Provenance study on Eocene–Miocene sandstones of the Rakhine Coastal Belt, Indo-Burman Ranges of Myanmar: geodynamic implications

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Abstract: The Indo-Burman Ranges (IBR) represent an accretionary wedge, which is the result of subduction of the Indian plate beneath the Asian plate. In the Rakhine Coastal Belt it comprises a thick stack of Cretaceous to Neogene turbiditic sediments and localized thrust sheets of oceanic plate mafics and pelagic sediments. We investigate Eocene–Miocene sandstones, aiming to reveal the provenance of the detrital material using modal framework grain, heavy mineral and detrital zircon analysis (U–Pb laser ablation ICP-MS dating, Hf isotope geochemistry and typology). The results show a predominant derivation of the clastic material from: (i) Late Cretaceous to Oligocene igneous rocks, which are often bimodal with a low number of zircons spanning the Cretaceous–Palaeogene boundary, and (ii) recycled orogenic terrane sources comprising ophiolitic rocks. Age corrected Hf isotope ratios confirm subduction-related mixed mantle-crust sources. We also observe minor reworking of older magmatic zircons. By comparing our obtained petrographic parameters and zircon characteristics with potential Himalayan, Indian continent and Burman margin sources we conclude a Burman margin and arc origin provenance. With regard to hydrocarbon exploration in the IBR, a forearc and trench basin system model linked with the Burman arc appears more appropriate for evaluating the petroleum system.

Supplementary material: sandstone modal composition, heavy mineral contents, detrital zircon U–Pb LA-ICPMS dating and hafnium results are available at http://www.geolsoc.org.uk/SUP18651

Myanmar (Burma) is located on the eastern edge of the zone of Himalayan convergence, south of the Eastern Himalayan Syntaxis where the west–east-striking Himalayan structures rapidly change to a north–south orientation (e.g. Brunnschweiler 1974; Bender 1983; Barley et al. 2003; Searle et al. 2007; Allen et al. 2008) (Fig. 1a). Myanmar is the region of transition between the main Himalayan collision belt and the Andaman arc, where, since the Cretaceous, the Indian plate has been subducting under Asia (e.g. Curray et al. 1979; Mitchell 1993; Pivnik et al. 1998; Curray 2005). This subduction is also driving the spreading in the Andaman Sea to the south (Stephenson & Marshall 1984). Since the Late Cretaceous the Burma micro-plate is assumed to have moved northward, relative to SE Asia, about 1000 km (Mitchell 1993). Chakraborty & Khan (2009) call it a sliver plate forming through right-lateral displacement between the subduction zone and the Sagaing fault, and its southward continuation in the Andaman Sea and offshore Sumatra. The Burma micro-plate constitutes part of the larger...
Indo-Burma–Andaman micro-continent, which rifted and drifted apart from Australia in the Late Jurassic and Early Cretaceous (Acharyya & Lahiri 1991; Acharyya 1998). It was brought into collision with the Asian continent in the Late Oligocene, and the northern tip of the micro-plate collided with India in the Mio-Pliocene (Acharyya 1998).

The Indo-Burman Ranges (IBR) occupy western Myanmar, and geographically are subdivided into the Naga Hills, the Chin Hills and the Arakan Yoma Range (Fig. 1a). They formed along the western margin of the Burma micro-plate buttress (Fig. 1). The IBR represent an active, west-verging accretionary complex encroaching on the Bengal basin from the east due to subduction of the Indian plate beneath the Burman micro-plate (e.g. Bender 1983; Sengupta et al. 1990; Mitchell 1993). Lithologically, the IBR consist of a complex tectonic collage of obducted metamorphic schists, marbles, ophiolites, Mesozoic and Cenozoic oceanic sediment, basin pelagic as well as turbiditic series (e.g. Brunnschweiler 1974; Bender 1983; Mitchell 1985; Pivnik et al. 1998).

The origin of the Cenozoic syn-accretionary turbiditic material of the IBR is a matter of debate. A current interpretation is that the west-verging range consists of off-scraped material of a proto-Bengal fan, which has been continuously incorporated into the IBR (Curry et al. 1979; Bender 1983; Hutchinson 1989; Curray 2005) since the Eocene. Hence, the clastic material could have been supplied from early collision events of the Indian plate with the Eurasian plate, dating back nearly 60 Ma sourced from the Himalaya (e.g. Curry et al. 1979; Alam et al. 2003). In Burma, evidence for that Himalayan-sourced sediment is poor and based only on preliminary field observations given by Brunnschweiler (1974) and Bender (1983). Allen et al. (2008) also found a source for the Palaeogene turbiditic sandstones in the Myanmar orogenic belt and arcs.

After recent large gas discoveries (e.g. Shwe discovery), western Myanmar and the Rakhine coastal basin have attracted renewed interest in hydrocarbon exploration. It is now one of the last unexplored deep-water basin systems in the world. In earlier days, when exploration focused close to the onshore, the lack of reservoirs was a main cause of failure, and the long distance to the Bengal Fan sandstones was used to explain this unfavourable situation. The results of the present study open new opportunities for exploration with reservoirs which did not form from Himalayan clastic material, but a close link of the petroleum system with the Burman arc and forearc, including the Myanmar Central Basin, appears promising.

The present work aims to increase our knowledge of the Cenozoic geological history of Myanmar by investigating the provenance of Eocene–Miocene sandstones tectonically included in the accretionary wedge in the Rakhine coastal area. The recognition of the detrital source rocks of the sandstones is of great importance for tracing the geodynamic evolution of the eastern Himalayan area in general. We investigate the modal framework grain composition of the sandstones and their heavy mineral associations to characterize the lithologies present in the source areas. Detrital zircon ICP-MS laser ablation U–Pb dating is used to evaluate the crystallization age ranges in the source regions and/or recycling of older crustal elements. Hf isotope ratios of the dated detrital zircons are applied to infer the origin of the syn-sedimentary magmas. Additional zircon typology arguments are used to constrain magma types present in the source regions. Finally, we are able to discuss the sourcing of the detrital material from Burman arc and continental plate rocks. Himalayan-sourced rocks, which one would expect to find in pre-Bengal fan deposits, were not identified.

Geological framework of Myanmar

Tectonically as well as geomorphologically, Myanmar can be subdivided into four major, north–south-orientated provinces (Fig. 1). These are, from east to west: (1) the Eastern Highland or Shan Plateau and Scarps, (2) the Central Lowlands or Myanmar Central Basin, (3) the Indo-Burman Ranges, and (4) the Rakhine Coastal Belt.

The Eastern Highland is a part of the Shan–Thai Block, a large continental domain of the Asian plate that connects to the Pacific plate in the east. It represents the southern continuation of the Quingtang and Lhasa blocks of Central Tibet (Searle et al. 2007). In central and southern Myanmar the Shan Scarps and Mogok Metamorphic Belt (MMB) occupy the western margin of the highlands. It is a high-grade metamorphic belt of Precambrian to Palaeozoic rocks (marbles, schists and gneisses), which show evidence of Cretaceous to Miocene magmatism and metamorphism (e.g. Mitchell 1993; Barley et al. 2003; Searle et al. 2007). Age and petrological features suggest that the MMB represents the remainder of an Andean-type subduction
margin, which developed prior and during collision of India with the Eurasian continent (Barley et al. 2003). In the Shan Scars and Plateau the unconformable contact between Proterozoic sandstone and shale formations, with overlying Cambrian to Early Cretaceous series, is observed (Mitchell 1993).

The Myanmar Central Basin (MCB) is located between the major dextral Sagaing Fault in the east and the Kabaw Fault and IBR in the west (Fig. 1). It hosts thick series of mixed marine and fluvial sediments of Late Cretaceous through Plio-Pleistocene age. They unconformably overlie underplated oceanic rocks in the west (belonging to the IBR) and metamorphic/crystalline continental basement and cover of the Burma plate (Fig. 1b). This lowland hosts a series of en echelon depocentres with Cenozoic basins presently affected by active tectonic inversion (Pivnik et al. 1998). According to Bertrand & Rangin (2003), basin inversion started in the Middle to Late Miocene. The central volcanic line is interpreted as the modern volcanic arc separating the thick forearc basin in the west from the relatively thin back-arc basin in the east. The back-arc basin is underlain by Mid- to Late Cenozoic non-marine sediments, and the forearc basin is filled by thick latest Cretaceous to Quaternary marine and non-marine sediments (Mitchell 1993; Pivnik et al. 1998).

The IBR is a prominent geotectonic element extending from the Naga Hills in the north to the Arakan Yoma Range in the south. From the Cape of Negrais southwards the IBR is submerged in the Bay of Bengal. The IBR of western Myanmar is considered an active accretionary wedge linked to the eastward subduction of the Bengal basin oceanic lithosphere beneath the Burmese platelet (e.g. Sengupta et al. 1990; Mitchell 1993; Searle et al. 2007). It terminates northward at the Eastern Himalayan Suture and extends southward to Sumatra. The IBR is composed mainly of Late Cretaceous–Neogene turbiditic and hemipelagic series. Along its eastern margin, the IBR includes a prominent range composed of Upper Triassic turbiditic (‘flysch-like’) sandstones and marbles in tectonic contact with a belt of mica schists (Mt Victoria). The tectonically associated ophiolitic rocks are thought to be the southern continuation of the Indus Tsangpo suture zone (Mitchell 1993; Searle et al. 2007). In the IBR major tectonic and subduction events have been identified from late Early Cretaceous to Middle Miocene and Quaternary periods (Mitchell 1993; Allen et al. 2008). The western edge of the IBR is represented by the Kaladan thrust fault along the Naga Hills in the north and the Arakan Yoma Hills in the south (Fig. 1a). All or most of the IBR sediments are widely interpreted either as accreted sediments of the early Bengal Fan, derived from the India–Asia collision belt (Mitchell 1974; Curray et al. 1979; Bender 1983; Hutchinson 1989; Curray 2005), or as sediments shed from the Indian margin (Sengupta et al. 1990), or from the Burman active volcanic arc (Allen et al. 2008).

The Rakhine Coastal Belt (RCB), including several offshore islands, is geologically part of the IBR because of a similar origin, but with an increased presence of Miocene sediments west of the Kaladan Fault (Fig. 1). The RCB developed between the subduction zone in the west and the main Arakan Yoma ranges in the east (IBR). In the southern part of the area, structural deformation is greater than further north. The Miocene sequences in this area are frequently tilted and faulted, locally folded and overthrust. The pre-Miocene turbiditic formations occurring below the transgressive Miocene beds show the intense folded structural style of the IBR (Bender 1983). Further north, the Miocene series exhibit relatively minor tectonic deformation. In Bangladesh, the RCB has its continuation in the Chittagong–Tripura Fold Belt.

Samples and methods

We undertook provenance analysis on Eocene to Miocene sand and sandstone samples from the onshore section of the Rakhine Coastal area. GPS and geographical locations, chronostratigraphic age correlations, lithologies of the samples and methods applied are given in Table 1.

Modal framework grain analysis of the sandstones was performed on thin sections stained for feldspars and carbonates (Dickson 1966; Norman 1974). Monomineralic grains (quartz and feldspars) and lithic rock fragments were distinguished according to the standard method of Dickinson & Suczak (1979) and Dickinson (1985). At least 300 grains were point-counted in the feldspar-stained thin sections. The percentages of various combinations of grains are plotted in standard triangular diagrams used to classify the sandstones (Folk 1974) and to interpret the geodynamic setting of the source terranes (Dickinson 1985).

To obtain the transparent heavy mineral fractions, the sandstones were crushed by SelFrag laboratory batch equipment using high voltage pulse power technology to liberate morphologically intact minerals. In the <2 mm sieved fraction the carbonate was dissolved in warm (60–70 °C) 10% acetic acid. The heavy minerals were extracted in separation funnels (Mange & Maurer 1992) from the 0.063–0.4 mm sieve fraction using bromoform (density 2.88). The bulk heavy mineral fractions were mounted in piperine (Martens 1932) between glass slab and cover. Mineral identification and
### Table 1. Geographical location and characteristics of the analysed samples in the Rakhine Coastal Belt

| Sample no. | Latitude (deg: min: sec) | Longitude (deg: min: sec) | Location | Formation | Stratigraphical age | Lithology | Analysis |
|------------|--------------------------|---------------------------|----------|-----------|---------------------|-----------|----------|
| 10TTN01    | 18°27′11″                | 94°17′44″                 | Near Thandwe airport Sabarkyi village | Ngapali | M. Eocene | Turbiditic shale, dark-grey, moderately hard | HM |
| 10TTN02    | 18°27′58″                | 94°17′17″                 | Sabarkyi village | Ngapali | M. Eocene | Turbiditic sandstone, brownish-grey, associated with carbonaceous shales | HM, MFA |
| 10TTN03    | 18°29′38″                | 94°16′18″                 | Gaw village | Ngapali | M. Eocene | Turbiditic sandstone, dark-grey, medium-grained | DZA, HM |
| 10TTN04    | 18°22′11″                | 94°23′01″                 | Andrew Bay (Abe hill) | Ngapali | M. Eocene | Turbiditic sandstone, dark-grey, medium-grained | DZA, HM, ZT |
| 10TTN05    | 18°22′07″                | 94°23′02″                 | Andrew Bay | Ngapali | M. Eocene | Turbiditic sandstone, dark-grey, medium-grained | HM, MFA |
| 10TTN06    | 18°18′31″                | 94°19′38″                 | Aungkyawmaw Chaing | Ngapali | M. Eocene | Very coarse turbiditic sandstone, brownish-grey to grey | DZA, HM, ZT |
| 10TTN09    | 17°59′37″                | 94°29′44″                 | Thandwe-Gwa Rd. (MP 51/4) | Yenandaung | M. Miocene | Turbiditic sandstone, light-grey | DZA, ZT, HM, MFA |
| 10TTN10    | 18°08′20″                | 94°26′37″                 | Htudu Island | Yechangyi | Oligocene | Turbiditic sandstone, reddish-brown, medium-grained | DZA, ZT, HM, MFA |
| 10TTN12    | 18°00′13″                | 94°26′16″                 | Kyeintali North | Yechangyi | Oligocene | Turbiditic sandstone, light-grey, medium-grained | HM, MFA |
| 10TTN13    | 18°00′11″                | 94°26′14″                 | Kyeintali North | Yechangyi | Oligocene | Turbiditic sandstone, yellowish-brown, medium-grained | DZA, ZT, HM |
| 10TTN14    | 18°00′11″                | 94°26′14″                 | Kyeintali North | Yechangyi | Oligocene | Turbiditic sandstone, yellowish-brown, medium-grained | HM, MFA |
| 10TTN16    | 19°52′55″                | 93°12′05″                 | East Bayonga Island | Yechangyi | Oligocene | Turbiditic sandstone, medium-grained | DZA, ZT, HM, MFA |
| 10TTN18    | 19°20′27″                | 93°29′34″                 | Kalaba to Gortu Leikkammaw | Leikkammaw | L. Miocene | Sandstone, soft | DZA, ZT, HM, MFA |
| 10TTN20    | 18°54′01″                | 93°57′16″                 | Kyauknimaw | Yechangyi | Oligocene | Turbiditic sandstone, fine- to medium-grained | DZA, ZT, HM, MFA |

Analysis key: DZA, detrital zircon age; ZT, zircon typology; HM, heavy minerals; MFA, modal framework grain analysis.
| Age     | Formation | Samples | Lithological/sedimentary features                                                                 |
|---------|-----------|---------|----------------------------------------------------------------------------------------------------|
| PLIO    | LEIKKAMAW | TTN18   | Mud flat, tidal channel and raised beach deposits                                                  |
|         |           | TTN09   | Sandstones intercalated with shales, gritty sandstones and conglomerates; shallow marine            |
| MIOCENE | YENANDAUNG| TTN16   | Turbiditic sandstones interbedded with sandy clays and gritty sandstone                            |
| Early   |           | TTN20   |                                                                                                   |
| Oligo-  | YECHANGYI | TTN12/14| Shales intercalated with highly carbonaceous siltstones                                            |
|ocene    |           | TTN10/13|                                                                                                   |
| Late    | SINBOK    | TTN06   | Blocks of micritic and nummulitic limestones (olistoliths)                                       |
|         |           | TTN05   | Laminated carbonaceous shales interbedded with turbiditic sandstones                              |
|         |           | TTN04   |                                                                                                   |
| Eocene  | NGAPALI   | TTN03   | Shales interbedded with thin highly carbonaceous sandstones                                       |
| Middle  |           | TTN02   | Turbiditic sandstones interbedded with carbonaceous shale                                         |
|         |           | TTN01   |                                                                                                   |
| Early   | GWA       |         | Shales interbedded with sandstone beds; Olistoliths of micritic limestones and basalts           |
| Cretaceous|           |         | Hard shale and thin turbiditic sandstones, black shales; reworked coal in places                |
|         |           |         | Basalt and micritic limestone blocks & pieces                                                     |
|         |           |         | Pillow lava basalt and chert blocks                                                              |
|         |           |         | Chaotic complex: blocks of micritic limestone with *Globotruncanca* fossils; reddish, black, brown radiolarian chert; brown mudstone with manganese nodules; serpentine and vesicular basalt slivers, boudinaded sandstones and shales |

Fig. 2. Idealized stratigraphic section (not to scale) of the Rakhine coastal area with comments on lithological and sedimentary features (see also Hutt et al. 2010). The sedimentary series are overthrusting an ophiolitic mélange, including oceanic plate pelagic sediments. In the turbiditic formations a general decrease in tectonic deformation is observed upsection, including several unconformities. The Late Miocene shallow-marine and coastal Leikkammaw Formation unconformably overlies the stack of turbiditic and hemipelagic sediments.
quantification was carried out under the petrographic microscope (e.g. Mange & Maurer 1992) using the mid-point ribbon and fleet counting methods, and a minimum of 200 grains was counted per sample.

Detrital zircons were extracted from c. 2 kg of sandstone using standard mineral separation techniques, including SelFrag high voltage fragmentation, dissolution of carbonate cement in cold hydrochloric acid and heavy liquid (methylene iodide, density 3.32) separation. Handpicked zircons were mounted in epoxy blocks and polished down to expose their core. Cathodoluminescence (CL) imaging was carried out on all analysed zircons in order to determine the internal structures of the grains prior to isotopic analysis. For typological analysis (Pupin 1980), only euhedral zircon crystals were investigated. Zircon classification was based on the zircon typology method of Pupin (1980). Each crystal was described in terms of relative development of the crystal surfaces. The distributions of morphological types are shown on the typology diagram for each sample.

In laser ablation ICP-MS U–Pb dating of detrital zircons, only magmatic domains were investigated, and we did not date inherited cores. Laser ablation ICP-MS analyses were performed on an Elan 6100 DRC instrument coupled to an in-house-built 193 nm Excimer laser at the Institute of Geochemistry and Petrology, ETH Zurich. Helium gas (1.1 l/min) was used as the carrier gas in the ablation cell. The laser was run at pulse rate of 10 Hz with energy of 0.5 mJ/pulse and a spot size of 40 μm. The accuracy and reproducibility within each run of analysis were monitored by periodic measurements of the GJ-1 external standard, with 207Pb/206Pb age of 608.5 ± 0.4 Ma (Jackson et al. 2004). Data reduction was performed using the GLITTER software (Van Achterberg et al. 2001) to calculate the relevant isotopic ratios, ages and errors. Concordia and frequency probability diagrams were performed using ISOPLOT v.3.0 (Ludwig 2003).

The in situ Hf isotopic analysis on dated zircons was carried out using the Nu plasma MC-ICP-MS (Nu instrument Ltd) attached to a 193 nm UV ArF excimer laser, at the Institute of Geochemistry and Petrology, ETH Zurich. Laser repetition rate of 5 Hz, spot size of 60 μm and He carrier gas (0.8–1.11 l/min) were applied. The energy density used in this study was 10–20 J cm⁻². Standard Monastery zircon was used for external correction. Each ablation was preceded by a 30 s-on mass background measurement, and ablated zircon was measured within 60 s. For the accurate measurement of Hf isotope ratios in zircon, the isobaric interference of 176Yb and 176Lu on 176Hf were corrected by measuring 171Yb (176Yb/171Yb = 0.897145) and 175Lu (176Lu/175Lu = 0.026549), respectively. The calculation of epsilonHf(T) (time-corrected) values was based on zircon U–Pb ages and the chondritic values (176Hf/177Hf = 0.282772, 176Lu/177Hf = 0.0332; Blichert-Toft & Albarède 1997).

Lithostratigraphy and sedimentology

The present lithostratigraphic subdivision and chronostratigraphic correlations apply the scheme developed in Htut et al. (2010). The Eocene–Miocene sediment sequence regionally is thrust over a chaotic complex (tectonic mélangé) comprising single blocks of variable size, which are embedded in a thoroughly deformed and slightly metamorphosed matrix of shales and boudinated sandstone beds (Fig. 2). The blocks comprise sediments of Cretaceous age (Brunnischweiler 1974; Bender 1983) and include white and red Globotruncanana-bearing micritic limestones, reddish-brown to black radiolarian cherts and shales, and brown manganosferous shales. Radiolarian cherts are also found to be intruded by vesicular basalts. Volcanic elements of the basal complex are blocks of dark-brown basalts with pillow and columnar structures.

The Gwa Formation (Fig. 2) is the lowest exposed formation in the surveyed area. It may reach a thickness of 3400 m and even more in the Rakhine Yoma area. The lower part of this formation is faulted in many places; boudinated sandstone beds and pencil structures are characteristic. The unit is composed mainly of indurated, carbonaceous bluish-grey shales, occasionally intercalated by thin grey to dark grey turbiditic sandstone beds. Flute, groove and load casts are common along bedding planes. Blocks of micritic limestones, radiolarian chert and (pillow) basalts are assembled at various basal levels of the formation.

The Ngapali Formation (Middle Eocene) has a compound thickness of c. 5500 m. In its lower part, it comprises fine- to medium-grained, hard, brown to grey bedded turbiditic sandstones, which are intercalated by laminated, grey to greenish-grey and highly carbonaceous shales. Occasionally, limestones, conglomerates and thin sandstones are intercalated into the generally massive sandstone-rich lower part. The middle and upper parts of the formation are dominated by grey to pinkish-grey, very hard and compact siltstone and thin highly carbonaceous turbiditic sandstone beds.

The Sinbok Formation (Late Eocene), with an estimated sediment thickness of about 2200 m, consists of three members. The Lower Shale Member is made up of highly fractured carbonaceous blue shales with some intercalated Nummulite-bearing turbiditic sandstone beds. The Middle Sandstone
Member comprises fine- to medium-grained yellowish-brown to dark brown thick sandstone beds (up to 13 m) with minor shale intercalations. The Upper Shale Member is composed of thinly to thickly laminated dark grey shales, which contain brown sandstone and limestone blocks with Nummulite and Globotruncana fossils, respectively. The shale members represent mainly pelagic sedimentation with intercalations of low-density turbidite deposits.

The Yechangyi Formation (Oligocene) measures about 2000 m and unconformably overlies the Sinbok. It consists of fine- to medium-grained, thick-bedded light grey turbiditic sandstones interbedded with sandy shale and shale beds. The shales are grey to greenish-grey, fairly soft, and micaceous. This formation clearly shows less tectonic deformation than the underlying ones.

The Yenandaung Formation (Early–Middle Miocene) amounts to about 850–2000 m in thickness. In turn it unconformably overlies the Yechangyi Formation in most locations, and is faulted in only a few places. It consists of a lower Shale and Siltstone Member and an upper Sandstone and Shale Member. The lower member comprises blue and light grey thick shale beds and highly carbonaceous siltstone beds of turbiditic facies. The upper member is made up of thick sandstone beds associated with sandy shales and gritty sandstone beds.

The Leikkamaw Formation (Late Miocene), with a sediment thickness of about 700 m, again unconformably overlies the Yenandaung Formation. It comprises an irregular succession of fine- to medium-grained, hard, well-bedded yellowish-brown sandstones intercalated with shale and gritty sandstone beds. Individual sand-rich sequences are up to 120 m thick. The formation was deposited in a near-shore environment. Pliocene and Pleistocene mud flat, tidal channel and beach sediments conformably overlie the Leikkamaw Formation.

Thrusts and folds in the sediment series show generally west-vergence with an obvious decreasing trend of tectonic deformation going from older to younger formations (Htut et al. 2010). Eocene and Oligocene sediments are highly folded and thrusted, whereas the Miocene series are gently folded or only tilted. Turbiditic and hemipelagic sedimentation is characteristic for the Eocene, Oligocene and Early–Middle Miocene formations (‘flysch-type’ sediments; Bender 1983; Allen et al. 2008). The Late Miocene and Pliocene show a shallowing trend from shelfal to coastal facies, generally classed as ‘Molasse-type’ facies by Bender (1983) and Htut et al. (2010). The presence of the ophiolite-bearing mélangé at the base of the folded and thrusted, west-verging turbiditic sediment pile suggests an accretionary wedge tectonic environment (e.g. Curray et al. 1979; Mitchell 1981; Alam et al. 2003; Allen et al. 2008), which comprises oceanic basement and pelagic cover, deformed trench and slope basin deposits, which finally were topped by shallow shelf and coastal deposits (Htut et al. 2010).

Results

The modal framework grain analysis classifies the sandstones mostly as feldspathic litharenite and litharenite within the QFL diagram of Folk (1974) (Fig. 3a). However, the amount of quartz grains is variable, showing that Oligocene and Miocene sandstones tend to have higher quartz content. Plagioclase is the dominating feldspar (Fig. 3b) but all Oligocene sandstones in addition reveal c. 5–15% K-feldspar. Comprising at least c. 50% of the sandstones, volcanic hypabyssal rock fragments prevail over metamorphic and sedimentary lithic ones (Fig. 3c). Also with regard to the inferred plate tectonic setting of the source terranes (e.g. Dickinson & Suczek 1979; Dickinson 1985) some trend is visible. A recycled orogenic nature of the source terranes is inferred for the majority of Oligocene and Miocene sandstones (Fig. 3d). An exception is represented by the Middle Eocene sample 10TTN03, which also shows particularly heavy mineral and detrital zircon age distributions compatible with recycling of older crustal elements (see below). Otherwise, most Eocene and parts of Oligocene and Miocene sandstones plot in the dissected and transitional magmatic arc fields.

Heavy mineral distributions show varying amounts of the ultrastable minerals zircon, tourmaline and rutile (ZTR, 23–60%) (Fig. 4). A variable contribution from continental crust rocks (including granitoids and volcanics) and recycled sandstones is implied (Mange & Maurer 1992). Garnet, epidote, zoisite and chloritoid grains show a wide range, from 2% up to 75%, which suggests variable detrital sources in medium-grade metamorphic rocks, or alternatively in garnet-bearing leucocratic intrusives. Oligocene sandstones generally show increased amounts of metamorphic minerals (up to 60–70%). The abundance of chromian spinel (up to 40%) suggests partly important sourcing of the detrital material from exhumed ophiolitic rocks (e.g. serpentinites and ultrabasic rocks). The Middle Eocene Ngapali Formation. turbidites, and the Middle and Late Miocene sandstones (10TTN09, 10TTN18) show highest chromian spinel occurrences (Fig. 4). The opposite trends of ultrastable mineral ZTR and the chromian-spinel/zircon index (Morton & Hallsworth 1999; Fig. 5), and the low to moderate ZTR index (20–60) implies that the observed heavy mineral distributions are due less to weathering but rather are source-rock
controlled. However, with regard to the marked surface etching of the garnets, the strong variations of the garnet–zircon index (Fig. 5) may partly be derived from temperature/burial controlled intrastratal dissolution (Morton & Hallsworth 1999).

Detrital zircon laser ablation ICP-MS U–Pb dating work in the Rakhine area revealed, in the majority of sandstone samples (8 out of 9), a dominant age distribution peaking in the mid- and Late Cretaceous, and in the Cenozoic (Fig. 6). Only concordant and quasi-concordant ages were used, as indicated by the presented concordia diagrams with each age probability curve (Fig. 6). The Cretaceous peaks are located between 123 and 66 Ma. With Palaeogene peaks ranging from 54 to 30 Ma, a nearly continuous supply of detrital zircons into the Cenozoic is documented. However, the Cretaceous–Cenozoic transition is often accentuated by a diminished number of grains. With two exceptions (10TTN16 and 10TTN03), only sparse reworking of Early Mesozoic, Palaeozoic and Precambrian zircons is observed. The Oligocene sandstone 10TTN16 reveals a prominent Cambrian–Early Ordovician population. The Middle Eocene sandstone 10TTN03 is characterized by a very broad and flat pre-Cambrian zircon distribution, together with rare Cambrian, Ordovician and Triassic grains. Even with the low number of zircons (54) available for dating in this sample, the absence of Cretaceous and Palaeogene age zircons in the source area is apparent, with similar cases of lower statistical rates still showing clear peaks for these periods. Together with the very high ZTR amount, prominent recycling of older sediments in the source area of this sandstone is corroborated.

Where possible, depending on the grain size of the zircons and their internal growth structure (supported by CL imaging), Hf isotope ratios were measured only on dated zircons. Hf isotopes in zircon of known U–Pb ages may provide information on the source of the magma from which the zircon crystallized (e.g. Griffin et al. 2004). The $e^{176}Hf(\varepsilon_\text{r})$ data from zircons older than Permian

**Fig. 3.** Detrital composition and classification of the sandstones examined in the Rakhine coastal sections.
Fig. 4. Heavy mineral composition of the sandstones of the Rakhine coastal area arranged in reconstructed stratigraphical order.
are not shown in Figure 7 because of their low relevance for the present purpose. In this plot, a majority of the $^{176}\text{Hf}(T)$ values are located between the Depleted Mantle and Chondritic Uniform Reservoir (CHUR) lines, whereas few grains fall below the CHUR line. They show a large spread in $^{176}\text{Hf}(T)$ from $+15.7$ to $+22.6$. Close to depleted mantle values are noticeable in Cretaceous ($\text{mean } +12.6$ $^{176}\text{Hf}(T)$) and Palaeogene zircons ($\text{mean } +13$ $^{176}\text{Hf}(T)$). The Hf isotopic composition for detrital zircons, clustering at 145–66 Ma shows $^{176}\text{Hf}(T)$ values of $+15.7$ to $+14.5$ with a few negative ones down to $-2.6$. In general, Cretaceous and Palaeogene detrital zircons scattered $^{176}\text{Hf}(T)$ values suggest a mixed mantle-crust origin as for subduction-related magmatism (Patchett & Tatsumoto 1980; Patchett 1983). There are no visible strong trends of change in $^{176}\text{Hf}(T)$ correlated with age. With some caution a negative trend of $^{176}\text{Hf}(T)$ values from the Cretaceous to the Palaeogene may be inferred for samples 10TTN04 and 10TTN09, which had detrital sources in magmatic arc environments (see above). A slight trend to increased continental crust involvement in magma generation may be inferred.

Aimed at determining the type of magmatic melts from which the detrital zircons were derived, a typology study (Pupin 1980) of euhedral zircons was carried out (Fig. 8), although a small number of grains was analysed. With this approach we assume that this analysis grasps in particular idiomorphic Late Cretaceous and Palaeogene zircons and, therefore, gives information on the types of arc magmas in which the reworked zircons have crystallized. In the basic typology diagram (Fig. 8) comparing the alkalinity (I.A) and temperature (I.T) indices, the zircon crystals show broader distributions with some clustering. The zircon suites contain mainly S-type crystals, with $\{100\}$, $\{110\}$ prisms and $\{101\}$, $\{211\}$ pyramids. They cluster particularly in the S7, S12, S13, S17, S18 and S19 fields, which characterize zircons derived from diorites, quartz gabbros, tonalites and granodiorites. According to the petrogenetic classification scheme for non-granitic rocks, zircons of the dominant populations (see Fig. 9) fall into the field of tonalites (T) and calc-alkaline series rhyolites (C.A.R), which figure at the commencement of hybrid granite development. The typology of the detrital zircons indicates that the parental magmas had a hybrid origin, which occurs by mixing of mantle and crustal material in the melts (Pupin 1980). This observation corroborates the $^{176}\text{Hf}(T)$ results, which were carried out on dated Cretaceous–Palaeogene zircons.

**Discussion**

The provenance of the detrital material in the Indo-Burman accretionary range in the larger framework of Himalayan orogeny has been a subject of research for decades. Often, the sediments of the IBR were reported to represent accreted deposits of the palaeo-Bengal fan (e.g. Brunnschweiler 1974; Bender 1983; Curray et al. 2003). DSDP and ODP drillings have penetrated the in-situ Bengal fan sediments, finding the last 15–17 Ma recorded (e.g. Thompson 1974; Yokoyama et al. 1990; Amano & Taira 1992; Curray et al. 2003) above a major hiatus with the pre-fan pelagic sediments of the Indian oceanic plate. The missing sediment record of the palaeo-Bengal fan was suggested to have been scraped off by eastward subduction and incorporated into the Indo-Burman accretionary wedge (IBR) and Andaman–Nicobar ridge (Curray et al. 2003). This is also a current model for inferring earlier, Eocene and younger Himalayan collisional events between the north-drifting Indian plate and Eurasia.

However, different potential sediment sources in this orogenic suture at the eastern termination of the Himalayan range can be considered: (1) the eastern Himalaya thrust belt situated on the Indian plate, (2) the Cretaceous–Cenozoic Trans-Himalayan volcanic arc situated on the Asian plate, or (3) the Burmese active (subduction) continental margin.
BUR04-09: Liang et al. (2008) Late Miocene "pre-Irrawaddy River" 82 zircons 3 > 1200 Ma youngest 49.5 ± 1.44

10TTN10: Oligocene 106 zircons 8 > 1200 Ma youngest 43.3 ± 1.24

10TTN18: Miocene 52 zircons 2 > 1200 Ma youngest 49.5 ± 1.44

10TTN09: Miocene 108 zircons 8 > 1200 Ma youngest 32.8 ± 0.63

10TTN16: Oligocene 162 zircons 30 > 1200 Ma youngest 28.1 ± 0.43

10TTN06: M. Eocene 148 zircons 3 > 1200 Ma youngest 46 ± 0.63

10TTN03: M. Eocene 54 zircons youngest 238.3 ± 2.29

10TTN04: M. Eocene 129 zircons 1 > 1200 Ma youngest 39.2 ± 1.06

10TTN013: Oligocene 56 zircons 4 > 1200 Ma youngest 28.1 ± 0.43

10TTN01: Oligocene 97 zircons 0 > 1200 Ma youngest 43.3 ± 1.24

10TTN013: Oligocene 56 zircons 4 > 1200 Ma youngest 28.1 ± 0.43

10TTN01: Oligocene 97 zircons 0 > 1200 Ma youngest 43.3 ± 1.24

10TTN013: Oligocene 56 zircons 4 > 1200 Ma youngest 28.1 ± 0.43

10TTN01: Oligocene 97 zircons 0 > 1200 Ma youngest 43.3 ± 1.24

10TTN013: Oligocene 56 zircons 4 > 1200 Ma youngest 28.1 ± 0.43

10TTN01: Oligocene 97 zircons 0 > 1200 Ma youngest 43.3 ± 1.24

10TTN013: Oligocene 56 zircons 4 > 1200 Ma youngest 28.1 ± 0.43

10TTN01: Oligocene 97 zircons 0 > 1200 Ma youngest 43.3 ± 1.24

10TTN013: Oligocene 56 zircons 4 > 1200 Ma youngest 28.1 ± 0.43

10TTN01: Oligocene 97 zircons 0 > 1200 Ma youngest 43.3 ± 1.24

10TTN013: Oligocene 56 zircons 4 > 1200 Ma youngest 28.1 ± 0.43

10TTN01: Oligocene 97 zircons 0 > 1200 Ma youngest 43.3 ± 1.24

10TTN013: Oligocene 56 zircons 4 > 1200 Ma youngest 28.1 ± 0.43

10TTN01: Oligocene 97 zircons 0 > 1200 Ma youngest 43.3 ± 1.24

10TTN013: Oligocene 56 zircons 4 > 1200 Ma youngest 28.1 ± 0.43

10TTN01: Oligocene 97 zircons 0 > 1200 Ma youngest 43.3 ± 1.24
comprising Cretaceous–Neogene magmatic arc rocks in the Mogok metamorphic belt, Shan Scarps and in the Central Basin (see also discussions in Najman et al. 2008; Allen et al. 2008). The present results need to be evaluated and compared with provenance data from coeval sediments in these various settings (Table 2).

Contemporaneous Eocene to Miocene Himalayan foreland basin series (Table 2) show common Himalaya-derived heavy mineral associations (zircon, tourmaline, epidote, zoisite, garnet) in addition to high-grade metamorphic minerals staurolite, kyanite and sillimanite at the detriment of chromian spinel (Szulc et al. 2006; Ravikant et al. 2011). Detrital zircon U–Pb age results from the Eocene–Pleistocene Subathu sub-basin in NW India reveal a dominant population ranging from 1000–900 Ma, with subordinate populations of 650–500, 1900–1800 and 2500 Ma (Ravikant et al. 2011). In the Nepalese thrust belt, the coeval Bhainskati and Dumri foreland basin formations are characterized by prominent detrital zircon age populations peaking at 500–600, 1000–1300, c. 1800 and c. 2500 Ma, which are correlated with sources in the Tethyan, Greater and Lesser Himalayan units, respectively (DeCelles et al. 2004). The three Neogene–Holocene megasequences of the Hatia Trough and Chittagong Hill tracts in Bangladesh with a presumed Himalayan provenance (Najman et al. 2012) are characterized by quartz-rich sandstones of recycled orogenic sources and, similar to the Siwlaik foreland basin series, showing high-grade metamorphic detrital grains of staurolite, kyanite and sillimanite.

Fig. 7. Diagram of epsilon Hf(T) (time-corrected) v. $^{206}\text{Pb}/^{238}\text{U}$ zircon age for sandstones of the Rakhine coastal area. Corrections in function of age are based on chondritic values (CHUR) from Blichert-Toft & Alabare`de (1997), which become the reference value. The depleted mantle evolution trend is from Griffin et al. (2002) values. The results are compared with relevant data derived from Transhimalayan batholiths, pre-Irrawaddy River (Late Miocene) and Modern Irrawaddy River sediments, as reported by Chu et al. (2006) and Liang et al. (2008). Timescale after Ogg et al. (2008).

Fig. 6. Detrital zircon laser ablation ICP-MS U–Pb age distributions (histograms and relative-age-probability curves, red), and concordia diagrams measured in sandstones of the Rakhine Coastal Belt. Discordant to slightly discordant age data are plotted. Also given is the number of zircons with ages $>1200$ Ma and the youngest dated grain in samples. For comparison (upper left) the detrital zircon ages of a Late Miocene–Pliocene transition sandstone from the Pegu Group in the Myanmar Central Basin (Liang et al. 2008). Timescale after Ogg et al. (2008).
Fig. 8. Typological distribution of zircons from sandstones of the Rakhine Coastal Belt, according to the alkalinity (IA) and temperature index (IT) of Pupin (1980). See also Figure 9.
In the Bengal Basin, including the Syleth, Surma, Assam and Haita sub-basins, Palaeogene and Neogene sandstones exhibit variable sources (Table 2). The Eocene Kopili and Oligocene Barail formations in the Syleth Trough contain low amounts of the stable minerals tourmaline, garnet, rutile, zircon and chromian spinel, leading Uddin & Lundberg (1998) to infer Indian craton derivation and intense chemical weathering. However, due to the coeval presence of pronounced c. 500 Ma and c. 1000 Ma detrital zircon age populations, and the prevalence of ZTR heavy mineral associations, Najman et al. (2008) inferred a major Trans-Himalayan arc source for the Kopili and Barail formations of the Surma Basin since about 38 Ma. In the Miocene Surma Group and younger formations, detrital staurolite, kyanite, sillimanite and orthopyroxene grains clearly suggest a prevailing Himalayan source (Uddin et al. 2007; Najman et al. 2012).

Autochthonous Bengal fan turbidites were analysed for heavy mineral contents in various sites of DSDP Leg 22 (218, 217, 219) and ODP Leg 116 (717–719). The results document Himalayan material reworking since 17–15 Ma (Langhian, Middle Miocene) (Table 2). In Leg 22 the main mineral phase is calcic amphibole (20–60%) with a general increasing trend towards younger ages, associated with epidote, garnet, ZTR, scarce tremolite, and clinopyroxene in the youngest part (Thompson 1974; Yokoyama et al. 1990). Staurolite, kyanite and sillimanite were observed in minor proportions and chromian spinel occurs very sporadically, thus correlating with other results from the Bengal Basin (Table 2). In the Bengal fan ODP sites 117–119 from the Middle Miocene upwards (c. 15 Ma to present day), the variations in calcic amphibole amounts (c. 25% and higher) and other minerals allow discrimination of Higher Himalaya from Indian subcontinent detrital sources (Amano & Taira 1992).

The eastern Trans-Himalayan batholiths in Lhasa and South China blocks like the Gangdese, Figure 9. Summary typological distribution of zircons of the Rakhine Coastal Belt in the Typological Evolutionary Trend (TET) model of Pupin (1980). The mean points of the samples plot in the hybrid field of tonalites and calc-alkaline series rhyolites of mixed crust and mantle origin.
Table 2. Compilation of provenance indicators from the Indo-Burman Ranges and contemporaneous basins in the larger Himalayan system

| Units | Indo-Burman Ranges | Indo-Burman Ranges | Central Burma Basin | Bengal Basin | Haita Trough and Chittagong Hills | Bengal Fan (DSDP/ODP drill sites) | Himalayan Foreland Basins |
|-------|-------------------|-------------------|---------------------|--------------|----------------------------------|----------------------------------|--------------------------|
| Authors | Present study | Allen *et al.* (2008) | Liang *et al.* (2008) | 1. Syleth Trough Uddin & Lundberg (1998), 2. Surma Basin Najman *et al.* (2008), 3. Assam Basin Uddin *et al.* (2007) | Najman *et al.* (2012) | 1. Thompson (1974), 2. Yokoyama *et al.* (1990), 3. Amano & Taira (1992) | 1. India/Siwalik: Szulc *et al.* (2006); 2. India/Subathu: Ravikant *et al.* (2011); 3. Nepal/Bhainskati-Dumri fms: DeCelles *et al.* (2004) |
| Sample ages | Middle Eocene–Miocene | 1. Palaeogene bed rock and river sands; 2. Neogene bed rock and river sands | Late Miocene (c. 10–5 Ma) | 1. Late Eocene–Pliocene/Pleistocene, 2. Late Eocene–Early Miocene, 3. Oligocene–Pliocene/Pleistocene | Pliocene–Recent (4–0 Ma) | 1. last 12 Ma (Late Miocene–Recent), 2 and 3. last 17 Ma (Early Miocene–Recent) | 1. ≈ 16–1 Ma (Middle Miocene–Pliocene); 2. ≈ 55–12 Ma (Eocene–Miocene); 3. Eocene–Early Miocene (≈ 55–15 Ma) |
| Sandstone provenance (Dickinson 1985) and heavy minerals | Dissected/transitional magmatic arc and recycled orogenic provenance; HM: ZTR, chromian spinel, garnet, epidote, chloritoid | 1. Magmatic arc provenance (minority); 2. Recycled orogenic provenance (majority); 1. and 2. HM: ZTR, epidote, garnet, chromian spinel (particularly in Palaeogene), traces of staurolite and amphibole (in Neogene) | 1. HM: ZTR, garnet, AKS, epidote, amphibole; 2. Recycled orogen source; 3. HM: pre-Miocene ZTR, Miocene and younger ZTR, AKS, garnet, epidote, hornblende, chloritoid | Recycled orogen provenance (quartz-rich); HM: ZTR, garnet staurolite, kyanite, amphibole, epidote, chloritoid | 1. HM: hornblende dominates and increases upsection, pyroxene, minor kyanite and sillimanite; 2. HM: calcic amphibole, epidote, garnet, clinopyroxene, ZTR; 3. HM: calcic amphibole, epidote, garnet, AKS, ZTR, pumpellyte | 1. Recycled orogenic provenance; HM: garnet; ZTR, epidote group, chloritoid, staurolite, kyanite and sillimanite, minor chloritoid; 2. no data available; 3. Cratunal and recycle orogen provenance |
| Detrital zircons – concordant ages (U–Pb) | Major populations: 123–66 Ma (Cretaceous) and 54–30 Ma | 1. c. 110–40 Ma populations dominate, minor 500–650 Ma, 2. Cretaceous–Cenozoic (younging upward) combined with minor peaks at ≈ 500–600 Early Palaeozoic–Precambrian zircons dominate with prominent | Nearly continuous Cretaceous–Cenozoic record: 110–70 Ma and | 1. no data available; 2. population 1000–900 Ma most prominent, others 650–
**Zircon Hf isotope ratios**

| General range: | eHf(T) c. 16–1; 145–66 Ma: eHf(T) c. +15.7 to +3.8, 65–32 Ma: eHf(T) c. +14.5 to +0.97 |

**Authors' interpretation of provenance**

| Burman Late Cretaceous–Palaeogene magmatic arc, accreted ophiolitic series and continental margin basement and cover (i.e. Mogok Metamorphic Belt, Shan Plateau) | Palaeo-Irrawaddy River draining the Trans-Himalayan region |
| Process | Palaeogene: significant Burmese arc-derived input; Neogene: Himalayan source predominates |
| Predominating Himalayan source | 1. pre-Miocene from Indian continent, Miocene and younger from Himalayan orogenic front; 2. pre-Late Eocene from Indian continent, Late Eocene and younger: Himalayan source with a Burman arc component; 3. eastern Himalayas and Indo-Burman Ranges |

**Abbreviations:** Heavy minerals, HM; ultrastable heavy mineral assemblage zircon–tourmaline–rutile (may also include brookite, anatase and sphene), ZTR; high-grade metamorphic association andalusite–kyanite–sillimanite, AKS.
Bomi–Chayu and Dianxi–Burma batholiths are characterized by prominent zircon age populations of Jurassic, Cretaceous and early Palaeogene (up to 45 Ma) age (Liang et al. 2008); Bomi–Chayu age ranges: 140–105 Ma, 65–50 Ma; Dianxi–Burma batholith: 140–110 Ma, 90–50 Ma; Gangdese batholith: 210–180 Ma, 115–75 Ma and 55–45 Ma (Chu et al. 2006). These batholith age clusters show some incomplete overlap with the present detrital zircon ages of the Rakhine Coastal Belt, and closest similarity exists with the Gangdese Cretaceous and Palaeogene zircons with respect to the reported eHf(t) values (Chu et al. 2006) (Fig. 7). However, the Jurassic component is much more developed in that batholith than in the detrital zircons described here. The Bomi–Chayu and Dianxi–Burma granitoids certainly can be excluded as important sources, because these batholiths systematically reveal low and negative eHf(t) values between +5 and −15 in Cretaceous and Palaeogene zircons (Fig. 7; Liang et al. 2008). With regard to the detrital zircon populations documented in the Rakhine Coastal Belt, there is an affinity observed with the age distributions reported by Liang et al. (2008) from a Late Miocene Pegu Group sandstone in the Myanmar Central Basin (Fig. 6; Table 2). However, Palaeogene (c. 65–23 Ma) zircons from this sandstone reveal a strong negative eHf(t) component (≈ −4 to −18; Liang et al. 2008), and a similar igneous rock sourcing of coeval sandstones in the Rakhine coastal area (Fig. 7) can be excluded. Presumably, this sandstone of the Pegu Group is younger than our youngest sample (base of Late Miocene, 10TTN18) and the detrital contribution from the Dianxi–Burma and Bomi–Chayu batholiths, as inferred by Liang et al. (2008), occurred later, or the river drainage pattern was more complex.

In Myanmar, Jurassic to Neogene subduction-related magmatic activity is documented in the MMB, in the Shan Scarps and in the Myanmar Central Basin (Mitchell 1993; Barley et al. 2003; Mitchell et al. 2007; Searle et al. 2007). Unfortunately, few occurrences have been dated with modern geochronological methods yet (Mitchell et al. 2007). The MMB represents a middle to lower crustal section giving evidence of magmatization and magmatic intrusions at different times, and final high-grade metamorphism lasting at least from c. 43–29 Ma (Searle et al. 2007). The MMB was exhumed prior to displacement along the right-lateral, brittle Sagaing fault, which initiated between 22 and 16 Ma (Barley et al. 2003; Bertrand & Rangin 2003; Searle et al. 2007). Mid-Cretaceous to Eocene I-type granitoids (c. 120–50 Ma) in the MMB suggest that in eastern Myanmar and southern Eurasia in general (from Pakistan to Sumatra), an Andean-type subduction margin existed during that time (Barley et al. 2003). Oligocene zircons were identified in the Mandalay Hill area (c. 30 Ma) and Yesin dam area (c. 25 Ma; Barley et al. 2003). During the youngest metamorphic event (c. 45–24 Ma), hybrid mantle-derived hornblende syenites (35–23 Ma) and crust-derived leucogranite melts (24.5 Ma) were produced (Searle et al. 2007).

South of Mandalay, a granitic belt cuts the Shan Scarps. The plutons intrude Late Jurassic–Early Cretaceous sediment series and provided age ranges from Early Cretaceous to Early Miocene (Darbyshire & Swainbank 1988). Cobb et al. (1992) concluded hybrid magma sources from mantle and crust for these intrusions. In the Myanmar Central Basin (Banmauk area) the allochthonous Maunggyi andesites and ophiolitic suites (Salingyi) are intruded by c. 106–91 Ma old (K–Ar data) granodioritic/tonalitic plutons. According to Mitchell (1993), intrusive rocks covering the Cretaceous–Cenozoic age transition (60–70 Ma) are rather rare. This author suggests that subduction and magmatism ceased during ophiolite obduction in the Indo-Burman Ranges during this period. Subduction and magmatic activity resumed in the Palaeogene. Eocene (c. 50–53 Ma, K–Ar) magmatic activity (andesites and quartzdiorites) is documented in the Shangalon porphyry copper near the Banmauk granodioritic plutons (Mitchell 1993). In the Myanmar Central Basin an Oligocene age (c. 38 Ma) of a granodiorite intrusion is also reported from the area south of Banmauk. Neogene–Quaternary volcanic arc activity is represented by the porphyry copper at Myunywa north of Salingyi and the Mt Popa volcano (Mitchell 1993).

According to the various provenance aspects, the investigated sandstones of the Rakhine coastal area show nearly coherent results. With one exception (Middle Eocene 10TTN03), a dominant sourcing of sediment from Late Cretaceous and Palaeogene (tonalitic/calc-alkaline rhyolitic) magmatic rocks is indicated. The bimodal occurrence of Late Cretaceous and Palaeogene detrital zircons (Fig. 6) is in agreement with Burman arc age distributions as suggested by Mitchell (1993). A mixed mantle–crust origin of the magmas is inferred by the measured Hf isotope ratios and the typology of the zircons. Consequently, a Late Cretaceous and Palaeogene subduction-related volcanic arc environment for the sediment source areas is evident. Ultrastable heavy mineral associations (ZTR), chromian spinel and variable medium-grade metamorphic mineral spectra show that the magmatic rocks were associated with sources that comprised continental and oceanic basement and older sedimentary rocks. This is in agreement with the development of the arcs in the Mogok Metamorphic Belt, the Shan Scarps and Central Basin (Mitchell
The nearly ubiquitous chromian spinel in the heavy mineral fractions implies additional sources from accreted ophiolitic rocks in the Indo-Burman Ranges. Unless some heavy mineral grains, such as garnet or epidote, stem from leucocratic intrusions (frequent in the MMB; Barley et al. 2003) or from the alteration of basic volcanic rocks, the heavy minerals in general indicate the erosion of metamorphic basement as exposed today at the eastern edge of the Shan Plateau. Discriminating key heavy minerals indicative for Himalayan sources, such as calcic amphiboles, clinopyroxene, staurolite, andalusite, kyanite and sillimanite, were not detected by us in sandstones of the Rakhine coastal area. Almost all Middle Eocene to Late Miocene sandstones show only subordinate – in the range of single grains – reworking of older Mesozoic, Palaeozoic and Precambrian zircons (Fig. 6). One Oligocene sandstone (10TTN16) sample indicates a strong sourcing in Cambrian–Ordovician rocks or their secondary reworking. Further noticeable Proterozoic detrital zircon populations, as typically reported from Himalayan foreland basins (Table 2 and references therein), are not obvious.

Two sandstones reveal a different composition and need further discussion. The Middle Eocene sample 10TTN03 shows a high recycling index in the heavy minerals (ZTR >60) and a broad Early Palaeozoic–Proterozoic detrital zircon age distribution, including one Triassic zircon, with only one to two grains per age bin (bin width is 15 Ma; Fig. 6). A dominant source area in exhumed Triassic flysch-like turbidite series (e.g. Mitchell 1993) in the IBR or in Palaeozoic sandstone formations in the Shan Plateau and Scars could explain these very restricted grain and age spectra. The other example, the Oligocene sandstone 10TTN16, is characterized by the co-existence of prominent bimodal Burman arc zircon populations with a strong Cambrian peak and a broad Proterozoic age spectrum. A c. 500–650 Ma age population, together with a strong Proterozoic component, could be indicative of a Himalayan source, as revealed in the foreland basins (DeCelles et al. 2004; Ravikant et al. 2011). Therefore, the origin of the detritus of this sandstone may be the confluence of Burman and Himalayan material in a turbiditic source. However, with respect to the overall picture, a secondary supply from reworking Cambrian sediments and Proterozoic basement in the Shan Plateau and Scars (Mitchell 1993) could be an explanation.

The presence of detrital zircon populations of 450–600 and 700–1200 Ma in river sands and bedrocks in the Arkhan coastal area was used by Allen et al. (2008) to infer a predominant Himalayan sourcing from the Neogene onwards. However, these heavy mineral assemblages (available from Allen et al. 2008, supplements) do not correlate with such ones reported from coeval sediments in the larger Bengal Basin and Himalayan foreland basin occurrences (Table 2), instead showing qualitatively similar assemblages, as in the present study. Finally, we arrive at the similar conclusion that during the Palaeogene the dominating source of detrital material was the Burman arc and crystalline basement with sedimentary cover. The studied Neogene deposits do not yet show strong influx from the Himalayan range, but some river confinement and consequent mixing of provenances may have occurred.

The results of the present study have interesting implications for hydrocarbon exploration in the larger area of Myanmar, because they suggest reservoirs were not only derived from the Himalayas but predominantly from the Burman active margin, and a connection with the development of the Myanmar Central Basin may be established. There exist good correlations of thick sediment packages of pre-Eocene, Eocene and post-Eocene age in the Rakhine Coastal Belt with those of the MCB and deep-water offshore areas. Further investigation of the stratigraphy and the history of deposition in the MCB will be fundamental. This in the context of an already proven petroleum system along the Rakhine coastal area, with significant hydrocarbon discoveries in stratigraphic traps or combined structural–stratigraphic traps, infallible indications of gas seepages from mud volcanoes, and local oil production from hand-dug wells.

Conclusions

The present work evaluated the provenance of detrital material in Middle Eocene to Late Miocene sandstones comprised in the Indo-Burman Ranges distal accretionary wedge and slope series in the Rakhine Coastal Belt. This area has large-scale significance for the reconstruction of the collisional history of the Indian with the Eurasian plate and Himalayan orogenesis in general. Our results do not confirm opinions that the Rakhine detrital sediments were supplied from Himalayan sources, and therefore, could be proxies of the palaeo-Bengal fan and early (≤ 55 Ma, Eocene and younger) collisional events in the Himalaya.

Our results, from framework grain, heavy mineral and detrital zircon analyses suggest a principal sourcing of the detrital material from the Late Cretaceous and Palaeogene magmatic arc and basement formations related to the Indian plate subduction beneath the Burman margin, as earlier suggested by Mitchell (1993). Ophiolitic series in
the Indo-Burman Ranges, and metamorphic complexes in the larger Mogok Metamorphic Belt and Shan Plateau area represent secondary sources.

With regard to the sediment routing from the Burman active margin, we infer that, from Middle Eocene to Middle Miocene, the erosional scree presumably was transported by rivers and subsequent turbiditic flows into the Burman trench and slope basins to be accreted and uplifted in the external Indo-Burman Ranges. Unconformable Late Miocene shallow shelf sediment shows a similar volcanic arc signature. However, in the light of the limited geographical and stratigraphic distribution of our sample material, we cannot exclude the merging of Burman arc and Himalayan sources in places closer to the Himalayan Range to the north and in younger times.

A better understanding of the provenance of the sandstones in the Rakhine Coastal Belt may also contribute to the reassessment of the petroleum system in the Indo-Burman Ranges and Myanmar Central Basin.

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