⁎, Peter A. Hochuli a, Sabine Méhay b, Hugo Bucher a, Thomas Brühwiler a, David Warea, Michael Hautmann a, Ghazala Roohic, Khalil ur-Rehman c, Aamir Yaseenc

a Institute and Museum of Palaeontology, University of Zurich, Karl Schmid-Str. 4, 8006 Zurich, Switzerland
b Massachusetts Institute of Technology, Department of Earth, Atmospheric, and Planetary Sciences, 45 Carleton Street, Cambridge, MA 02139, USA
c Pakistan Museum of Natural History, Garden Avenue, Islamabad 44000, Pakistan

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1. Introduction

Mass extinctions have attracted the attention of scientists for a long time starting with the recognition of mass extinctions by Cuvier in the mid 18th century. The search for the possible causes of the most severe end-Permian mass extinction resulted in a number of
hypotheses. Most of them call for a single overarching cause (e.g. the Siberian Traps) to account for the extinction. The cascading effects triggered by such a single cause are still not fully understood, thus a complex combination of several causes was also proposed for the drastic reduction of biodiversity (e.g. Berner, 2002; Erwin, 2006 and references therein). One of the most frequently proposed cause is a global superanoxic event, which is thought to have affected shallow water areas of the equatorial Tethys and Perigondwanan shelf between the latest Permian and the basal Triassic (Griesbachian). Evidence for this superanoxia has been postulated for the Pakistani sections (Wignall and Hallam, 1993; Wignall and Twitchett, 2002), which are the subject of the present study. These sections offer the rare opportunity to study excellently preserved organic matter (OM), which allows identifying oxygenation conditions and testing the anoxia hypothesis for the entire Early Triassic. Furthermore, in this well preserved OM, lipid biomarkers could be used to support results of the optically investigated POM results. Lipid biomarkers are biosynthetic molecules preserved in the form of resistant hydrocarbon skeletons in sedimentary rocks over geologic time. The molecules represent more or less specifically their precursor organisms and the diagenetic transformations they might have undergone. Hence, they are useful palaeoenvironmental and palaeobiological indicators (Eglinton and Eglinton, 2008; Peters et al., 2005).

During the last decade the focus of research expanded from the extinction event to the time during which life recovered, asking for the duration and key factors of the recovery and the environmental conditions throughout this time. During the initial recovery phase, the diversity of marine ecosystems was still low and only few marine clades recovered rather quickly, such as ammonoids and conodonts (Brayard et al., 2006, 2009; Goudemand et al., 2008; Orchard, 2007). Radiation of these groups started in the Dienerian and continued in discontinuous steps up to the Spathian. Both ammonoids and conodonts suffered a major extinction event in the latest Smithian. Benthic organisms (bivalves and gastropods) recovered at least locally already in the late Griesbachian (Hautmann and et al., in press), but it appears that this early recovery pulse did not form the basis of the main recovery phase that started towards the end of the Smithian (Posenato, 2008). In the Spathian, the number of newly evolved bivalve genera doubled without indication of a major crisis at the Smithian–Spathian boundary (Hautmann et al., 2008). Recovery of reef building organisms such as corals was relatively slow and did not start until the Middle Triassic (Flügel, 2002; Posenato, 2008).

Several proxies indicate profound environmental changes affecting marine ecosystems. Sedimentary records of the basal Triassic show unusual facies development such as microbialites, seafloor precipitated carbonate fans, flat pebble conglomerates and wrinkle structures, which can be explained by unusual ocean chemistry (Baud et al., 2007). These peculiar facies are known from various parts of the globe such as South China (Galletti et al., 2008), Turkey, and the United States (Pruss et al., 2006).

Within the Early Triassic, changes of ammonoid diversity in time and space have been interpreted as fluctuations of the global temperature gradient (Brayard et al., 2005, 2006). Palynological records from Norway and Pakistan suggest changing climatic conditions during the Griesbachian (Hochuli et al., 2010a,b) and a distinct climatic change from humid to dryer climate around the Smithian–Spathian boundary (Galletti et al., 2007a; Hermann et al., 2008). Multiple perturbations of the carbon cycle began with the negative spike at the Permian Triassic boundary and continued throughout the entire Early Triassic indicating changes in the biogeochemical cycles (e.g. Atudorei, 1999; Baud et al., 1996; Galletti et al., 2007a,b; Payne et al., 2004; Richoz, 2006). Positive shifts coincide with substage boundaries such as the Dienerian–Smithian boundary, the Smithian–Spathian boundary, and the Spathian–Anisian boundary. A major negative excursion is recorded in the middle Spathian.

This study contributes to the understanding of the Early Triassic marine and non-marine palaeoenvironment including new data from lithology, particulate organic matter (POM), total organic carbon (TOC), and biomarker analysis. Sedimentological successions are presented together with sequence stratigraphic interpretation and biostratigraphic dating. New C-isotope data (organic and inorganic) allow the correlation with other Tethyan sections (e.g. South China). The Early Triassic in the Salt Range and Surghar Range is a unique archive providing excellent data on C-isotopes, biostratigraphy, sequence stratigraphy, palynofacies, and palynology. This strengthens the role of these areas as a reference area for the study of the Tethyan Early Triassic record.

2. Palaeogeography

The Late Permian palaeogeography with the supercontinent Pangaea located approximately symmetrical to the equator continued to exist throughout the Early Triassic (Stampfli and Borel, 2002; Ziegler et al., 1983). The Salt Range and Surghar Range were situated on the Northern Gondwana shelf about 30° south (Golonka and Ford, 2000; Smith et al., 1994). The Salt Range and Surghar Range are located about 200 km southwest of Islamabad (Fig. 1a). The mixed silicilastic–carbonate shelf of the Indian subcontinent was bordered to the northeast by the Neotethyan rift zone. As a result of the Himalayan orogeny, the sequence of Cambrian to Pliocene rocks was thrust southward along the Salt Range Thrust onto late Tertiary sediments during Quaternary times (Gee, 1989). Permian and Early Triassic rocks are exposed in a structural stack of narrow thrust plates more or less parallel to the Salt Range and the Surghar Range, respectively. The dextral Kalabagh strike-slip fault separates the Salt Range East of the Indus from the Surghar Range West of it. Permian and Triassic strata are well exposed in several transecting gorges, among which Chitta–Landu and Narmia in the Surghar Range and Nammal and Chhidru in the Salt Range were studied (Fig. 1).

3. Methods

3.1. Carbonate and organic carbon isotopes

To analyse the carbon isotope composition (δ13C) of bulk carbonate, 128 samples (51 Nammal, 31 Chhidru, 46 Chitta–Landu) were drilled to produce a fine powder, which was treated with phosphoric acid at 90°C in a Kiel IV preparation device and Gasbench, respectively. Subsequently, the liberated CO2 gas was analysed with a Thermo-Scientific Delta V Plus isotope ratio mass spectrometer. For determination of δ13C of bulk organic carbon, 280 silt and shale samples (100 Nammal, 68 Chhidru, 98 Chitta–Landu, and 14 Narmia) were powdered with a ceramic mortar and treated with 1 N and 3 N hydrochloric acid respectively for at least 24 h to remove carbonates. The homogenised residue was analysed with an elemental analyser connected in continuous flow to a Thermo-Finnigan MAT 253 isotope ratio mass spectrometer. Average reproducibility of analyses, based on repeated measurements of laboratory standards calibrated to NBS 22 (δ13C = −30.03‰), is better than ±0.2‰.

All δ13C values are expressed in the standard δ notation in per mil (%) relative to the international VPDB isotope standard.

3.2. Palynofacies

184 cleaned samples (67 Nammal, 59 Chhidru, 7 (+2) Narmia, and 49 (+3) Chitta–Landu) were crushed and weighed (5–25 g) and subsequently treated with hydrochloric and hydrofluoric acid according to standard palynological preparation techniques (Traverse, 2007). The residues were sieved with an 11 μm mesh screen and mounted for the analysis of the POM. From stew mounts a
minimum of 250 particles per sample were counted. Selected samples have also been used for palynological analysis.

Palynofacies reflects the depositional environment based on the total assemblage of POM (Tyson, 1995). Relative abundances of the different groups reflect firstly oxygenation condition and are therefore a tool to discriminate between oxic and anoxic conditions and secondly document changes in sea level (transgressive–regressive trends), and distance from the shore (Combaz, 1964; Tyson, 1993, 1995). Palynofacies has also been used as a tool for sequence stratigraphic interpretations (e.g. Götz et al., 2008; Pittet and Gorin, 1997; Steffen and Gorin, 1993). To analyse palynofacies patterns, the POM has been grouped into a fraction of terrestrial derived particles and a marine fraction. The terrestrial POM includes translucent and opaque woody particles, intertinite, cuticles, membranes, spores, striate and non-striate bisaccate pollen, and other pollen. The marine fraction includes amorphous organic matter (AOM), acritarchs, foraminiferal test linings, prasinophyceae and other algal remains. AOM consists of structureless aggregates of organic matter derived from phytoplankton or bacteria (Lewan, 1986; Tyson, 1995); dissolved organic matter and faecal pellets can also contribute to the formation of AOM (Tyson, 1995).

3.3. Total organic carbon and Rock Eval analysis

48 samples from the Nammal site were analysed for total organic carbon (TOC) content. Total carbon contents were measured on a CNS Elemental Analyser (Carlo Erba Instruments) and the inorganic carbon contents were determined using a UIC CM 5012 Coulometer. The TOC was calculated from the difference between total and inorganic carbon contents. 12 samples from Nammal were pulverised, and about 100 mg were analysed using a Rock Eval/TOC analyser. The samples are pyrolysed at 300 °C for 3–4 min, and the amount of hydrocarbon liberated from 1 g of rock is noted as parameter S1. Parameter S2 is the amount of hydrocarbon released during the temperature programmed pyrolysis (300–600 °C). TOC is determined by oxidising the pyrolysis residue in a second oven (600 °C in air).

3.4. Lipid biomarkers

Three samples from Nammal were ground to a fine powder (60 to 70 g), and organic compounds were extracted with an accelerated solvent extractor (ASE 200/DIONEX) using a solvent mixture of dichloromethane (DCM) and methanol (MeOH) (4:1, v/v) at 1500 psi and 100 °C. The solvents were evaporated to dryness under a gentle stream of nitrogen (N₂) at 35 °C. The total lipid extracts were fractionated by column chromatography (SiO₂). Two fractions were recovered by the elution of DCM and a mixture of DCM/MeOH (1:1, v/v), respectively. The first fraction was further separated by column chromatography (SiO₂) to yield a fraction of apolar hydrocarbons by elution with n-hexane. Elemental sulphur was removed with activated copper. The apolar hydrocarbon fractions were then analysed by gas chromatography coupled to mass spectrometry (GC–MS).

GC–MS analysis was carried out on a Hewlett Packard 6890 gas chromatograph equipped with an on-column injector and coupled to a Hewlett Packard 5973 Mass Selective Detector mass spectrometer operating in the electron impact mode (70 eV). A fused silica capillary column (DB5ms 30 m × 0.32 mm, 0.25 μm film thickness) with helium as a carrier gas was used. The samples were injected at 70 °C. The GC oven temperature was subsequently raised to 120 °C at 10 °C/min and to 300 °C at 4 °C/min, followed by 20 min isothermal.
4. Stratigraphy and lithology

4.1. Terminology

The geology of the Salt Range and Surghar Range was first studied in the early and middle 19th century. The most important contributions defining and describing the Early Triassic stratigraphy of the Salt Range were published by Waagen (1891, 1895). His lithological subdivision of the Early Triassic sequence is extremely robust and is used in this paper with the minor modifications suggested by Guex (1978) and Kummel and Teichert (1966) (Fig. 2). Subsequent studies concerning brachiopods, conodonts, palynology and ammonoids contributed detailed information about Early Triassic fauna and flora, and the Permian–Triassic-boundary (PTB) interval (Balme, 1970; Kummel, 1966; Kummel and Teichert, 1970; Pakistani-Japanese Research Group (PJRG), 1985; Schindewolf, 1954; Teichert, 1966).

The Early Triassic Mianwali Formation has been divided into six lithological units, which include in ascending order: the Kathwai Member, Lower Ceratite Limestone (LCL), Ceratite Marls (CM), Ceratite Sandstone (CS), Upper Ceratite Limestone (UCL) including the Bivalve Beds (BB), Niveaux Intermédiaires (NI) and Topmost Limestone (TL). The lithologies of the Mianwali Formation consist of siltstones, sandstones, limestones and dolomites. LCL, UCL and TL include several packages of carbonate beds, whereas the CM, CS and the NI are dominated by silt- and sandstones. Several erosional horizons underlined by intraformational breccias and karst features can be observed in some restricted intervals of the sequence. The base of the Middle Triassic (Landa Member) has been included in this study. All formation names for the Early Triassic series are used in the sense of Kummel and Teichert (1966), with modifications of Guex (1978). The latter restricted the Narmia Member to the Topmost Limestone and excluded the Niveaux Intermédiaires and Bivalve Beds (Fig. 2). The "Chidru beds" of Waagen (1891) were formally named Chhidru Formation by Teichert (1966).

The PTB is officially defined by the FAD (First Appearance Datum) of the conodont *Hindeodus parvus* at the GSSP in Meishan, South China (Yin et al., 2001). Based on the lithological succession and the fossil content, the PJRG (1985) split the Kathwai Member in the Salt Range into three units considering the lowermost brachiopod bearing unit of the Kathwai Member as latest Permian in age. This part of the section at Nammal and Chitta–Landu has been eroded. Biostratigraphic and C-isotope data allow dating and precise correlation of the sections.

4.2. Sections and correlation

Three main sections have been studied, namely Nammal and Chhidru in the Salt Range, and Chitta–Landu and a smaller stratigraphic interval at Narmia in the Surghar Range (Fig. 1b). All lithological units (Fig. 2) can be traced laterally across the Salt Range and Surghar Range, with the exception of Chhidru where Triassic rocks younger than the BB have been eroded. Biostratigraphic and C-isotope data allow dating and precise correlation of the sections.

### Mianwali Formation

<table>
<thead>
<tr>
<th>Age</th>
<th>Formation</th>
<th>this study</th>
<th>Guex, 1978</th>
<th>Waagen, 1895</th>
</tr>
</thead>
<tbody>
<tr>
<td>T.</td>
<td>Tredian Fm.</td>
<td>Landa Member</td>
<td>Landa Member</td>
<td>Topmost Limestone</td>
</tr>
<tr>
<td>N.</td>
<td>Narmia Member</td>
<td>Topmost Limestone</td>
<td>Niveaux Intermédiaires</td>
<td>Dolomite Beds</td>
</tr>
<tr>
<td>E.</td>
<td>Mianwali Fm.</td>
<td>Niveaux Intermédiaires</td>
<td>Upper Ceratite Limestone</td>
<td>Bivalve Beds</td>
</tr>
<tr>
<td>M.</td>
<td>Mittiwalli Member</td>
<td>Bivalve Beds</td>
<td>Ceratite Sandstone</td>
<td>Upper Ceratite Limestone</td>
</tr>
<tr>
<td>K.</td>
<td>Kathwai Member</td>
<td>Ceratite Sandstone</td>
<td>Ceratite Marls</td>
<td>Ceratite Marls</td>
</tr>
<tr>
<td>C.</td>
<td>Chhidru Fm.</td>
<td>Lower Ceratite Limestone</td>
<td>Lower Ceratite Limestone</td>
<td>Lower Ceratite Limestone</td>
</tr>
<tr>
<td>P.</td>
<td>Kathwai</td>
<td>Kathwai</td>
<td>Upper Productus Limestone</td>
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*Fig. 2. Lithological units of the Mianwali Formation.*
from this area and other Tethyan sections (Baud et al., 1996; Galletti et al., 2007b; Payne et al., 2004) could be confirmed and precisely positioned in the lowermost bed of the BB in all three main sections (Nammal, Chhidru, and Chitta–Landu, Figs. 3–6). The *Procolumbites* fauna near the top of the BB indicates an early middle Spathian age for the top of the UCL; therefore the Smithian/Smithian boundary lies in the topmost metres of the UCL. Comparison of the ammonoid faunas of Nammal and Chitta Landu above the positive carbon isotope shift indicates a sedimentary gap between the lowermost BB and the *Procolumbites* fauna at Nammal. The *Procolumbites* fauna in the BB at Nammal is younger than the *Tirolites* fauna from the base of the NI at Chitta–Landu. At Nammal, age control of the transition to the Landa Member is suggested by increasing δ13Corg values.

In the Nammal section several reworking horizons and erosional surfaces with characteristic relief and tool marks have been observed. The bases of the CS and TL and the sandstone bed in the middle NI are marked by erosional surfaces (Fig. 3). Reworking horizons mark the base of the UCL and the top of the BB and the TL. Some of these horizons have also been recognised at Chhidru and Chitta–Landu. Erosional features at the base of the CS recognised at Nammal are clearly developed at Chhidru but have not been observed at Chitta–Landu. The erosional surface between the sandstone beds of the topmost NI and the first TL limestone bed at Nammal is also recognisable at Chitta–Landu within the uppermost few sandstone beds of the NI.

4.2.2. Chhidru gorge

At Chhidru the thickness from Kathwai Member to top of BB is about 90 m, thus slightly thicker than the corresponding interval at Nammal (72 m). Here, the Triassic units above the BB are not preserved. At Chhidru it was possible to take palynological samples from the Chhidru Formation up to the BB. Palynological assemblages from the Chhidru Formation indicate a Permian age for this interval. The presence of *Ophiceras* in the Kathwai Member indicates a basal late Griesbachian age; the early Griesbachian ammonoid zone (*Otoceras woodwardi*) has not been recognised. At Chhidru, the transition from LCL to CM is marked by two successive ammonoid-rich limestone beds bracketing the Dienerian–Smithian boundary. The lower limestone bed yields a Dienerian *Prionolobus* fauna and the upper bed a Smithian fauna including *Flemingites planatus* (Brühwiler et al., in press and Fig. 3). The Smithian ammonoid faunal succession with *Flemingites* beds, *Nanomalites* beds, *Priorites* beds and *Glyptophiceras* beds in the CS and the UCL allows for precise biostratigraphic correlations with the Nammal section (Fig. 3).

4.2.3. Chitta and Landu gorges

The two sites are located close together on the southern slope of the Surghar Range. The lithological succession can be traced laterally bed by bed. Here we present a composite section of the two sites. The section ranges from the Chhidru Formation up into the Landa Member. Only minor faults and folds disrupt the section and a total thickness of 155.3 m has been logged, of which 130.3 m are Early Triassic in age. Sampling focused on the Spathian–Anisian interval. In contrast to Chhidru gorge, palynological assemblages from the uppermost part of the Chhidru Formation at Chitta–Landu have a Griesbachian affinity (high abundance of spores, esp. *Krauseelsaporites* sp.) thus indicating regional diachronism of the lithological boundary between the Chhidru Formation and the Kathwai Member. At Landu gorge ammonoids of the top limestone bed of the BB at Nammal (Dienerian). In contrast to Nammal, the topmost thin-bedded limestone of the LCL at Chitta–Landu and Narmia contains *Kashmirites* sp. indicating a Smithian age hence indicating diachronism of the LCL–CM transition. Remarkable is a yellow weathering dolomitic limestone below the *Kashmirites* bed, which is also found at Nammal in the same stratigraphic position but within the CM.

The ammonoid fauna found below the BB (*Anasibirites* sp./ *Wasatchites* sp. and *Glyptophiceras* sp.) allow for a correlation of the upper UCL with the UCL at Nammal. The *Procolumbites* fauna present in the upper BB at Nammal has not been found in the upper UCL at Chitta–Landu. The *Anasibirites/Wasatchites* fauna is followed by a *Tirolites* fauna in the lower part of the NI, which is older than the *Procolumbites* fauna of the BB at Nammal (Fig. 3). In the absence of an adequate ammonoid record in this part of the section, age control for the Early/Middle Triassic boundary is given by palynological assemblages and cheemostratigraphy.

4.2.4. Narmia gorge

At Narmia sampling focused on the Permian–Triassic transition ranging from the upper Chhidru Formation to the basal part of the CM. Total thickness of the studied part of the Narmia section is 17.80 m, with 10.30 m of Early Triassic sediments. Similar to Chitta–Landu, palynological assemblages from the Chhidru Formation have a Griesbachian affinity. Ammonoids in the top limestone beds of the Kathwai Member indicate a Dienerian age. Similar to the Chitta–Landu section, the Dienerian–Smithian boundary falls within the topmost beds of the LCL in Narmia.

4.3. Lithological description

The following lithological description focuses on the Nammal section, complemented and compared to the Chhidru and Chitta–Landu sections.

4.3.1. Chhidru Formation

The uppermost part of the Late Permian Chhidru Formation consists of sandy limestone and white sandstone (White Sandstone Unit of Kummel and Teichert, 1970). The limestones are rich in fossils, especially brachiopods and gastropods. At Chhidru the White Sandstone Unit is replaced by a 3.5 m thick succession of alternating white to grey parallel laminated sandstone and dark grey siltstone. At Chitta–Landu, the transition to the Kathwai Member is marked by a 10 cm thick white medium-grained parallel laminated sandstone with two thin siltstone bands at its bottom and top. At Narmia, the topmost metres of the Chhidru Formation consist of an alternation of limestone and light coloured medium grained sandstone. Two horizons of dark coloured fine-grained sandstone could be sampled, one 80 cm below the Kathwai Member and the other between the Chhidru Formation and Kathwai Member. About seven metres below the Kathwai Member a poorly preserved organic rich siltstone lens is intercalated into the sandstone–limestone alternations (for OM description see Section 5.2.10).

4.3.2. Kathwai Member

At Nammal the 3.5 m thick Kathwai Member consists of sandy dolomites, limestones and calcareous sandstones. The lower part is dominated by brown to rusty weathering, middle to coarse grained and poorly bedded dolomite. Several coquinoid layers can be observed. In all sections the lower dolomitic part exceeds in thickness than the upper part. Towards the top of the Kathwai Member the thickness of beds decreases and the grey to greenish fine grained sandstones/limestones are better sorted. Glaucocite can be observed macroscopically. Sandstone occurs in the Kathwai Member of the Salt Range but not in the Surghar Range sections. At Chhidru the 3.8 m thick Kathwai Member comprises intercalated siltstone lenses in the dolomitic beds. Alternating sandstones and sandy dolomites represent the upper part of the member. In the sections of the Surghar Range the thickness of the Kathwai Member is reduced to about 2 m. Here, the massive orange to purple weathering dolomite is directly overlain by limestone (20 cm at Chitta–Landu and 70 cm thin-bedded at Narmia).
Fig. 3. Sedimentology and age of the sections and sequence stratigraphic interpretation. Prh = Prohungarites (occurrence after Guex, 1978); ammonoid faunas (Brühwiler et al., in press and ongoing work of Ware et al., Bucher et al.); Tr = Tirolites; Pcl = Procolumbites; Gly = Glyptophiceras; Ana = Anasibirites/Wasatchites; Prnt = Prionites; Nam = Nammalites; F. pl. = Flemingites planatus; F. fl. = Flemingites flemingianus; Ver = Vercherites; Cly = Clypeoceras; Par = Paranorites; Kash = Kashmirites; F. bh. = Flemingites bhargavai; Prlb = Prionolobus; Cyr = Gyronites (occurrence after Guex, 1978); Oph = Ophiceras. Additionally the Smithian-Spathian positive peak of the carbonate carbon isotope record is marked.
Fig. 4. Lithology, δ13Corg and δ13Ccarb data and palynofacies record of the Nammal and Chitta–Landu section. Pattern/colour code: see Fig. 5. For lithological signature: see Fig. 3.
Fig. 5. Lithology, $\delta^{13}$C$_{org}$ and $\delta^{13}$C$_{carb}$ data and palynofacies record of the Chhidru and Narmia sections. For lithological signature: see Fig. 3.
4.3.3. Lower Ceratite Limestone (LCL, Mittiwalli Member)

At Nammal the LCL is 1.2 m thick and consists of greyish–greenish limestone with coquimoid lenses, and intervals with alternating sandstone and limestone beds. Stylolites are present in the lower limestone beds. A 5 to 15 cm thick breccia layer is intercalated approximately 0.8 m above the base of the LCL. This grey to rusty coloured breccia consists of angular, poorly sorted limestone intraclasts of up to 10 cm in diameter and shell fragments. Coquimoid limestone beds of bivalve shell fragments, ammonoids and gastropods are found above the breccia. In these beds ammonoids are very unevenly distributed and show obvious signs of mechanical accumulations. In some cases the limestone is laminated. Small scaled burrowing is abundant on several bedding surfaces. At Chhidru, sandstone beds with intercalated limestone layers and lenses characterise the base and the top of the LCL, whose thickness locally amounts to 4.5 m. The middle limestone interval is about 1.2 m thick and contains common bivalve fragments and ammonoids. At Chittal-Landu and Narmia the LCL thickness ranges between 2.4 m and 3.1 m, respectively. It consists of thin limestone beds alternating with dark grey siltstone and lenses with high organic content. Rounded intraclasts and irregular distribution of the OM indicates reworking of these sediments (see Section 5.2.10). The top of the LCL is formed
by two prominent marker beds; the lower one is a yellow weathering dolomitic limestone and the upper one consists of thin-bedded limestone, which can both be traced laterally throughout the entire Surghar Range.

4.3.4. Ceratite Marl (CM, Mittiwali Member)

At Nammal the thickness of the CM reaches 29.7 m. The term “marls” is somewhat misleading as the CM also contain a large proportion of thin bedded siltstone and fine grained siltstone. These commonly display either parallel lamination or cross bedding. Greenish to grey limestone lenses and lenticular limestone beds are intercalated in the lower third of the CM. Sandy layers can be observed at the upper and lower boundaries of most limestone beds. These sandstone layers are laminated or even cross bedded. The carbonate beds vary laterally in thickness. In the lower CM the lenticular limestone beds contain numerous, sometimes imbricated ammonoids. Bivalves are also common. Few coquidoid layers with distinct bases and fining upward grain size are observed. In some horizons small gastropods and fish scales are present. The middle part of the CM is dominated by siltstone; the upper part consists of siltstone with intercalated beds and lenses of micaceous sandstone; carbonate almost become absent as micaceous sandstone appear in the CM. The sandstones and siltstones are mostly of dark grey or green colour. Occasionally plant remains can be found in the siltstones. The much thicker CM at Chhidru (36.1 m) are dominated by grey siltstone with few limestone and sandstone lenses at its base. Additionally the overall grain size appears to be finer and the section comprises only few limestone and sandstone beds (Fig. 3). At Chitta–Landu the CM consist of a 20.4 m thick, dark grey siltstone with fine- to medium-grained mica-rich sandstone lenses. In contrast to Nammal there are no limestone beds in its lower part at Chitta–Landu. Compared to Nammal and Chhidru the thickness of the unit is considerably reduced. Macrofossils are rare except for a few horizons rich in ammonoids and small plant remains in the lower part. Similar lithologies are found at Narma. There, intercalated mica-rich fine-grained sandstones display occasionally hummocky cross stratification.

4.3.5. Ceratite Sandstone (CS, Mittiwali Member)

At Nammal the contact between CM and CS is characterised by a sudden increase in thickness and frequency of sandstone beds. Total thickness of the CS at Nammal is 20.9 m. Medium grained sandstone alternates with fine grained sandstone and siltstone. Sandstone beds are massive, parallel laminated or show trough cross stratification; in a few beds of the upper part hummocky cross stratification is observed. The basal sandstone interval reaches 2.2 m in thickness; individual sandstone beds are up to 50 cm thick. Some beds including the basal one show erosional bases and are followed by fining upward cycles (1.5–2 m scale). Small sized bioturbation is present in several horizons. In the upper part of the CS the few limestone beds/lenses are intercalated. Fossils are generally rare, ammonoids and fish teeth (Colobodus) were found in the upper part. Small plant remains can be commonly found in the siltstones. A bivalve rich layer occurs 1 m below the top of the CS. At Chhidru the CS is slightly thicker (23.6 m). The thickness of individual sandstone beds and overall sandstone content is lower than at Nammal. The sandstone beds are parallel laminated or show trough cross stratification. A few sandy limestone beds are rich in fish remains. 7 m above the base of the unit two successive sandy limestone beds contain abundant specimens of the bellerophonid gastropod Warthia (Kaim and Nützel, 2010, which were referred to as “Stachella” since Waagen, 1895).

The lower part of the 18.5 m thick CS at Chitta–Landu is characterised by an alternation of siltstone, thin sandstone, and limestone beds and limestone lenses. The sandstone beds are massive, parallel laminated or display hummocky cross stratifications; some are bioturbated. The sandstone fraction is increasing towards the top of the unit with alternating fine-grained and medium-grained sandstones. Hummocky cross stratification, trough cross stratification and parallel lamination characterise the sandstone beds. The thickness of individual sandstone beds increases upward. Compared to the Narmal section the thickness of the CS and thickness of individual beds is lower at Chitta–Landu. Another distinct feature of the latter section is the occurrence of several limestone beds in the lower part of the CS and the abundance of sandstone in the upper part. This differs from Nammal, where the CS includes only very few limestone lenses and horizons of diagenetic limestone pods within the sandstones, with a sandstone–siltstone ratio decreasing upward.

4.3.6. Upper Ceratite Limestone (UCL, Mittiwali Member)

At Nammal the 16.4 m thick UCL shows a sharp contact with the underlying CS. At some but not all places, this boundary is and erosional surface marked by a bed containing rusty brown to purple coloured clasts (up to 10 cm in diameter), abundant small bivalve fragments, and small gastropods. The lower part of the UCL is dominated by limestone beds, which are interrupted by few sand- and siltstone beds. The limestone is light grey in colour showing in parts nodular structures. Several coquidoid bivalve layers are present with imbricated shells. The silt- and sandstone fraction is increasing in the middle part of the unit. The upper part of the UCL, termed Bivalve Beds, is again dominated by limestone alternating with sandstone. Here, the limestone is nodular except for the lowestmost bed of the BB which is distinguished by its sharp and planar boundaries. Numerous coquidoid bivalve layers (bivalve backstone) are present; in some beds bioclasts show graded bedding. The BB can be recognised in all sections where the UCL has been studied. At the top of the BB an 20 cm thick limestone bed contains pebbles and clasts in Nammal. In some outcrops of the Nammal gorge this bed shows a corroded surface with open dissolution cavities lined with iron oxides. Elongated but smooth clasts with the same coating of iron oxides occur in this bed. They represent the residues of this karstification episode. At Narma the top of the BB is also affected by karstification. There, the decimetric karstic cavities are filled with sparry calcite (Supplementary Fig. 1). In the lower part of the UCL ammonoids are common; gastropods occur as accumulations in thin (few centimetres thick) layers. Crinoids and plant remains are rare, the latter increase towards the top. Compared to the Narmal section the UCL is thicker at Chhidru (20.6 m). Limestone, sandy limestone and sandstone are alternating with fine-grained parallel laminated sandstone and siltstone. The sandstone fraction is generally higher at Chhidru than at Nammal. The top of the lowestmost bed of the BB is characterised by wave ripples at Chhidru. At Chitta–Landu the UCL is as thick as at Chhidru (20.6 m). Here, the UCL consists of thin-bedded limestone with some massive or parallel laminated sandstone beds. The intercalated siltstones contain limestone lenses, and, in a restricted interval just below the BB, siltstone lenses with high OM content. A thin section of this interval reveals that the OM is distributed heterogeneously and lamination is not orientated uniformly, which indicates reworking (see also Section 5.2.10). Within the BB bivalve shells are more fragmented at Chitta–Landu and of smaller size than at Nammal. Compared to the Salt Range sections, the Surghar Range sections show a higher proportion of limestone throughout the UCL, and the sandstone fraction is reduced. If the uppermost part of the BB is clearly karstified at Narma, no such indication was found in Landu. This suggests a deeper bathymetry in Landu than in Narma.

4.3.7. Niveaux Intermédiaires (NI, Mittiwali Member)

At Nammal the 31.4 m thick NI overlay the Fe-crusts of the topmost UCL beds. Their lower part consists of siltstone, sandy siltstone and sandstone beds (20–30 cm thick). At Nammal an approximately 2.5 m thick prominent sandstone bed with limestone pods containing ammonoids occurs in the middle of the NI, showing tool marks at their base. It is partly massive or shows trough cross
stratification. This bed marks the transition to the upper part of the NI where the sandstone beds increase in number and thickness. The uppermost parts of the NI contain sandstone pebbles of few centimetres in diameter. Here, the grey, yellow, rusty and purple colour of the beds show a variety of sedimentary features such as parallel lamination, channel fills, soft sediment deformation structures and hummocky cross stratification. The thickness of the beds varies laterally. The 35.1 m thick NI consists of an alternation of silstone and sandstone beds; the latter showing parallel lamination, trough cross stratification and bioturbation. A few intercalated calcareous sandstone lenses and diagenetic pods in the lower part contain abundant bivalve corquina and ammonoids. A thick sandstone bed at the boundary with the TL displays a marked erosional base. In the intercalated silstones of the upper NI in both sections, plant remains up to 3 cm in length have been observed.

4.3.8. Topmost Limestone (TL, Narmia Member)

At Nammal the contact between the 14.8 m thick TL and the underlying NI shows an erosional surface. The TL is dominated by an alternation of silstone and sandstone with few limestone intervals. The fine to medium grained sandstone beds are massive or show occasional hummocky cross stratification. Some of the sandstone beds have an erosional base; large scaled bioturbation occurs frequently. Fining upward cycles can be observed on a 0.2–1.0 m scale. The base and the top of the TL are dominated by limestone. Several coquinoind layers (bivalve fragments) and bioturbated horizons have been observed. At Nammal a characteristic feature of the TL is a red coloured bed containing crinoid columnalia within the lower limestone interval. This 15 to 20 cm thick bed can be traced throughout the entire Nammal gorge area. The uppermost part of the TL consists again of a 1 m thick limestone/dolomite bed with a reworking horizon at its top, containing crinoid columnalia accumulated in relief lows. They are associated with common rounded limestone pebbles up to 2 cm in diameter. Ammonoids are found only in the lowest part of this unit together with bivalve remains and crinoid stems. Plant fragments are present in the silstones. At Chitta–Landu, the TL is twice as thick (31.9 m) as at Nammal. The top and base of the unit are dominated by limestone, whereas the middle part of the TL is represented by a mixture of prominent limestone beds, silstone and few sandstone beds. The lowermost limestone interval is thin–to medium bedded and in parts nodular. Bedding surfaces of the limestone are rich in glauconite. Crinoid columnalia, bivalves and bioturbation are concentrated in some horizons, ammonoids are rare. Silstones are intercalated between the larger limestone intervals. Towards the top of the TL the limestone beds show increased evidence for reworking with rounded limestone pebbles and shell fragments. Bedding surfaces often show bioturbation. The upper limestone interval consists of a pisolithic limestone with common occurrence of brachiopods.

4.3.9. Landa Member

The 23.5 m of the Landa Member, studied in Chitta and Landa, respectively, consists of sandstone and silstone. The medium–to coarse-grained sandstone beds are mostly purple, sometimes orange in colour. They are parallel or cross laminated; in one case they display pillow–structures. Most sandstone beds are bioturbated. The thickness of the individual sandstone beds increases towards the top of the Landa Member. The dark grey mica-rich silstones and fine-grained sandstones contain abundant and large plant remains (up to 3 cm). The unit is overlain by the white sandstone of the Khaktiara Member. The transition is marked by a distinct unconformity.

4.4. Interpretation

The most obvious difference between the studied sections is the variability of the thickness of the individual units. A difference is most distinctly expressed in the CM with a variation of 16 m between Chitta–Landu and Chhidru. The CM has similar thicknesses in all sections. Compared to Nammal, the UCL is slightly expanded at Chhidru and Chitta–Landu. The NI and the TL are thicker at Chitta–Landu. Therefore, accommodation rate obviously varied in space and time. The Chitta–Landu section is generally more expanded and probably represents a more distal setting. This is in agreement with the smaller size of the bivalve fragments in the BB and the absence of karstification at the top of the same unit at this locality.

The disappearance of marine invertebrate fossils and the presence of large plant remains in the TL and the Landa Member indicate a general shallowing upward trend, reflecting several sequences with distinct boundary surfaces (SRT1–SRT9, Fig. 3). Not all mentioned erosional surfaces and reworking horizons necessarily correspond to higher-order sequence boundaries (SB). The distinct breaks in sedimentation associated with significant lithological changes are interpreted as SB, such as at the lithological change around the boundary between the Chhidru Formation and the Kathwai Member (SRT1), the sedimentary gap within the BB at Nammal and the lithological change between the BB and the NI at Chitta–Landu (SRT5), and the boundary between the TL and the Landu Member (SRT8). Other surfaces represent SB of undetermined order. All recognised sequences are listed and described in Table 1.

5. Carbon isotopes and organic matter

5.1. Carbonate and organic carbon isotopes

The detailed C-isotope data of the individual sections are displayed in Figs. 4 and 5 and in the Supplementary Table 1. Carbonate carbon isotopes (\(\delta^{13}C_{\text{carb}}\)) of Nammal and Landu have been reported by Atudorei (1999) and Baud et al. (1996). Our data confirm these earlier values and allow a detailed biotstratigraphic positioning of the positive \(\delta^{13}C_{\text{carb}}\) shift at the Smithian–Spathian boundary. The mixed siliciclastic–carbonate lithology of the Mianwali Formation allows only for a fragmentary \(\delta^{13}C_{\text{carb}}\) record. The more complete organic carbon isotope (\(\delta^{13}C_{\text{org}}\)) record features the same chemostratigraphic signal (positive shifts at the Dienerian–Smithian, Smithian–Spathian and Spathian–Anisian boundary, negative shifts in the Smithian and Spathian) as many other C-isotope records (e.g. Galletti et al., 2007a,b; Payne et al., 2004; Richoz, 2006).

5.2. Palynofacies

Palynofacies analysis has been performed on 184 samples and the results for Nammal, Chhidru and Chitta–Landu are shown in Figs. 4 and 5, and in more detail in Fig. 7 and Supplementary Figs. 2 and 3. The POM is well preserved and shows no indication of thermal alteration (alteration scale 2, Batten, 1996).

5.2.1. Chhidru Formation

Appropriate lithologies for palynological studies are rare in the upper Chhidru Formation. The oldest palynofacies (probably late Wuchiapingian age according to PJR, 1985) of this study is reported from the Chhidru site (Supplementary Fig. 2). The POM assemblages are dominated by translucent and opaque phytokecists; additionally bisaccate pollen grains (up to 11% at Chhidru) and spores (up to 15% at Chitta–Landu) occur in most samples. Marine influence is minor, only few acritarchs could be observed at Narmia and Chitta–Landu.

5.2.2. Kathwai Member

The palynofacies of the Kathwai member has also a terrestrial character. In contrast to the Chhidru Formation, the assemblages appear to be clearly dominated by opaque phytokecists (Fig. 7). Palynomorphs are absent from most samples. Only a few remains of bisaccate pollen have been observed in one sample from Chhidru. The assemblages at
Table 1

<table>
<thead>
<tr>
<th>Sequence</th>
<th>Lithological units</th>
<th>Sequence boundary (SB)</th>
<th>Dating of SB and sequence</th>
</tr>
</thead>
<tbody>
<tr>
<td>SRT8</td>
<td>Landa Member</td>
<td>Lithological change from pisolithic limestone of the TL to dark grey siltstone–purple sandstone alternation at Chitta–Landu</td>
<td>Early middle Spathian</td>
</tr>
<tr>
<td>SRT7</td>
<td>TL</td>
<td>Erosional relief between NI and TI, reworking at the base of the TL, change from sandstone to limestone</td>
<td>Smithian</td>
</tr>
<tr>
<td>SRT6</td>
<td>Upper part of the NI</td>
<td>Erosional surface at the base of the thick sandstone bed in the middle NI at Nammal.</td>
<td>Spathian</td>
</tr>
<tr>
<td>SRT5</td>
<td>BB (including lower part of NI at Chitta–Landu)</td>
<td>Significantly increased sandstone ratio up-section at Chhidru and Chitta–Landu.</td>
<td>Smithian</td>
</tr>
<tr>
<td>SRT4</td>
<td>UCL including lowermost bed of the BB</td>
<td>Reworking horizon (reworked bioclasts) at the base of the UCL</td>
<td>Smithian</td>
</tr>
<tr>
<td>SRT3</td>
<td>CS</td>
<td>AOM, acritarchs and translucent phytoclasts increase reaching values up to 30%. A single sample from the lowest part of the CM at Narmia displays reduced AOM contents (34%) associated with increased acritarch percentages</td>
<td>Permian–Griesbachian</td>
</tr>
</tbody>
</table>

5.2.3. Lower Ceratite Limestone (UCL)

The POM composition described for the Kathwai Member of Narmal and Chhidru continues unchanged into the LCL at Narmal and Chhidru. In contrast to that the palynofacies of the LCL at Narmia and Chitta–Landu in the Surghar Range is characterised by high abundances of AOM. At Narmia AOM contributes up to 88% to the POM. In these two localities minor components are acritarchs, sporomorphs, and phytoflagellates (Fig. 5 and Supplementary Fig. 3).

5.2.4. Ceratite Marl (CM)

At Narmal and Chhidru the lithological change from LCL to CM coincides with a distinct change in palynofacies (Supplementary Figs. 2, 3). The lower third of the CM at Narmal shows a clear predominance of AOM (up to 82%, mean value 54%), acritarchs and sporomorphs occur regularly. The main interval of the CM at Narmal is characterised by minor traces of AOM. Here, the composition of the POM is strongly dominated by phytoflagellates (mainly translucent woody particles), sporomorphs and acritarchs (mean 21%). In the topmost CM samples acritarch contents increase abruptly to about 50%. At Chhidru (Fig. 5, Supplementary Fig. 2) the predominance of AOM is not as pronounced as in Narmal. Peak values of AOM reach 30% at the very base of the CM and decrease rapidly. Acritarchs and translucent phytoflagellates are prominent elements of the POM composition and sporomorphs occur in limited quantities in this lower part. With decreasing AOM content, acritarchs increase and remain at mean values of 29%. The spore content (up to 26%) increases up-section. The POM of the basal CM at Chitta–Landu is characterised by translucent woody particles. Acritarchs, spores and bisaccate pollen are common, whereas AOM represents only a minor component. Compared to the high AOM values in the UCL the sample from the lowest part of the CM at Narmia displays reduced AOM contents (34%) associated with increased acritarch percentages. Membranes of uncertain origin amount to 17% of the POM.

5.2.5. Ceratite Sandstone (CS)

At Narmal the basal interval of the CS is dominated by acritarchs (mean 35%) without changes to the topmost CM. Phytoflagellates and sporomorphs are regularly recorded. Towards the top of the CS acritarch numbers decrease and phytoflagellates are dominant. At Chhidru the POM assemblage of the CS is comparable to those of the underlying CM. Remarkable is an acritarch spike in the lower third of the CM with values up to 47%. Towards the top of the CS the spore content increases reaching values up to 30%. A single sample from the CS from Chitta–Landu shows a similar composition to those from the basal CM.

5.2.6. Upper Ceratite Limestone (UCL)

The dominance of phytoflagellates continues into the UCL at Narmal. Spores reach their highest abundances in the lower half of the UCL (up to 40%). The upper half of the UCL displays slightly increased values of acritarchs, bisaccate pollen and minor amounts of AOM. At Chhidru the assemblage shows relatively high spore abundances at the base of the UCL decreasing slightly towards the top of the unit. Two acritarch spikes (up to 42%) have been recorded within the UCL. The POM in the lower part of the UCL at Chitta Landu consists of nearly equal parts of spores (up to 33%), opaque and translucent woody and acritarchs together with low percentages of AOM and membranes. The upper part of the UCL is characterised by increased numbers of AOM and a distinct membrane spike, which ranges up to the UCL–NI transition.

5.2.7. Niveaux Intermédiaires (NI)

The NI are characterised by almost exclusive terrestrial palynofacies assemblages with high abundances of translucent and opaque palynomorphs, and sporomorphs. The NI between Narmia and Chhidru are almost identical. No appropriate samples could be obtained at Chitta Landu and Narmia.
Fig. 7. Lithology, TOC and detailed palynofacies record of the Nammal section. Relative abundances of individual particle categories are displayed together with TOC values. The ratio of terrestrial to marine (black to white) is given together with the sequence stratigraphic interpretation. Dashed grey lines in SRT1 indicate lower order sequences. LST = lowstand system tract, TST = transgressive system tract, HST = high stand system tract. For lithological signature: see Fig. 3.
phytoclasts and sporomorphs. Due to coarser grained lithologies and therefore poor preservation some assemblages consist exclusively of plant debris. Low percentages of acritarchs and AOM are found in the basal and topmost part of the Nammal section and throughout the NI of the Chitta–Landu section (Supplementary Fig. 3).

5.2.8. Topmost Limestone (TL)

The palynofacies of the TL of Nammal and Chitta–Landu is dominated by woody particles. Sporomorphs are common and acritarchs occur in smaller numbers (8% at Nammal and 15% at Chitta–Landu).

5.2.9. Landa Member

The POM of the Landa Member at Chitta–Landu is dominated by translucent and opaque woody particles. Among the sporomorphs the non-striate bisaccates are well represented in all samples (Supplementary Fig. 3). In all samples spores are present with variable abundances. Large particles of cuticles with well preserved stomata are a distinct feature of the POM samples of the upper part of the Landa member. Acritarchs are only few but do persist throughout the member. A single sample from the basal part of the Landa Member at Nammal shows a similar POM composition as described above.

5.2.10. Organic rich rocks

A 30 cm thick, continuous, organic rich siltstone horizon within the late Smithian has been traced laterally from Landu to Narma. It is bracketed by the Wasatchites fauna below and the Glyptophiceras fauna above. It was previously recognised by the PJRG (1985) only in Narma area and was then referred to as “coaly shale”. Thinner and laterally more limited similar organic rich siltstone horizons also occur in the LCL of Narma and Landu area. These horizons may locally contain angular to subangular extraclasts several millimetres in size. Two samples from Narma and three samples from Chitta–Landu have been prepared in order to determine the POM. The POM is characterised by strongly degraded amorphous organic matter. The dark brown colour of the POM corresponds to a medium stage of maturation (oil window level). Phytoclasts and palynomorphs are very rare and badly preserved. Hence, it is difficult to determine the age of the OM. The difference in thermal alteration and preservation of this OM contrasts with that from enclosing rocks. All of the available evidence suggests that this OM is laterally transported and even this OM contrasts with that from enclosing rocks. Due to coarser grained lithologies and therefore poor preservation some assemblages consist exclusively of plant debris. Low percentages of acritarchs and AOM are found in the basal and topmost part of the Nammal section and throughout the NI of the Chitta–Landu section (Supplementary Fig. 3).

5.3. Interpretation of palynofacies results

Sedimentary sequences are reflected in changing dominance of the two main POM groups (marine/terrestrial ratio, Fig. 7, Supplementary Figs. 2 and 3) and the characteristic succession of POM assemblages within the sequences. Lowstand deposits are characterised by predominance of terrestrial POM. Increasing marine POM indicates transgressive deposits and early highstand deposits. Highstand deposits show decreasing marine influence up-section and corresponding increase of terrestrial POM. Palynofacies together with sedimentological observations can be used to describe the succession of sequences in a sedimentary sequence. However, the amount and composition of the terrestrial POM fraction depend not only on sea-level but also on hydrological features of the catchment area. For the studied sections measurements of palaeocurrent directions (ranging between NNE and NW at Nammal, Narma, and Chhidru) indicate relatively stable directions of sediment supply (PJRG, 1985). In our material several minima and maxima of the marine/terrestrial ratio coincide with, or occur close to lithological changes.

In the studied sections the composition of the POM indicates a general shallowing upward trend (Fig. 7, Supplementary Figs. 2 and 3), which is modulated by smaller transgressive-regressive cycles (Fig. 7, Supplementary Figs. 2 and 3). The distribution of the POM indicates lowstand deposits of the lowermost sequence SRT1 in the Kathwai Member and the LCL of Nammal and Chhidru, which are characterised by phytoclast dominance. A rapid transgression is expressed by high abundances of AOM in the LCL of Chitta–Landu and Narma and the lowermost CM at Nammal and Chhidru. In the studied sections the abundance of AOM indicates not only the strongest marine signal but also dysoxic conditions. The proportion of AOM in shallow shelf settings is generally low but increases towards offshore basinal settings (Leckie et al., 1990; Steffen and Gorin, 1993; Tyson 1993, 1995; Wood and Gorin, 1998). During transgressions, shelf areas may fall below the storm wave base, and thus reducing water mass mixing. This may allow the establishment of water mass stratification and increase the preservation for AOM (Tyson, 1995). The top of SRT1 is indicated by the decrease of the AOM content at Nammal and Chitta–Landu coinciding with a yellowish dolomitic limestone bed near the Dienerian–Smithian boundary. At Nammal the high sampling resolution allows for discrimination of lower order sequences within SRT1. The corresponding sequence boundaries are reflected in the higher terrestrial to marine ratios of the sampled limestone beds (NA 201AI0 and NA208AI0); they are marked with dashed grey lines in Fig. 7. At Chhidru the top of SRT1 is not clearly expressed (Fig. 5). Here AOM is gradually replaced by acritarchs and therefore the marine POM signal remains constant throughout the CM. Dysoxic conditions can be inferred for the late Dienerian of Nammal, the Dienerian of Narma and Chitta–Landu and to a lesser degree in the lowermost Smithian of Chhidru. AOM is rapidly degraded under oxic conditions (Jannasch, 1991; Pilskaln, 1991) but is well preserved under dysoxic to anoxic conditions in modern marine settings and ancient anoxic basins (Tissot and Pelet, 1981; Tyson, 1993, 1995). Since the establishment of dysoxic to anoxic conditions depends on various factors, such as climate and oceanographic constraints, we also consider sedimentological features to support our palynofacies interpretation. One of the possible processes that could lead to enhanced deposition of AOM during stable sea-level could be the upward migration of the oxygen minimum zone, thus providing the low oxygen environment for the preservation of AOM. Considering the sedimentary record of the aforementioned intervals (Sections 4.3) the increase in AOM abundance is accompanied by the disappearance of glauconite, as observed in the Kathwai Member, by a change to fine-grained lithologies as in the LCL (Chitta–Landu and Narma) and lower CM (Chhidru and Nammal), and the absence of sedimentary structures indicating shallow water settings, such as hummocky cross stratification as documented in the CS. Therefore, in these cases the high AOM abundances can be directly related to deepening of the depositional environment.

At Nammal and Chhidru, the subsequent SRT2 is characterised by reduced marine influence in the middle part of the CM (lowstand) followed by acritarch dominated transgressive and highstand deposits in the uppermost CM. Hence, AOM occurs only in minor percentages or is absent in the POM up-section. Together with rich marine faunas, the absence of AOM indicates well oxygenated conditions from the middle CM onward. As indicated by sedimentological data, the base of the CS marks the SRT3 SB. The palynofacies data show increased terrestrial influence in this interval (Fig. 7, Supplementary Fig. 2). At Nammal and Chhidru, the POM lowstand signal of SRT4 can be observed in the lowermost UCL corresponding to the sedimentologically defined sequence in the UCL. Transgressive and highstand deposits are indicated by increased acritarch abundances in the UCL at Nammal, Chhidru and Chitta–Landu (Fig. 7, Supplementary Figs. 2 and 3). Because of the sedimentary gaps within the BB the SRT5 SB is weakly expressed in the palynofacies record since lowstand and transgressive deposits are missing; the signal of SRT 4 highstand is immediately followed by the SRT 5 highstand. Because of the terrestrial POM predominance in the NI SRT5, SRT5/1, and SRT6 are undistinguishable based on palynofacies data. Transgressive and highstand deposits of SRT6 are expressed in the
TL at Nammal and Chitta–Landu by increased numbers of acritarchs. Compared to Nammal the N at Chitta–Landu displays a higher acritarch content, which confirms a more distal position of the latter section. The SRT8 SB is close to the lithological change between the TL and the Landa Member coinciding with the SB defined by lithology.

5.4. Total organic carbon (TOC) content and Rock Eval analysis

The TOC values at Nammal are generally low, ranging between 0.07% and 0.61%, with a mean value of 0.28% (Fig. 7). TOC normally correlates well with the relative abundance of AOM in the sediment (Tyson, 1995), meaning that high TOC values usually indicate the presence of well preserved AOM. Therefore, the very low TOC values of the AOM-rich intervals (Dienierian) of the section are unexpected. Three samples from the AOM-rich interval have also been analysed using Rock Eval pyrolysis (Table 2). These TOC values of the AOM-rich interval range between 1.63% and 3.08%, which is in agreement with using Rock Eval pyrolysis (Table 2). These TOC values of the AOM-rich intervals (Dienerian) of the section are unexpected.

The major molecular features are summarised in Table 3.

Table 2

<table>
<thead>
<tr>
<th>Lithological unit</th>
<th>Sample</th>
<th>TOC [%]</th>
<th>HI [mg HC/g TOC]</th>
<th>OI [mg CO2/g TOC]</th>
<th>Tmax [°C]</th>
<th>S1 [mg HC/g]</th>
<th>S2a [mg HC/g]</th>
<th>S2b [mg HC/g]</th>
<th>S3</th>
</tr>
</thead>
<tbody>
<tr>
<td>Niveaux Intermédiaires</td>
<td>NA102</td>
<td>0.37</td>
<td>3</td>
<td>105</td>
<td>437</td>
<td>0.00</td>
<td>0.01</td>
<td>0.00</td>
<td>0.39</td>
</tr>
<tr>
<td>NA100</td>
<td>0.53</td>
<td>11</td>
<td>83</td>
<td>439</td>
<td>0.00</td>
<td>0.06</td>
<td>0.00</td>
<td>0.00</td>
<td>0.45</td>
</tr>
<tr>
<td>NA85</td>
<td>0.25</td>
<td>0</td>
<td>161</td>
<td>437</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.40</td>
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<tr>
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<td>167</td>
<td>437</td>
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<td>0.00</td>
<td>0.00</td>
<td>0.38</td>
</tr>
<tr>
<td>NA77</td>
<td>0.45</td>
<td>62</td>
<td>92</td>
<td>438</td>
<td>0.00</td>
<td>0.28</td>
<td>0.00</td>
<td>0.00</td>
<td>0.41</td>
</tr>
<tr>
<td>NA73</td>
<td>0.15</td>
<td>0</td>
<td>208</td>
<td>437</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.30</td>
</tr>
<tr>
<td>Ceratite Sandstone</td>
<td>NA57</td>
<td>0.37</td>
<td>39</td>
<td>112</td>
<td>437</td>
<td>0.00</td>
<td>0.14</td>
<td>0.00</td>
<td>0.42</td>
</tr>
<tr>
<td>Ceratite Marls</td>
<td>NA55</td>
<td>0.10</td>
<td>0</td>
<td>231</td>
<td>437</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.23</td>
</tr>
<tr>
<td>NA213</td>
<td>0.16</td>
<td>0</td>
<td>196</td>
<td>429</td>
<td>0.02</td>
<td>3.54</td>
<td>0.00</td>
<td>0.00</td>
<td>0.32</td>
</tr>
<tr>
<td>NA301</td>
<td>0.27</td>
<td>263</td>
<td>25</td>
<td>429</td>
<td>0.02</td>
<td>5.44</td>
<td>0.00</td>
<td>0.00</td>
<td>0.52</td>
</tr>
<tr>
<td>NA202</td>
<td>0.08</td>
<td>443</td>
<td>14</td>
<td>429</td>
<td>0.18</td>
<td>13.61</td>
<td>0.00</td>
<td>0.00</td>
<td>0.42</td>
</tr>
<tr>
<td>NA200</td>
<td>1.63</td>
<td>213</td>
<td>59</td>
<td>441</td>
<td>0.04</td>
<td>3.47</td>
<td>0.00</td>
<td>0.00</td>
<td>0.97</td>
</tr>
</tbody>
</table>

In NA202, the apolar hydrocarbon fraction is dominated by \( \text{C}_{28} \) and \( \text{C}_{29} \) homologues, dominated by the immature stereochemistry of \( \text{R} \) and \( \text{S} \) stereomers (up to 0.62 for highly mature OM, Peters et al., 2005). The ratios calculated for our samples (Table 3) indicate that the ratio \( \text{S}/(\text{S}+\text{R}) \) for \( \text{C}_{27} \) is not detectable. However, the \( \text{C}_{22} \) 22S/(22S+22R)-hopanes ratio is extremely high compared to steranes (up to 15 times). The hopane distributions are dominated by the \( \text{C}_{29} \) homologue and show a progressive decrease with increasing lateral chain-length up to the hopanes in \( \text{C}_{32} \). There is a slight odd-over-even predominance within the long-chain n-alkanes as shown by the CPI (Table 3). Some isoprenoids were detected in low abundance, with Ph->Pr (Table 3). An unresolved complex mixture (UCM) observed in the mass range from 254 (n-C14) to 324 (n-C21) suggests the presence of numerous isomers of short-chain branched alkanes. In samples NA 55 and NA 88 the relative abundance of hopanes is extremely high compared to steranes (up to 15 times). The hopane distributions are dominated by the \( \text{C}_{29} \) homologue and show a progressive decrease with increasing lateral chain-length up to the hopanes in \( \text{C}_{32} \). \( \text{C}_{29} \) and \( \text{C}_{30} \) hopane configurations are observed in equal abundance. The maturity configuration of the \( \text{C}_{27} \) hopane, i.e. \( \text{Tm} \), is not detectable. However, the \( \text{C}_{25}/(\text{C}_{25}+\text{C}_{22}) \)-hopane ratio is about 0.55 (Table 3). No tricyclic and tetracyclic terpanes were found. The sterane distributions mainly consist of the \( \text{C}_{27} \) homologues and diasteranes were not detected.

5.6. Interpretation of biomarkers

During OM diagenesis the original biological configuration of some biomarkers evolves to thermally stable stereochemistries. Thus ratios of isomers give an insight into the level of maturity reached in the studied samples. The maturity of the OM is most reliably reflected in the ratio \( \text{SI}/(\text{SI}+\text{RI}) \) for \( \text{C}_{15} \)-hopanes since isomerisation at the \( \text{C}_{27} \)-R position occurs very early during OM diagenesis, leading to a mixture of R and S stereomers (up to 0.62 for highly mature OM, Peters et al., 2005). The ratios calculated for our samples (Table 3) indicate that they have been exposed to a light thermal stress, with the slightest...
Fig. 8. Total ion currents of the apolar hydrocarbon fraction of the samples NA 202 (a), NA 55 (b) and NA 88 (c) from the Nammal sections. UCM = unresolved complex mixture. For each hopane homologue, the αβ-stereochemistry is followed by the βα-stereochemistry.
In NA 202, the high abundance of Pr and Ph, often considered as derived from the side chain of the chlorophyll, is likely to rework porphyrin in the organic extracts and the absence of a significant amount of porphyrin in the organic extracts and the presence of isoprenoids, suggest that one or more of their precursors might have disappeared from the basin after the Dienerian. Given the decrease of the marine fraction in the POM assemblages after the Dienerian, the isoprenoid hydrocarbons in NA 202 are most likely of marine origin.

The relative amount of hopanes compared to steranes is a generalised proxy for the relative contribution of bacterial versus eukaryotic biomass. Compared to NA 202 this ratio is about five times higher in NA 55 and about 15 times higher in NA 88. High bacterial input has also been recorded in sediment preceding the end-Permian mass extinction event (Cao et al., 2009). Bacteria are ubiquitous in marine and terrestrial environments. However, a high hopane/sterane ratio is more indicative of a terrigenous or microbially reworked OM. In addition, some specific characteristics in the hopane distribution are usually associated with a land plant dominated sedimentary environment such as low tricyclic terpanes compared to hopanes, a high C31/C32-hopane ratio, and as in NA 55 and NA 88, a high C30/C30 hopane and Ts/(Ts+Tm) hopane ratios (Fig. 8 and Table 3). However, these ratios might have also been affected by the lithology (Peters et al., 2005). Even if the bacterial input is confirmed by the biomarker record its origin (marine or terrestrial) cannot be confidently determined.

At low level of OM maturity, the predominance of short chain n-alkanes, slight in NA 202 and stronger in NA 55 and NA 88, suggests a substantial contribution of marine OM. However, according to the palynofacies study, the marine contribution is expected to be higher in NA 202. Short-chain n-alkanes are most commonly interpreted as being synthesised by marine algae, but bacterial precursors are also known (Collister et al., 1994). Given the high hopane/sterane ratio in NA 55 and NA 88, bacteria, whether terrestrial or marine, might be the main precursors of the short-chain n-alkanes.

When exclusively derived from the lateral chain of chlorophyll, the Pr/Ph is used to describe redox conditions in the depositional environment (Didyk et al., 1978; Peters et al., 2005). However, in immature samples this ratio is believed to be unreliable to assess redox conditions (Alexander et al., 1981; Volkman, 1986) but mainly depends on the origin of the OM. Therefore, in NA 202 the high Pr/Ph is unlikely to reflect oxic palaeoenvironmental conditions. The detection of hopanes up the C31 homologue (Fig. 8) indicates instead favourable conditions of OM preservation.

In NA 55 and NA 88 low amounts of Pr and Ph associated with a low Pr/Ph suggest low oxygen levels, although dependence on source and maturity cannot be ruled out. However, the detection of hopanes only up to C31 suggests a less well preserved OM in these two samples compared to NA 202.

Although the OM is well preserved in the three samples there is no indication for anoxia. Biomarker distribution indicates that the level of oxygen might have been slightly higher during deposition of samples NA 55 and NA 88.

### Table 3

<table>
<thead>
<tr>
<th>Samples</th>
<th>CPI</th>
<th>Pr/Ph</th>
<th>Ts/(Ts+Tm)</th>
<th>C27/C29</th>
<th>C23 αβ C22S/(22R+22S)-hopanes</th>
<th>C29 αβ C22S/(22R+22S)-sterane</th>
<th>C28/C29 αβ C22S/(22R+22S)-sterane</th>
</tr>
</thead>
<tbody>
<tr>
<td>NA 202</td>
<td>1.29</td>
<td>2.7</td>
<td>0.17</td>
<td>0.60</td>
<td>0.25</td>
<td>0.46</td>
<td>0.79</td>
</tr>
<tr>
<td>NA 55</td>
<td>1.31</td>
<td>0.6</td>
<td>n.a.</td>
<td>1.27</td>
<td>0.48</td>
<td>0.56</td>
<td>0.71</td>
</tr>
<tr>
<td>NA 88</td>
<td>1.35</td>
<td>0.6</td>
<td>n.a.</td>
<td>1.18</td>
<td>0.47</td>
<td>0.55</td>
<td>0.71</td>
</tr>
</tbody>
</table>

n.a. not applicable.

a CPI = Carbon preference index Bray and Evans, 1961 defined as:

\[
\text{CPI} = \frac{1}{2} \times \left( \frac{\text{C}_{27} + \text{C}_{29} + \text{C}_{31} + \text{C}_{33}}{\text{C}_{25} + \text{C}_{27} + \text{C}_{29} + \text{C}_{31} + \text{C}_{33}} \right)
\]

The index was calculated based on the area of the peaks of the n-alkanes measured on the ion current m/z 57.

b The ratios of hopanes were calculated based on the corresponding peak areas measured on the ion current m/z 191.

c The ratios of hopanes were calculated based on the corresponding peak areas measured on the ion currents m/z 268 and 282.
6. Discussion

6.1. Sequence stratigraphy (see Table 1)

Both sedimentological data and palynofacies analysis indicate changes in sea-level during deposition of the Early Triassic sequence in the Salt Range and Surghar Range. In the studied sections nine sequences have been identified. A comparison of these sequences with the global sequence framework of the Triassic (Embry, 1997) and the sequence interpretation of the Early Triassic in the Namal gorge after Haq et al. (1988) shows their position within the global sea-level curve (Fig. 9). Gianolla and Jacquin (1998) compared the detailed sequence stratigraphy of the boreal realm with the framework of the Tethyan realm, which allows linking our sequences to a global sequence stratigraphic scheme. Two second order and two third order sequences can be identified with relatively high confidence. The significance of five additional sequences is less clear. They might represent fourth order sequences or parasequences.

In the Canadian Sverdrup Basin Embry (1997) described four sequence boundaries between the Permian–Triassic boundary and the Spathian–Anisian boundary. These sequences could be recognised in many other sections such as Svalbard, Germany, Italian Alps, East Siberia, Northern Himalayas, and Western USA and are therefore interpreted as sequence boundaries of global significance (see also Gianolla et al., 1998; Gianolla and Jacquin, 1998; Skjold et al., 1998). The basal Triassic second order sequence is marked globally by a major depositional shift at the PTB. In the Pakistani sections the transition from the Chhidru Formation to the Kathwai Member near the PTB is marked by an erosional surface and a change from a sandstone–limestone alternation to the dolomite lithologies and sandstone–siltstone succession (SRT1, Table 1 and Fig. 9).

The third order sequence boundary in the latest Dienerian or at the Dienerian–Smithian boundary, respectively, is marked by a change in depositional regime or conformable transgressive surfaces (Embry, 1997). This sequence boundary, which corresponds to the European T1–3 or to the In 2 sequence boundary (Gianolla and Jacquin, 1998, 1999), can be recognised in the studied sections.

Fig. 9. A) Sequences stratigraphy of the studied sections in comparison with sequence stratigraphy of a) Sverdrup Basin (Embry, 1997); b) Barents Sea (Gianolla and Jacquin, 1998; Skjold et al., 1998); c) Western Tethys (Gianolla and Jacquin, 1998; Gianolla et al., 1998); D) Biotic and climatic events d) ammonoid and conodont diversification (Brayard et al., 2006; Orchard, 2007); e) benthic diversification (Hautmann et al., 2008), f) climatic change in Pakistan (Hermann et al., 2008); see Fig. 3 for lithological legend.
see also Fig. 9) is less distinct in the sedimentological record of the Salt Range and Surghar Range. However, the yellow weathering dolomitic limestone bed below the early Smithian *Kashmirtes* fauna, near the top of the LCL at Chitta–Landu, and in the same biot stratigraphic position in the CM at Nammal, coincide with a significant change from AOM dominated to more terrestrial POM dominated assemblages. This coincidence suggests that the global third order sequence boundary near the Dienerian–Smithian boundary (T1–3, In2) is expressed in the two sections (SRT2, Fig. 9). At Chhidru the sequence boundary could not be located with certainty. It may correspond to the local LCL–CM transition.

Another global third order sequence boundary is reported from the latest Smithian along with a moderate shift in the depositional regime of the above mentioned areas (Embry, 1997); it corresponds to the European T2–1 and OI 1 (Ganolla and Jacquin, 1998) respectively. At Nammal and Chitta–Landu, a sequence boundary could be discriminated above the lowermost bed of the BB and at the top of the BB, respectively (SRT5, Figs. 3, 9).

The second order sequence boundary near the Spathian–Anisian corresponding to the European T2–3 and OI 4 (Ganolla and Jacquin, 1998) is marked globally by a distinct change in depositional regime, and in some cases the development of this sequence boundary seems to be tectonically influenced (Sverdrup Basin and Italian Alps; Embry, 1997). At Nammal and Chitta–Landu a marked lithological change occurring between the TL and the Landa Member probably corresponds to this sequence boundary (SRT8, Fig. 9).

The identification of the SB SRT3, SRT4, SRT5/1, SRT6 and SRT7 recognised in sequence stratigraphic framework of the Tethyan and boreal basins is less clear. The boreal *romunduri* zone corresponds approximately to the ammonoid faunas found in the topmost part of the CS and the lower UCL. Therefore, the reworking at the base of the UCL (SRT4) could correspond to the T1–4 sequence boundary described from the boreal realm (Skjold et al., 1998). T1–5 coinciding with the *tardus* zone in the boreal realm could not be recognised in the Pakistani sections. Two of the erosional surfaces described in the Smithian of Nammal might correlate with the OI 2 and OI 3 sequence boundaries in the western Tethys. However, this correlation is rather tentative because of the loose ammonoid age control in the NI and TL in the studied sections.

A first sequence stratigraphic interpretation of the Nammal gorce has been established by Haq et al. (1988) based on the lithological description biot stratigraphy of the PJRG (1985). Haq et al. (1988) described three sequences within the sedimentological record of the Nammal gorce. The first includes the Chhidru Formation, the Kathwai Member, the LCL and the lower part of the CM. The second one ranges from the upper part of the CM up to the middle NI. The third sequence includes the upper part of the NI and the TL. According to Haq et al. (1988), the lowermost SB in the sedimentary succession of the Nammal section is located in the middle CM and is indicated by “increased terrigenous input of coarser grained sandstones”, which, however, disregards the discontinuous record and the depositional gaps near the Permian–Triassic boundary (SRT1), which after Haq et al. (1988) corresponds to a transgressive surface and not to a SB. Limestone beds are absent in the upper part of the CM and are substituted by rare fine-grained sandstone beds and lenses; the mentioned coarse-grained sandstones or even a dominance of sandstones could not be confirmed. Therefore, a more reasonable interpretation of the upper CM would be HST rather than LST deposits. The second sequence of Haq et al. (1988) ranges up into the middle NI ignoring the base of the CS (SRT3), the base of the UCL (SRT4), the sedimentary gap in the BB (SRT5), and the emersion features in the BB (SRT5/1); according to Haq et al. (1988) the BB represent early HST deposits. The base of the UCL (SRT4) is again interpreted as transgressive surface and not as a SB. The SB in the middle part of the NI of Haq et al. (1988) probably corresponds to SRT6. The third SB after Haq et al. (1988), at the top of the Mianwali Formation, correspond to the second order SB near the Spathian–Anisian boundary of Embry (1997) and our SRT8. The statement of Haq et al. (1988) that the Tredian Formation consists of continental deposits could not be confirmed. Samples of the Landa Member (lower Member of the Tredian Formation) contain marine palynomorphs (*Micrhystridium* spp.) confirming a marine depositional environment.

Therefore, the interpretation of the sedimentological record of the Nammal gorce by Haq et al. (1988) contrasts with the sequence framework of Embry (1997) and the results and sequence interpretation of this study. Our results suggest that the transgressive surfaces of Haq et al. (1988) should be reinterpreted as SB.

Baud et al. (1996) subdivided the Early Triassic succession of the Salt Range and Surghar Range into three third order sequences (SB, S7, and SB) placing the SB of S6 at the transition from the Chhidru Formation to the Kathwai Member similar to our interpretation (SRT1). The SB of S7 is located at the base of the CS corresponding to our SRT3. The position of the SB to S6 is not clear as it occurs on top of BB or within the BB (Figs. 4 and 5 of Baud et al., 1996); therefore, it might correspond to our SRT5 or to SRT7/1.

### 6.2. Palaeoenvironment

Palaeoenvironmental studies on the Permian–Triassic sections of the Salt Range and Surghar Range (Nammal, Chhidru, Narmia and Zaluch) have previously been carried out by Wignall and Hallam (1993). These authors argued in favour of an anoxic event in this region, which they regarded as the principal kill mechanism of the end-Permian extinction event (Wignall and Twitchett, 2002). Three of their four sections are included in the present study. The stratigraphic framework and lithological units of Wignall and Hallam (1993) are based on the study of the PJRG (1985) but have been inaccurately implemented. The PJRG (1985) applied the stratigraphic framework and lithological units of Waagen (1895) and Guex (1978) but named the lithological units differently. Unit 1, in the sense of the PJRG (1985) corresponds to the LCL in the sense of Waagen (1895) and Guex (1978). However, the PJRG (1985) attributed Unit 1 to the “Griesbianchian *Gyronites frequens* zone”, which contradicts the standard ammonoid biostratigraphy. The *Gyronites frequens* zone represents the basal Dienerian (e.g. Guex, 1978); therefore Unit 1 is of Dienerian age. Fig. 3 of Wignall and Hallam (1993) implies that the authors apply Unit 1 in the sense of the PJRG (1985); however, they included parts of the CM (unit 2 of the PJRG, 1985) into their Unit 1 (Figs. 5 and 7 of Wignall and Hallam, 1993) and attributed Unit 1 to the Griesbianchian *Ophiceras* Zone in Fig. 3, erroneously citing the PJRG (1985). Hence, Wignall and Hallam (1993) extended their “Griesbian- chian” up into the CM, thus in fact up to the top of the ammonoid–defined Dienerian.

According to previous documentation and our findings the Kathwai Member in the Salt Range contains *Ophiceras connectens* and Schindewolf and therefore can be attributed to the mid to early Griesbianchian (Kummel, 1970; Schindewolf, 1954). So far *Otoceras* sp., which would indicate an early to middle Griesbianchian age has not been reported from the area. Therefore, either the early Griesbianchian is absent or local environment was inappropriate (too shallow) for ammonoids.

The LCL (or Unit 1 in the sense of the PJRG, 1985) can be attributed to the Dienerian at Narmia, Chitta–Landu, and Chhidru. It is dated as early Dienerian at Nammal based on the presence of *Gyronites frequens* (e.g. Guex, 1978 and ongoing studies of D. Ware and H. Bucher). At Narmia, Chitta–Landu, and Chhidru the basal part of the overlying CM is of Smithian age and of late Dienerian at Nammal (Brühwiler et al., in press). Therefore, the onset of anaerobic conditions proposed by Wignall and Hallam (1993) would not correspond to a Griesbianchian event but fall within the Dienerian or late Dienerian at Narmal. Our palynofacies data show that the late
Due to selective preservation of refractory terrestrial material (e.g. POM composition. However, the AOM abundance decreases abruptly in the record in Figs 4 and 5. When using organic C-isotope records as a correlation tool. In this study the C-isotope data is shown together with the palynofacies study when the composition of the POM and the negative shift in the marine OM around the Lowermost CM) at Chhidru and the Dienerian (LCL) at Chitta-Landu and Narmia are marked by POM assemblages with increased AOM contents, reflecting poor oxygenation. The relatively high TOC values (Fig. 8) and the rare occurrences of laminated beds suggest oxygen deficient conditions during the Dienerian and the earliest Smithian time interval. Biomarker distribution confirms good preservation conditions of the OM in the late Dienerian sample (NA 202) implying low oxygen levels. Palynofacies samples from the Chhidru Formation (at Chitta-Landu and Narmia) containing palynological assemblages with Griesbachian affinity document well oxygenated conditions. Thus the hypothesis of Wignall and Hallam (1993) that the end-Permian extinction event was delayed until the late Griesbachian in the Salt Range and Surghar Range and caused by the combination of a flooding event (Kathwai Member) and a severe and widespread anoxia in the late Griesbachian is not supported by our data.

6.3. The carbon isotope record

Our data from the Early Triassic of Pakistan (sedimentology, C-isotopes, POM, TOC and biomarkers) allow linking recovery events to coeval environmental changes. For the correlation with other sections we use our new organic C-isotope record. Bulk organic C-isotope data are known to be strongly dependent on the composition of the organic matter and physical factors during plant growth, such as temperature, water availability, insolation etc. (Jahren et al., 2008). Therefore, it is essential to know the composition of the OM as well as the isotopic end-members of marine and terrestrial OM respectively. Our studies of OM across the PTB from Norway show that the Early Triassic primary C-isotope signal of terrestrial OM varies around –22‰ and marine OM around –32‰ (Hermann et al., 2010). This contrasts with the C-isotope signal of recent marine organic matter (–21‰) and terrestrial C3 plants (–28‰) (Derry and France-Lanord, 1996). Therefore, the composition of the organic matter has to be considered when using organic C-isotope records as a correlation tool. In this study the C-isotope data is shown together with the palynofacies record in Figs 4 and 5.

In the studied sections several shifts in the δ13Corg can be observed. In the Nammal section the positive shift along the Dienerian–Smithian boundary is not related to a corresponding change in the POM composition. However, the AOM abundance decreases abruptly near the Dienerian–Smithian boundary, contrasting with the rather protracted δ13Corg change (Fig. 4). There is also no correlation between the composition of the POM and the negative shift in the Smithian and the positive shift at the Smithian–Spathian boundary. At Nammal the positive shift at the Smithian–Spathian boundary is slightly obscured by four single data points around –26‰, which are probably due to poor preservation. Detailed re-sampling of this interval in a nearby section resulted in a coherent curve. The positive shift at the Spathian–Anisian boundary cannot be caused by the demise of marine OM, since marine OM is already reduced in the Spathian and the positive shift is also seen in other Tethyan C-isotope records (Fig. 6). The two most negative data points in the carbonate carbon isotope record in the CM of the Nammal section (Fig. 4) have not been recorded in the Chhidru and Chitta-Landu sections. These samples are relatively heterogeneous therefore biogenic carbonate might have obscured the isotope signal.

According to several authors effects of diageneis on the variability of the bulk organic carbon isotope values are thought to be minor (~1‰, e.g. Freudenthal et al., 2001, Meyers et al., 1995). However, changes of the early diageneric conditions such as the variability of productivity, and changing redox conditions of bottom waters can have significant impact on the preservation of the isotope signal of organic matter (Freudenthal et al., 2001). As a consequence several studies documented significant changes in δ13Corg values (up to 4‰) due to selective preservation of refractory terrestrial material (e.g. Dean et al., 1986) or of non-metabolisable organic matter during anoxic anaerobic degradation (e.g. Benner et al., 1987). However, since the isotopic end-members of marine and terrestrial organic matter differ significantly, one of the most important factors that could mask δ13Corg records are changes in the composition of the POM as discussed above. Thus, the parallel study of bulk organic carbon isotopes and POM allows a good control on possible compositional effects on the carbon isotope record.

In chemostratigraphic studies of Cretaceous successions a close relationship between sea-level changes and carbon isotope records has been demonstrated (e.g. Jarvis et al., 2006, Weissett et al., 1998). Sea-level rises, as documented in the chalk series of southern and eastern England, correspond to positive shifts in the carbonate carbon records of these series. Regressions match with negative isotope shifts. These correspondences are induced by changes in the partitioning of the carbon between the two major carbon sinks — the carbonate carbon sink and the organic carbon fixation by marine phytoplankton and terrestrial vegetation. During transgressive phases the area of flooded shallow seafloor is increased, sediments and soils are reworked and promote increased nutrient availability, thus enhancing productivity and burial of isotopically light organic matter (Jarvis et al. 2002). During sea level fall these areas are exposed to erosion and oxidation, which induces the release of 13C depleted carbon into the carbon cycle.

In this study we identified several concurrences between sea-level changes and the isotope record. The positive shift at the Dienerian–Smithian boundary correlates with a sea-level rise (transgression). A similar correlation is observed during the Spathian–Anisian transition, where a positive isotope shift across the TL–Landu Member boundary concurs with a sea-level rise. However, the hypothesis fails to explain the negative–positive carbon isotope excursion couplet at the Smithian–Spathian transition. The negative shift in the Smithian might correspond to a regressive phase, however, the following positive shift does not coincide with a sea-level rise. Jarvis et al. (2002) noted that the relationship between sea-level changes and carbon isotope curve is not always straightforward. In Cretaceous records intervals with a mismatch of the two records have been assigned to a release of mantle derived light CO2 during enhanced sea floor spreading (e.g. Gröcke et al., 1999). Changes in the marine primary productivity or changes in the preservation of organic matter might also influence the isotopic composition independent of the sea-level curve (Kump and Arthur, 1999). Faster processes, such as rapid releases of CO2, e.g. from metamorphosed organic rich sediments during the Siberian Trap emplacement (Payne and Kump, 2007; Svensen et al., 2009), could explain why the couplet of rapid negative and positive isotope excursion across the Smithian–Spathian transition might be unrelated to the development of the sequences.

6.4. Comparison with other sections

The aforementioned considerations suggest that the organic C-isotope record of the studied sections reflects real global changes independent of the OM composition. Therefore, we can use the organic C-isotope chemostratigraphy to compare our results with other sections such as the Jinya/Waili area in South China, the Losar section in North India (Galletti et al., 2007b), and the Tulong area in South Tibet (Brühlwiler et al., 2009).

Recent studies demonstrate that the early/middle Smithian and Spathian time intervals in the Loulou Formation were times of deposition of well oxygenated carbonates (Nanpanjiang Basin, Guangxi Province, South China (Galletti et al., 2008)). The same time intervals are marked by increased rates of diversity and abundance of skeletal material (Galletti et al., 2008) and coincide with the highest values in taxonomic richness of ammonoids (Brayard et al., 2006; Brühlwiler et al., 2010) and conodonts (Orchard, 2007). In contrast, the Dienerian and late Smithian time intervals are marked by predominantly mixed
silliclastic–carbonate depositional environments with dark, suboxic, laminated mudstones and organic-rich shales and limited bioturbation, suggesting an overarching causal mechanism leading simultaneously to low sea water oxygenation levels and slow downs or even set backs of the course of the biotic recovery (Galfetti et al., 2008).

Similar to the South China and Indian records our data demonstrate that dysoxic conditions, suggested by AOM abundance and high TOC, prevailed during the Dienerian interval and coincided with low diversity of benthic organisms (Hautmann et al., 2008). From the Smithian onward oxygenated conditions prevailed in the Mianwali Formation of the Salt Range and Surghar Range. Global diversity increase of benthic organisms in the Smithian (Hautmann et al., 2008) coincided with the pronounced recovery phase of ammonoids (Brayard et al., 2006; Brühwiler et al., 2010) and conodonts (Orchard, 2007). Additionally, the abundance of hopanes in the Smithian (NA 55) and Spathian samples (NA 88) indicates proliferation of bacteria from the Smithian onward. However, the decline of isoprenoids from the Dienerian (NA 202) to the Smithian sample (NA 55) indicates that the precursor organisms, probably of marine origin, disappeared or became reduced after the Dienerian. In contrast to South China and the Spiti area, theoxic conditions in the Mianwali Formation were not interrupted during the late Smithian. The continuous shallowing upward trend within the Mianwali Formation probably prevented the formation of a dysoxic water–sediment interface during Early Triassic. Well oxygenated conditions during late Smithian are also reported from the Tulong area in South Tibet (Brühwiler et al., 2009). However, in the Spiti area dysoxic conditions are documented during the late Smithian (Galfetti et al., 2007b). During Early Triassic, the Salt Range and Surghar Range, the Tulong and Spiti area were all situated on the northern margin of the Indian subcontinent.

The identical and synchronous changes in the Early Triassic records from Spiti on the northern Gondwanan margin and from Guangxi–Guizhou Province in the equatorial, palaeogeographically isolated South China block strongly suggest that the absence of late Smithian dysoxic conditions in the Salt Range and Surghar Range may result from the local Early Triassic tectonic configuration of this area (Galfetti et al., 2007b). The unusually thick and clastic-dominated nature of the Spathian record of the Salt Range also concurs with this interpretation.

The above mentioned recovery of ammonoids and conodonts in the Smithian was interrupted by an extinction event at the Smithian–Spathian boundary (Brayard et al., 2006; Orchard, 2007). Our sequence stratigraphic interpretation demonstrates that this extinction event is closely associated to a sea-level low stand near the Smithian–Spathian boundary (SRT5), which is recognised globally (Embry, 1997; Gianolla and Jacquin, 1998; Gianolla et al., 1998; Skjold et al., 1998). It also coincides with a global positive shift of C-isotopes (Galfetti et al., 2007a,b), and climatic changes inferred from ammonoid distribution (Brayard et al., 2005, 2006) and palynology (Hermann et al., 2008; Fig. 9). Benthic faunas are still not well enough documented to distinguish between extinction or absence of extinction during the late Smithian global disruption of the recovery (Hautmann et al., 2008).

Early Triassic terrestrial ecosystems are thought to be heavily affected by the end Permian extinction event (Benton and Twitchett, 2003; Retallack et al., 1996). It has been proposed that the recovery of equatorial conifer forests was delayed until the beginning of the Middle Triassic, whereas Early Triassic terrestrial ecosystems were continuously dominated by lycopsids (Looy et al., 1999). However, recent studies demonstrate that terrestrial ecosystems recovered rather quickly already in the Griesbachian of Norwegian sections (Hochuli et al., 2010b). In the Pakistani palynofacies data, the continuous record of sporomorphs and plant debris reflects prolific ecosystems in the hinterland of the Salt Range and Surghar Range during Early Triassic times. Within the sporomorph group the continuous abundance of striate and non-striate bisaccate pollen suggests that pteridosperms and conifers proliferated on the Indian subcontinent already during the Early Triassic.

7. Conclusions

The interpretation of the succession of marine palaeoenvironments of the Early Triassic sections in the Salt Range and Surghar Range are based on sedimentological observation combined with palynofacies data, C-isotopes records, TOC, and biomarkers.

(1) Sea level changes are reflected in the sedimentological and palynofacies records. Two second order SB (SRT1 near the Permian–Triassic boundary, SRT8 near the Spathian–Anisian boundary) and two third order SB (SRT2 near the Dienerian–Smithian boundary, SRT5 near the Smithian–Spathian boundary) have been identified. Six SB of undetermined order (in the Smithian SRT3 at the base of S and SRT4 the base of UCL in the Spathian SRT5/1 at the top of the BB, SRT6 in the middle NI and SRT7 at the base of TL, and in the Anisian SRT9 at the top of the Landa Member) could be differentiated.

(2) The sequence boundaries defined in this study differ from the interpretation of the Nammal section of Haq et al. (1988), but are consistent with the global sequence boundaries established by Embry (1997) and Gianolla and Jacquin (1998).

(3) The composition of the POM reflects changing ecological conditions such as a protracted shallowing upward combined with short-termed sea-level changes, oxygenation conditions, and proximity to the shore. These results are supported by the molecular biomarker results. Dysoxic conditions are recorded in the Dienerian of Chitta–Landu and Narmia, in the late Dienerian at Nammal, and in the earliest Smithian at Chhidru. The results show that the patterns of the Early Triassic recovery were closely linked to the prevailing environmental conditions, such as carbon cycle, sea-level and oxygenation levels.

(4) The POM suggests well oxygenated conditions in the Late Griesbachian. The proposed late Griesbachian shallow water anoxia as kill mechanism for the Permian fauna as postulated for the Salt Range and Surghar Range (Wignall and Hallam, 1993) is refuted.

(5) The observed low oxygenation levels in the Dienerian are also documented from coeval marine sediments of South China and North India (Galfetti et al., 2007b, 2008) and are probably responsible for low diversity of benthic organisms during this time interval.

(6) From the Smithian onward well oxygenated conditions prevailed in the Early Triassic of Pakistan coinciding with the onset of the explosive radiation of ammonoids (Brayard et al., 2006, 2009; Brühwiler et al., 2010, in press) and conodonts (Orchard, 2007) as well as a diversity increase of benthic organisms, and the proliferation of bacteria as inferred from the biomarker data.

(7) The positive shift of C-isotopes and the SB SRT5 near the Smithian–Spathian boundary coincides with a major extinction of ammonoids and conodonts and a change from humid to dryer climate (Brayard et al., 2005, 2006; Galfetti et al., 2007a; Hermann et al., 2008). Accumulation of OM-rich shales during the late Smithian, as documented from other Tethyan sections, is not recorded in the Salt Range and Surghar Range. The OM rich siltstone horizon deposited in the Surghar Range during this time interval probably originates from an accumulation of older OM that has been reworked during this regression episode.

(8) Early Triassic terrestrial ecosystems in the hinterland of the Salt Range and Surghar Range proliferated and constantly shed OM onto the Northern Gondwanan shelf. Continuous terrestrial primary production is documented for the entire Early Triassic.


