



universität
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DISSERTATION / DOCTORAL THESIS

Titel der Dissertation / Title of the Doctoral Thesis

„Stratigraphic architecture and depositional sequence
stratigraphy of the Jurassic Datta Formation in the low
latitude Tethyan Salt Range Pakistan“

verfasst von / submitted by

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angestrebter akademischer Grad / in partial fulfilment of the requirements for the degree of
Doktor der Naturwissenschaften (Dr. rer. nat.)

Wien, 2018 / Vienna, 2018

Studienkennzahl lt. Studienblatt /
degree programme code as it appears on the student
record sheet:

A 796 605 426

Dissertationsgebiet lt. Studienblatt /
field of study as it appears on the student record sheet:

Erdwissenschaften

Betreut von / Supervisor:

Ao. Univ. Prof. Dr. Michael Wagreich

Mitbetreut von / Co-Supervisor:

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ABSTRACT

The Triassic–Jurassic boundary (TJB) represents an interval of a major biotic crisis in the Earth’s Phanerozoic history and coincides with a prominent sea-level fall. The nature and severity of the biotic crises and the magnitude and extend of sea-level fall at the boundary interval have remained the focus of many studies. Both the events are linked to the Central Atlantic Magmatic Province (CAMP) basaltic activity and the associated hypotheses of extreme greenhouse warming is considered as a favoured mechanism for widespread extinctions in the marine and terrestrial realms. However, most of these evidences so far are based on studies focused on the European basins including the western Tethys, and North America, the then northern hemisphere (Laurasia). No significant data is available from the then southern hemisphere (Gondwana). In the western Salt Range and Trans–Indus ranges of Pakistan, the thick dolomite of the Upper Triassic Kingriali Formation and the overlying siliciclastic rich strata of the Lower Jurassic Datta Formation represent the Triassic–Jurassic transition. The present study uses outcrop data, framework composition (petrography), bulk mineralogy (XRD), clay mineralogy, bulk sediments geochemistry and heavy mineral analysis to unravel the depositional palaeoenvironments, spatio-temporal facies architecture, and source to sink analyses of the Datta Formation. The palaeoclimatic conditions during the Triassic–Jurassic transition have been worked out in the area to test the hypothesis of global extreme greenhouse conditions. Palynological analysis has been used to establish the biostratigraphic positions of the Kingriali and Datta formations.

Following the demise of the Upper Triassic platform carbonates of the Kingriali Formation; terrestrial settings established in the Salt and Trans–Indus ranges. A major continental, fluvial-deltaic depositional system was established in the area and it deposited the Lower Jurassic Datta Formation in the western Salt and Tran–Indus ranges. In the western Salt Range mostly channel belt, channel margin/floodplain and lagoonal/marginal marine facies associations were deposited by this system. However, in the Trans–Indus ranges the lagoonal/marginal facies association pinches out and dominantly continental-fluvial facies associations were deposited in these areas.

During the deposition of the Kingriali Formation arid-semiarid hothouse palaeoclimatic conditions prevailed in the area. The illite rich clay mineralogy (>90% illite), presence of gypsum beds, low values of chemical index of alteration (CIA; average 77) and other

geochemical proxies for the upper part of the Kingriali Formation provide conclusive evidences for such palaeoclimate. Abnormally high total rare earth elements content (REE > 6000 ppm in Nammal Gorge) of the topmost green shale of the Kingriali Formation provides a potential correlation horizon between this and the volcanism related to the Central Atlantic Magmatic Province (CAMP) activity. Following this the Datta Formation was deposited in an overall hot and humid greenhouse conditions. The kaolinite rich clay mineralogy (>90% kaolinite), supreme mature quartz arenites, boehmite, haematite rich laterites, very high CIA values (average > 92) and combined results of the geochemical proxies all advocate deposition under greenhouse conditions.

Exhumation and subsequent exposure of the Permian and older siliciclastics of the Salt Range and areas further east –southeast in Indian craton, in a shoulder rift tectonic setting, provided the source for the siliclastic detritus. These sediments were transported and reworked by the fluvial-deltaic system in a west-northwestward direction to the depositional site of the Datta Formation. The quartz arenite nature, very high ZTR-index (> 95) and geochemical proxies all favour reworking and recycling of sediments under greenhouse conditions following the late Triassic aridity. These proxies provide coherent evidences for sediments derivation from a dominantly acidic igneous source exposed towards the east-southeast in the Indian craton; the Aravalli–Malani ranges.

The dinoflagellate cysts *Rhaetogonyaulax*, *Suessia* and *Baumontella* confirm a Rhaetian age for the Kingriali Formation. The first occurrence of *Cerebropollenites thiergatii* within the lower part of the Datta Formation indicates the approximate position of the base of the Hettangian. Thus the Salt and Trans–Indus ranges of Pakistan preserve a complete record of sedimentation across the Triassic–Jurassic transition. This study presents one among the first outcrop areas of the TJB interval from the Gondwana region and may open new area for the understanding of the Triassic–Jurassic transition.

ZUSAMMENFASSUNG

Die Trias-Jura-Grenze stellt einen Zeitraum einer großen Krise in der Erdgeschichte dar und geht einher mit einem ausgeprägten Meeresspiegelabfall. Die Natur und Schwere der Organismenkrise und die Magnitude und Ausmaß des Meeresspiegelrückganges an dieser Grenze sind Ziel vieler wissenschaftlicher Untersuchungen. Beide Ereignisse stehen in Zusammenhang mit dem basaltischen Vulkanismus der Zentralatlantischen Magmatischen Provinz (Central Atlantic Magmatic Province, CAMP) und der damit verbundenen Hypothese extremen Treibhausklimas, die als der bestimmende Prozess gesehen wird, der zu den weitverbreiteten Aussterbeereignissen im marinen und terrestrischen Bereichen geführt hat. Allerdings sind die meisten Studien bisher in europäischen Sedimentarchiven und der West-Tethys sowie Nordamerika durchgeführt worden, also auf die nördliche Hemisphäre (Laurasia) beschränkt. Es gibt kaum signifikante Daten aus der Südhemisphäre (Gondwana). In Pakistan, in der westlichen Salt Range und den Trans-Indus Ketten, wird der Trias-Jura-Grenzintervall durch die mächtigen Dolomite der obertriassischen Kingriali-Formation und den überlagernden siliziklastikreichen Schichten der unterjurassischen Datta-Formation gebildet. Die vorliegende Arbeit benutzt Aufschlussdaten, Modalzusammensetzungen (Petrographie), Gesamtgesteinsmineralogie (XRD), Tonmineralogie, Gesamtgesteinschemismus und Schwermineralanalysen um die Ablagerungsmilieus, Faziesarchitektur in Raum und Zeit, und die Sedimentquellen und Sedimentfallen der Datta-Formation zu analysieren. Die paläoklimatischen Bedingungen während des Trias-Jura-Überganges werden herausgearbeitet um die Hypothese eines globalen extremen Treibhausklimas zu testen. Palynologische Untersuchungen wurden unternommen, um die biostratigraphische Einzeitung der Kingriali- und Datta-Formationen zu rekonstruieren.

Nach dem Niedergang der obertriassischen Karbonatplattformen der Kingriali-Formation etablierten sich terrestrische Ablagerungsbedingungen in der Salt- und der Trans-Indus Range. Ein ausgedehntes fluviatil-deltaisches Ablagerungssystem entstand mit der unterjurassischen Datta-Formation. In der westlichen Salt Range treten Rinnenfüllungen, Rinnenrandsedimente-Überflutungssedimente und lagunäre-randlich marine Ablagerungen auf. In den Trans-Indus Ketten keilen die lagunär - randlich marine Ablagerungen aus, und vorwiegend kontinental-fluviatile Faziesvergesellschaftungen wurden abgelagert.

Während der Ablagerung der Kingriali-Formation herrschte ein arid-semiarides Treibhausklima. Belegt wird dieses Paläoklima durch die Illit-reiche Tonmineralzusammensetzungen (>90% Illit), das Auftreten von Gips, niedrige Werte des Chemical Index of Alteration (CIA; Mittelwert bei 77) und andere geochemische Signale aus dem oberen Abschnitt der Kingriali-Formation. Abnormal erhöhte Gehalte an Seltenen Erden (REE > 6000 ppm in der Nammal Schlucht) der obersten grünlichen Tonsteine der Kingriali-Formation sind als potentieller Korrelationshorizont zu dem Vulkanismus der Central Atlantic Magmatic Province zu interpretieren. Die folgende Ablagerung der Datta-Formation fand unter einem heißen und humiden Treibhausklima statt. Kaolinit-reiche Tonmineralogie (>90% Kaolinit), superreife Quarzarenite, Böhmit- und Hämatit-reiche Laterite, und sehr hohe Werte des Chemical Index of Alteration (Mittelwert größer 92) und andere geochemische Signale zeigen alle Ablagerung unter Treibhausbedingungen.

Das siliziklastische Material wurde durch Aufsteigen und Freilegung permischer und älterer siliziklastischer Sedimente im Liefergebiet der Salt Range und in Gebieten weiter östlich und südöstlich am Indischen Kraton, in einem extensionellen Regime mit Grabenbrüchen mit randlichen Erhebungen, bereitgestellt. Diese Sedimente wurden durch ein westlich bis nordwestlich orientiertes fluviatil-deltaisches System wiederaufgearbeitet und zum Ablagerungsgebiet der Datta-Formation transportiert. Das Auftreten von Quarzareniten, der sehr hohe ZTR-Index der Schwerminerale (>95%) und weitere geochemische Signale weisen alle auf Wiederaufarbeitung der Sedimente in einem Treibhausklima nach dem ariden Spättriasklima hin. Ein dominant saures magmatisches Liefergebiet im Ost-Südosten im Indischen Kraton, in den Aravalli-Malani-Ketten, kann nachgewiesen werden.

Das Rhätium-Alter der Kingriali-Formation wird durch die Dinoflagellatenzysten *Rhaetogonyaulax*, *Suessia* und *Baumontella* bestätigt. Das Erstauftreten von *Cerebropollenites thiergatii* im unteren Teil der Datta-Formation belegt die Basis des Hettangiums. Damit wird auch ein komplettes Sedimentarchiv über die Trias-Jura-Grenze biostratigraphisch belegt. Diese Arbeit präsentiert damit eine der ersten Trias-Jura-Sedimentabfolgen aus Gondwana und kann daher zu einem neuen und tieferen Verständnis der Trias-Jura-Grenze beitragen.

ACKNOWLEDGEMENT

I am thankful to my supervisor Ao. Univ. Prof. Dr. Michael Wagreich for his kind nature, professional guidance, prompt availability and support (both professional and personal) during the study. I pay special thanks to the Higher Education Commission (HEC), Pakistan for financially supporting the study under the OSS-II/HEC Overseas Scholarships MS/MPhil leading to PhD in Selected Fields, Phase II. The Austrian Agency for International Cooperation in Education and Research (OeAD) handled the official and financial aspects of the scholarship. I am thankful to Ao. Univ. Prof. Dr. Susanne Gier for her help during the XRD, clay mineralogy, petrographic and heavy minerals analyses. Special thanks to Prof. Dr. Leopold Krystyn, Prof. Dr. Patrick Meister and Dr. Benjamin Sames for their help and discussion on various aspects of the study. I am thankful all my co-authors for their time and support. Prof. Dr. Wolfram Michael Kürschner (University of Oslo, Norway) is thanked for his help in the palynological analysis. I am thankful to Claudia Beybel, Ilka Wünsche and Leopold Slawek for their technical support in thin section preparation. Sabine Hruby-Nichtenberger is thanked for her help in the heavy minerals samples preparation. I am also thankful to Beatrix Bethke for her help during the laboratory analyses. Ali Ahmad Gomaa, Erik Wolfgring, Maria Meszar, Shaha Liu, Dua Fathy and Seyed Hamidreza Azami are thanked for their help.

I thank Dr. Irfanullah Jan, Dr. Muhammad Hanif, Shahid Hussain, Mukhtiar Ghani, Syed Irfan Hashmi and Nasar Khan for their help during the laborious fieldwork. IGCP-609 financed the fieldwork and geochemical analysis and National Centre for Excellence in Geology (NCEG), University of Peshawar, Pakistan provided logistic support during the fieldwork under the NRPU project awarded to Dr. Irfanullah Jan.

Special thanks to my parents, brothers and sisters who helped me at every stage of my life and to my lovely wife Mehwish Bibi for providing a discussion platform at home and sharing ideas on various aspects of my study. Thanks to my daughter Harram Fatima Shahid and son Razaan Shahid for being the source of my relaxation in the stress times.

CHAPTER 1

INTRODUCTION

1.1 Introduction and research question

The Triassic–Jurassic boundary (TJB) represents one of the five major mass extinctions in the Earth’s Phanerozoic history (Raup and Sepkoski, 1982). Considering the ecological severity, this event is ranked as the third most severe crisis of the Phanerozoic (McGhee et al., 2004, 2013) with an estimated 40–73% loss in biodiversity (Sepkoski, 1996; Alroy, 2010). Many of the traditionally used marine stratigraphic markers fossil groups like ammonites, radiolaria, articulate brachiopods, gastropods and bivalves suffered heavily and conodonts perished completely (Lindström et al., 2017). In total, 103 families went extinct, 53% of all genera and 76% of species became extinct (Sepkoski, 1996). The event also had strong impact on terrestrial fauna and flora. Vertebrates suffered and tetrapods including *Pseudosuchia* (diapsida), synapsids (first mammals) and *temnospondyls* (amphibians) became extinct (Sepkoski, 1996; Colbert et al., 2001; Lindström et al., 2017). The flora suffered significantly and in Europe 95% of the flora went extinct. The gymnosperm forests in Europe were for a spell replaced by fern and fern-associated vegetation, an assemblage common in disturbed ecosystems (van de Schootbrugge et al., 2009). In addition, this interval is also characterised by a prominent sea-level fall (Hallam, 2001; Haq, 2017). Both the biotic crises and the sea-level fluctuations at the boundary interval have remained the focus of many studies. Both events are linked to a Large Igneous Province (LIP) volcanism, i.e. the Central Atlantic Magmatic Province (CAMP) basaltic activity (e.g. Marzoli et al., 1999; Olsen, 1999; Pálfi et al., 2001; Wignall, 2001; Galli et al., 2005), including hypotheses of extreme greenhouse warming as the cause for widespread extinctions in the marine and terrestrial realms. For the reasons mentioned above, the Late Triassic (Rhaetian) and Early Jurassic (Hettangian) strata have been explored in detail in many parts of the world. However, most of these studies have focused so far on the European basins including the western Tethys, and North America, the then northern hemisphere (Laurasia). No significant data is available from the then southern hemisphere (Gondwana).

The Global Stratotype Section and Point (GSSP) for the base of the Hettangian (and thus Early Jurassic) is located in the Karwendel Mountains, Northern Calcareous Alps, Austria (Morton, 2012; Hillebrandt et al., 2013). The stratigraphic succession includes the Eiberg

Member of the Kössen Formation (Rhaetian) and the Tiefengraben Member of the overlying Kendlbach Formation. The initial strong negative $\delta^{13}\text{C}_{\text{org}}$ corresponds to the boundary between the Eiberg Member (Kössen Formation) and Tiefengraben Member (Kendlbach Formation) while the Schattwald Beds display a shift to more positive $\delta^{13}\text{C}_{\text{org}}$ values. The main isotopic excursion of negative $\delta^{13}\text{C}_{\text{org}}$ values starts at the transition of the Schattwald Beds to the proximate Tiefengraben Member (Ruhl et al., 2009; 2011; Hillebrandt et al., 2013). This corresponds to the Triassic–Jurassic boundary and the “Golden Spike” was fixed at Kuhjoch East (47°29′02″N, 11°31′50″E). The TJB corresponds to the first occurrence (FO) of the ammonite *Psiloceras spelae tirolicum* (Hillebrandt and Krystyn, 2009). Other bio-events include the FO of *Praegubkinella turgescens* (aragonitic foraminifer), and of diverse ostracod species 1.0–3.40 cm below the FO of *P. spelae*. The continental palynomorph *Cerebropollenites thiergartii* occurs 3.2 m below *P. spelae*, the closest event to the FO of *P. spelae* and allows a correlation with terrestrial sediments (Hillebrandt et al., 2013). The radiometric investigations of the TJB indicate an age about 201.36 Ma (Schoene et al., 2010; Marzoli et al., 2011; Guex et al., 2012; Wotzlaw et al., 2014). A correlation between the important boundary sections on either sides of the CAMP activity is recently provided by Lindström et al. (2017).

A prominent sea-level fall is a common feature of many TJB sections (Hallam, 1997, 2001). In the central Rhaetian Sea the permanent shallow-marine stratification during the middle Rhaetian (Blain et al., 2007) abruptly ended in the latest Rhaetian (Blumenberg et al., 2016). This was triggered by an extensive regressive phase (Hesselbo et al., 2004), probably caused by glaciation as result of volcanic ash and SO₂ exhalation during the CAMP activity that led to the development of a “volcanic winter” (Schoene et al., 2010). This shallowed the Rhaetian Sea (Schoene et al., 2010), thereby terminating the stratification. Shortly after the TJB interval the sea level re-established in the lowest Jurassic (Richoz et al., 2012; Jaraula et al., 2013) and evidences for a progressing sea-level rise exists in the following Early Jurassic (Hallam, 1997; Hallam and Wignall, 1999).

In the Salt and Trans–Indus ranges of Pakistan, a thick, ridge forming dolomite of the Kingriali Formation represents Upper Triassic strata (Iqbal et al., 2014a, b; 2015b). The siliciclastic rich Early Jurassic Datta Formation overlies the Kingriali Formation in the area. Unfortunately, the stratigraphic positions of both the Kingriali and Datta formations have not been established in detail so far due to the lack of such studies. Both are very tentatively

dated based on their relative stratigraphic positions. Nor there has been any study conducted to understand this transition from the Upper Triassic dolomite to the overlying Lower Jurassic siliciclastics.

The present study aims at unravelling the TJB interval in a poorly known area from the low latitudes of the Southern Hemisphere. It is an attempt to work out the depositional palaeoenvironments, depositional sequence stratigraphy and spatio-temporal facies architecture, and source to sink analyses of the Lower Jurassic Datta Formation in the Salt and Trans-Indus ranges, Pakistan. Furthermore, an attempt has been made to understand the palaeoclimatic conditions during the Triassic–Jurassic transition in the area to test the hypothesis of global extreme greenhouse conditions.

The stratigraphic position of the Upper Triassic–Lower Jurassic succession has been worked out and reliable ages for the two formations have been established during this study (Kuerschner et al., 2017). A regional and global correlation of the succession and the global events, including the end-Triassic extinction event (ETE), the sea-level fall, and the base of the Hettangian as defined by the GSSP in Kuhjoch, Austria (Hillebrandt et al., 2013) has been established with India, the European basins and areas around the CAMP. The study uses outcrop data, framework composition (petrography), bulk mineralogy (XRD), clay mineralogy, bulk rock geochemistry, heavy mineral analysis and palynology to achieve the above mentioned targets. This study presents one among the first outcrop areas of the TJB interval from the Gondwana region and may open new area for the understanding of the Triassic–Jurassic transition.

1.2 Location and accessibility

The Salt Range, with its Trans–Indus extension, forms the southern boundary of the Kohat–Potwar Plateau (Iqbal et al., 2015 a, b; Fig. 1.1). The plateau is a part of the foreland zone of the Himalayan Fold and Thrust Belt (HFTB). The nearly EW trending Salt Range (Fig. 1.2) demarcates the surface expression of the Salt Range Thrust (Kazmi and Jan 1997; Jadoon et al., 1997; 2015 and Iqbal et al., 2015a). The study area is bounded by longitudes 70°30' - 73°30' East and latitudes 32° - 33° North (Figs. 1.1 and 1.2). The studied sections are located in the western Salt Range and adjoining Trans–Indus ranges (Fig. 1.2 and Table 1.1). Nearly all parts of the Salt Range are accessible through metalled and unmetalled roads. Some

parts like Khewra are also approachable by train. Ideal time for fieldwork in the area is November to mid-March as summers are very hot. Field work was mainly performed during years 2014 (December - January) to 2016 (December - January) based on former field campaigns of earlier workers and reconnaissance fieldwork during 2012 – 2013 (October – March).

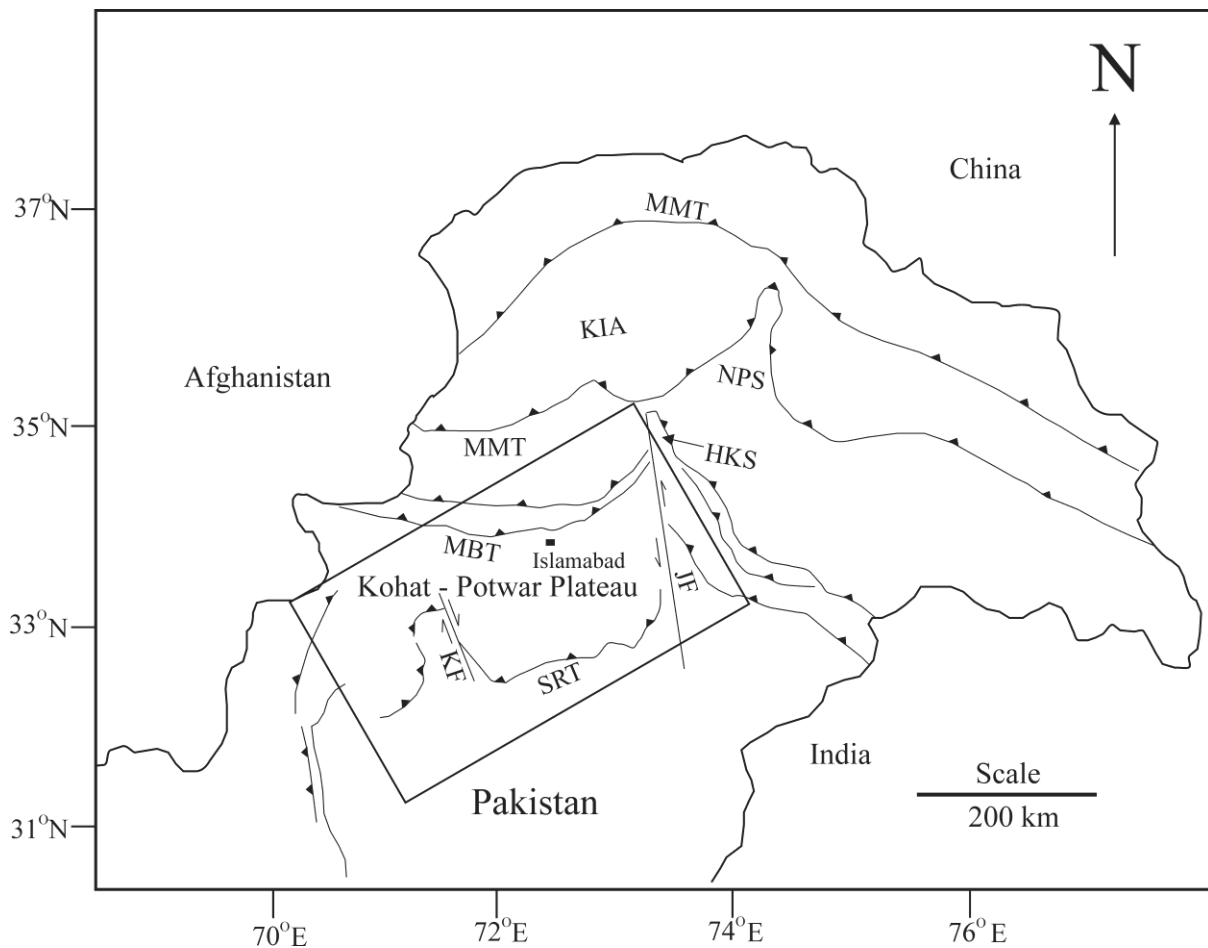


Figure 1.1 Tectonic map of north Pakistan showing major tectonic units and structures. (MKT = Main Karakoram Thrust, KIA = Kohistan Island Arc, MMT = Main Mantle Thrust, MBT = Main Boundary Thrust, SRT = Salt Range Thrust, JF = Jhelum Fault, KF = Kalabagh Fault, NPS = Nanga Parbat Syntaxis, HKS = Hazara Kashmir Syntaxis. The rectangle highlights the study area (after Iqbal et al., 2015a).

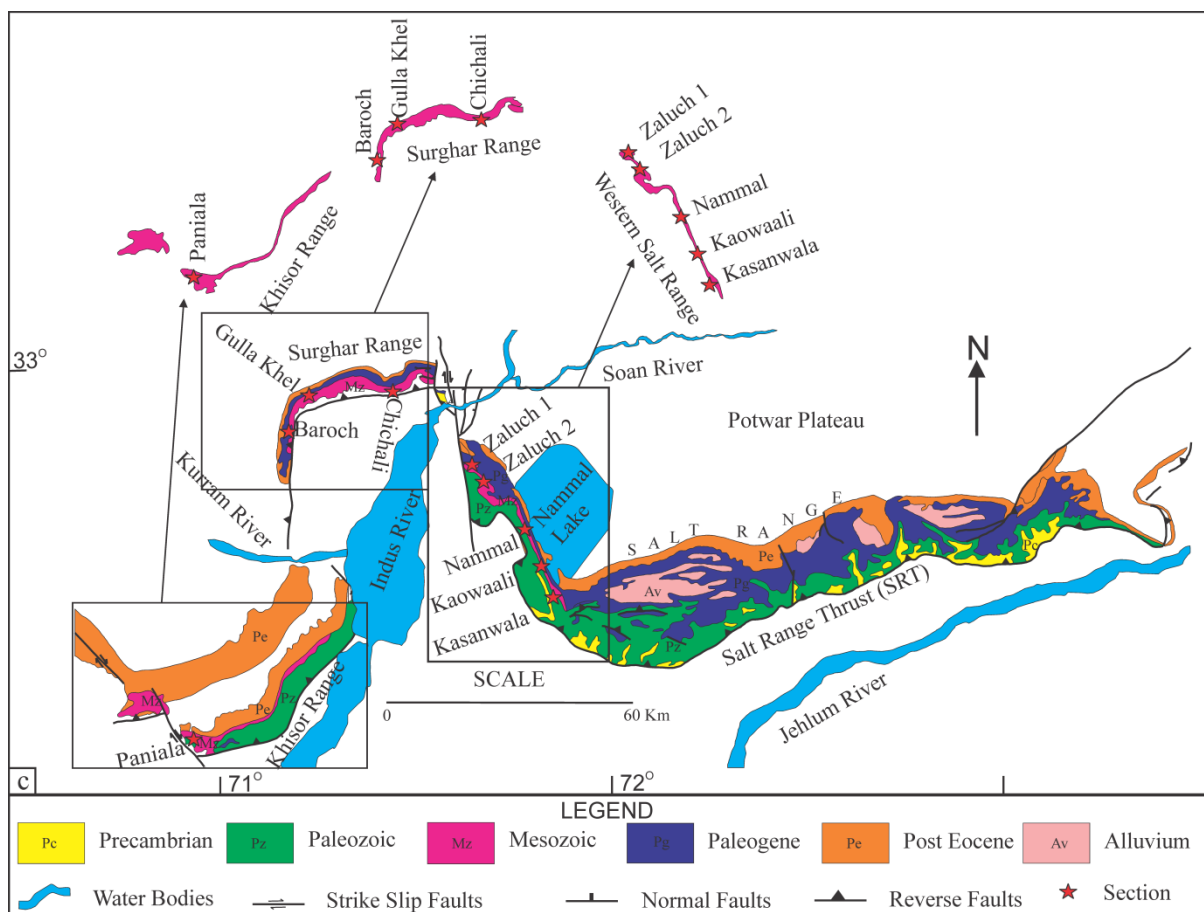


Figure 1.2. Geological map of the Salt Range and Trans-Indus ranges showing locations of the studied sections. The Western Salt Range, Surghar Range, Khisor Range and the studied sections are highlighted (Iqbal et al., 2018(under review)).

Table 1.1 Geographic positions of the studied sections (see Fig. 1.2 for the locations of the studied sections).

S. No	Locality Name	Area	Latitude	Longitude
1	Kasanwala Nala	Western Salt Range	32°36'14.17"N	71°49'45.63"E
2	Kaowaali		32°37'1.28"N	71°49'13.00"E
3	Nammal Gorge		32°39'43.83"N	71°47'48.23"E
4	Zaluch Nala1		32°47'14.95"N	71°39'11.67"E
5	Zaluch Nala2		32°46'1.45"N	71°39'54.06"E
6	Chichali Nala	Surghar Range	32°59'57.34"N	71°24'9.66"E
7	Gulla Khel		32°58'28.80"N	71°13'23.17"E
8	Baroch Nala		32°55'5.71"N	71°9'1.53"E
9	Paniala	Khisor Range	32°13'25.19"N	70°53'49.33"E

1.3 Methods

1.3.1. Seismic sections and borehole logs

Balkassar area, Potwar Plateau was selected for the understanding of the subsurface geology and structural style of the area. Both the seismic and borehole data are available in the public domain of the Directorate General Petroleum Concessions (DGPC) Pakistan for academic purposes. As a commitment, the area was selected to exclude the classified information on the Jurassic strata, major reservoir in the study area and the subject under discussion. A total of eleven 2D post-stack (migrated) seismic lines, containing 24-fold seismic data were used. These lines have a total length of 152.6 km and cover an area of 350 Km². Oil and Gas Development Company Limited (OGDCL) acquired and processed the data in 1981. Vibrosise survey was conducted for data acquisition. Well tops, density (RHOB), sonic (DT) and acoustic impedance (AI) logs of Bal-OXY-01, Bal-07, Bal-08 and Bal-A-01 were included, and summary sheets of these wells were also incorporated. Detail information about the seismic data and borehole logs is provided in Iqbal et al., (2015a).

1.3.2. Fieldwork and sampling

The upper part of the Upper Triassic Kingriali Formation and the entire Lower Jurassic Datta Formation were studied in the western Salt Range, Surghar Range and Khisor Range (Fig. 1.2). Detailed sections were measured at Kasanwala, Kaowaali, Nammal Gorge, Zaluch Nala – 1, Zaluch Nala – 2 (western Salt Range), Chichali Nala, Gulla Khel, Baroch Nala (Surghar Range) and Paniala anticline (Khisor Range). The sections are stretched over a length of around 300 km and covering an area of around 10,000 km². Details of the fieldwork are given in the manuscript (Iqbal et al., 2018 (under review)). In total 400 samples were collected covering all the lithological/facies variations of the upper part of the Kingriali Formation and the entire Datta Formation in different parts of the area. The samples were properly labeled and archived (Selley, 2000; Boggs, 2009).

1.3.3. Petrographic studies

With the exception of loose shale/mudstone thin sections were made for all the siltstone, sandstone, conglomerates, carbonates and evaporites samples. In total 300 thin sections were prepared at the Department of Geodynamics and Sedimentology, University of Vienna thin section petrography lab. These thin sections were studied systematically using Leica DM2700 P microscope with Leica MC170 HD camera to find out the framework

composition, matrix and cement types, depositional setting and diagenetic history of the strata. Details of the petrographic studies are provided in the articles/manuscripts (Iqbal et al., 2015b; 2018 (under review)).

Point counting was performed (Appendix 1) for all the siltstones, sandstones and conglomerates (Dickinson, 1970; Graham et al., 1976; Ingersoll et al., 1984; Pettijohn et al., 1987). The data was then used in Quartz – Feldspars – Rock Fragments (QFR) plot (Dickinson and Suczek, 1979; Suttner and Dutta, 1986).

1.3.4. Bulk mineralogy/XRD analysis

Based on fieldwork observations and petrographic analysis, 100 samples were selected for the bulk mineralogy using XRD from different parts of the sections covering all the lithological/facies variation in the strata. Powder XRD was used for the determination of the bulk mineralogical composition (Appendix 2a) of the selected samples (Iqbal et al., 2018 (under review)). XRD analysis was conducted using PANalytical X'Pert Pro diffractometer (CuK α - radiation (40 kV, 40 mA), step size 0.0167, 5 s per step) at the Department of Geodynamics and Sedimentology, University of Vienna.

1.3.5. Clay mineralogy/XRD separation

30 samples were investigated for clay mineralogy. The samples were treated with Hydrogenperoxyde (H₂O₂) to remove the organic matter/carbon (Van Langeveld et al., 1978; Mikutta et al., 2005). After complete dissolution of the organic content the samples were dispersed with a 400 W ultrasonic probe. The less than 2 μ m fraction was separated using Atterberg cylinders. Oriented clay samples were prepared by pipetting 1 ml of suspension (10 mg/ml) onto a glass slide. XRD mounts were analysed, saturated with Mg and K ions and after saturated with glycerol (Mg-samples) and ethylene-glycol (K-samples) to separate smectite and vermiculite. Finally, the samples were heated to 550 °C (Moore and Reynolds, 1997). The XRD patterns were interpreted (Appendix 2b) according to Moore and Reynolds (1997); Brindley and Brown (1980) and quantified using the correction factors of Schultz (1964).

1.3.6. Bulk geochemistry

Various geochemical proxies have reliable palaeoclimatic and palaeoenvironmental significance and are widely used for such interpretations (e.g. Climap, 1976; Nesbitt and Young, 1982; Roser and Korsch, 1988; Wefer et al., 1999; Hofmann et al., 2001; Zachos et al., 2001; Fischer and Wefer, 2012). In the present case 150 samples were powdered and

homogenised for bulk rock geochemistry (Appendix 3). The bulk rock geochemistry analyses were conducted in Acme Analytical Laboratories Ltd. Canada (now Bureau Veritas) using Inductively Coupled Plasma Emission Spectroscopy/Mass Spectrometry (ICP- ES/MS). The analyses include oxides, major, minor, trace and rare earth elements.

The bulk geochemical composition of each sample was measured using a combination of fusion and Aqua Regia digestion on sample splits. A split of 0.2 mg powdered and homogenised sample was fused in Lithium metaborate/tetraborate followed by nitric acid digestion to extract Si, Nb, and Rb and measured by Inductively Coupled Plasma Optical Emission Spectrometry (ICP-OES; Spectro Ciros Vision). In addition 0.5 g sample was digested in 95 °C Aqua Regia and analysed for Al, Fe, Mg, Ca, Na, K, Ti, P, Mn, Cr, Ba, Co, Sr, Th, U, V, W, and Zr by ICP-MS (Perkin Elmer ELAN 9000, Sciex). Both splits were calibrated against reference materials SO18, DS 10, and OREAS 45 EA. Total C (TOT/C) and total S (TOT/S) were detected by LECO analysis. 13 samples were analyzed in duplicates. The average analytical error in the analyses is 0.18%.

1.3.7. Heavy minerals analysis

30 sediment samples (Appendix 4) from the siltstones and sandstones were selected from the strata after detail petrography for heavy mineral analysis. The samples were crushed to separate the constituent particles. A solution of 3 : 2 distilled water and 80% acetic acid was prepared at the Department of Geodynamics and Sedimentology, University of Vienna. The samples were decalcified in this solution. After the complete dissolution of the carbonate content the samples were sieved and 0.4 mm to 63 µm fraction was used for the heavy minerals separation. This above particle diameter range was selected firstly to avoid the separation funnel throat from blockage by coarse particles, secondly as it is the grains diameter range used in similar studies (e.g. Stern and Wagreich, 2013). Heavy minerals by gravity settling in Low-toxicity inorganic solutions (LST), based on tungsten compounds lithium heteropolytungstates (LST) with a specific gravity of 2.85 g/cm³ (Mange and Maurer, 1992; Koroznikova et al., 2008). The grain concentrates containing the whole heavy mineral fraction were embedded in epoxy resin (Canada balsam mount).

1.4 Previous work

Numerous studies dealing with different aspects of geology have been undertaken in Salt Range. Publications of the Geological Survey of Pakistan (GSP) have summarised the earlier contributions of then British India geologists. Most of these earlier contributions dealt with the stratigraphic, paleontological or structural aspects or with the economic geology of the region. Information is available in the publications of Wadia (1966), Fatmi (1974) and Shah (1977; 2009). From amongst the later workers, the contribution of Gee is significant (e.g. Gee, 1980; Gee and Gee, 1989) and a geological map (6 sheets) on 1: 50,000 scale was published by him in collaboration with GSP.

Other contributions are summarised in Kadri (1995), Khan and Bhutta (1983, 1985 and 1987), Chaudhry et al., (1994a, b) and Kazmi and Jan (1997). The Datta Formation has recently yielded hydrocarbons and has been investigated in parts of the Salt Range and adjacent Surghar Range by Ashraf et al. (1976), Rafique et al. (1976) Mujtaba and Memon (1990), Fatmi et al. (1990), and Obaid et al. (2005) thereby providing information about the lithofacies, sand body geometry, depositional and diagenetic environments. Detailed information on the Kingriali and Datta formations is provided in the stratigraphy of the area (Chapter 2) and also at appropriate places in the published articles and submitted manuscripts (Chapter 3 – 6).

CHAPTER 2

STRATIGRAPHY

This chapter deals with the lithostratigraphy of the Salt and Trans–Indus ranges. Generalised information on all the stratigraphic units is provided (Fig. 2.1, Table 2.1). The Kingriali and Datta formations are explained in detail (Fig. 2.2).

2.1 Generalised stratigraphy

Excellent exposures of sedimentary rocks ranging in age from Precambrian to Recent occur in the Salt and Trans–Indus ranges (e.g. Kadri, 1995; Shah, 2009; Iqbal et al., 2014a). Precambrian and Paleozoic strata are widely exposed in the eastern and central Salt Range where Mesozoic rocks are missing (most probably due to Early Jurassic and Pre-Paleocene uplift of these parts; Iqbal et al., 2015a). In the western Salt Range and Trans–Indus ranges the Mesozoic and Cenozoic are well developed to go with the older strata (Fig. 2.1; Shah, 2009). A generalised stratigraphic record, summarising the different lithostratigraphic units in the area, is given in Table 2.1. All these names have been approved as formal names or provisionally accepted by the Stratigraphic Committee of Pakistan (SCP) as stated in the publications of Geological Survey of Pakistan (e.g. Shah, 1977; 2009).

The Precambrian – Eocambrian Salt Range Formation is the oldest rock unit in the Salt, Trans–Indus ranges, Potwar Plateau (Asrarullah, 1967; Gee and Gee, 1989; Shah, 2009). It is mostly exposed along the southern fault scarp of the Salt Range Thrust (SRT). Gypsiferous marl and salt are the dominant lithologies in this more than 800m thick formation at the type locality of Khewra. Besides, dolomite and clays are also important constituents. In addition, a layer of weathered igneous rock (Khewraite) occurs in the upper part. Deposition occurred in an enclosed basin under arid conditions (Shah, 1980), in intertidal to supratidal environments (Kemal, 1992).

The Paleozoic rocks in the area belong to the Jhelum (Cambrian), Nilawahan and Zaluch (both Permian) groups. The Jhelum Group, named after Jhelum area, includes the Cambrian Khewra Sandstone, the Kussak Formation, the Jutana Formation and the Baghanwala Formation. In the Khisor Range the Khisor Formation replaces the stratigraphic equivalent Baghanwala Formation. The type locality is Khewra Gorge in eastern Salt Range. The Khewra Sandstone is composed of shale and siltstone in the lower part and medium to coarse

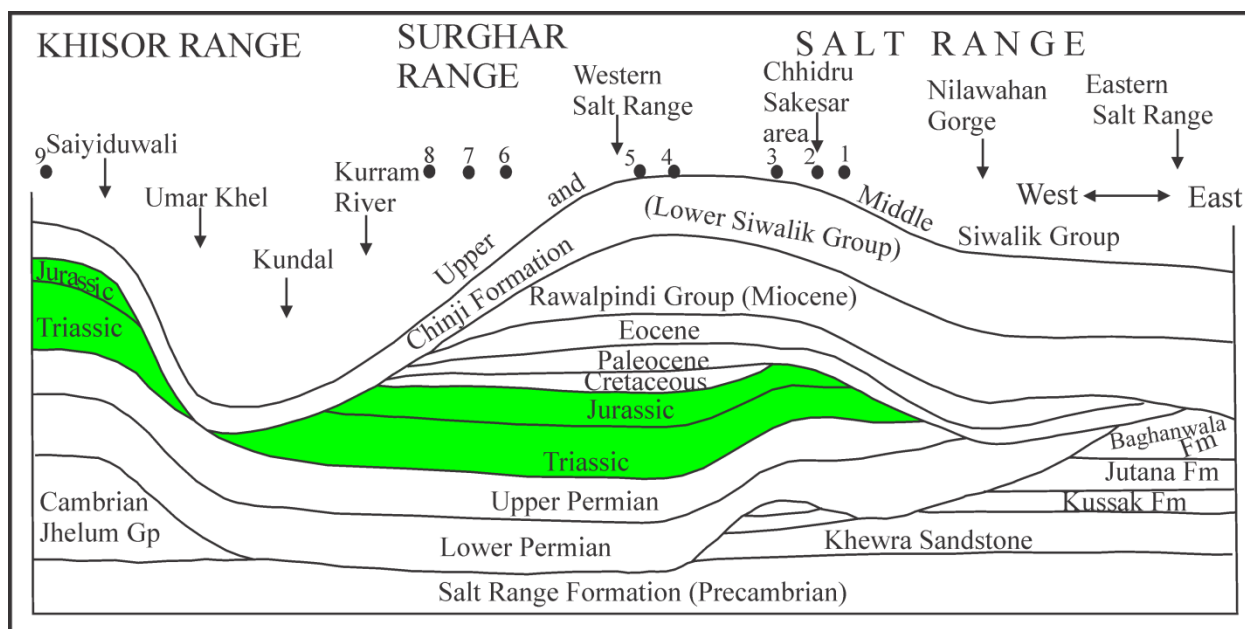


Figure 2.1 Generalised cross section of the Salt and Trans-Indus ranges showing the distribution of lithostratigraphic units in different parts of the area. The studied sections are numbered 1 – 9 (see Fig. 1.2 and Table 1.1 for detail; Kadri, 1995).

grained sandstones occur as dominant lithological unit of the formation and was deposited in fluvial–continental setting (Jehangiri et al., 2015). The formation has a disconfirmable upper contact with the Kussak Formation. The Kussak Formation consists of green, gray, brown shale, siltstone with minor, fine to medium grained sandstone deposited in restricted, arid lagoonal conditions (Shah, 2009). The formation has a conformable upper contact with the Jutana Formation which is composed of light gray, yellowish gray dolomite, shale, sandy dolomite deposited in shallow marine, tidal setting (Shah, 2009). The overlying Baghanwala and Khisor formations consist of strata typical of arid – restricted lagoonal environments (Table 2.1).

Ordovician to Lower Carboniferous strata are not developed in any part of the Salt and Trans-Indus ranges, nor reported from the subsurface in any part of the Indus Basin (e.g. Shah, 2009). The Cambrian strata are overlain by the Upper Carboniferous – Lower Permian Nilawahan Group, named after the Nilawahan Gorge in the central Salt Range (Fig. 2.1), composed mainly of continental siliciclastics. The formations constituting the Nilawahan Group in ascending order are the Tobra Formation, the Dandot Formation, the Warchha Sandstone and the Sardhai Formation. The Tobra Formation includes conglomerates, sandstone, siltstones and shale deposited in glacio-fluvial environments (Jan et al., 2017). The conformably overlying Dandot Formation consists of shale/mudstones, siltstones and

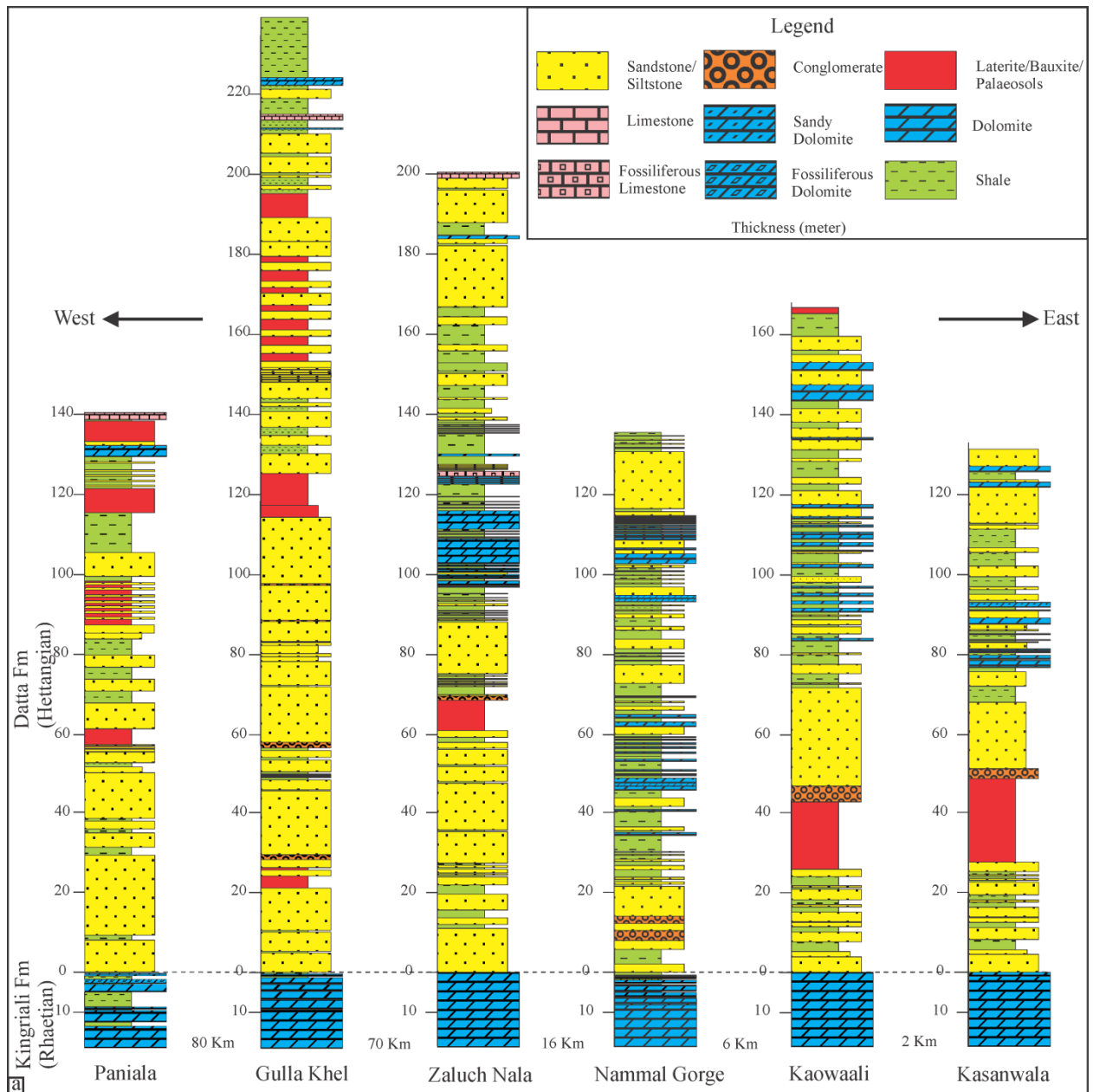


Figure 2.2. Representative stratigraphic sections of the upper part of Kingriali Formation and Datta Formation from Salt and Trans-Indus ranges (distance between sections in km).

sandstones and was deposited in tidal flat, estuarine and shelf setting (Jan et al., 2016). The Warchha Sandstone confirmably overlies the Dandot Formation and is composed of medium-coarse, arkosic sandstone, pebbly sandstones and conglomerates and interbedded shale deposited in arid fluvial – floodplain settings (Ghazi and Mountney, 2009; 2011). The Sardhai Formation consists of varicoloured clays deposited in shallow marine environments (Jan et al., 2009).

The Upper Permian Zaluch Group, named after the Zaluch Nala in western Salt Range, confirmably overlies the Nilawahan Group and includes the Amb Formation, the Wagal Limestone and the Chhidru Formation in ascending order. The group represents thick carbonates succession deposited as platform carbonates on the north-western margin of the Indian Plate (Mertmann 2003; Jan and Stephenson, 2011). The upper contact of the Chhidru Formation with the overlying Triassic Mianwali Formation correlates to the Permo-Triassic Boundary (Iqbal et al., 2014a).

More than 1000 m thick Mesozoic strata occur in the western Salt Range and Trans-Indus ranges (Fig. 2.1). The Triassic strata are named Musa Khel Group after the Musa Khel village in western Salt Range. It consists of the Mianwali Formation, the Tredian Formation and the Kingriali Formation in ascending order. The mixed fine grained siliciclastics and carbonates of the Mianwali Formation and the thick continental sandstones of the Tredian Formation represent a prograding deltaic deposition (Iqbal et al., 2014a). The overlying Kingriali Formation (especially the upper part) is discussed in detail in the present study (Iqbal et al., 2018 (under review)).

The Jurassic Baroch Group, named after the Baroch Nala in Surghar Range, consists of the Datta, Shinawari and Samana Suk formations. The Lower Jurassic Datta Formation is the main subject of the present study (Iqbal et al., 2015b) and is described in a separate section. The Shinawari Formation (Toarcian) is composed of limestone, shale and sandstone with siltstone at the top. The Shinawari Formation was deposited in shallow marine, lagoonal and intertidal settings (Mertmann and Ahmad, 1994). The overlying Samana Suk Formation (Callovian) consists dominantly of limestone with interbeds of shale in the lower part deposited in shallow, subtidal, lagoonal environments (Mertmann and Ahmad, 1994).

The Surghar Group, named after the Surghar Range, consists of the Chichali, Lumshiwal and Kawagarh formations. The Chichali Formation consists of glauconitic shales and sandstones of Oxfordian to Valanginian age deposited in deep marine environments (Hallam and Maynard, 1987; Shah, 2009). The Lumshiwal Formation (Hauterivian to Aptian – early Albian) consists dominantly of sandstones with thin glauconitic shale at the base. The formation was deposited in swampy, flood plain and deltaic conditions (Fatmi, 1974; Hallam and Maynard, 1987). The Kawagarh Formation consists of micritic limestone, marl and shale of Coniacian to Campanian and was deposited in open marine settings (Latif, 1970; Shah, 2009).

The younger rock units, their names, dominant lithologies and inferred environments are shown in Table 2.1. The description to follow deals with the Kingriali and Datta formations investigated in the present study.

2.2. Kingriali Formation

Gee (1945) used the name “Kingriali Dolomite” for a succession of dolomite, dolomitic shale, and sandstone in the upper part that was formalised as Kingriali Formation after the Kingriali Peak in the Khisor Range (Lat. 32°27'16" N, Long. 71°09'59" E).

The lower part of the formation consists of light gray, yellowish gray, brownish gray, thick bedded, massive dolomite (Fig. 2.3a). The dolomite displays oolitic texture. The exposed surfaces are jointed and have fragmental surficial appearance. Cross bedding and laminites are the common sedimentary structures (Fig. 2.3b). The middle and upper part of the formation consists of dolomite and dolomitic limestone (Shah, 2009). Thin interbeds of greenish gray, brownish gray and gray shales, dolomitic shales and marl appear in the upper part. The shales gets thicker and more frequent upsection and form medium to thick bedded dark brown and black shale in the upper most part of the formation (Fig. 2.3c). Around 1 m below the formation top, an intraclastic dolomite bed/facies consisting of aligned dolomite clasts occurs at Kasanwala, Kaowaali, Nammal Gorge and Zaluch Nala section (Fig. 2.3d). This bed shows bivalves and gastropod rich horizons (Fig. 2.3e). The same bed at the Paniala anticline section is rusty brown and fossiliferous (Fig. 2.3f). The top of the formation is marked by green shale in the Nammal Gorge, Gulla Khel, while this shale bed is brownish black and carbonaceous at the Paniala anticline section. This shale has a 6 – 20 cm thick gypsum bed at Nammal Gorge and Paniala anticline (Fig. 2.3g) while gypsiferous shale occurs at the same level in Gulla Khel section that is rusty brown in the lower part (Fig. 2.3h).

The formation is widely distributed in the western Salt Range, Trans–Indus ranges, Kohat, Kala Chitta, and Hazara. Excellent exposures occur at Nammal Gorge, Zaluch Nala in the western Salt Range, Landa Nala, Gulla Khel in the Surghar Range and Tapan Wahan, Paniala anticline and Gori Tang Nala in the Khisor Range. In the western Salt Range, the Kingriali Formation is 80 m thick at Nammal Gorge and 85 m in Zaluch Nala. In the Surghar Range it is 92 m in the Trakai Nala, 108 m in the Narmia Nala and 104 m in the Landa Nala section. In

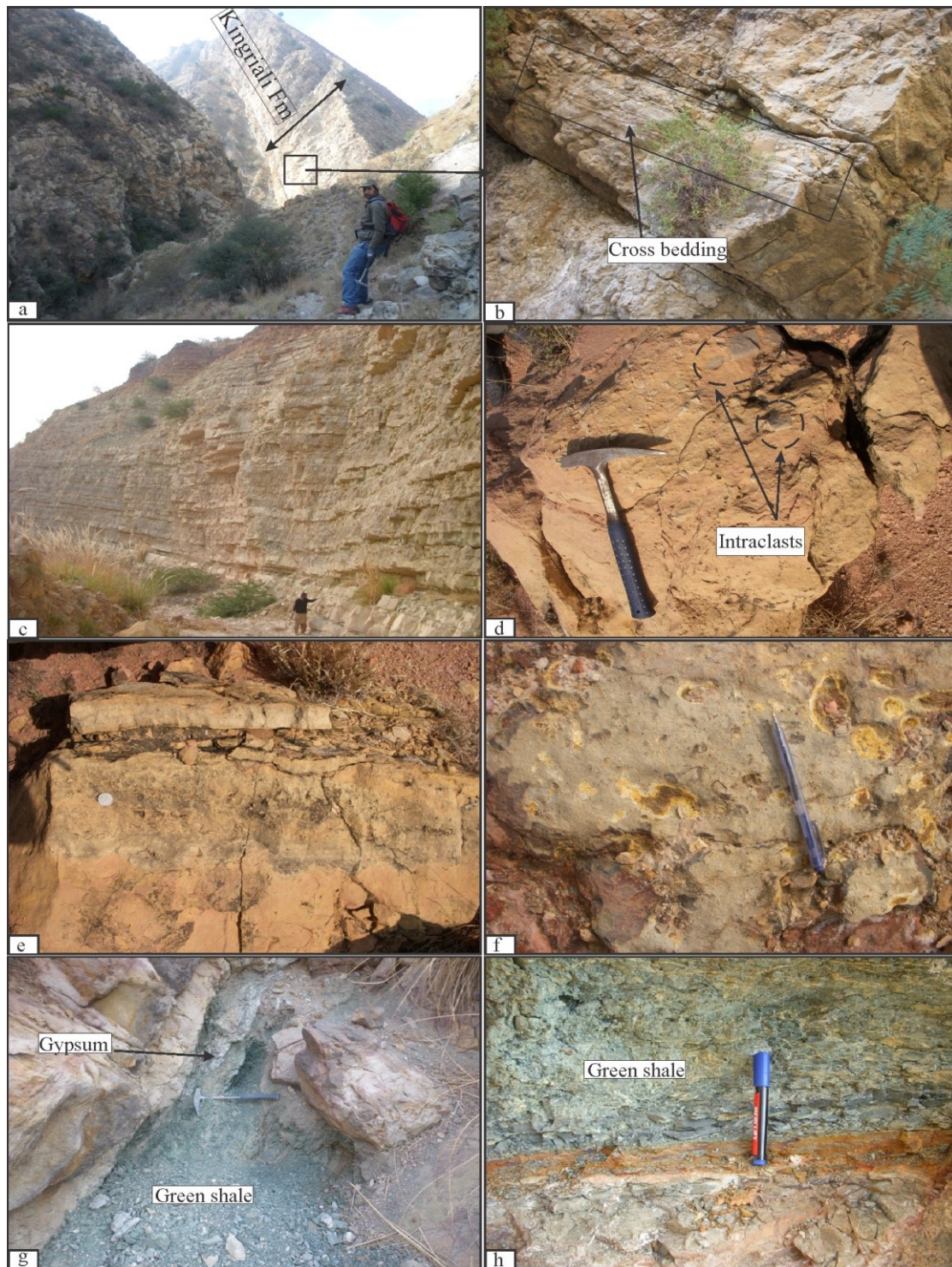


Figure 2.3 (a-h). a) Lower and middle part of the Kingriali Formation in Nammal Gorge section, b) Cross bedding in the dolomite in the Nammal Gorge, c) Middle and upper part of the Kingriali Formation in Paniala anticline section, d) Intraclast bearing dolomite bed in the upper part of the Kingriali Formation at Kasanwala, e) Fossiliferous dolomite in the upper part of the Kingriali Formation at Kasanwala, f) Rusty brown, fossiliferous dolomite bed in the upper part of the Kingriali Formation in the Paniala anticline area, g) Green shale with dolomite at the top of the Kingriali Formation in the Nammal Gorge, h) Topmost shale bed at the contact between Kingriali and Datta formations at Gulla Khel section.

the Kala Chitta Range it is 91 m thick.

In the Salt and Trans–Indus ranges the lower contact is conformable with the Tredian Formation and the upper contact is with the Datta Formation (discussed in detail in the present study). In the Kala Chitta area is the formation conformably overlies the Chak Jabbi Limestone (stratigraphic equivalent of the Tredian Formation).

Overall, the fossils are rare and poorly preserved. However, brachiopods, bivalves and crinoidal remains have been reported (Shah, 2009) and confirmed in the Present study. Furthermore, palynology results from the present study confirm Rhaetian age of the upper part of the formation.

The formation can be correlated with the upper part of the Wulgai Formation of the lower Indus Basin. The lower part of the formation displays strong similarities to the Hauptdolomit while the upper part of the Formation is more similar to the Kössen Formation of the Northern Calcareous Alps (Austria).

2.3 Datta Formation

The formation was described as variegated beds by Wynne (1878, 1879); Kioto Limestone by Middlemiss (1896) and Cotter (1933). Cox et al., (1930) introduced the term “samana beds” in the Kohat area. Later on, it was named as variegated stage (Gee, 1945) and Maira formation (Davies and Gardezi, 1965). Danilchik (1961) and Danilchik and Shah, (1967) introduced the name Datta Formation. The name was formalised by stratigraphic committee of Pakistan after the locality of Datta Nala (lat. 33°00′N: long. 71°19′E) in the Surghar Range.

In the western Salt Range and Trans–Indus ranges the formation consists of conglomerates, sandstones, siltstones, shale, dolomite and laterite horizons (Fig. 2.4a). Detail discussion on the lithofacies variation is given in the following lines:

Salt Range

Kasanwala and Kaowaali sections

In the Kasanwala and Kaowaali sections the lower part of the formation consists of fine to medium grained white, light gray, pink, red to brown sandstone and interbedded red, brown, yellowish brown shale (Fig. 2.4a). This is overlain by thick, dark red, purple laterite/bauxite interval (14 – 21 m; Fig. 2.4b). Quartz rich oligomictic conglomerates and pebbly sandstones of white, light and dark gray colours overlie the interval in the two

sections (Fig. 2.4c). These conglomerates and pebbly sandstones are pink and red at places and are overlain by red, brown fine to medium grained sandstone. Trough cross bedding, normal graded bedding and clasts alignment are the common sedimentary structures. Plant remains and coal patches are common (Fig. 2.4d). Ripple marks, mudcracks, Raindrop imprints (Fig. 2.4e), and bioturbation (Fig. 2.4f) are the common sedimentary structures. Light gray, brown and red shales overlie the succession (Fig. 2.4g).

This succession is overlain by interbedded sandstones, siltstones, shales and carbonates. The sandstones are fine to medium grained and compositionally quartz arenites. At places calcareous sandstones occur. Cross bedding, ripple marks, bioturbation and palaeosoil horizons are common. The shales are dark gray, brown and black in colour (Fig. 2.4h). Carbonaceous horizons occur and yellow colour due to sulphur can be observed with these horizons. The carbonates are yellowish orange and light brown in colour and include both pure and arenaceous dolomites (Fig. 2.4g). Planar and wavy bedding planes occur and ripple marks are also observed.

This part is overlain by interbedded sandstones and shales similar to those mentioned earlier. As we go upsection the sandstones becomes coarser and the shales becomes rich in kaolinite. The kaolinite rich horizons are mined in the upper part of the formation at both Kaowaali and Kasanwala sections. The uppermost part of the formation in the two sections is composed of interbedded red, brown sandstone and shales.

Nammal Gorge

The lower most fine to medium grained sandstone and shales typically exposed at Kasanwala and Kaowaali section (to the east of Nammal Gorge) are not developed at the Nammal Gorge (Fig. 2.5). Similarly the overlying thick laterite/bauxite, exposed in the two eastern sections is not developed at the Nammal Gorge. This package is replaced by a 2 m thick red, white, light gray fine to medium grained quartz rich sandstone and overlying 3.2 m light gray, dark gray and brown shale. The rest of the succession is similar to that observed in the eastern sections. The only difference is the increase in thickness and frequency of carbonaceous shales. The uppermost part of the formation at the Nammal Gorge also has much thicker, coarse grained, trough cross bedded sandstone and has more plant remains than anywhere else in the formation. Another marked difference is the absence of kaolinite mining in Nammal Gorge.

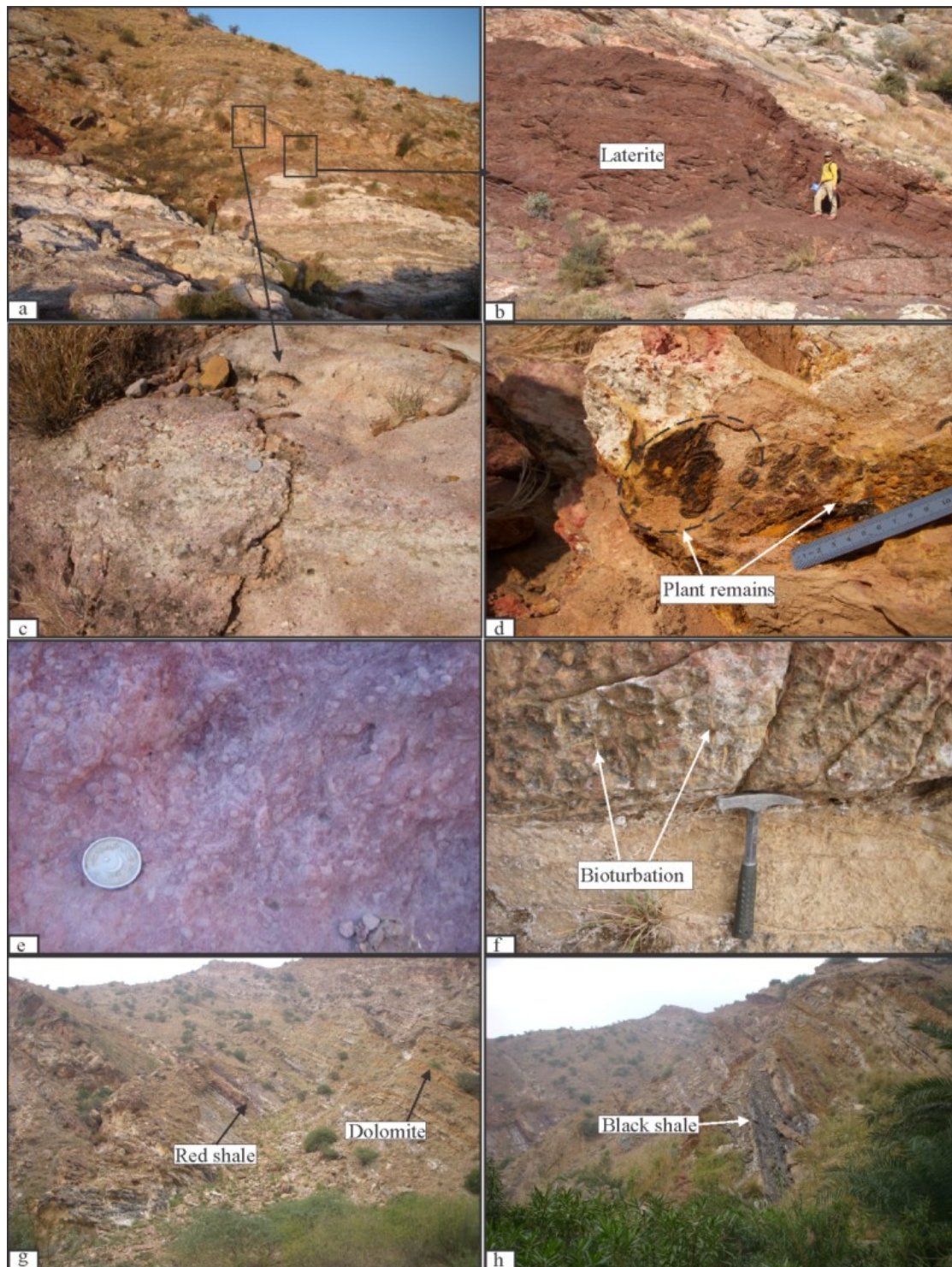


Figure 2.4 (a-h). a) View of the Datta Formation in the Kasanwala Section, b). Laterite in the Datta Formation at Kasanwala, c). Quartz rich conglomerate in the Datta Formation at Kasanwala, d). Plant remains in the Datta Formation at Kasanwala, e). Rain drop imprint in the fine to medium grained sandstone of the Datta Formation in Nammal Gorge, f). Bioturbation in the fine sandstones of the Datta Formation in Nammal Gorge, g). Red shales and dolomite in the Datta Formation in Nammal Gorge, h). Brown-black shale in the Datta Formation in Nammal Gorge.

Zaluch Nala

The Zaluch Nala sections (both 1 and 2) display similar lithofacies architecture (Fig. 2.6) to the eastern-most two sections (Kasanwala and Kaowaali sections). The lower white, red, brown sandstones and shales reappear at the Zaluch Nala after the disappearance in the east (at Nammal Gorge). Kaolinite is mined (Fig. 2.7a) in the area and the formation is thicker than the eastern sections.

Surghar Range

Chichali Nala

The lower contact of the Datta Formation is not exposed in the Chichali Nala. The rest of the succession exposed in Chichali is similar to that exposed above the thick laterite/bauxite horizons in Zaluch Nala (and two eastern most sections). In this respect the section more closely resembles the Nammal Gorge section. Also kaolinite is not mined here like Nammal Gorge. The carbonate content is less than the eastern sections and more shales occur in the upper part.

Gulla Khel

The Datta Formation in the Gulla Khel section and generally in the Surghar Range has more continental strata. Here it has more sandstone with interbedded red, continental shales. The conglomerates which are otherwise common in the Salt Range are replaced by coarse grained quartz rich sandstones. Similarly, coal patches occur and a 30 cm coal bed (Fig. 2.7b) occurs within the upper part of lower coarse sandstone unit. Palaeosoil horizons are more frequent in this section and large and extensive mudcracks are the typical features of the shales (Fig. 2.7c). In the upper middle part of the formation dark gray, black shales occur and include thin dolomite beds. This lithofacies is overlain by a thick interbedded red sandstone and shale unit of apparently continental origin, and has numerous laterites, palaeosoil horizons (Fig. 2.7d). Top of the formation here is marked by a 6 m thick, gray, brown, red shale that is overlain by the dominantly shale rich Shinawari Formation (Fig. 2.7e).

Baroch Nala

The lower part of the Datta Formation is not exposed in Baroch Nala and the exposed part is similar to the middle and upper parts of the Gulla Khel section. Here the carbonates are completely absent and the strata are completely continental in origin.

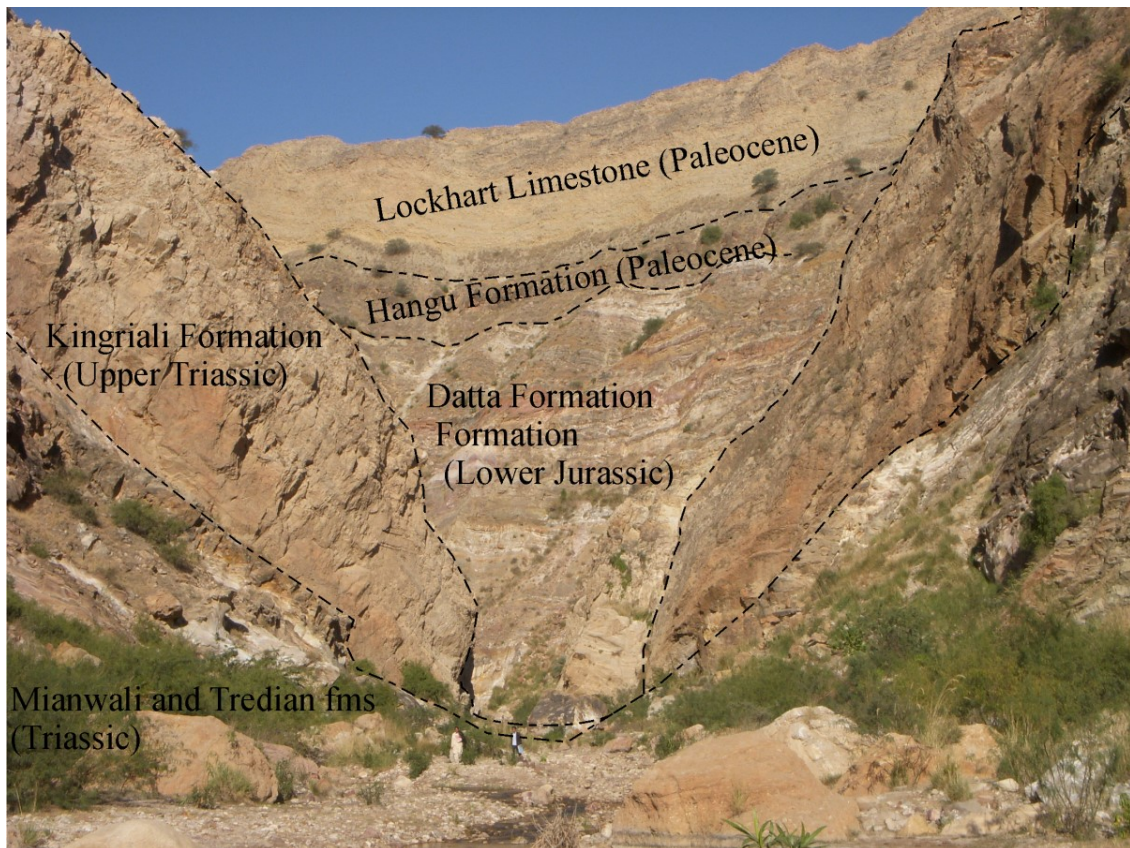


Figure 2.5. Field View of the Kingriali and Datta formations in Nammal Gorge. Note: Person standing in the mid of the gorge used as scale.



Figure 2.6. Field View of the Kingriali and Datta formations in Zaluch Nala. Car used as scale.

Khisor Range

Paniala anticline section

In the Khisor range the Datta Formation is more arenaceous. Medium to coarse grained quartz rich sandstones occur in the lower part. The sandstone is white, light gray in colour and is thick to very thick bedded. Cross bedding is the most common feature. Due to the high quartz (silica) content it is extensively mined for glass sand in the area (Fig. 2.7f). A typical feature of the lower sandstone is the occurrence of multiple 30 – 70 cm thick coal/carbonaceous horizons. Red, gray, brown shale horizons occur. In the upper part of the formation 6 m red, purple sandy laterite horizon occurs (Fig. 2.7g). Multiple kaolinite layers are mined in the section. The upper most part of the formation has a 6 m thick dolomite interbedded with gypsum (Fig. 2.7h).

Other areas

The Datta Formation has excellent outcrops in the Samana Range, along the Fort Lockhart road (Shah, 2009). The Datta Formation is composed of quartzose sandstone. In the Kala Chitta, Hazara area and Nizampur, the formation is represented by white, light gray, quartzose sandstone, red, purple, haematitic sandstone and fireclay horizons (Fatmi, 1974).

Distribution and thickness

The Datta Formation is widely developed in Kohat–Potwar Plateau, Nizampur and Hazara, Kala Chitta area. In the Salt Range excellent exposures occur in the western part at Kasanwala, Kaowaali, Nammal Gorge and Zaluch Nala. In the Trans–Indus ranges the best exposures are in the Surghar Range at Chichali Nala, Datta Nala, Gullah Khel, and Baroch Nala. The Formation has good exposures in the western Khisor Range and in the Sheikh Badin Hill. In the western Salt Range, the formation is 132 m thick at Kasanwala, 167 m at Kaowaali, 143 m in the Nammal Gorge, and 200 m in the Zaluch Nala. In the Surghar Range it is 150 m at Chihali Nala, 212 m at Datta Nala, 240 m at Gulla Khel, and 230 m in the Punnu Nala. In the Khisor Range it is 140 m in the Paniala anticline section. In the Sheikh Badin Hills the reported thickness is over 400 m but it apparently does not fit with the present measurement of 140 m at Paniala that is located just 2 – 3 km to the east of the Sheikh Badin. In Hazara the formation is 10 m in the Bagnotar section. In the Kala Chitta (Chak Dalla section) the thickness is 6 m while in the Nizampur it is 32 m and in the Samana Range it is 272 m near Shamshuddin village.

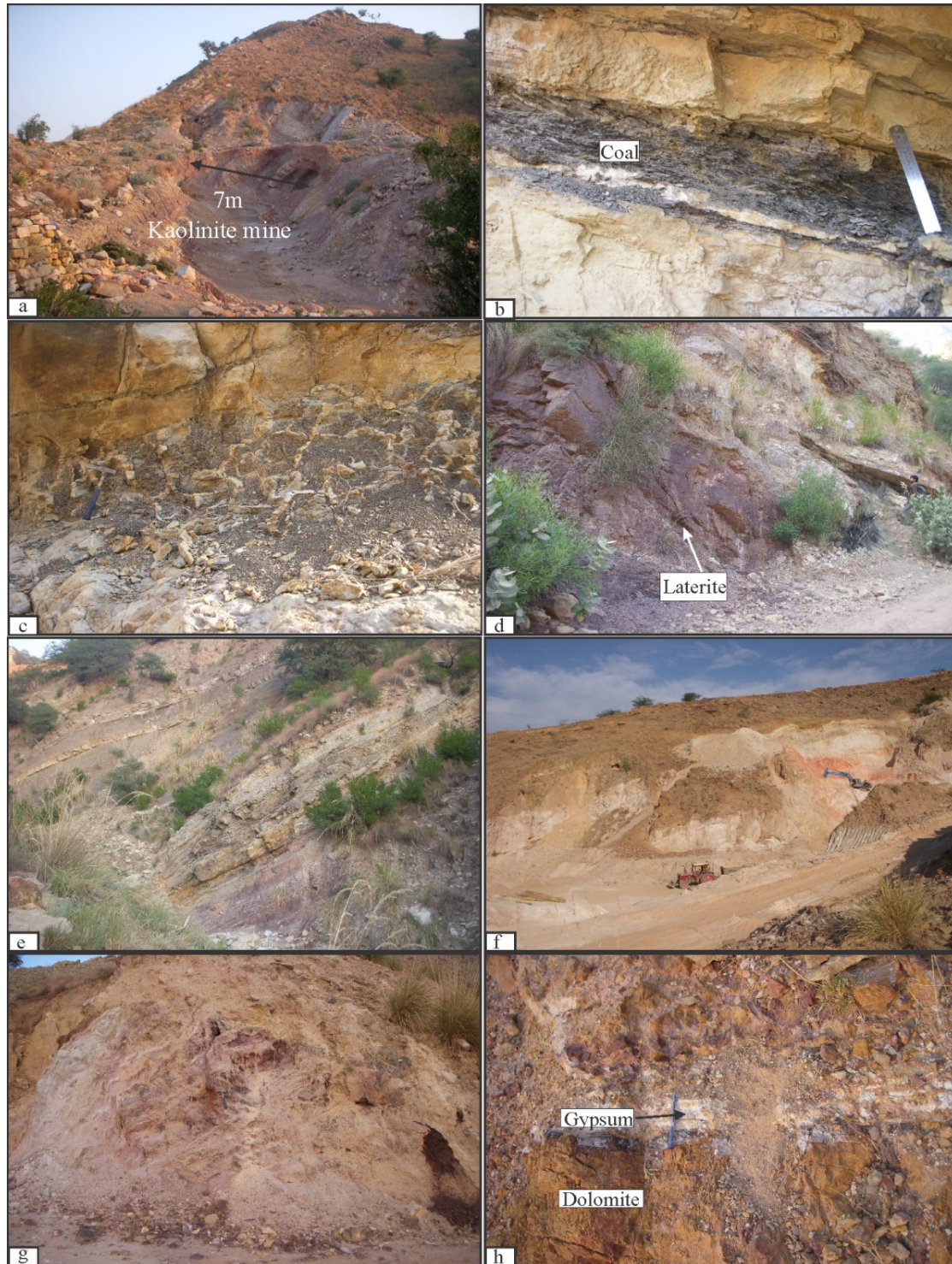


Figure 2.7(a-h). a). Kaolinite mine in the upper part of the Datta Formation in Zaluch Nala. b). Coal bed in the lower part of the Datta Formation at Gulla Khel, c). Mud cracks in the middle part of the Datta Formation at Gulla Khel, d). Laterite in the upper part of the Datta Formation at Gulla Khel, e). Upper contact of the Datta Formation with the Shinawari Formation at Gulla Khel. f). Silica sand mine in the lower part of the Datta Formation at Paniala anticline section, g). Laterite in the upper part of the Datta Formation at Paniala anticline section and, h). Interbedded dolomite and gypsum in the uppermost part of the Datta Formation at Paniala anticline section.

Contacts

In the Salt and Trans Indus ranges, Potwar, eastern Kohat, and Kala Chitta–Nizampur area the formation conformably overlies the Kingriali Formation. In Hazara, the formation unconformably overlies Precambrian Hazara Formation. The upper contact is unconformable with the Paleocene Hangu Formation at Kasanwala and Kaowaali. In the Nammal Gorge and Zaluch Nala section the formation is conformably overlain by the Samana Suk Formation. In the Trans-Indus ranges and Kohat area the formation has a gradational upper contact with the Shinawari Formation.

Age

The existing literature denies the occurrence of age-diagnostic fossils in the formation and a Pre-Toarcian age is assigned based on the stratigraphic position below the Shinawari Formation, which in its lower part has yielded lower Toarcian ammonites. However, the present study (Kuerschner et al., 2017) reports spore-dominated palynomorph assemblages. The first occurrence of *Cerebropollenites thiergartii* indicates the approximate position of the base of the Hettangian. Upsection pollen assemblage from the shales of the Datta Formation is dominated by *Classopollis turosus*. Thus the formation is Hettangian in age.

Table 2.1. Generalised stratigraphy of the Salt Range-Kohat-Potwar Plateau (after Shah, 2009). The Kingriali and Datta formations are highlighted in bold text.

AGE	Group/Formation		Lithology	Depositional environment.
Pleistocene		Kalabagh Conglomerate	Unsorted, mixed composition conglomerate.	Glacio-fluvial deposits.
Angular Unconformity				
Pliocene	Siwaliks	Soan Formation	Compact, massive conglomerate with subordinate, varicoloured sandstone, siltstone and/or clay.	Fluviatile.
		Dhok Pathan Formation	Gray, white and reddish brown sandstone and clay.	Fluviatile and Lacustrine.
		Nagri Formation	Greenish gray, light gray, medium to coarse grained and medium to thick-bedded sandstone.	Fluviatile-meandering and braided rivers.
Miocene	Rawalpindi	Chinji Formation	Red shales with greenish gray, micaceous sandstone	Fluviatile with ponded conditions.
		Kamlial Formation	Purple-gray and dark brick red sandstone with red shales and intraformational conglomerate.	Fluviatile and Lacustrine setting.
		Murree Formation	Dark gray, purple, cross-bedded sandstone with red shales.	Fluviatile and Lacustrine setting.
Major Unconformity				
Eocene	Charat	Chorgali Formation	Dolomitic limestone and shale in the lower part and dolomitic limestone in the upper part.	Low energy, tidal flat environment.
		Sakesar Limestone	Creamy white, nodular limestone with chert lenses and nodules.	Mostly lagoonal environment.
		Nammal Formation	Alum colour shale and marl in the lower part and light gray limestone in the upper part.	Low energy shallow marine conditions.
Paleocene	Makarwal	Patala Formation	Dark gary to black, carbonaceous shale and marl with limestone beds.	Lacustrine, organic rich lagoonal environment.
		Lockhart Limestone	Gray, nodular, fossiliferous limestone with gray marl and shale in the lower part.	Low energy marginal marine.
		Hangu Formation	Dark gray, black, carbonaceous shale and sandstone, with some limestone interbeds.	Terrestrial and shallow marine conditions.
Cretaceous-Paleogene boundary				
Cretaceous	Surghar Group	Kawagarh Formation	Light gray marl, calcareous shale and creamy white, micritic limestone.	Open marine
		Lumshiwal Formation	Greenish gray, yellowish gray, cross bedded sandstone	Marine swamps, deltaic.
		Chichali Formation	Dark green to black glauconitic shale and sandstone.	Marine (reducing).

Jurassic	Baroch Group	Samana Suk Formation	Gray to creamy white oolitic, dolomitic limestone.	Shallow, subtidal, lagoonal setting.
		Shinawari Formation	Light gray limestone with shale, siltstone and sandstone in the upper part.	Lagoonal, intertidal environment.
		Datta Formation	Dark gray, purple conglomerate in the lower part, light to dark gray quartzose, cross-bedded sandstone and shale, yellowish orange dolomite and dolomitic limestone, silica sand, fire clay and lateritic zones.	Deltaic, flood plain and alternating shallow marine (discussed in detail in the present study).
Triassic	Musa Khel	Kingriali Formation	Light gray massive dolomite and dolomitic limestone with shale in the upper part and gypsum/evaporate at the top.	Shallow marine (upper part discussed in detail in the present study).
		Tredian Formation	Dark, carbonaceous shale with sandstone in the lower part, thick bedded to massive, cross-bedded sandstone with some dolomite beds in the upper part.	Deltaic to fluvial settings.
		Mianwali Formation	Shale, cross bedded, rippled, light to dark gray sandstone, yellowish orange fossiliferous limestone and dolomite.	Shallow marine to prodelta settings.
Permo-Triassic Boundary				
Late Permian	Zaluch	Chhidru Formation	Light gray limestone with white calcareous sandstone, fossiliferous.	Marginal marine, estuarine or lagoonal environment.
		Wargal Limestone	Light to dark gray limestone, dolomitic at places and dark shale interbeds at the base, fossiliferous.	Mostly shallow marine environment.
		Amb Formation	Light to yellowish gray, Impure limestone and calcareous sandstone, fossiliferous	Shallow marine mostly with paralic conditions.
Early Permian	Nilawahan	Sardhai Formation	Bluish, greenish gray clay with subordinate sandstone	Lacustrine environment.
		Warchha Sandstone	Red, purple spackled, cross-bedded, arkosic sandstone, with thin shale interbeds at places.	Fluvial to localized channel environment.
		Dandot Formation	Light gray, olive green, yellowish sandstone with subordinate dark gray, greenish gray splintery shales	In the east marine and in the west tidal flat environment.
		Tobra Formation	Conglomerate (glacial), sandstone and some shale.	Glacio-fluival environment.
Major Unconformity				
Cambrian	Jhelum	Khisor Formation	White gypsum, finely crystalline dolomite and dolomitic shale	Lagoonal environment with evaporitic conditions.
		Baghanwala Formation	Red shale and clay alternating with flaggy sandstone having salt pseudomorphs.	Lagoonal environment with evaporitic conditions.
		Juttana Formation	Light green, hard, bedded to massive dolomite with dolomitic sandstone and subordinate shale.	Shallow marine conditions.

		Kussak Formation	Greenish gray, glauconitic, micaceous sandstone and siltstone.	Restricted to arid lagoonal environment.
		Khewra Formation	Red, flaggy shale in the lower part and thick bedded to massive, purple to brown and cross-bedded sandstone in the upper part.	Fluvial to lacustrine.
	Precambrian	Salt Range Formation	Red gypseous marl with thick salt seams in the lower part, dolomite and shale in the upper part and Khewra Trap, in upper part, in eastern Salt Range.	Intertidal-supratidal environment/ enclosed basin and arid condition.

PUBLICATIONS AND MANUSCRIPTS

The present PhD thesis consists mainly of two published research articles and a third manuscript currently under review. The articles are structured starting with structural geology followed by depositional sequence stratigraphy and facies architecture and palaeoclimatic reconstruction in a chronological order of the date of publication of the articles followed by the manuscript under review.

1. **Iqbal, S.**, Akhter, G., and Bibi, S. (2015a) Structural model of the Balkassar area, Potwar Plateau, Pakistan. *International Journal of Earth Sciences*, 104(8), 2253-2272. DOI: 10.1007/s00531-015-1180-4.

This article describes the structural geology of the Balkassar area, Potwar Plateau, a famous oil field in the area. The area provides an ideal location for the structural and stratigraphic comparison between the highly deformed North Potwar Deformation Zone (NPDZ) to the north and the relatively less deformed southern Potwar and the Salt Range to the south. The article is based on 2-D seismic reflection data and bore-hole logs. The data was initially including the nearby areas where the Datta Formation is a producing horizon, with the aim to model the lithofacies and hydrocarbon producing/potential zones within the Jurassic Datta Formation in the subsurface. However these zones were confidential and as a commitment data from Balkassar Area, Potwar Plateau was used to avoid the confidential Triassic–Jurassic strata both on seismic sections and well logs. Thus only the structural geology is the main focus of this article aiming a comparison between the highly deformed NPDZ and the much less deformed area to south fit for stratigraphic studies and facies analyses.

2. **Iqbal, S.**, Jan, I. U., Akhter, M. G., and Bibi, M. (2015b) Palaeoenvironmental and sequence stratigraphic analyses of the Jurassic Datta Formation, Salt Range, Pakistan. *Journal of Earth System Science*, 124(4), 747-766. doi:10.1007/s12040-015-0572-y.

This article describes the depositional palaeoenvironments and depositional sequence stratigraphy of the Lower Jurassic Datta Formation in the western Salt Range. The lithofacies and facies associations have been worked out in the Kasanwala and Kaowaali sections (eastern most sections) to understand the

response of the silicilastic system to fluctuating sea-level. This article provides the initial guideline to explore the Datta Formation in the western Salt Range and the Tans – Indus ranges further to the west.

3. **Iqbal, S.**, Wagreich, M., Jan, I. U., Kuerschner, W. M., and Gier, S. (2018) A southern hemisphere record of climate change during the Triassic–Jurassic transition: evidence for hothouse climate from the Salt Range, Pakistan. *Global and Planetary Change* (under review; manuscript number: GLOPLACHA_2018_92).

This manuscript uses multi-proxy data from the upper part of the Kingriali Formation and the Datta Formation to understand transition from the Late Triassic arid palaeoclimate to a hot and humid climate in the Early Jurassic. Sections from the western Salt Range, Surghar Range and Khisor Range have been included to provide a comparison for the Triassic–Jurassic transition. A regional and global correlation with classical Triassic–Jurassic boundary intervals (mostly northern hemisphere) has been made to elaborate the contemporaneous palaeoclimate in the southern hemisphere.

CHAPTER 3

STRUCTURAL GEOLOGY

In this chapter, a brief description of the tectonic setting and structural geology of the study area is provided. Tectonics and structural geology of the area are explained using 2D–Seismic reflection data and well logs in a published article (Iqbal et al., 2015a).

Structural model of the Balkassar area, Potwar Plateau, Pakistan.

Iqbal, S., Akhter, G., and Bibi, S. (2015a) International Journal of Earth Sciences, 104(8), 2253-2272. DOI: 10.1007/s00531-015-1180-4.

Author's comments

The article was written by Shahid Iqbal as corresponding author. The seismic and bore-hole log data was selected through mutual discussion of all the co-authors. Shahid Iqbal and Gulraiz Akhter reviewed the data. Sehrish Bibi loaded and did Quality Control (QC) of the data in Petrel software and prepared the time sections, time–depth tables and depth sections. Sehrish Bibi made the synthetic seismogram and Gulraiz Akhter and Shahid Iqbal refined the match between the seismic and the well logs. Shahid Iqbal did the final interpretation and structural modeling. Sehrish Bibi and Gulraiz Akhter provided the images of seismic sections and well logs. Shahid Iqbal refined the images and marked the images and produced the structural model in the desired format.

Structural model of the Balkassar area, Potwar Plateau, Pakistan

Shahid Iqbal^{1,2}  · Gulraiz Akhter¹ · Sehrish Bibi³

Received: 16 December 2014 / Accepted: 29 March 2015 / Published online: 18 April 2015
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Abstract Balkassar is an important hydrocarbon producing area of the Potwar Plateau, Pakistan. Two-dimensional seismic reflection data of the area revealed tectonically controlled, distinct episodes of (1) normal faulting in the basement followed by (2) reverse faulting in the cover sequence. Himalayan orogeny and associated diapirism of the Precambrian Salt Range Formation have produced many salt-cored anticlines in the Potwar Plateau, and one such salt-cored anticline is present in the Balkassar. This anticline has NE–SW-oriented axis, and both the SE and NW limbs are bounded by reverse faults. The basement normal faults indicate Jurassic rifting and splitting of Pangaea. We interpret reverse faults with dip angles of about 60°–75° in the cover sequence, having both hinterland and foreland vergence. Both NW- and SE-dipping faults are present in contrast to the only southward-directed thrusts of previous models. Duplexes and triangle zones, which are common in the northern part of the Potwar Plateau, are not developed in the Balkassar area due to comparatively less crustal shortening in the area. The present interpretation can help in understanding the complex structures in other parts of the Potwar Plateau for hydrocarbons exploration and also in deformed foreland basins worldwide that display similar characteristics but are considered to be dominated by low-angle thrust tectonics.

Keywords Balkassar · Potwar Plateau · Himalayan orogeny · Crustal shortening · Foreland basin

Introduction

The area of Balkassar, 105 km SW of Islamabad, lies in the central part of Potwar sub-basin that forms part of the Himalayan Foreland Fold-and-Thrust Belt (Tahirkheli 1979; Farah et al. 1984; Duroy et al. 1989; Kazmi and Jan 1997), and is located on the southern limb of the Soan syncline (Figs. 1 and 2). The Potwar sub-basin forms a part of the Kohat–Potwar Plateau and is comprised of low hills and dissected valleys. Average elevation of the region is about 500 m above mean sea level (Khan et al. 1986). The left lateral Jhelum Fault (Mukherjee submitted) marks the eastern boundary of the plateau; the western margins are delineated by the Kurram–Pezu Fault (Gee 1989). To the south, the plateau is bounded by the Salt Range Thrust, whereas the northern boundary is marked by the Main Boundary Thrust (Fig. 1; Rana and Asrarullah 1982; Jaswal et al. 1997; Kazmi and Jan 1997).

The Potwar Plateau is one of the oldest oil provinces of the world, where the first commercial discovery was made in 1914 at Khaur. Since then, many hydrocarbon fields have been discovered in different parts of the plateau (Khan et al. 1986). Most of these discoveries are associated with subsurface reverse faults, duplexes and associated anticlines (mostly salt-cored anticlines). About 150 exploratory wells have been drilled so far in the area; many of these were prematurely abandoned, as they could not reach their target depths due to structural complexities and associated technical drilling problems. Mostly, the surface geological features of the plateau do not represent the subsurface structures. This is because normally the exposed younger

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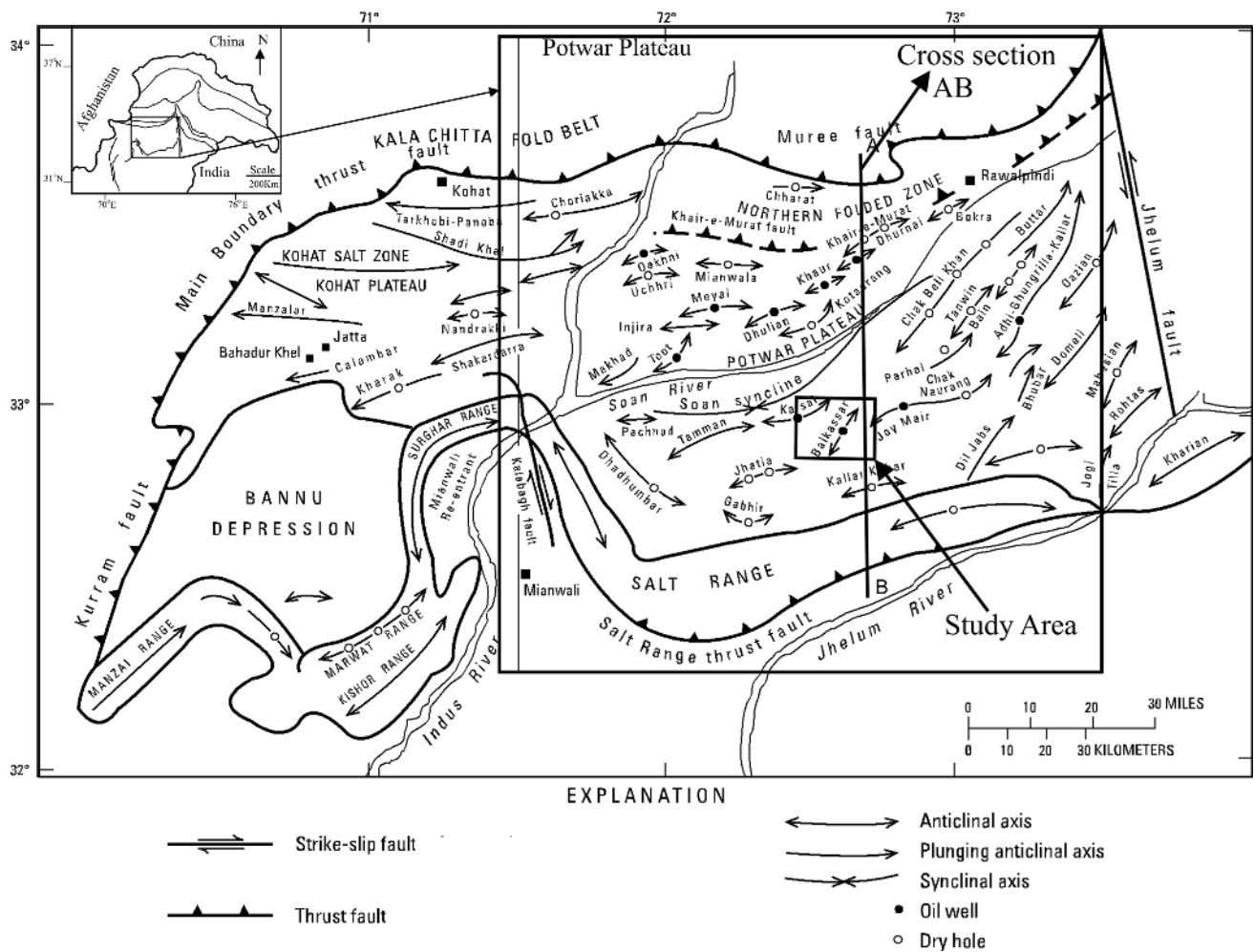


Fig. 1 Structural map of the Kohat–Potwar Plateau. The *inset* area is the Potwar Plateau and the *enclosed block* marks the study area of the base map. The cross section along the profile AB is given in Fig. 2a (modified after Khan et al. 1986; Gee 1989)

molasses (Siwaliks) display broad anticlines with gently dipping limbs, whereas in the subsurface (visible on seismic), these mostly overly the duplexes, triangular zones and salt-cored anticlines bounded by steep faults (e.g., Jadoon et al. 1999; Jadoon et al. in press).

Seismic data, its integration with wells and geological data, are a reliable tool for subsurface geological interpretation. The exploration industries use this integrated approach to subsurface interpretations. This helps in tracing the surface geological structures in the subsurface and establishing reliable correlations. An accurate representation of the subsurface geological horizons and structures is integral for hydrocarbon exploration. This provides an insight into the framework and structural styles of a basin.

In the present case, stratal architecture is a function of the complex relationships between compressional tectonics and basement slope. The uneven thickness of Precambrian evaporites also has its effects and has produced many salt-cored anticlines. The deposition of very thick molasse-type

sediments shed from the Himalayan orogeny has further complicated the situation by creating the overburden load (e.g., Kadri 1995; Shah 2009). Structurally, the Potwar Plateau is complicated, and surface features do often not reflect subsurface structures (Jadoon et al. in press). This is due to the presence of detachments at different levels (Jadoon et al. 1999). In such cases, it is necessary to integrate seismic data with geological information for an accurate delineation of subsurface structures (Moghal et al. 2007).

Balkassar area is important as it is host to proven hydrocarbon reserves (Kadri 1995). The area has 4000-m-thick Precambrian to Recent sediments. Among these, the Khewra Sandstone (Cambrian), Sakesar Limestone (Eocene) and Chorgali Formation (limestone; Eocene) are proven reservoirs (Khan et al. 1986). Geophysical data indicate the presence of reverse faults in the cover sequence (Precambrian–Recent), while the basement displays normal faults, which are also visible on seismic data (Lillie et al. 1987; Pennock et al. 1989; Moghal et al. 2007).

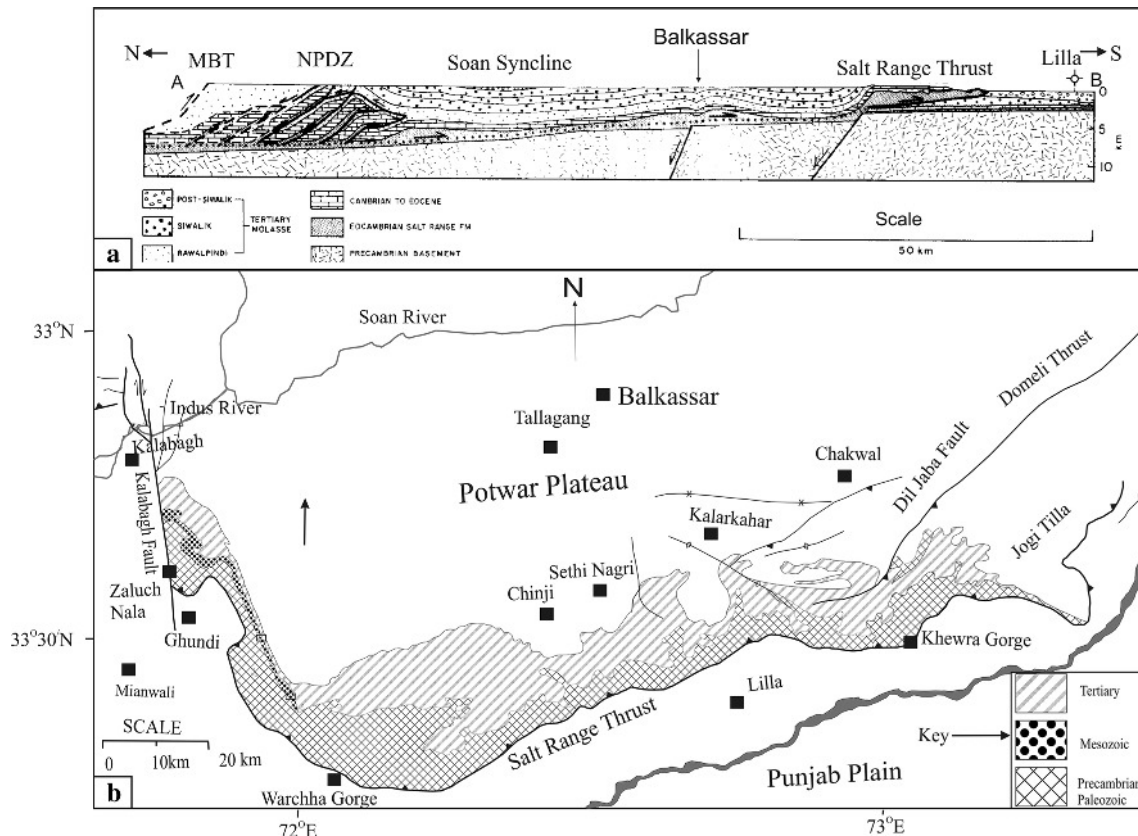


Fig. 2 **a** Structural cross section along profile AB (Fig. 1) showing different structural zones in the Potwar Plateau, MBT Main Boundary Thrust, NPDZ North Potwar Deformation Zone (after Baker 1987;

Baker et al. 1988; Pennock et al. 1989). **b** Geological map of the Salt Range showing the exposed Precambrian–Recent strata along the Salt Range Thrust (after Gee 1989; Iqbal et al. 2014a)

The present study applies an integrated approach using 2D seismic reflection and well logs data to interpret the subsurface geological features of the Balkassar area for hydrocarbons exploration. It further explains why the duplexes and triangle zones which are otherwise common in the northern part of the Potwar Plateau have not been formed in this area. The 2D seismic lines and wells are plotted on the base map (Fig. 3). This study uses interval, average and mean average velocities (from seismic data) and their correlation with synthetic seismogram (generated from well data) to mark the horizons on the time section and convert them into depth sections. The data are further used to generate structural model for the area. Similar approach, if applied to other parts of the Potwar Plateau, will help the exploration of regional subsurface structures.

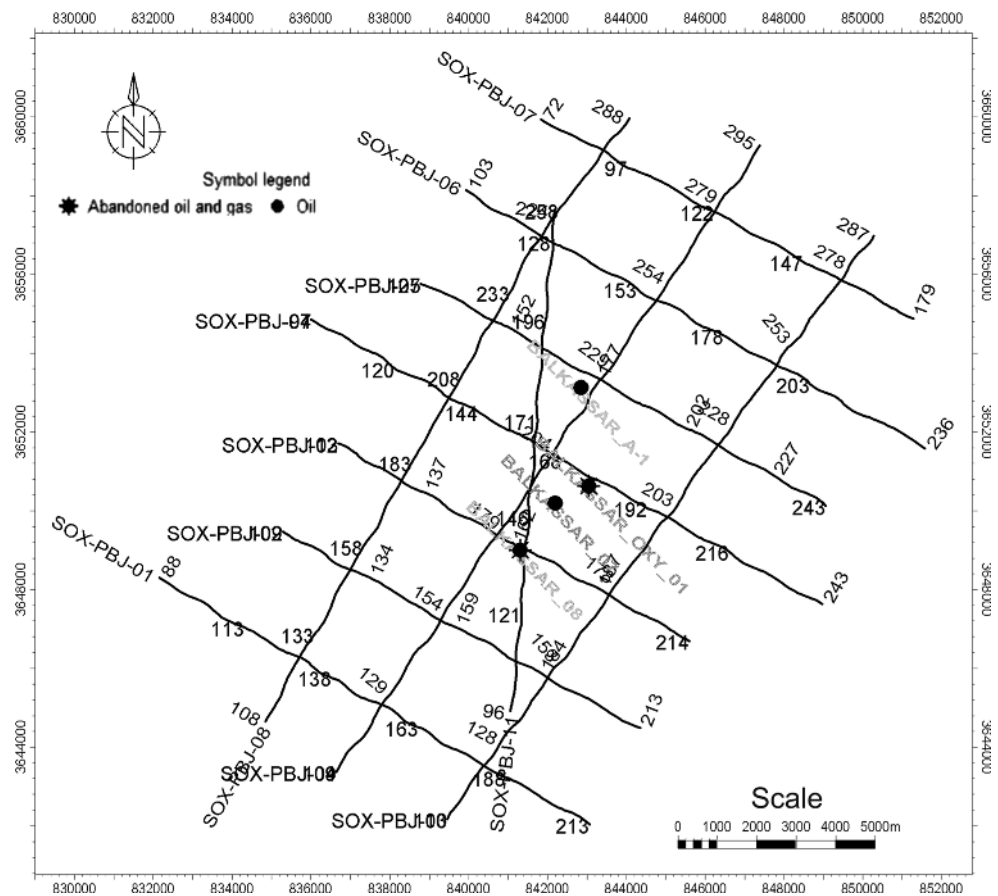
Geology of the area

The Potwar Plateau has 5000- to 1000-m-thick Precambrian to Recent sediments preserving the sedimentation, paleoenvironments, paleogeography and tectonic history of

the plateau in them (e.g., Tahirkheli 1979; Gee 1980, 1989; Kazmi and Jan 1997; Kazmi and Abbasi 2008; Shah 2009; Iqbal et al. 2014a, b). The area is known as field museum of dynamic geology due to its spectacular structures, strata exposures and fossils occurrence along the Salt Range Thrust, the southern boundary of the plateau (Fig. 2b and Tahirkheli 1979).

Stratigraphy of the plateau is well established along the Salt Range (Kazmi and Abbasi 2008; Shah 2009; Iqbal et al. 2014a, b; Fig. 2; Table 1). The Precambrian basement is composed of slates and phyllites (Davies and Crawford 1971). The oldest strata are the Precambrian Salt Range Formation that consists of evaporites dominated by rock salt, gypsum and marl with some dolomite and minor oil shale (Gee 1945, 1947). It is overlain by the Cambrian Jhelum Group that consists of Khewra Sandstone, Kusak Formation (sandstones and shale), Jutana Formation (dolomite) and Baghanwala Formation (sandstone, siltstone and shale). The group is unconformably overlain by the Lower Permian continental Nilawahan Group (Ghazi et al. 2012) and Upper Permian marine carbonates of the Zaluch Group (Shah 2009). The Mesozoic sediments are eroded/

Fig. 3 Base map showing orientation of the 2D seismic lines and well locations in the Balkassar area



not deposited in the eastern and central parts of the Salt Range due to Early Jurassic rifting (Kadri 1995; Iqbal et al. 2014b). A high (horst) is present near Sargodha located to the east of the Salt Range (Kazmi and Abbasi 2008; Shah 2009). Thick Mesozoic exposures occur in the western Salt Range (Fig. 2b). Shallow marine foraminiferal limestone and dark-gray shale dominate the Paleocene–Eocene succession (Hanif et al. 2014). The base is marked by a laterite bed (Hallam and Maynard 1987). During Middle to Late Eocene time, a regional emergence resulted in complete and permanent withdrawal of the sea. Oligocene sediments have not been found in the region. From the Miocene onward, only fluvial sediments were deposited (Khan et al. 1986).

The entire sedimentary succession is severely disturbed by the Himalayan orogeny between the northern margin of the Indian Plate and Eurasian Plate (Figs. 1 and 2) since about 55 Ma (Gansser 1979, 1980; Molnar and Tapponnier 1981; Valdiya 1984, 1997; Baker et al. 1988; Burbank and Reynolds 1988; Molnar 1990; Davis and Lillie 1994; Jadoon and Frisch 1997; Jadoon et al. 1997; Kazmi and Jan 1997; Mohadjer et al. 2010; Hasany and Saleem 2012; Jain 2014). Extensive work is conducted on this orogeny in India, Nepal and nearby areas (e.g., Le Fort 1975; Hodges

et al. 1992; Carosi et al. 1998; Upreti 1999; Hodges 2000; DeCelles et al. 2002; Lee et al. 2004; Yin 2006; Carosi et al. 2007; Wang et al. 2012; Robinson and Pearson 2013; Khanal and Robinson 2013; Mandal et al. 2014; Khanal et al. 2014). Further in Pakistan, the orogeny has remained the subject of many studies (e.g., Davis and Engelder 1985; Baig and Lawrence 1987; Butler et al. 1987; Baker et al. 1988; Jaumé and Lillie 1988; Grelaud et al. 2002).

The ENE trending Salt Range Thrust (Figs. 1 and 2), a surface expression of the leading edge of a regional décollement (Lillie et al. 1987), brings the entire sedimentary succession to the surface (Butler et al. 1987; Treloar et al. 1992). Structurally, the plateau is subdivided (Fig. 2a) into (1) the Northern Potwar Deformation Zone (NPDZ) and (2) the southern Potwar platform zone (Khan et al. 1986; Kadri 1995; Jadoon et al. 1999). Mainly, contractional folds and faults occur in the area; however, preexisting normal faults also occur in the crystalline basement (Pennock et al. 1989). Normally, duplexes and imbricate thrusts occur in the NPDZ (Jadoon et al. 1999; Ghazi et al. in press; Fig. 2a), and the southern part of the plateau is relatively less deformed (Jaswal et al. 1997; Fig. 2a). Cross-sectional balancing studies (e.g., Baker 1987; Leathers 1987; Abbasi and McEiroy 1991) show that thrust stacks of ramp

Table 1 Lithostratigraphic units of the Kohat–Potwar Plateau (after Fatmi 1973; Iqbal et al. 2014a)

Era	Period		Group	Formations	
CENOZOIC	Tertiary	Pliocene	Siwalik Group	Soan Formation	
				Dhok Pathan Formation	
				Nagri Formation	
		Miocene	Rawalpindi Group	Chinji Formation	
				Kamlial Formation	
				Murree Formation	
		Unconformity			
		Eocene	Cherat Group	Kohat Formation	
				Kuldana Formation	
				Chorgali Formation	
				Sakesar Limestone	
				Nammal Formation	
	Paleocene	Makarwal Group	Patala Formation		
			Lockhart Limestone		
			Hangu Formation		
Unconformity					
MESOZOIC	Cretaceous		Kawagarh Formation		
			Lumshiwal Formation		
			Chichali Formation		
	Jurassic	Surghar Group	Samana Suk Formation		
			Shinawri Formation		
			Datta Formation		
	Triassic	Musa Khel Group	Kingriali Formation		
			Tredian Formation		
			Mianwali Formation		
Permo-Triassic Boundary					
PALEOZOIC	Permian	Zaluch Group	Chhidru Formation		
			Wargal Limestone		
			Amb Formation		
		Nilawahan Group	Sardhai Formation		
			Warchha Sandstone		
			Dandot Formation		
	Unconformity				
	Cambrian	Jhelum Group	Baghanwala Formation		
			Juttana Formation		
			Kussak Formation		
Khewra Sandstone					
PROTE R OZOIC	Precambrian			Salt Range Formation	

anticlines have accommodated most of the shortening in the NPDZ. A triangle zone occurs in the foothill of Khair-i-Murat Ridge (Lillie et al. 1987; Baker 1987; Pennock et al. 1989; Jaswal et al. 1997; Jadoon et al. 1999; Moghal et al. 2007; Jadoon et al. in press). The central and western parts of the area have south-verging reverse faults, whereas, in the eastern part, NE–SW trending tight and occasionally overturned anticlines separated by broad synclines occur (Fig. 1). This difference may be related to lesser thickness of the Precambrian salt in the eastern area (830 m in the east and +2000 m in the west) and very low northward dip of the basement (1° – 1.5°) as compared to central and western Potwar (2° – 3°). Paleomagnetic studies have shown that originally, the structural trends developed perpendicular to the transport direction and subsequently acquired the present alignment because of the left lateral Jhelum strike-slip fault (McDougall and Khan 1990; Sercombe et al. 1998).

In the central Potwar, folds are bounded by thrusts and back thrusts, while at Dijaba, Bhubair, Domali and Rothas (Fig. 1), asymmetric anticlines are bound by a single fault according to Kazmi and Jan (1997). Subsurface mapping reveals that many surface folds are cored by both south- and north-vergent blind reverse faults (e.g., Qazian; Fig. 1), and some are fault propagation folds, e.g., Rothas and Mahesian (Fig. 1, Pennock et al. 1989; Jaswal et al. 1997; Moghal et al. 2007).

The Kalabagh Fault was formed by transpressive right-lateral strike-slip movement along the western Salt Range–Potwar Plateau (McDougall and Khan 1990). Lateral ramping from a décollement thrust along the Precambrian evaporites produced NNW to NW trending folds and NE- to N-dipping thrust faults in a topographically emergent zone up to 10 km wide (McDougall and Khan 1990). Piercing points indicate 12–14 km of middle to late Quaternary

Table 2 Detailed information about the geographic orientation and total length of the seismic lines used in the present study

Line name	Orientation	Nature (Dip/strike)	Total VP	Total length (Km)
PBJ-1	SE–NW	Dip	85–215	12.6
PBJ-2	SE–NW	Dip	105–215	10.4
PBJ-3	SE–NW	Dip	99–218	10.3
PBJ-4	SE–NW	Dip	92–250	14.9
PBJ-5	SE–NW	Dip	123–248	11.8
PBJ-6	SE–NW	Dip	101–239	13.4
PBJ-7	SE–NW	Dip	70–181	10.7
PBJ-8	NE–SW	Strike	106–290	18.0
PBJ-9	NE–SW	Strike	101–297	19.3
PBJ-10	NE–SW	Strike	99–289	18.4
PBJ-11	NS	Oblique	94–225	12.8
Total length of the lines (Km)				152.6

VP vibrosise points, Km kilometers

right-lateral offset of the displaced Surghar Fault along the Kalabagh Fault (Fig. 1), and the Kalabagh Fault terminates northward into the Kalabagh Fault zone where N-dipping thrust faults splay to the west (McDougall and Khan 1990).

Methods

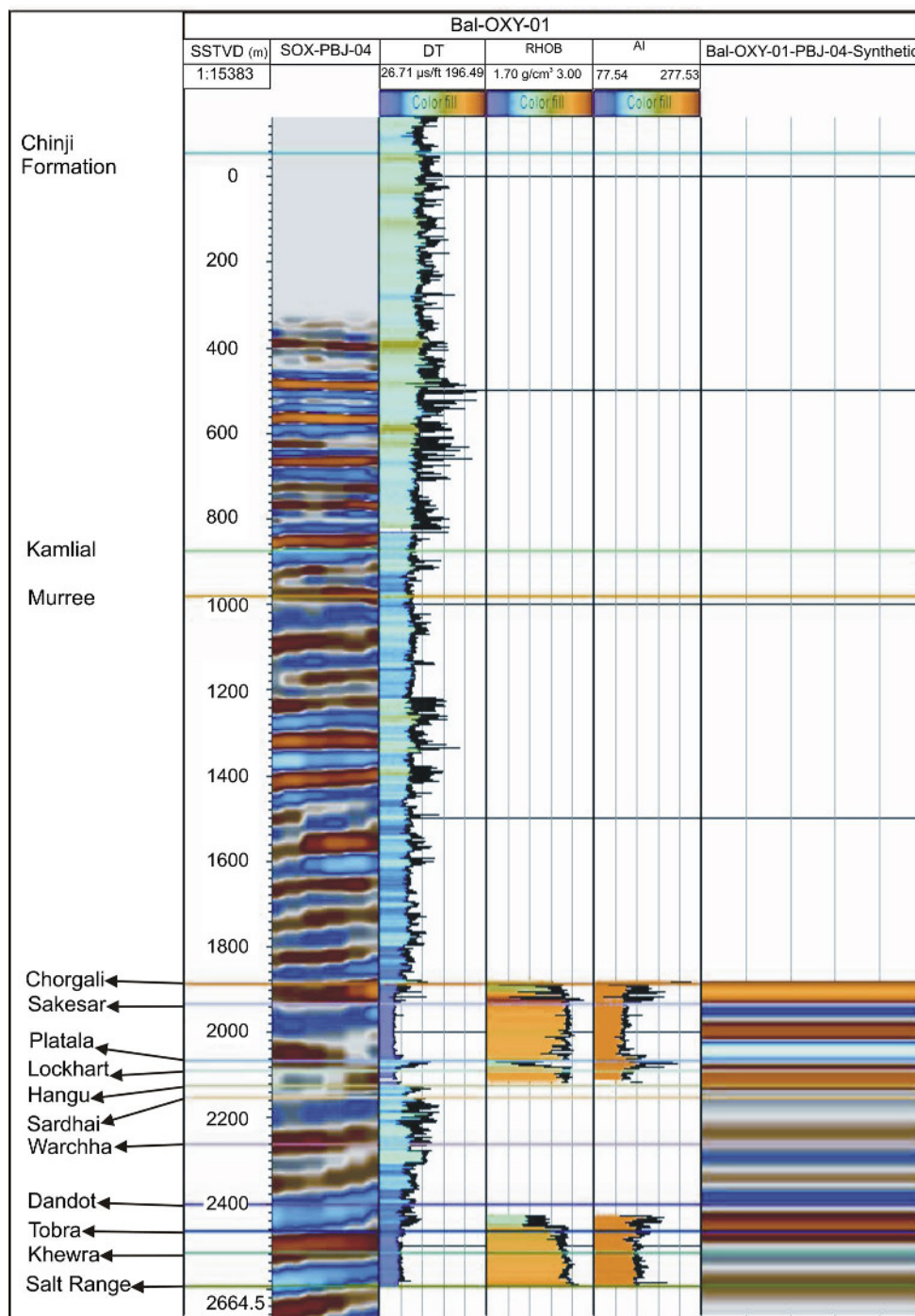
Eleven seismic lines containing 24-fold 2D post-stack (migrated) seismic data were used, covering a total length of 152.6 km (Table 2). Seven lines were dip lines, while three were strike lines and one line (PBJ-11) was oblique (Fig. 3; Table 2). The lines constitute a complete 2D seismic survey set (of the area) available in the public domain of the Directorate General Petroleum Concessions (DGPC) for academic and research purposes. No further seismic line is present (for the area) in the public domain of the DGPC. The dip lines cover both the limbs of the Balkassar anticline and the faults on the limbs. The strike line PBJ-08 covers the northwestern limb, while PBJ-9 and PBJ-10 cover the central part (fold axis) and southeastern limb, respectively. The dip and strike lines form the grid that covers the Balkassar anticline, and the oblique line PBJ-11 serves as an oblique traverse (from south to the north) through the structure. Removing any seismic line (both dip and strike) from the set will have a high negative impact on the time and depth contouring of the different horizons and thus the final interpretation. The data were acquired (through vibrosise acquisition survey) and processed by Oil and Gas Development Company Limited (OGDCL) in 1981. Detailed information about the orientation of these lines is given in Table 2, while the geographic orientation in the base map is shown in Fig. 2. Well tops, density

(RHOB), sonic (DT) and acoustic impedance (AI) logs of Bal-OXY-01, Bal-07, Bal-08 and Bal-A-01 were included, and summary sheets of these wells were also incorporated. A synthetic seismogram was generated for the seismic interpretation (Fig. 4). In seismic interpretation, a synthetic seismogram is a direct one-dimensional model of acoustic energy traveling through the layers of the Earth (Ganley 1981). RHOB and DT logs were used to calculate AI log (using Petrel). The synthetic seismogram was generated (Fig. 4) by convolving the reflectivity derived from digitized AI and RHOB logs of Bal-OXY-01 with the wavelet derived from seismic line PBJ-04 (Figs. 4, 7a). By comparing marker beds and other correlation points picked on well logs with major reflections on the seismic section, seismic data were interpreted (Figs. 5, 6, 7, 8). Time sections were marked for the interpretation based on the synthetic generated from well data, and time surfaces were generated (Figs. 7, 8, 9, 10).

Velocity analyses of the seismic sections were conducted, and interval, average and mean average velocities were used. Since Petrel software (Schlumberger) uses complex algorithms to perform these velocity analyses and produces the results in the form of depth surfaces, figures for only depth surface maps are incorporated (Fig. 10) and not the complex analyses. Domain conversion allows us to take data from one domain, typically seismic data in time, and convert it to another, typically depth, to correlate it with well data. The workflow of converting data between domains involves two processes: (1) making the velocity model, which will define how the velocity varies in subsurface, and (2) depth conversion, which uses the velocity model to move data between domains. Depth conversion was calculated vertically, starting from datum, progressing downward and taking one zone at a time. Time–depth relationship (TDR) was established by using time velocity pairs given on seismic section; this enabled well tops to be displayed on seismic section. The convergent interpolation method was used for the gridding (mapping) process. The convergent grider is a control point-oriented algorithm (rather than grid point), which will converge upon the solution, iteratively adding more and more resolution with each step (iteration). Based on the depth conversion of the time section, depth sections/surfaces were produced (Fig. 10).

Petrel was used for the interpretation of the seismic and well data. Faults were marked, and the validity of the present interpretation was confirmed using the existing literature and surface geological maps (Lillie et al. 1987; Pen-nock et al. 1989; McDougall and Khan 1990; Kadri 1995; Moghal et al. 2007). The depth sections/surfaces were then correlated with the well tops of Bal-OXY-01, Bal-07, Bal-08 and Bal-A-01 in order to establish pinned horizons. By using average velocities given on seismic sections, a TDR was established to visualize well tops (synthetic

Fig. 4 Synthetic seismogram for Bal-OXY-01 well. The synthetic is on the *right* and the seismic on the *left* (*SSTVD* sub-sea true vertical depth, *DT* depth vs. time (sonic log), *RHOB* density log, and *AI* acoustic impedance log)



seismogram) on seismic section (Fig. 7a) to validate the interpretation.

Using the fault polygon option of the Petrel software, faults were correlated on different seismic sections and used in the 2D and 3D contour maps. The contour maps were prepared for the marked horizons, faulted blocks and preparation of structural model. The interpretation was finally converted into depth by using the velocity model. Static modeling was performed by using these interpreted

horizons and faults. Once the static model was prepared, then it was converted into a depth structure model.

Results and interpretation

The Balkassar area is located in the central Potwar Plateau, on the southern flank of the Soan syncline (Figs. 1 and 2). The structures in and around the area are salt-cored

Fig. 5 Representative uninterpreted *dip lines*. **a** PBJ-04. **b** PBJ-05. **c** PBJ-06 and **d** PBJ-07

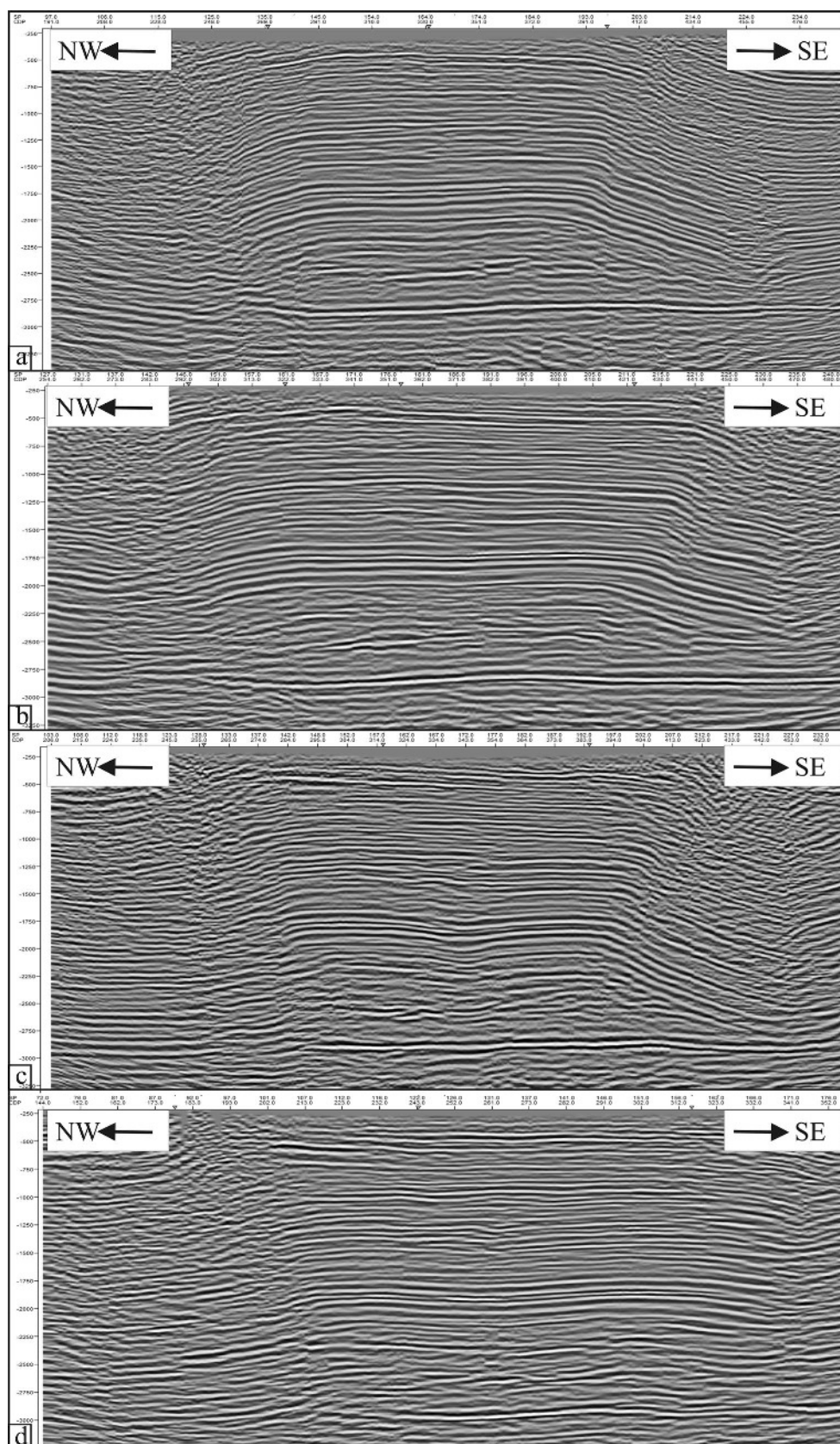


Fig. 6 Representative uninterpreted *strike lines*. **a** PBJ-08. **b** PBJ-09. **c** PBJ-10 and **d** PBJ-11

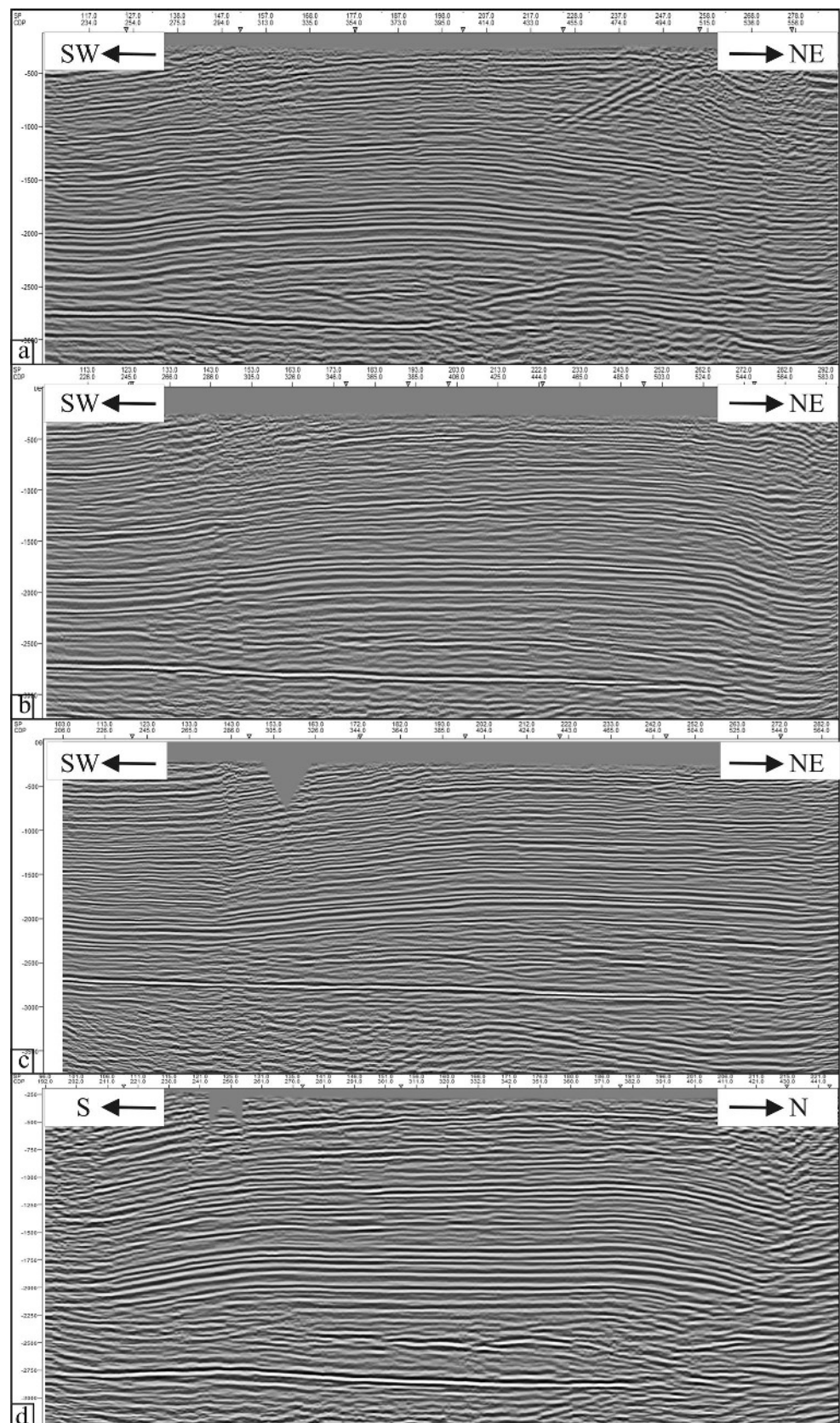


Fig. 7 Representative seismic dip lines. **a** PBJ-04 with synthetic of Bal-OXY-01 displayed at shot point (SP) 172 and horizons marked. **b** PBJ-05 with horizons and faults labeled. **c** PBJ-06 with horizons labeled; however, normal faults in the basement not visible. **d** PBJ-07 with the SE fault not visible. The (X) symbols represent the tie points. Please consider the fault positions as probable where the horizons are not disturbed

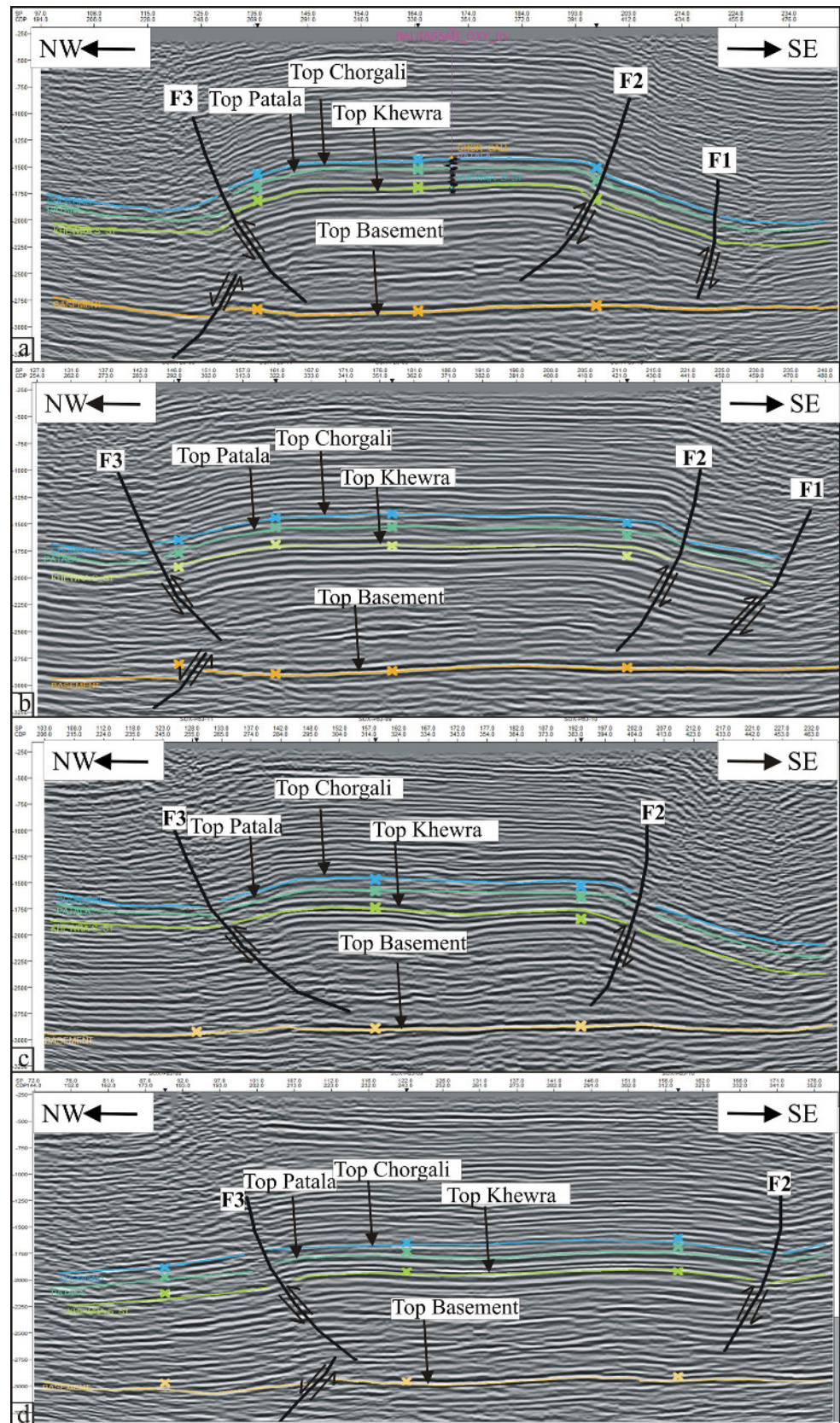


Fig. 8 Representative *strike lines*. **a** PBJ-08 with horizons including *marked*. **b** PBJ-09 with synthetic of Bal-OXY-01 well projected on the *interpreted line*. **c** PBJ-10 with horizons *marked*. **d** PBJ-11 with synthetic of Bal-OXY-01 well projected on the *interpreted line*. The *dashed lines* on the sections represent faults in the cover sequence, while faults in the basement are not visible. The (X) symbols represent the tie points on *each line*

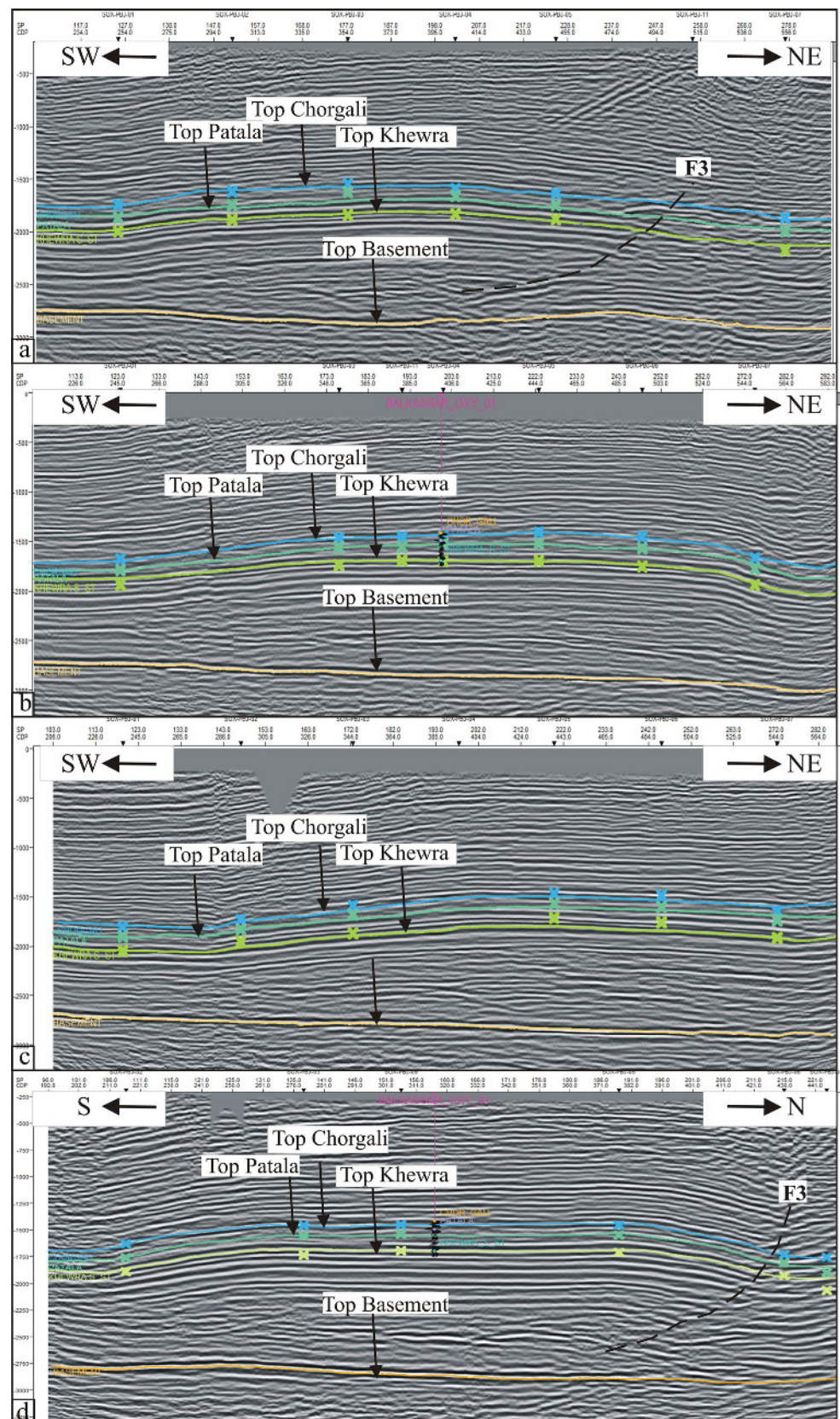


Fig. 9 Time surfaces of the top of the Chorgali Formation (a), Patala Formation (b) and Khewra Sandstone (c). The anticlinal structure has four-way dip closure with shallow time values in the central portion and NE–SW orientation of the reverse faults

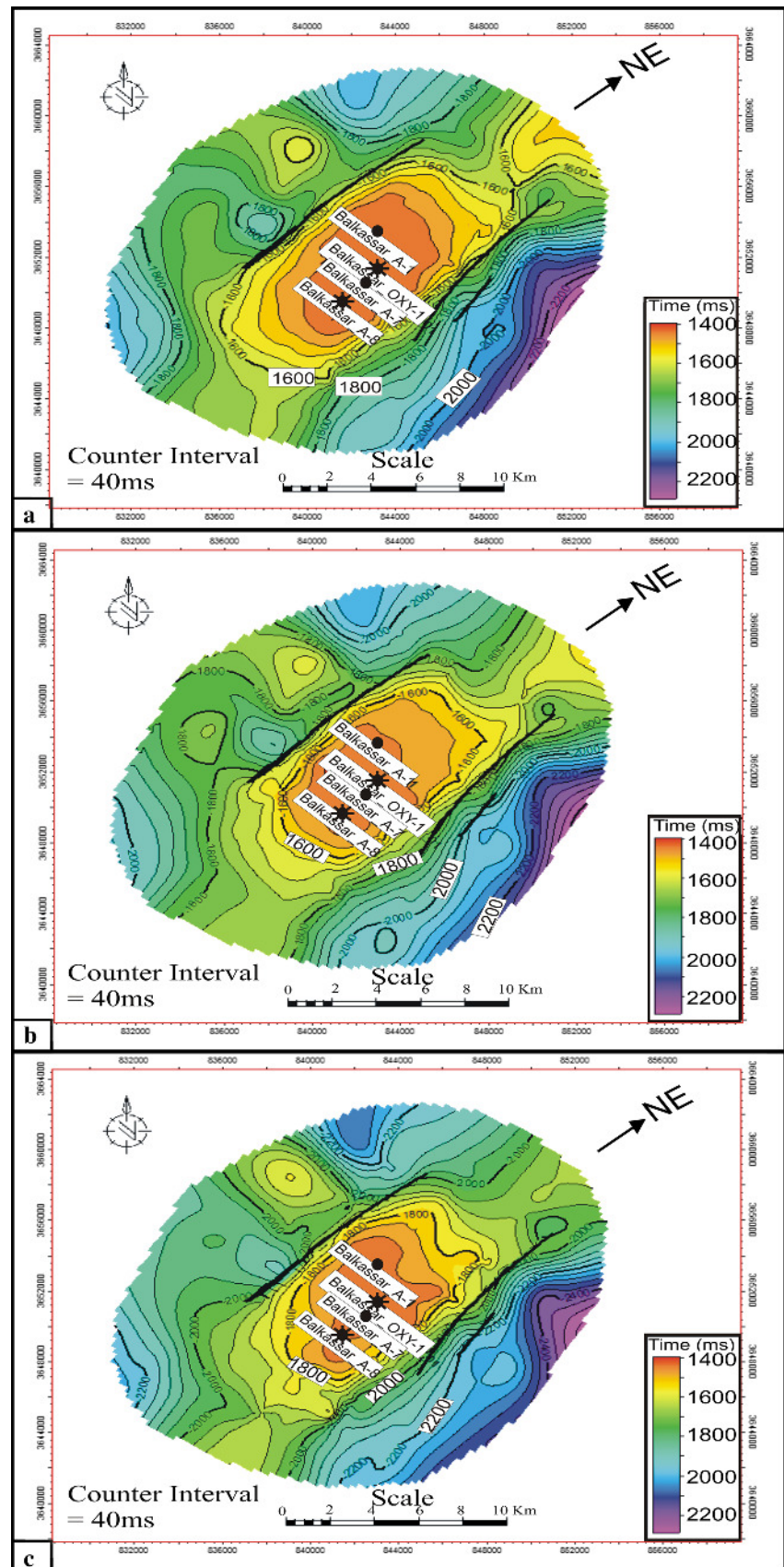
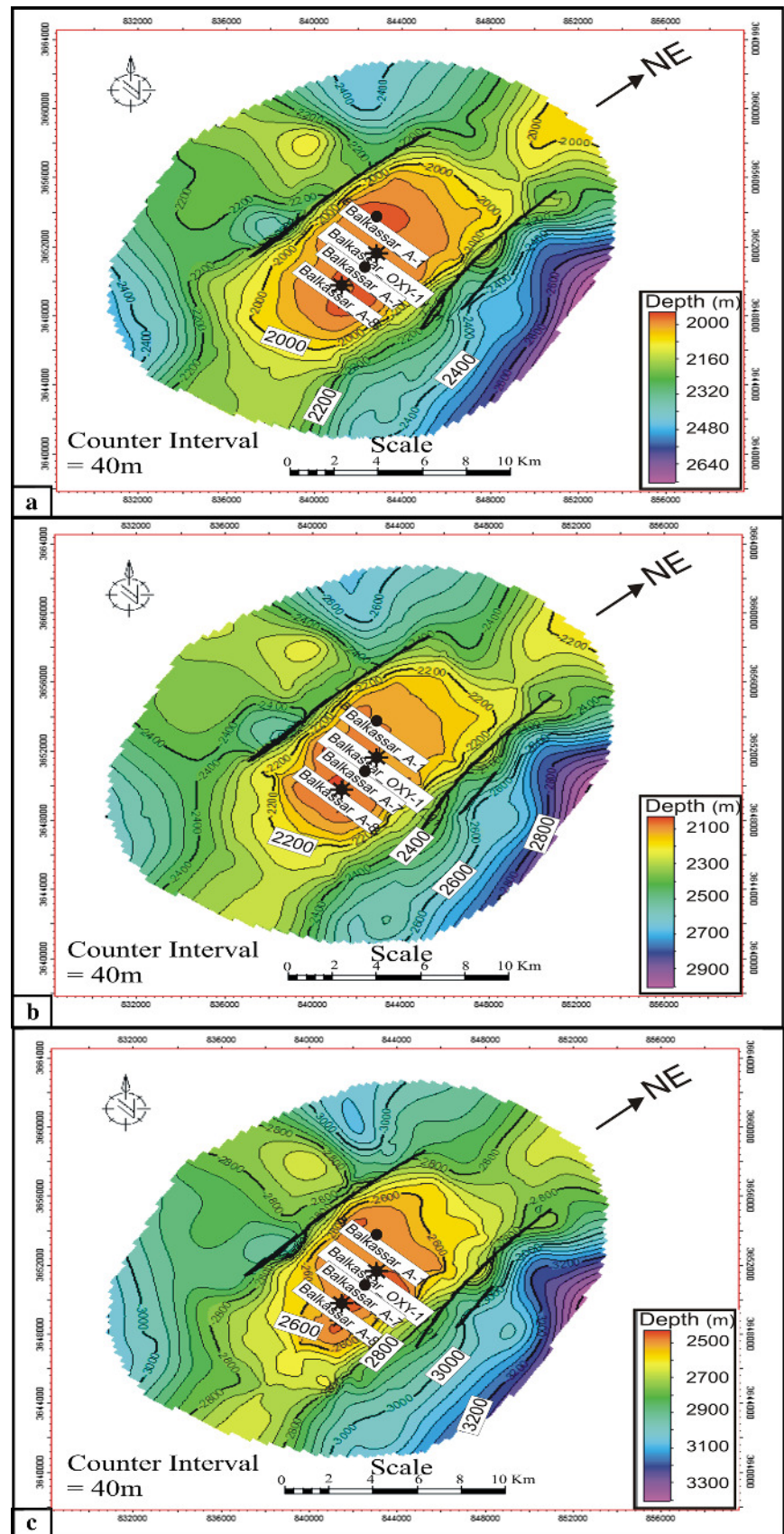


Fig. 10 Depth surfaces of the top of the Chorgali Formation (a), Patala Formation (b) and Khewra Sandstone (c). Shallow depths in the central part of each surface are indicative of the Balkassar anticlinal structure having four-way dip closure in each case. Note the NE–SW orientation of the faults on each surface



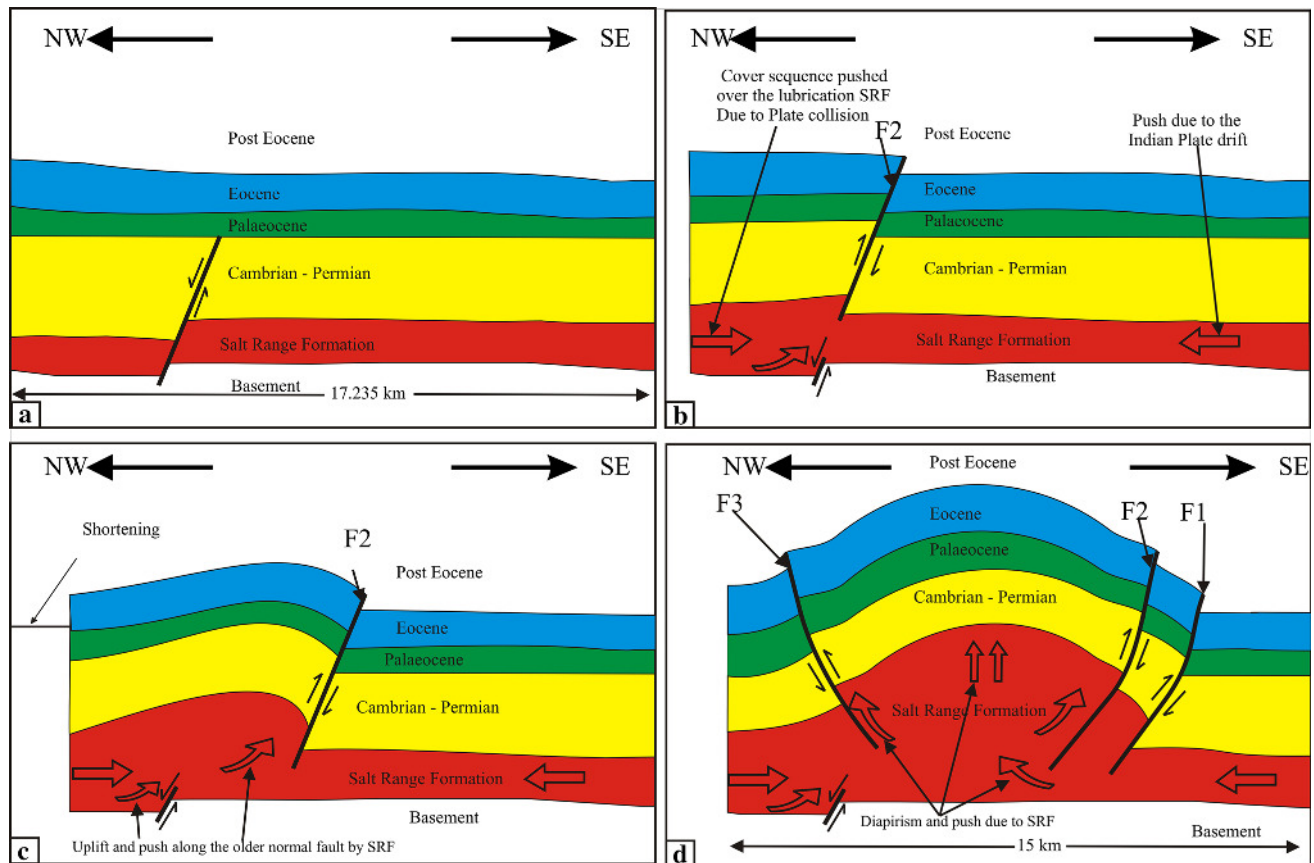


Fig. 11 Sequential development of the Balkassar Anticline. **a** Sedimentary package with normal fault before the onset of the Himalayan orogeny. **b** Conversion of the normal fault into reverse fault during the plate collision. **c** Southward shift of the thrust sheet during due

anticlines, which are bounded by reverse faults (Pennock et al. 1989; Jaswal et al. 1997). The gentle dipping (10° – 20° NW) Nagri Formation (Pleistocene) crops out along a stream cut (Jadoon et al. 1999). A NE–SW elongated anticlinal structure is visible, on the time and depth surface maps (Figs. 9 and 10), that narrows toward the NE.

The seismic grid of $2.5 \text{ km} \times 2.5$ – 3.5 km dimensions was used for the data acquisition in the area (Fig. 3). This grid is inconsistent due to the population and other inaccessible locations due to cultural limitations. The Balkassar anticline was mapped using this 2D seismic grid at prominent reflections corresponding to the top of the Eocene limestone (Bhadrar/Chorgali Formation), top of the Paleocene Patala Formation and top of the Cambrian Khewra Sandstone. These prominent horizons delineate the Balkassar anticline (Figs. 11 and 12).

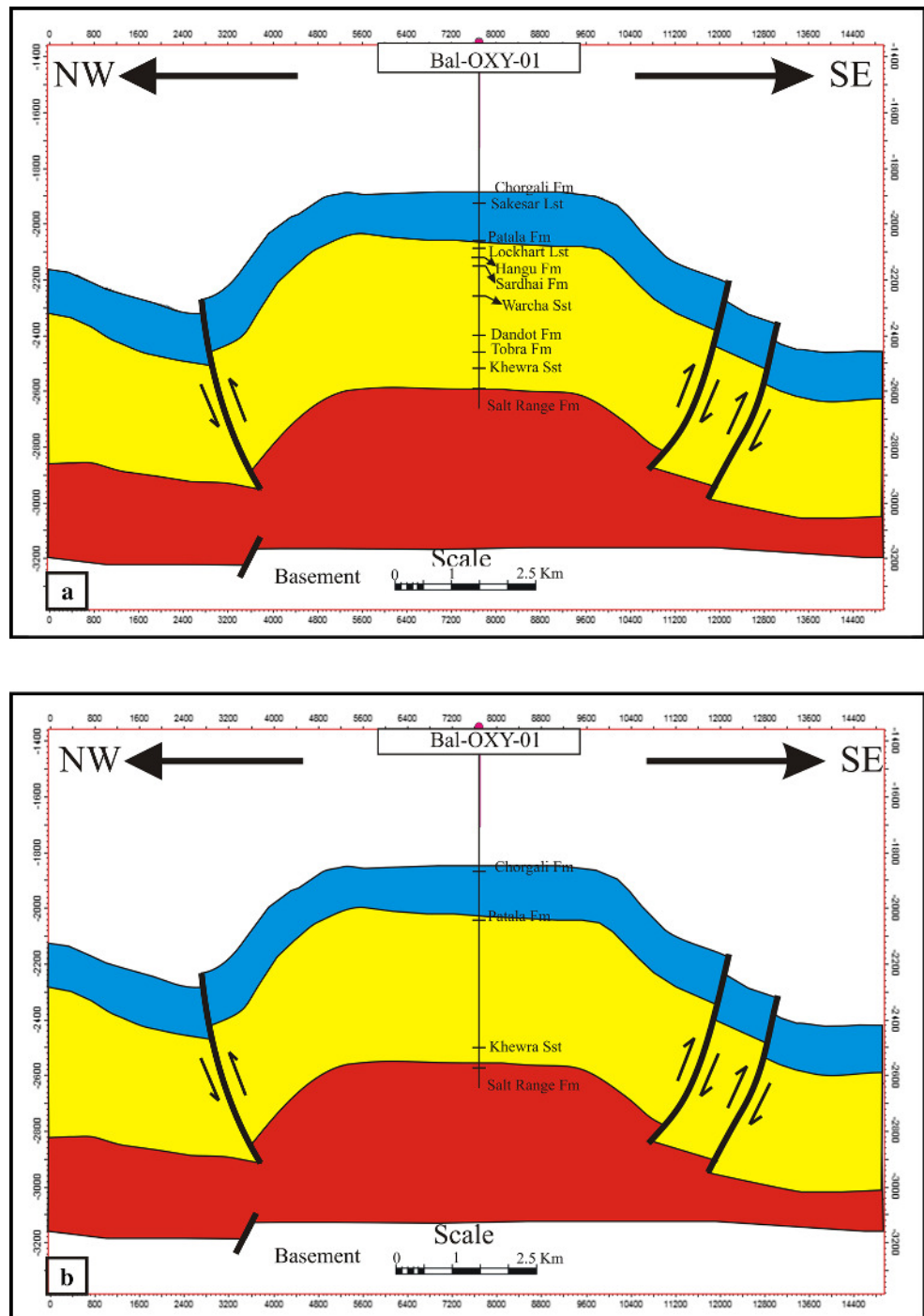
Synthetic seismogram

The synthetic seismogram was prepared using RHOB, DT and AI logs of the Bal-OXY-01 well. The RHOB and AI logs were run only in the Chorgali–Lockhart portion

to the push during plate collision and thickening of the Precambrian evaporites above the normal fault. **d** Present day shape of the Balkassar anticline. Note the crustal shortening during the formation of the Balkassar anticline (SRF Salt Range Formation)

(Eocene–Paleocene, respectively) and then Tobra–Khewra (Permian–Cambrian, respectively). It is an understood choice in well logs economics as these horizons are the reservoir rocks of the plateau (Fig. 4; Kadri 1995). In the present case, the Chorgali Formation (Eocene) is taken as the reference horizon. This horizon is the last marine rock unit (and logs are run in this formation) in the Potwar Plateau and is unconformably overlain by the Miocene and younger molasse sediments (e.g., Shah 1977, 2009). Further, the Chorgali Formation is the main reservoir rock (limestone and dolomite) in the Balkassar. At Chorgali level [1886 m sub-sea true vertical depth (SSTVD)], there is a good match in the synthetic seismogram and the surface seismic, as the well Bal-OXY-01 for which the synthetic seismogram is made is on line PBJ-04 (shot point 172), and the response of log curves is also prominent. On the basis of synthetic seismogram and well tops, Chorgali is marked on seismic section. TDR was established using well tops measured depth (MD), average velocities given on seismic section and two-way time (TWT). In the present case, MD was used as to bring the well data on the same datum (surface elevation) as the seismic data; otherwise, there would have

Fig. 12 Model of the Balkas-sar structure having anticline in the central part bounded by faults on its NW and SE flanks. **a** Well tops of all the formations encountered in Bal-OXY-01 well and tops of the interpreted horizons on seismic are displayed, **b** tops of the horizons interpreted on seismic are displayed and well tops of the same horizons from Bal-OXY-01 well are displayed showing as exact match between the two



been a mismatch between the two as originally the seismic data were using the surface elevation as datum while sea level was the datum for well data. This TDR was further used in velocity modeling.

Seismic interpretation (time sections)

The stratigraphy of the area, as it appeared on 2D seismic, was divided into four vertical successions (Figs. 5, 6,

7, 8). These successions were identified based on seismic reflection and the various interfaces (boundaries). From top to bottom (younger to older), these are as follows: (1) the molasse deposits semi-continuous, parallel reflections of moderate amplitude; (2) below the molasse, is a series of strong, parallel, continuous reflections of moderate to high amplitude, which represent the Cambrian to Eocene platform sequence; (3) a seismicity transparent zone corresponding to the Salt Range Formation; and (4) a locally

Table 3 Crustal shortening along each dip line and average shortening in the Balkassar area along the dip lines

Line name	Line length $L_{(s)}$ (Km)	Horizon length $L_{(h)}$ (Km)	Crustal shortening $CS = L_{(h)} - L_{(s)}$ (Km)	Shortening %
PBJ-1	12.6	14.377	1.777	13.9
PBJ-2	10.4	11.898	1.498	13.9
PBJ-3	10.3	11.804	1.504	14.3
PBJ-4	14.9	17.224	2.324	15.5
PBJ-5	11.8	13.629	1.829	15.3
PBJ-6	13.4	15.450	2.050	15.1
PBJ-7	10.7	12.284	1.584	14.8
Average shortening %				14.9

$L_{(s)}$ Seismic line length, $L_{(h)}$ Seismic horizon length, SC Crustal shortening

continuous, flat to sub-horizontal, strong reflection at the top of the basement. The available well data and velocity information were incorporated with seismic reflection data for the interpretation (Figs. 9, 10, 11, 12). The uninterpreted seismic lines are displayed first (Figs. 5, 6) followed by interpreted seismic lines (Figs. 7, 8) so that the readers can have their own interpretation and can observe the validity of the present interpretation.

Line PBJ-4 is in the middle of the area and was assumed to be a representative line (Fig. 3). The Bal-OXY-01 well, the deepest among the four wells (drilled to the Precambrian Salt Range Formation, 3209 m), was drilled on shot point (SP) 172 of line PBJ-04; therefore, the line holds a key position in interpretation and correlation of the well data (Figs. 4, 7a). A strong reflection was identified as top of the Eocene Chorgali Formation (Fig. 7a) on time level of 1400 ms at shot point 172 (well point) and has a shallowest to deepest time range of 1400–2030 ms, respectively (Figs. 5, 7, 8). Synthetic seismogram at the location of Bal-OXY-01 displayed a confirmation of the well tops of the Chorgali Formation, Patala Formation and Khewra Sandstone on the seismic section (Figs. 4, 7a). To validate the interpretation, all other intersecting seismic lines were tied with PBJ-04. The tops of the Chorgali Formation, Patala Formation, Khewra Sandstone and Basement were mapped and correlated.

Three prominent reverse faults are present in the sedimentary cover sequence, following prominent broken reflectors (Figs. 7, 8, 9, 10, 11, 12). These reverse faults are bounding an anticline where one reverse fault is on the NW limb of the anticline, while two are on its SE limb. All the reverse faults have strike parallel to the axis of the anticline (NE–SW). In addition, NW-dipping normal faults are present in the basement below the NW limb of the structure (Fig. 7a–d).

On seismic line PBJ-01, in the SW part of the study area (Fig. 3), the tops of Chorgali Formation, Patala Formation, Khewra Sandstone and Basement are identified.

An anticline having gently dipping limbs (10° – 20°) with no faults exists. This may be either due to the termination of the Balkassar anticline at this location or due to the low resolution of the seismic data. We interpret that line PBJ-01 lies too far to the SW of the Balkassar anticline. On the NE line PBJ-07 (Fig. 7c), the same reflectors are marked with two reverse faults on both sides (limbs).

The strike lines (Fig. 6) display the tops of the Chorgali Formation, Patala Formation, Khewra Sandstone and Basement. Synthetic seismogram of Bal-OXY-01 is posted on PBJ-09 and PBJ-11 (Fig. 8b, d). Seismic line PBJ-09 passes through the crest of the Balkassar anticline, and no fault exists on this line (Fig. 8a) as faults are on the limbs (Figs. 7, 12). Similarly, no fault is visible on PBJ-10. Only fault F3 is marked on PBJ-08 and PBJ-11 in dashed lines (Fig. 8a, d). We interpret an anticline on the strike lines (Fig. 6). The top of the anticline for the Chorgali Formation is at 1407 ms, for the Patala Formation is at 1528 ms, for the Khewra Sandstone is at 1702 ms and the Basement is marked at 2741 ms time level.

The interpreted seismic lines (Figs. 7, 8) and time structure maps (Fig. 9) display a broad, dome-shaped anticline, which is nearly flat on the top and the longitudinal axis trending in NE–SW direction (Figs. 9, 10). The anticline is bounded by one reverse fault (F3) on to the NW limb and two reverse faults (F1 and F2) on the SE limb (Figs. 7, 8, 9, 10). Normal drag is also observable in the interpreted figures (Fig. 7; Mukherjee 2014).

The length for each seismic line was known (Table 2), and the horizon length for each section was calculated by measuring the length of all the faulted segments (Table 3). The crustal shortening was calculated by subtracting the section length from the horizon length (Yalin et al. 2011). The average crustal shortening in the area is around 14.9 %. Detail analysis (Table 3) suggests maximum crustal shortening of 15.5 % along line PBJ-4, in the central part of the Balkassar anticline. The crustal shortening decreases to a minimum value of 13.9 % in the SW (PBJ-4). The

calculated crustal shortening in the NE is 14.8 % (along PBJ-7). It is important to mention here that only the dip lines were used for the calculation of the crustal shortening and not the strike lines. This is a general procedure in seismic analysis as the structural features are more prominently visible along the dip lines. In comparison, the strike lines display very minor and in many cases no structural variations (Robinson 1988; Robinson and Coruh 1988; Telford et al. 1999; Yilmaz 2001).

Time and depth surface mapping

The time and depth surface maps for the Chorgali Formation, Patala Formation and Khewra Sandstone are given in Figs. 9 and 10. In the previous section, the time sections have already been discussed in detail, and therefore, the depth surfaces/maps are discussed in detail in this section. The depth surface map for the interpreted Chorgali Formation horizon is displayed at a contour interval of 40 m (Fig. 10a). The shallowest depth of 1880 m is observed in the middle part of the map, while the deepest point for the Chorgali Formation is 2720 m at the margins of the map (Fig. 10a). Three faults are bounding the surface (Fig. 10a). A contour level of 1880 m for the Chorgali Formation is covering the top flat surface of the Balkassar anticline. The map displays a structure with four-way closure bounded by three reverse faults (Fig. 10a). In addition to the major anticline, two other small closures occur in the NW (Fig. 10a); however, these are pseudo-structures/features formed due to lack of data in those regions. The base map shows that due to poor coverage of seismic data, these small structures around the main structure are present. The map displays a NE–SW (fold axis) elongated anticline bounded by NE–SW trending faults on its limbs (Fig. 10a–c).

Similar depth surface maps with a contour interval of 40 m are modeled for the interpreted Patala Formation horizon (Fig. 10b) and the Khewra Sandstone horizon (Fig. 10c). These surfaces display similar trends (Fig. 10a–c).

Discussion

Based on the interpreted seismic sections (Figs. 7, 8), mapping and modeling of the horizons (Figs. 9, 10), three reverse faults bound the Balkassar anticline on the NW and SE limbs (Figs. 9, 10, 11, 12). All of these faults are reverse faults, with faults on the SE limb (F1 and F2) dipping toward NW, while the NW fault (F3) dips toward SE (Figs. 7, 9, 10). The faults F1 and F2 are fore-thrusts as these have foreland vergence, while F3 is a back-thrust due to its hinterland vergence (Fig. 7).

Normal faults in the basement, underlying the fault F3 in the cover sequence, are marked on the seismic sections. These normal faults are NW dipping and are attributed to the breakup of Pangaea, in Early Jurassic and subsequent rifting of the Indian Plate from the Arabian and African Plates (Kari 1995; Kazmi and Jan 1997; Rees et al. 2000; Damborenea 2002; Arias 2007). The Infracambrian evaporites of the Salt Range Formation allow the cover sequence on the NW margin of the Indian Plate to easily deform. A similar Jurassic salt mobilization phenomenon occurs in the Infracambrian–Cambrian Ara Formation of the Huqf Group on the Arabian Plate (Murriss 1980). Furthermore, similar mobilization of the Harmos and Namakdan salt diapirs is observed in the Persian Gulf (Mukherjee et al. 2010). A likewise deep erosion, faulting, tilting of strata are observed in the Indian basins that marks post-Early Norian hiatus (Mukhopadhyay et al. 2010; Iqbal et al. 2014b). Similar normal faults are present below: Rothas anticline, Domeli thrust in eastern Potwar (Pennock et al. 1989), Joya Mair–Chak Naurang anticline, Qazian–Bhubar anticline and Chank Beli Khan anticline in central Potwar (Lillie et al. 1987; Baker 1987; Baker et al. 1988). Normal faults in the basement could also be related to the flexural loading of the basement during the Neogene time as suggested by earlier workers (e.g., Lillie and Yousaf 1986; Baker 1987; Baker et al. 1988). However, in that case we should have observed these normal faults on seismic sections in the cover sequence. Further, the Mesozoic and younger rock should have had similar deposition on the eastern and western sides of the Sargodha High (horst), which is not the case (Kadri 1995). Moreover, normal faults are reported associated only with transtensional deformation around the right-lateral Kalabagh Fault also observable in outcrops (McDougall and Khan 1990). Thus, these normal faults are interpreted to have originated during the Jurassic rifting event and salt mobilization (Yeats and Hussain 1987; Pennock et al. 1989).

The ongoing Himalayan orogeny (Molnar 1990) uplifted the Precambrian–Cenozoic strata, along the Salt Range Thrust, over the molasse sediments of the Punjab Plain. During the orogeny, the Indian Plate, with its sedimentary succession, drifted northward (and later rotated NW in the study area). The Indian Plate started to collide with the Eurasian Plate (Kohistan Island Arc in north Pakistan) creating a compressive environment (e.g., Farah et al. 1984; Kazmi and Jan 1997; Jain 2014). The Precambrian evaporites of the Salt Range Formation provided the lubricating mechanism for the deformation (Fig. 11a), and thus, the basement normal faults remained unaffected (Pennock et al. 1999). This overprinted its effects in these normal faults, and these were converted into reverse faults (Fig. 11b) in the cover sequence (Kazmi and Jan 1997).

Reverse faulting in the Potwar and the Salt Range occurred during the southward advance of the thrust sheets (Fig. 11c, d) during the Neogene (Farah et al. 1984; Baker et al. 1988; Jaswal et al. 1997). This southward transport produced fault F2 on the SE limb of the Balkassar anticline (Fig. 11c). Later salt diapirism and push provided by the leading edge of the basement normal faults produced faults F1 and F3 (Fig. 11d). Compression during the Himalayan orogeny was followed by rotation and strike-slip faulting (Mukherjee submitted). This resulted in the rotation of the Potwar Plateau, and subsequently, all the structures (including F1, F2, F3 of the present study and other faults in the area) now display northeast–southwest orientation. This rotation of the plateau is demonstrated by the Jhelum and Kalabagh strike-slip faults (McDougall and Khan 1990; Mukherjee submitted). This is confirmed as the activity along the Kalabagh Fault is dated as 1.9–2 ma (Yeats and Hussain 1987). Activity along the Salt Range Thrust is dated to 10–5 ma, but activity up to 2 ma is reported (McDougall and Khan 1990) and confirmed by the disturbed Miocene–Recent molasse sediments throughout the plateau. Salt mobilization and diapirism during this activity has produced salt-cored anticlines and pop-up structures (Fig. 1; Pennock et al. 1989; Kadri 1995; Jaswal et al. 1997).

During the compression and uplift, the strata suffered crustal shortening (Table 3). The difference between the horizon lengths and the lengths of the seismic section reveals a crustal shortening of around 14.9 % (Table 3). Detailed shortening calculation indicates that maximum crustal shortening (15.5 %) occurred in the central part of the Balkassar anticline. The amount of crustal shortening decreases both to the SW and to the NE where the faults die out (Table 3). A sequential development of the Balkassar anticline and faults development on its limbs explain this shortening (Fig. 11). The calculated crustal shortening for the NPDZ is 19 % (Pennock et al. 1989). Thus, the crustal shortening in the study area is some 20 % less than that in the NPDZ. This supports the relatively less deformation to the south of Soan syncline. This is why there are only folds bounded by faults, whereas the duplexes and triangle zones, which are otherwise common in the NPDZ (Jadoon et al. 1999), are not developed in this part of the Potwar Plateau.

The Balkassar anticline is a salt-cored anticline that is producing hydrocarbons from the Eocene Chorgali Formation. The stratigraphic history of the area is well established in literature (e.g., Shah 1977, 2009; Kazmi and Jan 1997; Kazmi and Abbas 2001; Kadri 1995 and Kazmi and Abbasi 2008). The Eocene Chorgali Formation is the important producing horizon in the Balkassar area, and the Balkassar anticline at this level has a total surface area of about $11 \times 6.5 \text{ km}^2$ (Fig. 10a), while the estimated area of the

closure of the Balkassar anticline is 4000 m^2 and the vertical closure is 370 m approximately.

Conclusions

The Potwar Plateau, located in the North Pakistan, is a petroleum prolific basin, famous for its structural traps. Two-dimensional seismic interpretation in the Balkassar area, central Potwar Plateau, indicates the presence of a salt-cored anticline bounded by two reverse faults at its SE limb and a reverse fault at its NW limb. In addition, older, normal faults are interpreted in the basement. The reverse faults in the cover sequence are attributed to the compressive tectonics of the Himalayan orogeny (Eocene and younger in age), while the normal faults in the basement are related to the Jurassic splitting of Pangaea and associated rifting and opening of the Indian Ocean. The Precambrian Salt Range Formation appears to have controlled the structure in the Balkassar area and many other structures of the Potwar Plateau. Balkassar area has suffered 20 % less crustal shortening as compared to NPDZ, and hence, duplexes and triangles zones are absent in the area.

Acknowledgments The Authors are thankful to the Director General, Petroleum Concessions (DGPC), Pakistan, for the provision of data for this research work and Schlumberger, Pakistan, for providing the software facilities for the work. The authors highly appreciate the valuable remarks and suggestions of the reviewers Dr. Soumyajit Mukherjee and Dr. Afroz Ahmad Shah that greatly improved the manuscript. Prof. Michael Wägrich (University of Vienna) is thanked for his guidance during the work.

References

- Abbasi IA, McElroy R (1991) Thrust kinematics of the Kohat Plateau, Trans Indus Salt Range, Pakistan. *J Struct Geol* 13:319–327
- Arias C (2007) Pliensbachian–Toarcian ostracod biogeography in NW Europe: evidence for watermass structure evolution. *Palaeogeogr Palaeoclimatol Palaeoecol* 251:398–421
- Baig MS, Lawrence RD (1987) Precambrian to early paleozoic orogenesis in the Himalaya. *Kashmir J Geol* 5:1–22
- Baker DM (1987) Balanced structural cross-sections of the central Salt Range and Potwar Plateau of Pakistan: shortening and overthrust deformation (Doctoral dissertation Oregon State University). Department of Geology, Oregon State University, Corvallis, p 120
- Baker DM, Lillie RJ, Yeats RS, Johnson GD, Yousuf M, Zamin ASH (1988) Development of the Himalayan frontal thrust zone: salt Range. *Pakistan, Geol* 16:3–7
- Burbank DW, Reynolds RGH (1988) Stratigraphic keys to the timing of thrusting in the terrestrial foreland basins: applications to the northwest Himalaya. In: Kleinspahn KL, Paol C (eds) *Frontiers in sedimentary geology, new perspective in basin analysis*. Springer, Berlin, pp 331–351
- Butler RWG, Coward MP, Harwood GM, Knipe RJ (1987) Salt, its control on thrust geometry, structural style and gravitational collapse along the Himalayan mountain front in the Salt Range

- of northern Pakistan. In: Lerche I, O'Brien JJ (eds) *Dynamical geology of salt-related structures*. Academic Press, Orlando, Florida, pp 339–418
- Carosi R, Lombardo B, Molli G, Musumeci G, Pertusati PC (1998) The South Tibetan detachment system in the Rongbuk valley, Everest region. Deformation features and geological implications. *J Asian Earth Sci* 16:299–311
- Carosi R, Montomoli C, Visonà D (2007) A structural transect in the lower Dolpo: insights on the tectonic evolution of Western Nepal. *J Asian Earth Sci* 29:407–423. doi:[10.1016/j.jseae.2006.05.001](https://doi.org/10.1016/j.jseae.2006.05.001)
- Damborenea ES (2002) Jurassic evolution of Southern Hemisphere marine palaeobiogeographic units based on benthonic bivalves. *Geobios Mémoire spécial n° 24*, 51–71
- Davies RG, Crawford AR (1971) Petrography and age of the rocks of bulland hill, kirana hills, Sargodha district, west Pakistan. *Geol Mag* 08(03):235–246. doi:[10.1017/S001675680005158X](https://doi.org/10.1017/S001675680005158X)
- Davis DM, Engelder T (1985) The role of salt in fold-and-thrust belts. *Tectonophysics* 119(1–4):67–88. doi:[10.1016/0040-1951\(85\)90033-2](https://doi.org/10.1016/0040-1951(85)90033-2)
- Davis DM, Lillie RJ (1994) Changing mechanical response during continental collision: active examples from the foreland thrust belts of Pakistan. *J Struct Geol* 16:21–34
- DeCelles PG, Robinson DM, Zandt G (2002) Implications of shortening in the Himalayan fold-thrust belt for uplift of the Tibetan Plateau. *Tectonics* 21(6):1062–1087. doi:[10.1029/2001TC001322](https://doi.org/10.1029/2001TC001322)
- Duroy Y, Farah A, Lillie RJ (1989) Subsurface densities and lithospheric flexure of the Himalayan foreland in Pakistan. *Geol Soc Am Spec Pap* 232:217–236
- Farah A, Abbas G, De Jong KA, Lawrence RD (1984) Evolution of the lithosphere in Pakistan. *Tectonophysics* 105(1):207–227
- Fatmi AN (1973) Lithostratigraphic units of the Kohat-Potwar Province Indus Basin Pakistan. *Geol Soc Pak Mem* 10(3):22–28
- Ganley DC (1981) A method for calculating synthetic seismograms which include the effects of absorption and dispersion. *Geophysics* 46(8):110–1107
- Gansser A (1979) The Himalayas—a fascinating geological challenge. *Episodes* 4:17–21
- Gansser A (1980) The significance of the Himalayan suture zone. *Tectonophysics* 62(1):37–52
- Gee ER (1945) The age of the saline series of the Punjab and of Kohat. *Proc Indian Acad Sci Sect B* 14:269–310
- Gee ER (1947) Further note on the age of the saline series of the Punjab and of Kohat. *Proc Indian Acad Sci Sect B* 16:95–153
- Gee ER (1980) Pakistan Geological Salt Range Series: Directorate of Overseas Surveys. United Kingdom, for Gov Pak Geol Surv Pak 6 sheets, scale 1:50,000
- Gee ER (1989) Overview of the geology and structure of the Salt Range, with observations on related areas of northern Pakistan. In: Malinconico LL, Jr, Lillie RJ, (eds.) *Tectonics of the western Himalayas*. *Geol Soc Am Spec Pap.* 232: 95–111
- Ghazi S, Mountney N, Butt A, Sharif S (2012) Stratigraphic and palaeoenvironmental framework of the Early Permian sequence in the Salt Range, Pakistan. *J Earth Syst Sci* 121(5):1239–1255. doi:[10.1007/s12040-012-0225-3](https://doi.org/10.1007/s12040-012-0225-3)
- Ghazi S, Haroon AS, Sahraeyan M, Hanif T (in press) An overview of tectonosedimentary framework of the Salt Range, northwestern Himalayan fold and thrust belt, Pakistan. *Arab J Geosci*. DOI [10.1007/s12517-014-1284-3](https://doi.org/10.1007/s12517-014-1284-3)
- Grelaud S, Sassi W, Lamotte DF, Jaswal T, Roure F (2002) Kinematics of eastern Salt Range and Southern Potwar Basin (Pakistan): a new scenario. *Mar Petrol Geol* 19:1127–1139
- Hallam A, Maynard JB (1987) The iron ores and associated sediments of the Chichali Formation (Oxfordian to Valanginian) of the Trans-Indus Salt Range, Pakistan. *J Geol Soc* 144(1):107–114. doi:[10.1144/gsjgs.144.1.0107](https://doi.org/10.1144/gsjgs.144.1.0107)
- Hanif M, Imraz M, Ali F, Haneef M, Saboor A, Iqbal S, Ahmad S Jr (2014) The inner ramp facies of the Thanetian Lockhart Formation, western Salt Range, Indus Basin, Pakistan. *Arabian J Geosci* 7(11):4911–4926. doi:[10.1007/s12517-013-1099-7](https://doi.org/10.1007/s12517-013-1099-7)
- Hasany ST, Saleem U (2012) An Integrated Subsurface Geological & Engineering Study of Meyal Field Potwar Plateau Pakistan. Search and Discovery Article # 20151, 1–41
- Hodges KV (2000) Tectonics of the Himalaya and southern Tibet from two perspectives. *Geol Soc of Am Bull* 12:324–350. doi:[10.1130/0016-7606\(2000\)1122.0.CO;2](https://doi.org/10.1130/0016-7606(2000)1122.0.CO;2)
- Hodges KV, Parrish RR, Housh TB, Lux DR, Burchfiel BC, Royden LH, Chen Z (1992) Simultaneous Miocene extension and shortening in the Himalayan orogen. *Science* 258:1466–1470. doi:[10.1126/science.258.5087.1466](https://doi.org/10.1126/science.258.5087.1466)
- Iqbal S, Jan UI, Hanif M (2014a) The Mianwali and Tredian Formations: an example of the Triassic Progradational Deltaic System in the low-latitude western Salt Range, Pakistan. *Arab J Sci Eng* 39(7):5489–5507. doi:[10.1007/s13369-013-0836-2](https://doi.org/10.1007/s13369-013-0836-2)
- Iqbal S, Jan UI, Akhter G, Wagreich M, Hanif M (2014b) Triassic–Jurassic boundary: evidences from the Tethyan Salt Range, Pakistan and correlation with Europe. *J Himal Earth Sci* (abstract volume), 59
- Jadoon IAK, Frisch W (1997) Hinterland vergent tectonic wedge below the Riwar thrust, Himalayan Foreland, Pakistan: implications for hydrocarbon exploration. *Am Assoc Petrol Geol Bull* 81:438–448
- Jadoon IAK, Kemal A, Frisch W, Jaswal TM (1997) Thrust geometries and kinematics in the Himalayan foreland (North Potwar Deformed Zone), North Pakistan. *Geol Rundsch* 86:120–131
- Jadoon IAK, Frisch W, Jaswal TM, Kemal A (1999) Triangle zone in the Himalayan foreland, north Pakistan. *Geol Soc Am Spec Paper* 328:277–286
- Jadoon IAK, Hinderer M, Wazir B, Rizwan YR, Bahadar S, Hassan M, Abbasi ZH, Jadoon S (in Press) Structural styles, hydrocarbon prospects, and potential in the Salt Range and Potwar Plateau, north Pakistan. *Arab J Geosci*. doi:[10.1007/s12517-014-1566-9](https://doi.org/10.1007/s12517-014-1566-9)
- Jain AK (2014) When did India–Asia collide and make the Himalaya? *Curr Sci* 106(2):254–266
- Jaswal TM, Lillie RJ, Lawrence RD (1997) Structure and evolution of the northern Potwar deformed zone, Pakistan. *AAPG Bull* 81(2):308–328
- Jaumé SC, Lillie RJ (1988) Mechanics of the Salt Range–Potwar Plateau, Pakistan: a fold-and-thrust belt underlain by evaporites. *Tectonics* 7(1):57–71. doi:[10.1029/TC007i001p00057](https://doi.org/10.1029/TC007i001p00057)
- Kadri IB (1995) Petroleum geology of Pakistan. Pakistan Petroleum Limited, Karachi Pak, pp 40–61
- Kazmi AH, Abbas SG (2001) Metallogeny and mineral deposits of Pakistan. Graphic Publication, Karachi, pp 207–215
- Kazmi AH, Abbasi AI (2008) Stratigraphy & historical geology of Pakistan. National Centre of Excellence in Geology, Peshawar Publication, Peshawar, pp 181–185
- Kazmi AH, Jan MQ (1997) Geology and tectonic of Pakistan. Graphic publication, Karachi, pp 130–141
- Khan MA, Ahmed R, Raza HA, Kemal A (1986) Geology of petroleum in Kohat-Potwar depression, Pakistan. *AAPG Bull* 70:396–414
- Khanal S, Robinson DM (2013) Upper crustal shortening and forward modeling of the Himalayan fold thrust belt along the Budhi-Gandaki River, central Nepal. *Int J Earth Sci*. doi:[10.1007/s00531-013-0889-1](https://doi.org/10.1007/s00531-013-0889-1)
- Khanal S, Robinson DM, Mandal S, Simkhada P (2014) Structural, geochronological and geochemical evidence for two distinct thrust sheets in the ‘Main Central thrust zone’, the Main Central thrust and Ramgarh-Munsiari thrust: implications for upper crustal shortening in central Nepal. In: Mukherjee S, Carosi R, van der Beek PA, Mukherjee BK, Robinson DM (Eds.) *Tectonics of the Himalaya*. *J Geol Soc Lond Spec Pub* 412. doi:[10.1144/SP412.2](https://doi.org/10.1144/SP412.2)

- Le Fort P (1975) Himalayas: the collided range. Present knowledge of the continental arc. *Am J Sci A275*:1–44
- Leathers M (1987) Balanced structural cross-section of the western Salt Range and Potwar Plateau: Deformation near the strike-slip terminus of an overthrust sheet [M.S. thesis]: Corvallis, Oregon State University, 271 p
- Lee J, Hacker B, Wang Y (2004) Evolution of north Himalayan gneiss domes: structural and metamorphic studies in Mabja Dome, southern Tibet. *J Struct Geol* 26(12):2297–2316
- Lillie RJ, Yousaf M (1986) Modern analogs of some midcrustal reflections observed beneath collisional mountain belts. In: Barazangi, M, Brown L (eds.) *Reflection Seismology: The Continental Crust: Amm Geophysical Union Series*. 14: 55–65
- Lillie RJ, Johnson GH, Yousaf M, Zamin ASH, Yeats RS (1987) Structural development within the Himalayan foreland fold-thrust belt of Pakistan. In: Beaumont and Tankard (eds) *Sedimentary Basins and Basin forming Mechanism*. *Can Soc Petro Geol Mem* 12: 379–392
- Mandal S, Robinson DM, Khanal S, Das O (2014) Redefining the tectonostratigraphic and structural architecture of the Almora klippe and the Ramgarh–Munsiyari thrust sheet in NW India. *J Geol Soc Lond Special Publications*. In: (Eds.) Mukherjee S, Carosi R, van der Beek PA, Mukherjee BK Robinson DM, *Tectonics of the Himalaya*. *J of Geol Soc of LondSpec Pub* 412: doi:[10.1144/SP412.6](https://doi.org/10.1144/SP412.6)
- McDougall JW, Khan SH (1990) Strike-slip faulting in a foreland fold-thrust belt: the Kalabagh Fault and western Salt Range, Pakistan. *Tectonics* 9(5):1061–1075
- Moghal MA, Saqi MI, Hameed A, Bugti MN (2007) Subsurface geometry of Potwar Sub Basin in relation to structuration and entrapment. *Pak J Hydroc Res* 17:61–72
- Mohadjer S, Bendick R, Ischuk A, Kuzikov S, Kostuk A, Saydullaev U, Lodi S, Kakar DM, Wasy A, Khan MA, Molnar P, Bilham R, Zubovich AV (2010) Partitioning of India-Eurasia convergence in the Pamir-Hindu Kush from GPS measurements. *Geophys Res Lett* 37:L04305. doi:[10.1029/2009GL041737](https://doi.org/10.1029/2009GL041737)
- Molnar P (1990) A review of the seismicity and the rates of underthrusting and deformation at the Himalaya. *J Himal Geol* 1(2):131–154
- Molnar P, Tapponnier P (1981) A possible dependence of tectonic strength on the age of the crust in Asia. *Earth Planet Sci Lett* 52:107–114
- Mukherjee S (2014) Review of flanking structures in meso- and micro-scales. *Geol Mag* 151:957–974
- Mukherjee S, Talbot CJ, Koyi HA (2010) Viscosity estimates of salt in the Hormuz and Namakdan salt diapirs, Persian Gulf. *Geol Mag* 147:497–507
- Mukhopadhyay G, Mukhopadhyay SK, Roychowdhury M, Parui PK (2010) Stratigraphic Correlation between Different Gondwana Basins of India. *J Geol Soc* 76:251–266
- Morris RJ (1980) Middle East stratigraphic evolution and oil habitat. *AAPG Bull* 64:597–618
- Pennock ES, Lillie RJ, Zaman ASH, Yousaf M (1989) Structural interpretation of seismic reflection data from eastern Salt Range and Potwar Plateau, Pakistan. *AAPG Bull* 73(7):841–857
- Rana RA, Asrarullah (1982) Tectonic map of Pakistan. Government of Pakistan, Min. of Petroleum and Natural Resources
- Rees PM, Ziegler AM, Valdes PJ (2000) Jurassic phytogeography and climates: new data and model comparisons. In: Huber BT, Macleod KG, Wing SL (eds) *Warm climates in earth history*. Cambridge University Press, Cambridge, pp 297–318
- Robinson ES (1988) Geophysical prospecting. *Pub Lond* 76–78
- Robinson ES, Coruh C (1988) Basic exploration geophysics. Wiley, New York, pp 65–121
- Robinson DM, Pearson ON (2013) Was Himalayan normal faulting triggered by initiation of the Ramgarh–Munsiyari Thrust? *Int J Earth Sci* 102(7):1773–1790. doi:[10.1007/s00531-013-0895](https://doi.org/10.1007/s00531-013-0895)
- Sercombe WJ, Pivnik DA, Wilson WP, Albertin ML, Beck RA, Stratton MA (1998) Wrench faulting in the northern Pakistan foreland. *AAPG bull* 82(11):2003–2030
- Shah SMI (1977) Stratigraphy of Pakistan. *Mem Geol Sur Pak* 12:36–45
- Shah SMI (2009) Stratigraphy of Pakistan. Gov Pak Ministry of Petrol Nat Resour Geol Sur Pak 245–273
- Tahirkheli RAK (1979) Geology of Kohistan and adjoining Eurasian and Indo-Pakistan continent, Pakistan. *Geol Bull Univ Pesh* 11:1–30
- Telford WM, Geldart LP, Sheriff RE, Keys DA (1999) Applied geophysics. Cambridge University Press, London, pp 65–78
- Treloar PJ, Coward MP, Chambers AF, Izatt CN, Jackson KC (1992) Thrust geometries, interferences, and rotations in the northwest Himalaya. In: McClay KR (ed) *Thrust tectonics*. Chapman and Hall, London, pp 325–342
- Upreti BN (1999) An overview of the stratigraphy and tectonics of the Nepal Himalaya. *J Asian Earth Sci* 17:577–606. doi:[10.1016/S1367-9120\(99\)00047-4](https://doi.org/10.1016/S1367-9120(99)00047-4)
- Valdiya KS (1984) Evolution of the Himalaya. *Tectonophysics* 105(1–4):229–248. doi:[10.1016/0040-1951\(84\)90205-1](https://doi.org/10.1016/0040-1951(84)90205-1)
- Valdiya KS (1997) Himalaya, the northern frontier of East Gondwanaland. *Gondwana Res* 1(1):3–9. doi:[10.1016/S1342-937X\(05\)70002-2](https://doi.org/10.1016/S1342-937X(05)70002-2)
- Wang X, Zhang J, Liu J, Yan SY, Wang J (2012) Middle-Miocene transformation of tectonic regime in the Himalayan orogen. *Chin Sci Bull* 57(33):3162–3172
- Mukherjee S (Submitted) A review on out-of-sequence deformation in the Himalaya. In: Mukherjee, S., Carosi, R., van der Beek, P. A., Mukherjee, B. K., Robinson, D. M. (Eds) *Tectonics of the Himalaya*. *Geol Soc Lond Spec Pub* 412
- Yalin L, Chengshan W, Chao M (2011) Balanced cross-section and crustal shortening analysis in the Tanggula–Tuotuohe Area, Northern Tibet. *J Earth Sci* 22(1):1–10. doi:[10.1007/s12583-011-0152-2](https://doi.org/10.1007/s12583-011-0152-2)
- Yeats RS, Hussain A (1987) Timing of structural events in the Himalayan foothills of northwestern Pakistan. *Geol Soc Am Bull* 9(2):161–176
- Yilmaz O (2001) Seismic data analysis and processing, inversion and analysis of seismic data. *Soc Exp Geophys, Tulsa*, 265–353
- Yin A (2006) Cenozoic tectonic evolution of the Himalayan orogen as constrained by along-strike variation of structural geometry, exhumation history, and foreland sedimentation. *Earth Sci Rev* 76:1–131. doi:[10.1016/j.earscirev.2005.05.004](https://doi.org/10.1016/j.earscirev.2005.05.004)

CHAPTER 4

PALAEOENVIRONMENTS AND SEQUENCE STRATIGRAPHY

This chapter describes the depositional palaeoenvironments and depositional sequence stratigraphy of the Lower Jurassic Datta Formation in the western Salt Range. The lithofacies and facies associations have been worked out in the Kasanwala and Kaowaali sections (eastern most sections) to understand the response of the silicilastic system to fluctuating sea-level (Iqbal et al., 2015b).

Palaeoenvironmental and sequence stratigraphic analyses of the Jurassic Datta Formation, Salt Range, Pakistan.

Iqbal, S., Jan, I. U., Akhter, M. G., and Bibi, M. (2015b) Journal of Earth System Science, 124(4), 747-766. Doi:10.1007/s12040-015-0572-y.

Author's comments

Shahid Iqbal wrote the article and was the corresponding author. The studied sections were selected through mutual discussion of all the co-authors. Shahid Iqbal and Irfanullah Jan conducted the fieldwork and sample collection. Gulraiz Akhter made the thin sections and processed the X-Ray Diffraction (XRD) samples. Mehwish Bibi prepared the Quartz-Feldspars-Rock Fragments (QFRF) plots and CorelDraw maps/figures. Shahid Iqbal studied the thin sections, refined the figures and interpreted all the data sets. Shahid Iqbal prepared the sea-level curve and depositional model.

Palaeoenvironmental and sequence stratigraphic analyses of the Jurassic Datta Formation, Salt Range, Pakistan

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The Lower Jurassic Datta Formation, western Salt Range, Pakistan, comprises three facies associations: (1) channel belt facies association (CBFA), (2) channel margin, and overbank facies association (CMOFA), and (3) lagoonal facies association (LFA). A cyclic fining-upward trend in the succession is represented by basal quartzose conglomerate/pebbly sandstone, through coarse to fine quartzose sandstone to siltstone and shales/claystone, which contains some carbonate accumulation. Two prominent depositional sequences are recognized in the Datta Formation with the lower high and upper low magnitude cycles. The Datta Formation thus represents a thick sedimentary succession and in the study area, i.e., western Salt Range, mainly channel belt, flood plain and/or delta top facies are exposed. The palaeocurrent analysis shows that the source area with acidic plutonic rocks laid to S–SE in the Indian shield, aravalies or older sedimentary rocks of the Indus Basin (i.e., Khewra, Tobra and Warchha formations). A tentative stratigraphic correlation of the Datta Formation with the lower Jurassic Lathi Formation, India invites further work in parts of India, which will elaborate the extent of the Datta Formation in the Greater Indian peninsula and develop palaeogeographic setting for this Lower Jurassic deltaic rock unit.

1. Introduction

The Jurassic Datta Formation in the western Salt Range, Pakistan (figure 1a, b and c) represents a thick siliciclastic rock unit with some minor carbonate accumulation. The succession holds appreciable hydrocarbon, fire clay and silica sand deposits and thus has proved economically very significant. However, despite being economically important, the strata has received very limited research attention in the study area, i.e., the western Salt Range. Elsewhere, in the Salt Range and Trans-Indus ranges (figure 1b), the lithofacies, sand body geometry, and depositional palaeoenvironments have been investigated for this unit (i.e., Ashraf *et al.* 1976; Rafique *et al.* 1976;

Fatmi *et al.* 1990; Mujtaba and Memon 1990; Chaudhry *et al.* 1994; Obaid *et al.* 2005; Abbasi *et al.* 2012).

The type locality of the Datta Formation lies in the Surghar Range, where the reported thickness is 212 m (Fatmi 1974, 1977). The formation is widely distributed in the Salt Range, and Trans-Indus ranges, parts of Kohat Plateau, Kala Chitta, and Hazara areas. In the Salt Range (table 1), it is well-developed in the western part, i.e., Nammal Gorge section, showing a maximum recorded thickness, i.e., 150 m, with eastward decrease in thickness (figure 1b and c). In the study area, i.e., Kaowaali and Kasanwala sections, the thickness recorded is 121.5 and 83.7 m, respectively (figures 1c and 2).

Keywords. Lower Jurassic; Datta Formation; Salt Range; Indian shield; Indus Basin.

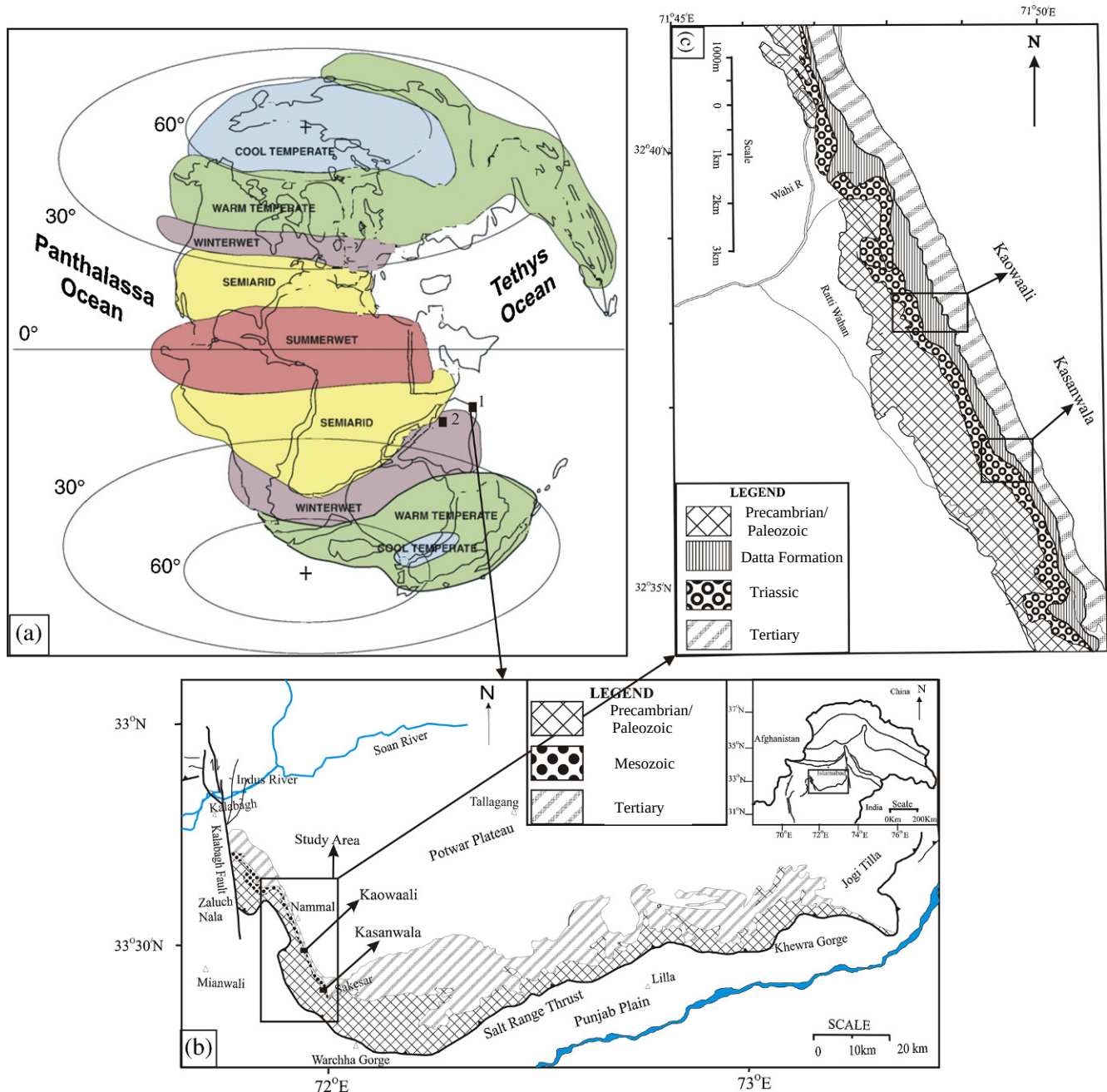


Figure 1. Current and palaeogeographic location of the Salt Range, Pakistan. (a) Early Jurassic palaeogeographic location of the study area, showing, 1. Salt Range, 2. Jaisalmer (India). (b) Location of Salt Range. The map (c) inset shows locations where the study was conducted (modified after Rees *et al.* 2000; Gee 1980, 1989).

The formation has a disconformable lower contact with the Triassic Kingriali Formation (figure 3a) and the upper contact with the Palaeocene Hangu Formation (figure 3b).

The Datta Formation represents first Jurassic sedimentation following the Late Triassic unconformity (Shah 1977, 2009; Abbasi *et al.* 2012). The present study aims to understand the depositional setting and sequence stratigraphic architecture of the Datta Formation in the western Salt Range, based on investigation in the field and petrographic studies in the laboratory (figures 3–6). This study interprets

the depositional palaeoenvironments of different lithofacies in the Datta Formation, establishes facies association for the formation, and explains its depositional sequence stratigraphic architecture. It further makes an effort to present a tentative palaeodepositional model for the formation based on interrelationship of the various lithofacies, facies association, and sequence stratigraphic architecture. This study may thus help in establishing the position of the Datta Formation as an important horizon for regional and global correlating, following the Late Triassic disconformity.

Table 1. Lithostratigraphic units of the Kohat–Potwar Plateau. The Datta Formation is highlighted in bold (modified after Iqbal *et al.* 2014).

Era	Period	Group (Formations of Baroch Group listed)		
CENOZOIC	Tertiary	Pliocene	Siwalik Group	
		Miocene		
		Unconformity		
		Eocene	Cherat Group	
Paleocene		Makarwal Group		
Unconformity				
MESOZOIC	Cretaceous	Surghar Group		
	Jurassic	Baroch Group	Samana Suk Formation Shinawari Formation Datta Formation	
	Triassic	Musa Khel Group		
	Permo-Triassic Boundary			
PALEOZOIC	Permian	Zaluch Group		
		Nilawahan Group		
	Unconformity			
	Cambrian	Jhelum Group		
PROTEROZOIC	Precambrian	Salt Range Formation		

2. Palaeogeography and tectonics

The Salt Range occupied a location at the north-western extremity of the Indian Plate, i.e., ~25°S, along the Tethyan margin during Early Jurassic (Rees *et al.* 1999, 2002; Damborenea 2002; Arias 2007; figure 1a). The Pangaea break up in the Jurassic and subsequent rifting of the Indian Plate from Africa and Arabia resulted in many conspicuous topographic features and in the Potwar Plateau many subsurface normal faults were produced, which are interpreted based on seismic data and geological sections (e.g., Kadri 1995; Kazmi and Jan 1997). Additionally, the low density evaporites of the Infracambrian Salt Range

Formation uplifted the northwestern margin of the Indian Plate, a phenomenon resembling the Jurassic salt mobilization in the Infracambrian to Cambrian Ara Formation of the Huqf Group on the Arabian Plate (Murriss 1980). This uplift resulted in the west-northwestward depression in the area and subsequent development of the fluvial system, which in Pakistan resulted in the deposition of the Datta Formation.

Likewise, in many Indian basins, deep erosion, faulting, tilting of strata mark a post-Early Norian hiatus. This represents a complete change of depositional setting and a sedimentation break between the underlying Late Triassic and Late Pliensbachian sediments within the Gondwana basins of India

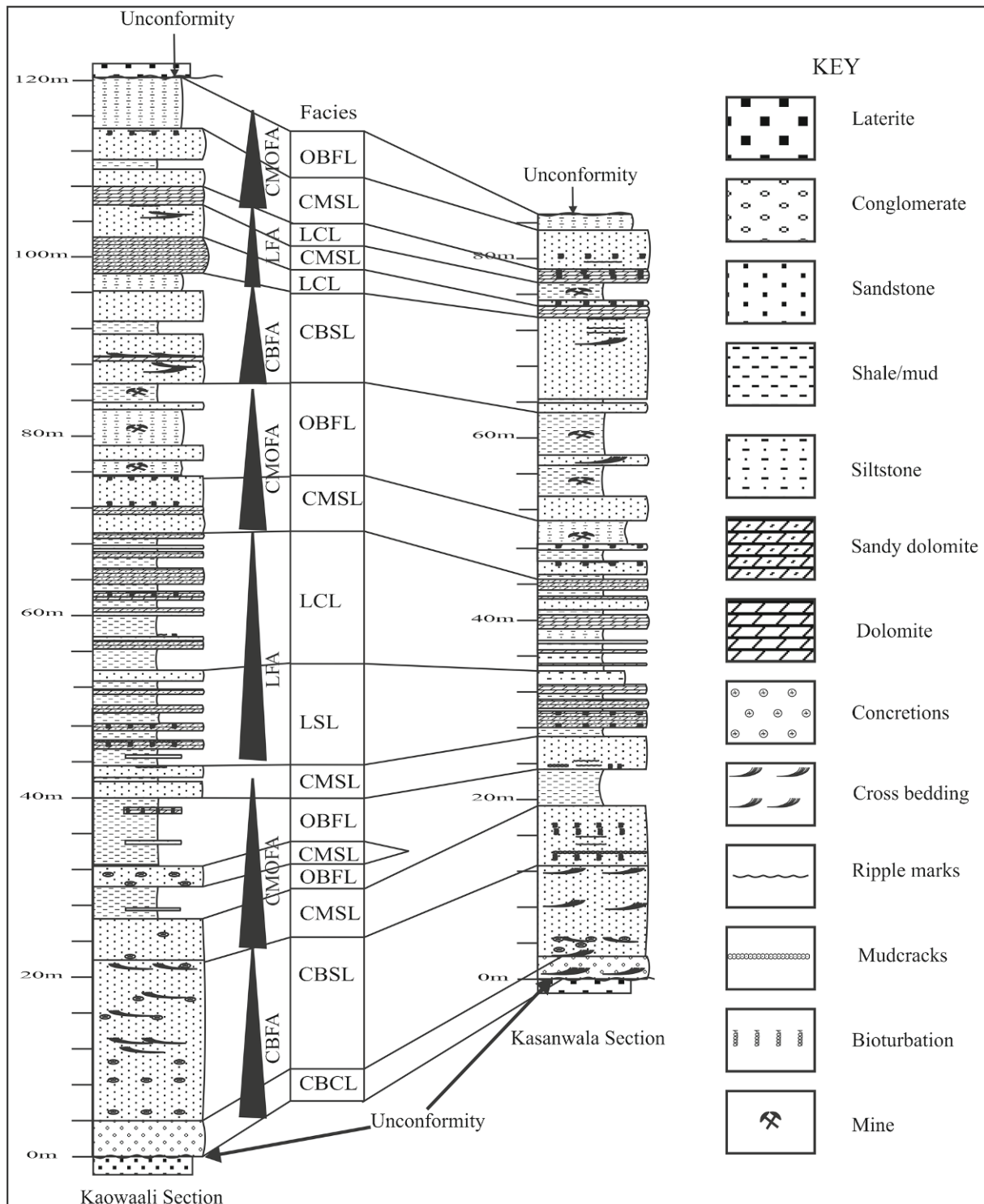


Figure 2. Stratigraphic sections measured for the Datta Formation at the studied locations. CBCL=Channel Belt Conglomerate Lithofacies, CBSL=Channel Belt Sandstone Lithofacies, CMSL=Channel Margin Sandstone Lithofacies, OBFL=Overbank/Floodplain Lithofacies, LSL=Lagoonal Shale Lithofacies, LCL=Lagoonal Carbonates Lithofacies, CBFA=Channel Belt Facies Association, CMOFA=Channel Margin and Overbank Facies Association and LFA=Lagoonal Facies Association.

(Mukhopadhyay *et al.* 2010). The ongoing Himalayan orogeny uplifted the Precambrian–Cenozoic strata, along the Salt Range Thrust, over the Molasse-type sediments of the Punjab Plain (Baker *et al.* 1988; Kazmi and Jan 1997).

3. Material and methods

The Lower Jurassic Datta Formation was studied in the western Salt Range (figure 1b and c). Sections of the Datta Formation were studied at the

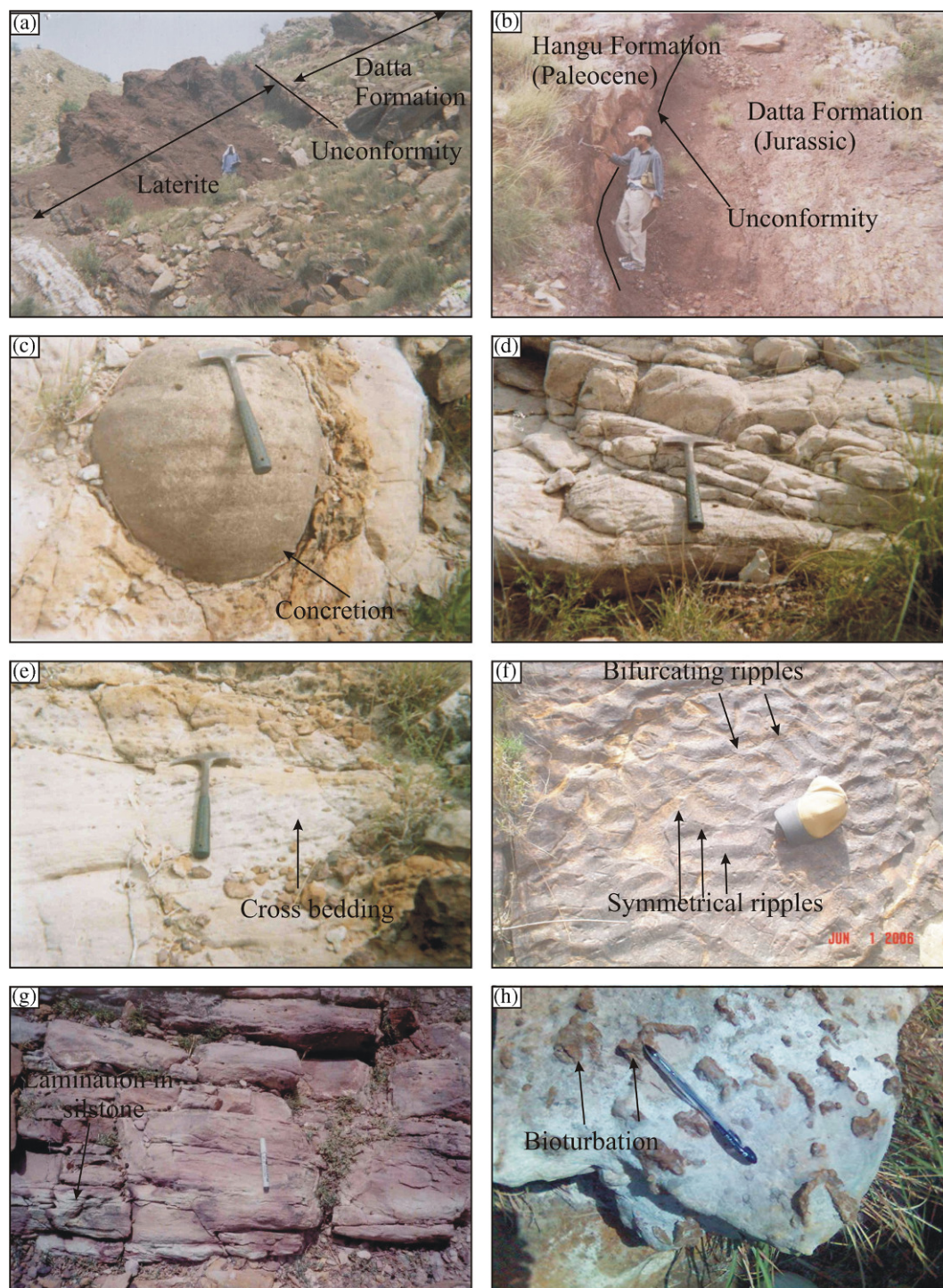


Figure 3. Characteristic lithofacies present in the Datta Formation of the Salt Range, (a) lower contact of the Datta Formation at Kasanwala section, (b) upper contact of the Datta Formation at Kaowaali section, (c) concretions in the CBSL, (d) quartz-rich sandstone of the CBSL, (e) cross-bedding in the CBSL, (f) ripple marks in CMSL, (g) oxidised sandstone in OBFL, and (h) bioturbation in OBFL.

Kasanwala Nala and Kaowaali section (figures 1b, c and 2). The two sections are easily accessible and the formation has excellent continuous exposures in both the sections (figure 2). Sedimentological analyses of the Datta Formation were carried out in the two sections; field observations (mesoscopic) and petrographic studies were performed. The lithofacies

(figure 2) were traced for vertical and lateral changes, i.e., variations in the bedding behaviour and bedding types (e.g., pinching out and thickening, coarsening and/or fining up). Marker horizons such as erosional surfaces, oxidized zones, palaeosoles, rippled and cross-bedded horizons and beds with plant remains were identified. Photographs of all

the important structures and features were taken (figure 3). Samples for petrography and grain size analysis from the formation were collected (43 from Kasanwala Nala and 59 from Kaowaali section). Olympus BX41 microscopic with DP12 camera was used for petrographic studies to determine the grain composition, size and shape parameters of the lithofacies in the Datta Formation (figure 4). The lithologic logs were prepared and positions of samples were plotted. 20 samples were selected for XRD analysis from the two sections (figure 7). X-Ray Diffractometer (JEOL) 1987 JDX-60PX was used for XRD analysis. The petrographic analyses combined with the field observations produced a refined interpretation of the lithofacies and facies associations. A schematic palaeodepositional model was suggested for the formation in the study area and various palaeodepositional subenvironments for the lithofacies were marked.

4. Sedimentary facies analysis

Six distinctive lithofacies are recognized in the Datta Formation in two sections (figure 2), making up various proportions. These are discussed as follows.

4.1 Channel belt conglomerate lithofacies (CBCL)

Description. The lithofacies comprises clast-supported conglomerate (Boggs 2006, 2009), dark to yellowish brown (i.e., at the Kaowaali section) and light to dark grey (at the Kasanwala section). The maximum thickness attained is 4 m (figure 2). The lithofacies is characterized by dominance of granule to pebble sized clasts (1.25–5 cm) and finer sediments also exist, showing good sorting. Rock fragments constitute 50–75% of the lithofacies. Pebble size quartz, chert, quartzite constitute more than 90% of the clasts while granitic rock fragments also occur (figure 4a). Occasional angular, poorly sorted dolomite and sandstone clasts are also present but these constitute less than 10% of the clasts but these (especially the angular, poorly sorted dolomite clasts) make significant proportion (20–30%) of the clasts in the lowermost bed at the contact with the underlying Kingriali Formation. The proportionally insignificant matrix consists of fine to coarse quartz rich sand and iron rich material (figure 4a). The effects of oxidation are prominent in the lithofacies at Kaowaali section only. Individual beds are 1–1.2 m thick and can be traced over 1000–1500 m, while lateral pinching/thinning is observed. Erosive basal bedding plans are commonly observed. Cross and graded bedding are the other important sedimentary structures,

which are present at places. Trough cross stratification is identified and the cross bed sets display fining-upward trends.

Most of the conglomerates are classified as quartzose conglomerates with ultrastable quartz, quartzite, and chert being the dominant rock fragments (Pettijohn *et al.* 1973, Folk 1974; Pettijohn 1975; Boggs 2009). However, the conglomerates in the lowermost part of the lithofacies that contain 20–30% angular and poorly sorted clasts of dolomite are classified as petromictic conglomerates (Boggs 2009). The clasts are mostly sub- to well-rounded, well- to very well-sorted and smooth margins are common (figure 4a). Long, linear contacts are common and concave-convex and sutured contacts though present, are rare. Fracturing is a common diagenetic phenomenon and haematite is common cementing material, which postdates fracturing (figure 4a). At places, second phase fine-grained quartz is introduced into the pore spaces between the grains. The lithofacies has an erosive basal plain and topographic variation of the palaeodepositional surface is observed.

Interpretation. The basal bed with unsorted and angular dolomite fragments indicates that the clasts were derived from the underlying Triassic Kingriali Formation, marking facies dislocation (Iqbal *et al.* 2014), which occurred due to a prominent relative sea-level fall (Reynolds 1994). The presence of petromict conglomerate containing highly soluble/unstable clasts of dolomite imply rapid erosion of elevated highlands (Smith *et al.* 2006; Boggs 2009). On the other hand, such unstable carbonate clasts may also represent accumulation under very cold condition where glacial activity provides the eroded clasts (Boggs 2009). In the present case, the thick laterite at the base of the Datta Formation and the dominance of ultrastable clasts in the quartzose conglomerates of the lithofacies favour clasts formation in humid palaeoclimate and inhibit the possibilities of occurrence of cold glacial palaeoclimate.

The individual beds of the lithofacies display fining upward trend, a feature generally not observed in sheet flood (braided stream) conglomerates, though the braided stream conglomerates generally display a fining upward trend that is observed in the present lithofacies (Boggs 2006, 2009; Smith *et al.* 2009; Lunt *et al.* 2013). On the other hand, the stream flow conglomerates (meander channel belt conglomerates) display well developed fining upward trends and commonly exhibit well developed cross beddings (Boggs 2009). Both these features are observed in the lithofacies. Further the presence of dominant finer material (i.e., siltstones and shale) in the overlying lithofacies indicates

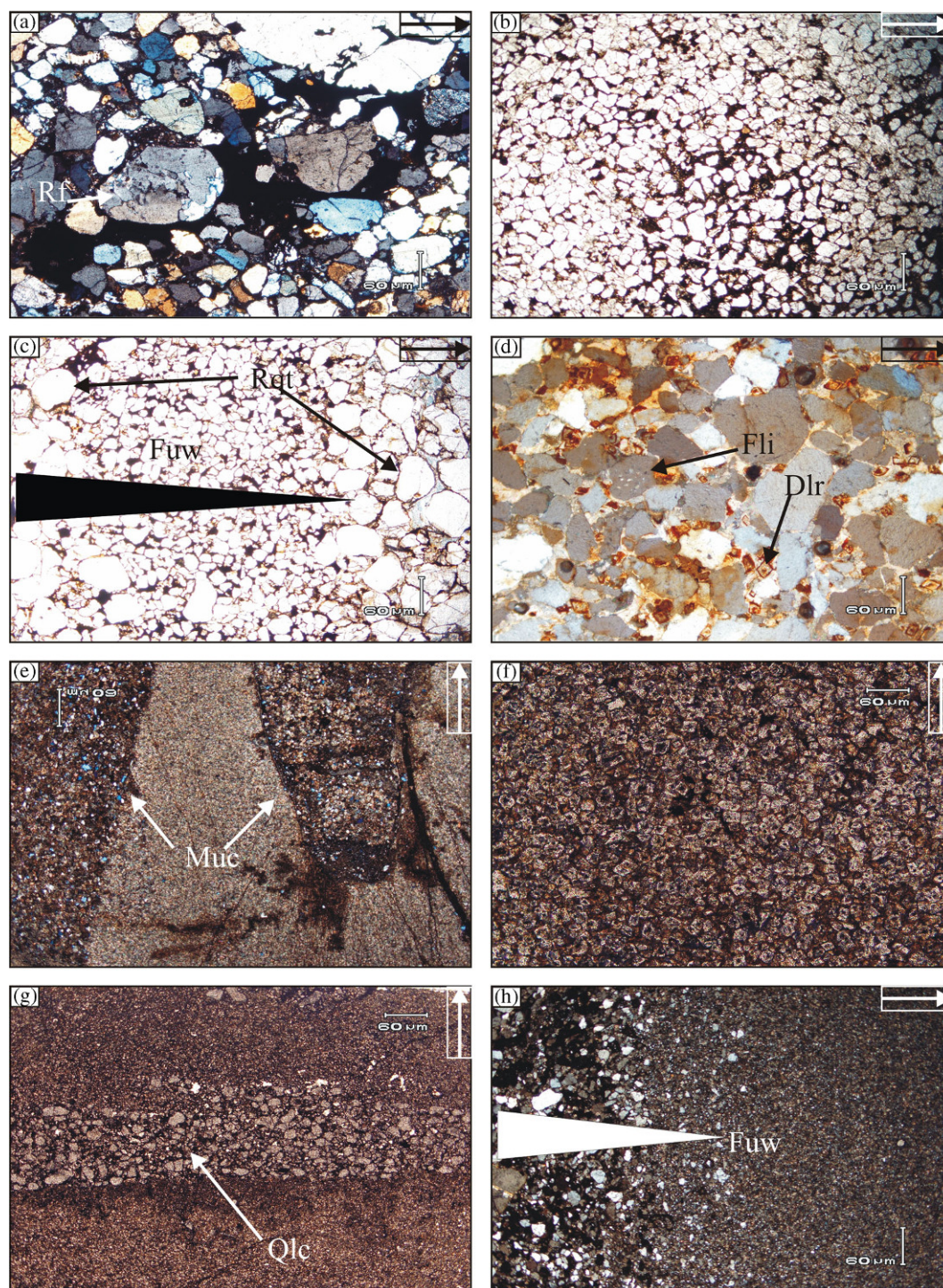


Figure 4. Representative photomicrographs of the lithofacies in the Datta Formation. The arrow in each case indicates the bed top. (a) Quart-rich conglomerate, (b) quartz-rich sandstone of CMSL cemented by iron oxide, (c) fining upward cycle in the quartz-rich sandstone of CBSL, (d) quartz-rich sandstone having dolomite rhombs in CBSL, (e) mudcracks in dolomite of LSL, (f) pure dolo-mudstone of LCL, (g) sharp base quartz-rich layer (pulse) in dolomite of LCL, and (h) fining upward clastics merging into carbonates in LSL. Rf: rock fragments, Rqt: rounded quartz, Fuw: fining upward, Fli: fluid inclusions, Dlr: dolomite rhombs, Muc: mudcracks, Qlc: quartz layer in carbonates.

that the lithofacies formed in a meandering stream. However, the upward decrease in the cross bedding in this lithofacies supports the deposition in a braided stream (Nichols 2000). In general, it is

difficult to assign this lithofacies to a braided or meandering river; however, as braiding can be a part of a large meander too, the most likely depositional setting is the point bar.

4.2 Channel belt sandstone lithofacies (CBSL)

Description. The lithofacies comprises of medium to thick and very thick bedded (Boggs 2006) quartz arenites (figures 3c–e and 4d), showing variable thickness (figure 2). The lithofacies occurs at several stratigraphic levels and shows a wide range of shades, i.e., light grey to dark grey, maroon, red and dark brown. The beds can be traced for few meters and lateral pinching/lenticular bedding (figure 3e) and lateral accretion is observed (Cadle and Cairncross 1993), especially in the lower part of the lithofacies (first occurrence of the lithofacies). Further erosive bedding plans/reactivation surfaces are observed in the beds and trough cross bedding and planar bedding plans (Boggs 2009) are more common in the upper part of the lithofacies. The sandstone is medium to coarse to very coarse-grained and some pebbly bases are also observed. An overall fining upward trend is observed in the lithofacies and the individual beds also display a fining upward trend. More than 90% of the quartz is moderate- to well-sorted and mostly rounded monocrystalline (e.g., figure 4b–d). Spherical concretions (figure 3c), ranging in diameter from 2 to 45 cm are common in the lower part of the lithofacies, which impart a red/maroon colouration to the lithofacies. These concretions are truncating the bedding plans and the bedding plans are not curved at their margins. Planar and trough cross bedding is observed in the lithofacies of which the trough cross bedding predominates. Large scale trough cross bedding (figure 3e) is common in the lithofacies, while planar cross bedding is also observable in the upper part.

The palaeocurrent direction is S–SE to N–NW (figure 5). Occasionally some ripples are found at or near the top of this lithofacies, which include bifurcating and symmetrical ripples (figure 3f) in

a ripple terrain (Reineck and Singh 1975; Boggs 2009). The absence of mudcracks, bioturbation and presence of very rare ripple marks differentiate this lithofacies from the overlying fine- to very fine-grained sandstones of the channel margin. Some scattered coal patches/carbonaceous material is also observed at the top of this lithofacies.

Interpretation. The large scale planar and trough cross beds are produced by the migration of megacurrent ripples (e.g., Reineck and Singh 1975). They indicate deeper depths of channels, i.e., more than 4.5 m (Freedman and Sanders 1978). Symmetrical ripples indicate relatively less sedimentation but pronounced reworking (Reineck and Singh 1975), conditions generally common in the upper flow regime during flood. Widespread ripple marks are generally present in the upper part of the bar sequences, natural levees, crevasse splays, and other flood plain deposits but are lacking here. The spherical concretions represent a diagenetic feature as the lamination/bedding planes of the sandstones are not inclined in the zones of concretions but are rather straight and undisturbed (Blatt 1982). An overall fining upward trend is present both in channel bar and point bar deposits, so is also cross bedding (Ferguson and Werritty 1983). In the present case, fining upward trend occurs at individual bed level, a feature generally not observed in channel bar deposits and common in point bars (Boggs 2009). Further trough cross beds that mostly characterize the lower part of the point bar sequence are observed (Massari 1983; Cadle and Cairncross 1993) and planar cross beds occur in the upper part of the lithofacies. However, the convolute bedding and slumping commonly observed in the upper part of channel bars are absent. On the other hand, the lenticular channelized bedding and lateral accretion observed in the lithofacies, is generally

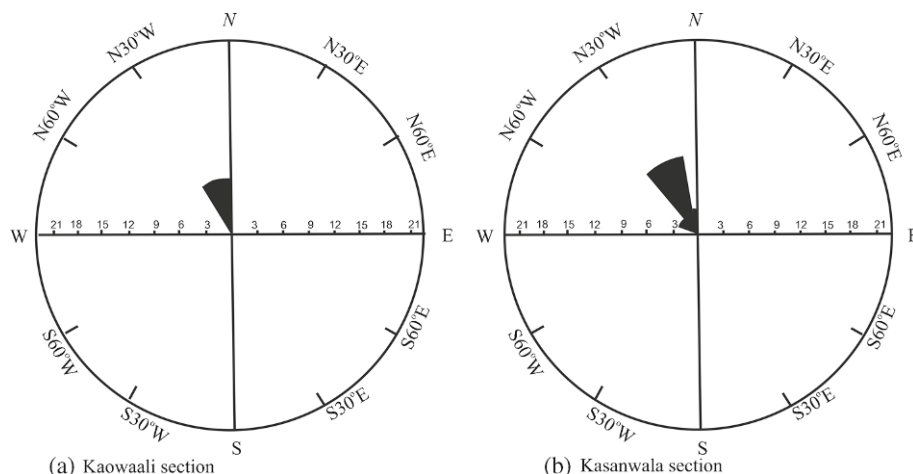


Figure 5. Palaeocurrent analysis of the Datta Formation in (a) Kaowaali and (b) Kasanwala sections. The readings on the E–W line indicate the number of data points.

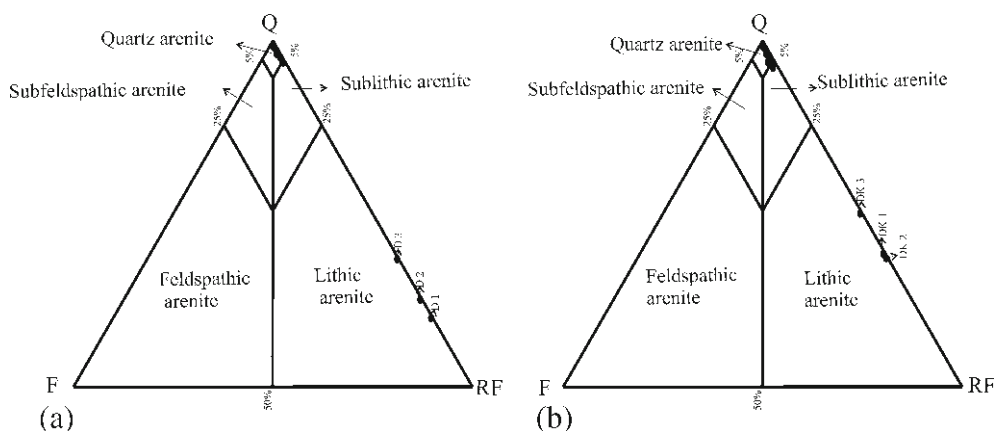


Figure 6. The QFR diagram of the Datta Formation and interpretation of the provenance. See figure 2 for key of samples; (a) Kaowaali section and (b) Kasanwala section (c.f. Pettijohn *et al.* 1973; Pettijohn 1974).

common in point bar deposits (Miall 1985; Willis 1989).

Based on the above discussion and relationship with the underlying CBCL, the lithofacies represents point bar deposition top of CBCL (Tucker 1986, 1992).

4.3 Channel margin sandstone lithofacies (CMSL)

Description. The lithofacies predominantly consists of thin-to-medium and occasionally thick-bedded sandstone. Individual beds can be traced in outcrop for few meters. Planar and wavy bedding planes and lateral pinching and lenticular bedding are observed. The sandstone is fine- to very fine-grained and pink, yellowish to dark brown (figure 3g). Fining upward cycles are represented and the sandstones grade first into siltstones and then to clay/shales. The associated shale is white, light grey, orange and occasionally dark brown to black in colour and shows lamination. Sedimentary structures include ripple marks, mudcracks, occasional bioturbation features, internal contortion, and raindrop imprints. Symmetrical, asymmetrical, and bifurcating ripples are observed (figure 3f). The symmetrical ripples have a wavelength of 10–15 cm and height of 3–5 cm. Absence of dominant large scale cross bedding, presence of ripple marks, mudcracks, bioturbation, raindrop imprints, and other similar features differentiate this lithofacies from the underlying CBSL lithofacies. Cross bedding is observed but the cross beds sets are thin as compared to those of the underlying CBSL lithofacies. A similar S–SE to N–NW palaeocurrent flow is indicated with some minor variations.

Petrographic and XRD studies suggest that the sandstones are composed of sub-angular to rounded and moderately-well sorted to very well-sorted quartz. More than 90% of the quartz are monocrystalline and have unit extinction. Iron oxide is the

common cementing material and most of the quartz grains float in iron oxide (haematite) and point contacts between the grains are the most (figure 4b). The grain margins are mostly smooth but corroded margins may also be observed. The sandstone is a quartz arenite (figure 6a and b). The quartz rich composition is also confirmed by the XRD analysis (figure 7).

Interpretation. The absence of large scale planar cross beds indicate the lacking of migrating mega-current ripples (e.g., Reineck and Singh 1975; Boggs 2006, 2009). Presence of symmetrical ripples indicates relatively less sedimentation and strong reworking (Reineck and Singh 1975; Cadle and Cairncross 1993). Such conditions are common in the upper flow regime during flooding in the upper part of the channel margin environments, natural levees, crevasse splays, and other flood plain deposits (Miall 1985; Willis 1989; Boggs 2009). The presence of mudcracks, bioturbation, and raindrop imprints suggest sub-aerial conditions (Remin *et al.* 2014). These inferences strengthen the deposition of this lithofacies in the channel margin near floodplain environments. The presence of associated channel belt facies of CBCL and CBSL, suggest a channel margin environment for this lithofacies with some effects of channel at the proximal extreme and floodplain environment at the distal extreme.

4.4 Overbank/floodplain lithofacies (OBFL)

Description. The lithofacies comprises of shale and siltstone with thin interbeds (3–10 cm) of sandstone (McKee and Weir 1953; Ingram 1954). This lithofacies is common in the middle and upper parts of the studied sections (figure 2), and is frequently underlain by channel margin sandstone

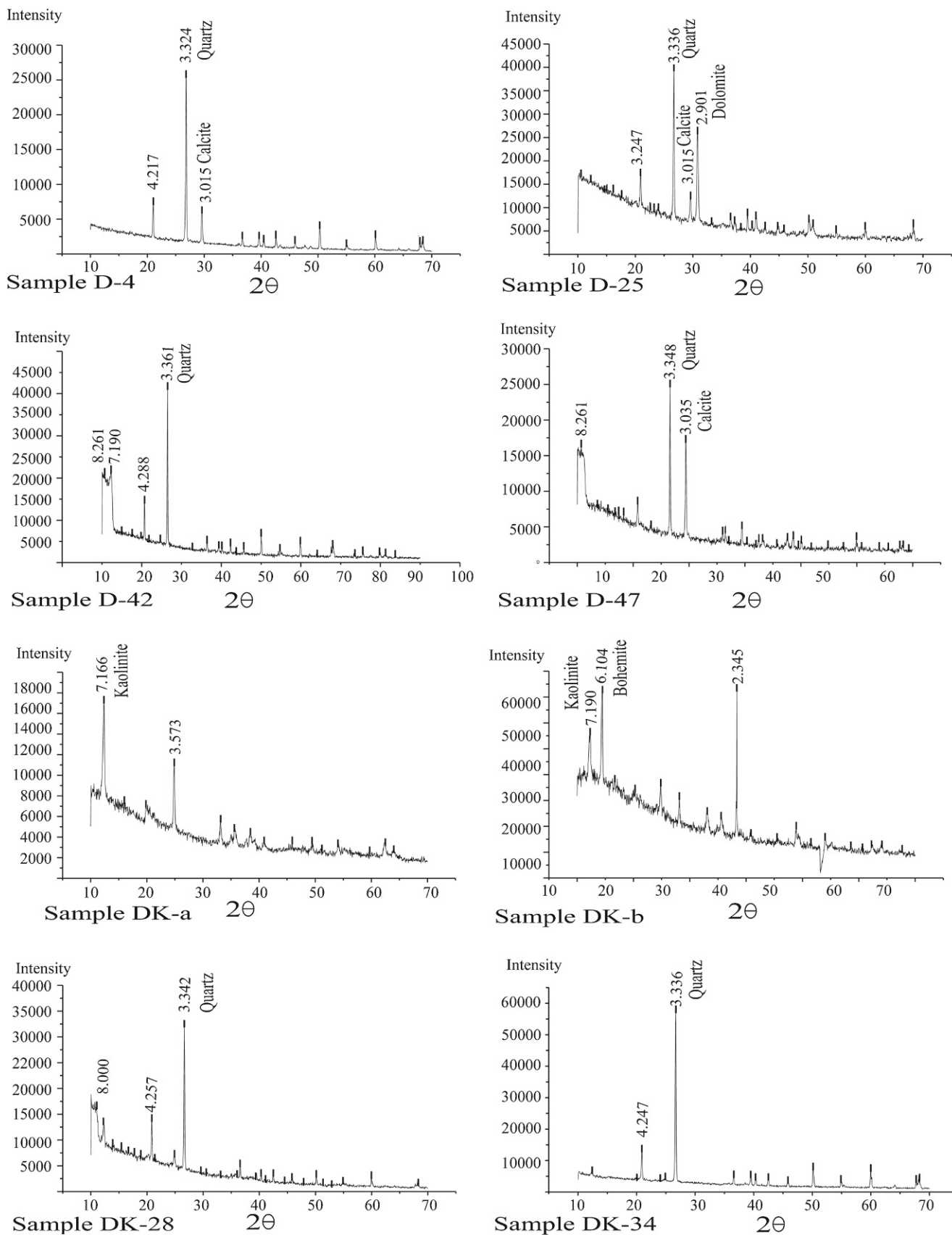


Figure 7. XRD analysis of the selected samples from the Datta Formation. Labelled peaks in each sample represent the dominant minerals in the samples.

lithofacies (CMSL) and only occasionally by channel belt lithofacies (CBSL). The white, light grey, pink and reddish brown shales are internally laminated.

Petrographic and XRD analyses indicated the presence of quartz only (figure 7). The siltstones and sandstones are varicoloured and thin- to thick-bedded. Oxidation effects are commonly observed in the siltstones and sandstones. The sedimentary structures observed are small scale cross-bedding (up to 10 cm), ripple marks, wavy bedding planes, and features interpreted as mudcracks. Climbing ripples and bioturbation features are also present. Plant remains are observed at the base of the sandstone beds. Oxidized surfaces with disintegration, dissolution, and features like mudcracks are common, showing palaeosols. The siltstones and associated shales contain deposits of fire clay (figure 2).

Interpretation. The fine-grained siliciclastic materials indicate deposition in low energy settings (Reineck and Singh 1975; Nichols 2000). The presence of fire clay indicates that parts of the lithofacies experienced weathering and soil forming processes (Tucker 1988, 1992). This also indicates a site of deposition that was not completely submerged but had undergone periodic exposure. Such conditions are usually observed in the fluvial channel margin and flood plain settings (Freedman and Sanders 1978) and also in the delta plain. The distinction between them is usually undertaken on the basis of the association of subenvironments that characterize the delta and the relationship to other facies in the stratigraphic section (Nichols 2000). The lithofacies commonly overlies the channel margin sandstone lithofacies (CMSL) and is occasionally associated with the channel belt sandstone lithofacies CBSL (figure 2). The presence of climbing ripples, plant remains, and other similar features suggest deposition in overbank/floodplain environments. The presence of rare fining upward trends in the associated CMSL supports the deltaic environment; the distributary channels (Nichols 2000). The lithofacies is overlain by lagoonal shale lithofacies, which also indicates the close relationship to deltaic/lagoonal subenvironments.

4.5 Lagoonal shale lithofacies (LSL)

Description. The lithofacies consists of predominant shale and subsidiary sandstone and siltstones. Minor, thin-bedded carbonates, 0.1–0.3 m thick, are also found associated with the shale (figure 2). The shale is carbonaceous, black and dark brown in colour in contrast to the white,

pink shale/clays of the OBFL but shows a similar quartz-rich composition. A 1–2.5 m thick coaly shale bed is observed in the middle part of this lithofacies. The associated sandstone/siltstones beds are light to dark grey and ash colour. Sharp basal bedding planes are observed and are fine- to very fine-grained.

Vertical and inclined burrows (figure 3h) are observed and mudcracks are also common. The associated sandstones/siltstones contain symmetrical ripples. The sandstones/siltstones show coarsening upward trend. Few planar cross beds of less than 10 cm in thickness are also observed at places.

The associated thin bedded carbonates are yellowish orange to light brown in colour and petrographic and XRD analyses show that most of the carbonates contain quartz grains. These can be termed as dolo-mudstones and arenaceous dolo-mudstones (Dunham 1962). Sharp based, gradually fining upward quartz pulses (figure 4g) are observed at many places and at places siliciclastic deposition gradually fining upward and finally merging into carbonate deposition is also observed (figure 4h).

Interpretation. The lithofacies is devoid of any fossils and thus hard to assign to an environment confidently. However, the associated lithologies, i.e., the interbedded sandstones and siltstones contain bioturbation, vertical, and inclined burrows. All the associated subenvironments of lagoons, including mudflats contain abundant bioturbation features, fauna, and organic material (Tucker 1992). The presence of coal supports deposition in a relative sea-level rise and flooding conditions, where poorly drained channels and flood plain are potential sites of such deposits (Emery and Myers 1996). The silty and muddy sediments can form on the lagoon bottom or in the low energy channels (Reineck and Singh 1975). A small scale fining upward trend may be present in the distributary channel. At the same time, the interdistributary bay contains thin sand/silt beds formed due to influx of sediment from the fluvial/delta top and symmetrical ripples due to wave action, but has a coarsening upward trend (Nichols 2000). Thus features of interdistributary bay are indicated in the present case.

Carbonate deposition favours settings of rather minor terrigenous input (Nichols 2000). However, the presence of thin layers of quartz in the carbonate (figure 4g) suggests some periodic influx of the siliciclastic grains. The fining upward of the siliclastics and finally merging into carbonate deposition (figure 4h) indicates facies mixing (Mount 1984), while fine grained siliclastics (shale) indicate deposition in less agitated environment, a condition generally observed in the interdistributary

bay (Nichols 2000). Source mixing (Mount 1984) may be a more appropriate reason for this change from shales to carbonates. The context of this lithofacies, i.e., association with OBFL (i.e., overbank/floodplain lithofacies) suggests a site having similar conditions as the delta plain or delta front environment. The shallow water environment inferred for the carbonates suggests a location not far from the land in an area where clastic input resulted in facies mixing in areas such as abandoned delta lobe or one of the interdistributary bay/lagoonal environments during a relative sea-level rise.

4.6 Lagoonal carbonate lithofacies (LCL)

Description. The lithofacies consists predominantly of carbonates with minor shale and very rare sandstone. The carbonates, mainly represented by yellowish brown dolomites, are commonly found associated with the shale and vary in thickness from 0.1–1.5 m. The petrographic and XRD studies show that most of the dolomites are pure (i.e., dolostones), while some contain significant amounts of quartz and impure carbonates (i.e., arenaceous dolo-mudstones; figures 4e–h and 7). The associated shale/clay is similar to that of LSL, but very minor sand/silt is associated in the present case. The sedimentary structures present in the sandstones are vertical to inclined burrows and some features interpreted as mudcracks also occur. The associated sandstones contain symmetrical ripples and show a fining upward trend.

Interpretation. Carbonate deposition, in general, favours settings of minor terrigenous input (Nichols 2000). However, the presence of shale and few quartz-rich sand beds (figure 4g) suggest some periodic siliciclastic influx. Thus facies mixing (Mount 1984) like the one discussed earlier is reported here. Most recently uplift and erosion of the underlying Late Triassic Kingriali Formation (carbonates) and carbonate clasts incorporation into the overlying beds of the Datta Formation along the Triassic–Jurassic boundary is reported in the area (Iqbal et al. 2014). A similar source mixing is also reported in the marginal marine deposits of the Mesozoic Gabriani Formation of Greece (Price 1977). The fining upward of the siliclastics indicates a waning siliclastic supply and calm deposition in less agitated carbonate environment, a condition generally observed in the interdistributary bay/lagoonal settings (figure 4h, Nichols 2000; Catuneanu 2006). The occurrence of this lithofacies in association with LSL suggests a site reflecting similar conditions as the delta plain or delta front environment. The shallow water environment

inferred for the carbonates suggests calm condition than that of LSL. Such an area could be delta plain where an abandoned delta lobe existed or on the interdistributary bay/lagoonal settings.

The bioturbation indicates a shallow, less agitated water environment of deposition for the lithofacies, while the occurrence of mudcracks represents subaerial exposure.

Seasonal/periodic sea-level fluctuation and/or sediment source mixing of the delta lobes were responsible for the deposition of the minor siliclastics.

5. Facies association

The distinctive lithofacies recognized are grouped into three associations. These facies associations belong to the delta channels, delta plain, and delta top subenvironments of the complex deltaic system. All the three facies associations are discussed in the following lines.

5.1 Channel belt facies association (CBFA)

The channel belt facies association (CBFA) comprises of channel belt conglomerate lithofacies (CBCL) and channel belt sandstone lithofacies (CBSL). This association primarily represents the basal part of the formation in both sections (figure 2) and is partly repeated at several levels up section (i.e., CBSL lithofacies). A general E–W increase in thickness is observed for this association. The association is typically recognized by the basal conglomerates of CBCL overlain by thick, cross-bedded sandstones of CBSL that contain prominent concretions (figure 3c). Mostly planar beds with planar bedding plans are observed, however erosional bedding plans are also observed especially at the base of the conglomerates. Cross bed sets are common in the sandstones. The association of CBCL and CBSL lithofacies suggests a bar deposit in the channel/distributary channel palaeoenvironment (Tucker 1992).

5.2 Channel margin and overbank facies association (CMOFA)

The channel margin and overbank facies association (CMOFA) comprises of the channel margin lithofacies (CMSL) and overbank/floodplain lithofacies (OBFL). This association overlies CBFA in the lower part of the formation and is overlain by the lagoonal facies association (LFA). In the middle and upper parts of the formation, it overlies LFA and is overlain by CBFA. The sandstones here are different to those of CBFA and contain mudcracks,

bioturbation features, and ripple marks which are not common in CBFA, while concretions which are common feature of CBFA are not present here. Similarly, trough cross bedding is lacking and planar cross bedding is observed at very few places. The sandstones are fine-grained as compared to those of CBFA.

5.3 Lagoonal facies association (LFA)

The lagoonal facies association (LFA) comprises of the lagoonal shale lithofacies (LSL) and lagoonal carbonate lithofacies (LCL). The shale in contrast to the oxidized shale of the CMOFA is mostly black, dark brown, and ash colour. Similarly, the sandstone/siltstones are thin-bedded, fine-grained and cross bedding is minor while mostly symmetrical ripples are observed. Bioturbation activity is also an important feature of the lithofacies.

6. Discussion

6.1 Palaeoenvironments and sequence stratigraphy of the Datta Formation

The Datta Formation represents an interval of massive siliciclastic deposition subsequent to Triassic dolomite deposition (e.g., Shah 1977). The presence of residual laterite zone indicates intense chemical weathering under humid tropical-near tropical conditions (Tucker 1992) and represents a candidate type 1 sequence boundary of higher hierarchical order (Posamentier *et al.* 1988; Posamentier and Vail 1988; Van Wagoner *et al.* 1990; Wood 1990; Wood *et al.* 1991; Westcott 1993; Emery and Myers 1996). The presence of boehmite, kaolinite, and other likewise minerals in the laterite zone (figure 7) indicate a warm, humid, tropical to near tropical climate at the onset of Datta deposition that prevailed throughout the formation (Abbasi *et al.* 2012).

6.1.1 Lower lowstand systems tract

The conglomerates (figures 2 and 8) of CBCL represent an abrupt juxtaposition of the fluvial/continental basal channel belt deposits on top of the underlying thick marine dolomites of the Triassic Kingriali Formation (figure 8), representing a facies dislocation in incised valley after the incised valleys are gradually filled by the fluvial channels during the slow relative sea-level rise stage of the lowstand system tract (Marzo *et al.* 1988; Van Wagoner *et al.* 1990). The presence of dolomitic fragments in the basal conglomerate bed indicate reworking and incorporation of the underlying/surrounding nearby Kingriali Formation, a

process in which a stream dissipates its energy by erosion of the channel banks, resulting in the widening of the stream channel and increasing width of the incised valley more than that of the fluvial channel itself. The width:thickness ratio is around 1:1000 that matches with the ratio for valley fills of Reynolds (1994) and is in contrast to his 1:100 ratio for channels.

The overlying (figure 2) channel belt sandstone lithofacies (CBSL) is characterized by medium- to coarse-grained trough cross-bedded sandstones with pebbly bases that is overlain by planar cross-bedded medium- to coarse-grained sandstone. This fining upward succession indicates a renewed alluvial aggradation, associated with initial slow relative sea-level rise, following a base level fall (figure 8). The relative sea-level rise rates are slightly higher compared to that of the basal conglomerates of the incised valley and thus deposit multistorey-multilateral sand bodies; however, these still represent the lower valley/channel fills (Emery and Myers 1996). These sands represent deposition in the relatively accelerated sea-level rise, during the lowstand systems tract (LST). Thin coal patches at the top of several beds within this unit indicate top of the episodic sea-level rise (Van Wagoner *et al.* 1990).

The overlying (figures 2 and 8) channel margin sandstones are fine- to medium-grained sheet sands having associated siltstone and shale/clays. These sediments indicate increasing relative sea-level rise and associated increase in the accommodation space resulting in decrease of the overall stream gradient and hence decrease in the fluvial capability to transport coarse sand and gravels. This resulted in an increase in the sinuosity of the channels, formation of meander channels and belts, and more effective subdivisions into channel and flood-plain components (Emery and Myers 1996). These fine-grained sands and associated silt and clays thus represent respective channel and channel margin sedimentation during accelerated relative sea-level rise and are candidates for marking top of the LST. The top of this unit probably marks a transgressive surface and hence, the top of the lowstand systems tract. Thus, the Datta Formation from its base (type 1 sequence boundary) up to this candidate transgressive surface represents lowstand systems tract (LST) deposits, which are 27-m thick in the north-northwest (Kaowaali section) and thins out to 19 m in the east-southeast (at Kasaanwala section).

6.1.2 Lower transgressive systems tract

The overbank/floodplain shale dominated unit overlies the thick LST succession. The lower boundary is interpreted as a transgressive surface (TS) and

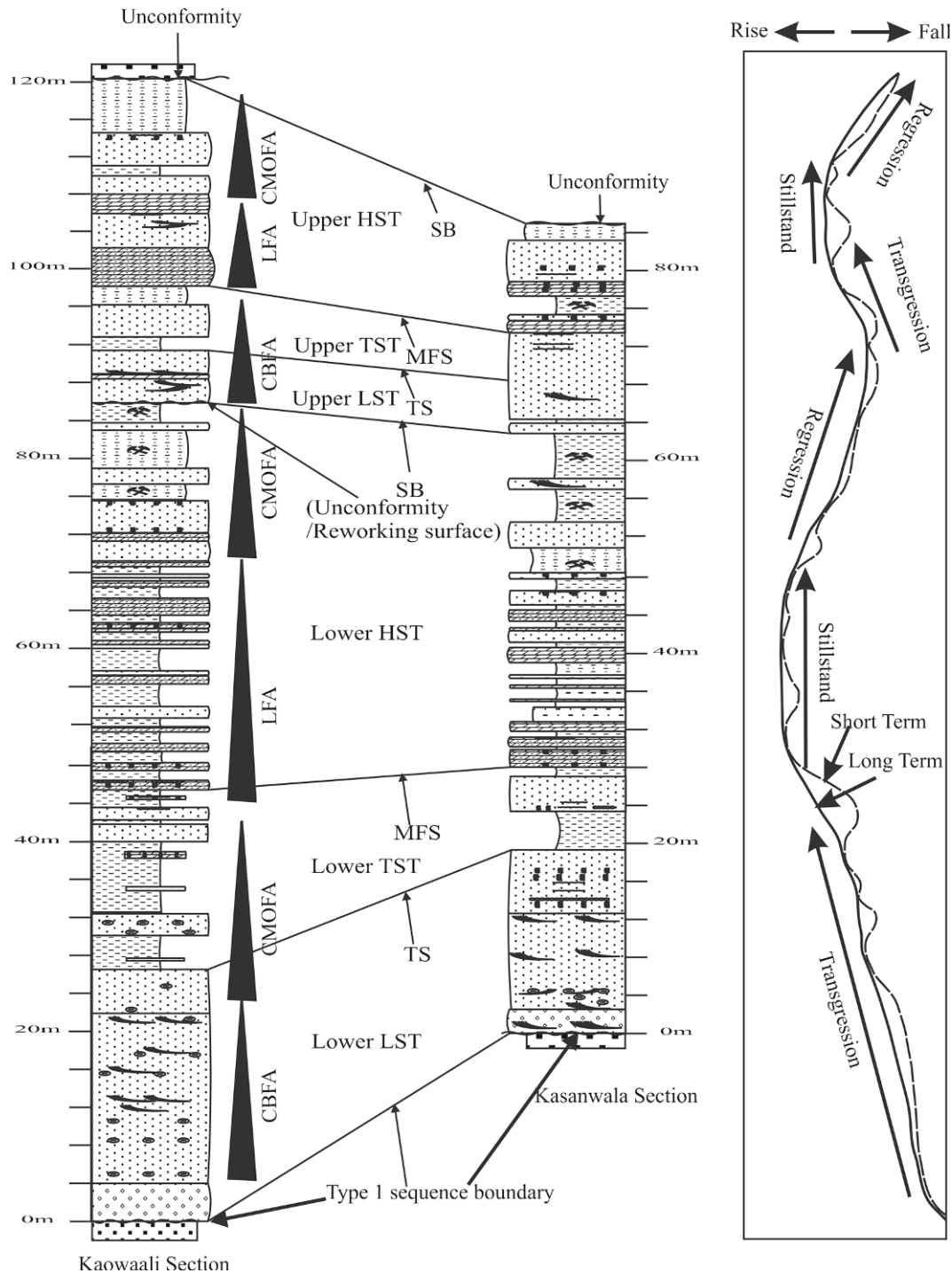


Figure 8. Distribution of different systems tracts and sequence boundaries. LST: Lowstand Systems Tract, TS: Transgressive surface, TST: Transgressive Systems Tract, MFS: Maximum Flooding Surface, HST: Highstand Systems Tract and SB: Sequence Boundary. A composite sea-level curve is suggested for both the sections in the right side of the figure. Please see figure 2 for key (modified after Haq *et al.* 1987).

the unit indicates a balance between sediment supply and relative sea-level rise. The dominance of shale/clays may indicate a reduction in the fluvial gradient as increase in the crevasse and similar channel margin deposits (figure 8) to promote an attempt to maintain grade with rising base level (Ryseth 1989; Tornqvist 1993). This unit

thus represents deposition in the basal part of transgressive systems tract (TST), following the transgressive surface.

The lagoonal shales/clays and the associated lagoonal carbonates (figure 8) indicate deposition in an accelerating relative sea-level rise. The succession indicates cyclic deepening and the presence of

possible flooding surfaces at the top of each cycle is indicated by the occurrence of shale. A tentative maximum flooding surface may be placed at the top of thickest shale unit which is 1–2.5 m and contains coal (Ryer 1981; Atkinson 1986; figure 8).

6.1.3 Lower highstand systems tract

Following the maximum flooding surface, deposition of the lagoonal carbonates and shale/clay continues and 4 m (Kasanwala section) to 10 m (Kaowaali section) of the same unit is deposited above the MFS. The uniform thickness of the interbedded shale/clays and carbonates (i.e., dolomites) indicate deposition in a relatively still stand of sea-level. This sequence is representing lower part of the highstand systems tract (HST), deposited following the MFS (Richards 1994). However, these facies show very close resemblance to the late TST deposits and differentiation between the two is difficult. The lower part of HST is overlain by fine-grained sandstones of the channel margin lithofacies marking a fall in the relative sea-level and the onset of channel margin/floodplain conditions. These in turn are overlain by the repetition of the overbank/floodplain lithofacies shale/clays marking a further drop in relative sea-level in the proximal locations during the upper part of the HST. However, it is noteworthy that the relative sea-level drop associated with later HST resulted in drop in the base level. The fluvial channel switching resulted in the establishment of fluvial channels and the overbank/floodplain suffered exposure (Shanley *et al.* 1993) and were subjected to humid, tropical–near tropical weathering. This is indicated by the presence of thick fire clay deposits in the upper, regressive part of the HST. The upper surface of the HST is marked by 10–30 cm pebbly sandstone bed with erosive base marking a sequence boundary and probably the onset of the next LST (Gowland and Riding 1991).

6.1.4 Upper lowstand systems tract

Overlying the HST is the sequence boundary and the next LST. The LST is characterized by the deposition of medium- to coarse-grained channel belt sandstones, while the basal channel conglomerates are absent. This LST appears to be associated with a minor relative sea-level fall than the one at the base of the formation. On the other hand, the possibility of the delta lobe switching, channel migration, and lateral accretion cannot be discarded as reasons for the onset of new fluvial conditions and deposition of channel belt sandstones on top of the HST deposits (Shanley *et al.* 1993).

Thus a candidate sequence boundary is marked below this LST rather than a proven one. The upper surface of this LST is marked in Kasanwala section by rippled, bioturbated fine-grained sandstone, whereas in Kaowaali section siltstone and shale/clay are present (figure 8). These mark the onset of new transgressive surface (TS) associated with new relative sea-level rise and/or delta lobes/channel switching. This is marked as second transgressive surface and boundary of the second LST.

6.1.5 Upper transgressive and highstand system tracts

The LST is again overlain by carbonates and shale/clays (figure 2) similar to those described earlier and are considered as deposits of the next but smaller TST. A tentative flooding surface is marked on top of the last dolomite beds in both the sections as black shale is not observed here. This may be the maximum flooding surface (MFS) and thus the boundary of the TST. This TST is finally overlain by the 12–14 m thick channel margin and overbank/floodplain facies, similar to those discussed in the earlier HST. This represents the second repetition of the HST.

The lateral channel variation, delta channel fluctuation may have played a dominant role in the distribution of various lithofacies in this upper part of the formation. The top of this HST is a major sequence boundary (Type 1 sequence boundary). This boundary marks the upper contact of the Datta Formation with the Paleocene Hangu Formation (figure 3b), a major unconformity along which the upper Jurassic and entire Cretaceous succession is missing (Shah 1977).

6.2 Palaeoclimate and sea-level changes

The Jurassic marks the time of significant breakup of Pangaea (Chandler *et al.* 1992; Moore *et al.* 1992; Valdes and Sellwood 1992; Chandler 1994; Fawcett *et al.* 1994; Valdes 1994). The associated northward drift of the Indian Plate during the Mesozoic into a warm-humid tropical to subtropical climatic zone at approximately 25°–30°S of the equator in the Early Jurassic, resulted in a substantial increase in humidity and rainfall compared to the hot and arid conditions in most of the Triassic period (Parrish *et al.* 1982; Ziegler 1982; Hallam 1984; Scotese 1994). This interplay between temperature and rainfall remained the main controlling factor of the Early Jurassic palaeoenvironments and sea level fluctuations (Hallam 1981, 1997, 2001). The study area, located at the northwestern margin of the Indian Plate evidenced likewise conditions. The warm and

humid conditions are represented at the unconformity below the formation characterized by residual laterite formation and presence of boehmite, kaolinite, and other similar minerals which are products of such conditions (Dera *et al.* 2009). The lower channel belt conglomerate facies and the adjoining channel belt sandstone lithofacies (figure 9a) are representative of deposition in warm, humid conditions and in a lowstand of relative sea-level. This indicates the most proximal deposition of the Datta Formation in an incised valley and associated meander channel belt (figure 9a and b). Minor relative sea-level fluctuations are indicated by the presence of coal patches at the top of many sandstone beds in this unit. These rocks are overlain by fine sandstones, siltstones and clays/shale indicative of warm tropical deltaic deposition in a rising stage of relative sea-level. These upsections merge into the lagoonal shale/clays and associated carbonates with features of warm, lagoonal, lake and near shore marine environments in a relative sea level rise conditions (maximum relative sea-level rise in the formation). This sequence is overlain by a number of palaeosols, laterites, and fire clay horizons, all indicative of warm and humid palaeoclimate under conditions of falling relative sea-level. Plant remains, bioturbation activity indicate moist conditions and water availability under warm conditions. The formation from its base up to this unit most probably represents a 3rd order depositional sequence that starts with the LST, merges into the TST, following a transgressive surface and HST deposition following the MFS (figure 8).

A distinct shift in the facies from overbank/floodplain deposits to more channel belt facies and disappearance of carbonates is observed on top of this unit. This indicates a second but minor episode of relative sea-level fall and onset of channel belt environment. Further, upsection in both the studied locations is overlain by similar carbonates and shale/clays and fine sandstone as discussed earlier. These features in the upper part of the formation indicate an interplay between sediment supply, and channel/delta lobes migration associated with very minor fluctuation in relative sea-level in proximal locations of the delta top/fluvial subenvironments (figure 9). This unit probably represents another 3rd order depositional cycle containing LST followed by a TST and HST (figure 8).

It is interpreted that the Jurassic time, as suggested by Hallam (1984) and in reconstructed rainfall maps for the Pliensbachian (Parrish *et al.* 1982), suggest that the climate was sufficiently uniform to establish a relatively uniform warm and humid palaeoclimate. Investigations of clay mineralogy (Dera *et al.* 2009) and palaeosols of the formation indicate a warm, humid climate (Graff-Petersen and Bondam 1963; Arndorff 1993).

6.3 Palaeo-depositional model and correlation

The present study suggests a tentative palaeo-depositional model for the Datta Formation in the study area based on the lithofacies distribution and palaeo-depositional environments interpretation (figure 9a and b). Based on the proportion of the various lithofacies/environmental indicators in the two sections, it is suggested that the Kasanwala section located in the S–SE of the study area represents more landward deposits whereas the Kaowaali section (located in the N–NW) represents a more basin ward part of the study area. An overall increase in thickness trend from S–SE (Kasanwala) to N–NW (Kaowaali) is observed (figure 2). A detail of the distribution of fluvio-continental lithofacies can be observed in figure 9(a). A fining upward vertical succession is elaborated that starts from coarse-grained Channel Belt Conglomerate Lithofacies (CBCL) overlain by Channel Belt Sandstone Lithofacies (CBSL) and Channel Margin Sandstone Lithofacies (CMSL). The first two lithofacies constitute the Channel Belt Facies Association (CBFA). This is overlain by Channel Margin and Over Bank Facies Association (CMOFA). This basal succession in both sections is interpreted as point bar deposits (figure 9a).

In the N–NW part of the study area, the presence of Lagoonal Shale Lithofacies (LSL) and Lagoonal Carbonate Lithofacies (LCL) is indicated (figure 9b). These two comprise the Lagoonal Facies Association (figure 9b) and presence of more basinal conditions with marine influence.

The palaeo-depositional model in general suggests that either the thick deposition occurred in N–NW part of the study area or the pre-Palaeocene erosion was more severe in the S–SE part. However, since the individual facies can be traced in both the sections and there is an overall decrease in the thickness of the individual lithofacies (figure 2), it is suggested that the first possibility appears to be a stronger reason for facies control of the Datta Formation in the area.

The Datta Formation is characterized by the occurrence of quartz-rich sands, fire clay and ferruginous horizons and is host to significant hydrocarbons in the Kohat–Potwar Plateau of north Pakistan (e.g., Kazmi and Abbas 2001). The formation provides excellent sections in the western Salt Range, Trans Indus ranges, and Kohat–Potwar–Hazara area for local and global correlation. Coarse-grained sandstones, conglomeratic at places and mixed with fine siliciclastics and some carbonates occur in the Kachh and Jaisalmer area of India (Gaetani and Garzanti 1991; Verma *et al.* 2012). The Datta Formation displays a good lithological and environmental correlation with the economically significant Lathi Formation of the

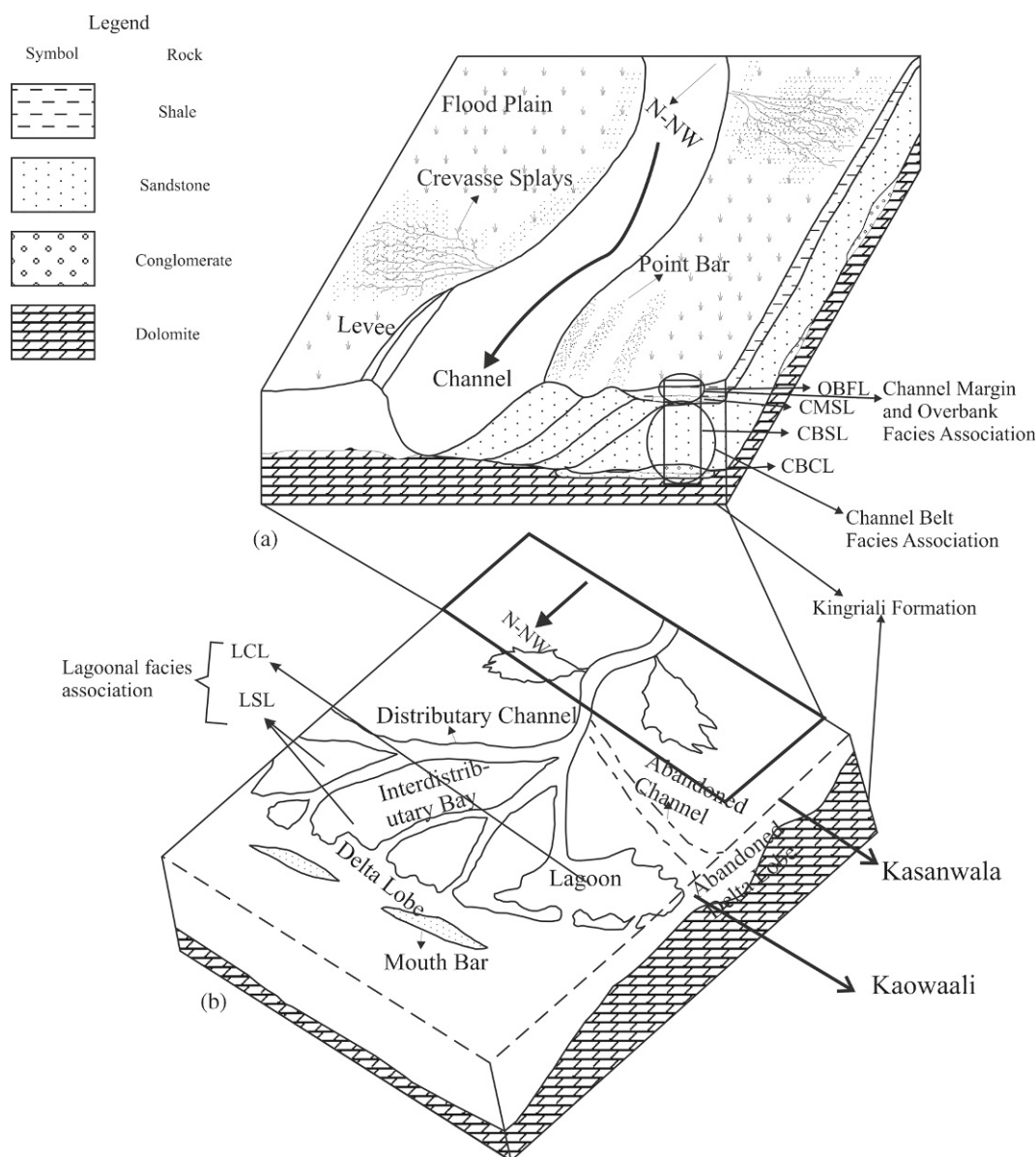


Figure 9. Palaeo-depositional schematic model for the Datta Formation. (a) Elaborated Channel Belt Facies Association consisting of Channel Belt Conglomerate Lithofacies (CBCL) and Channel Belt Sandstone Lithofacies (CBSL), Channel Margin and Overbank Facies Association (CMOFA) consisting of Channel Margin Sandstone Lithofacies (CMSL) and Overbank/Flood Plain Lithofacies (OBFL). (b) Composite model showing location of the two study sections and distribution of various subenvironments. Lagoonal Facies Association consisting of Lagoonal Shale Lithofacies (LSL) and Lagoonal Carbonate Lithofacies (LCL) are also marked.

adjoining India. However, robust biostratigraphic control is needed to establish this correlation and to establish the palaeogeography for this unit.

7. Conclusions

In the western Salt Range (Kaowaali and Kasanwala sections), the Datta Formation is represented by a thick siliciclastic unit having minor carbonates. The dominant lithologies include sandstone, shale/mud, dolomite (both pure and impure), siltstone and clast-supported conglomerate. Prior to

this work, no sedimentological studies have been conducted on the Datta Formation of this part of Salt Range.

The lower contact of the formation is disconformable with the Triassic Kingriali Formation and is marked by a residual laterite in both the sections. The basal conglomerate bed has angular fragments derived from the underlying Kingriali Formation. The Channel Belt Conglomerate (CBCL) is composed of pebble to granule and fine-grained particles and compositionally quartz rich oligomictic conglomerates. The Channel Belt Sandstones (CBSL) are medium to thick bedded,

fine- to coarse- and very coarse-grained and are quartz arenites. The carbonates include both pure and impure thin- to medium-bedded dolomite. The siltstone and shale/mud show laminations.

The rocks show colour variation from light grey, yellowish grey, to dark grey and even black. Sedimentary structures like cross- and graded-beddings, ripple marks, mudcracks and bioturbation feature are common. Apart from plant remains, no fossils were recognized. Planar and wavy bedding are the most common bedding types.

The Datta Formation is most probably deposited in fluvial-deltaic environments. Grading with good sorting and packing together with cross-bedding indicate that the lower half of the studied sections form part of an incised valley fill and prominent channel belt setting. A point bar succession is interpreted in the meander channel belt setting and consists of the channel belt facies association overlain by channel margin and overbank/flood plain facies association. Mixed lithologies of the lagoonal facies association in the upper half of the studied sections most likely represent lagoonal/lake-deltaic environment (delta plain-delta front). Apart from the channel belt conglomerate lithofacies, the other lithofacies are repeated at several intervals showing the repetition of the depositional cycles in the area.

Two main regression-transgression cycles are interpreted in the formation, of which the lower cycle indicates a more prominent relative sea-level regression and transgression following the Pre-Jurassic type-1 sequence boundary. Prominent LST, TST, and HST are recognized in this cycle. The overlying cycle is similar to the earlier cycle and tentative LST, TST, and HST are recognized. However, more influence of fluvial lateral and vertical migration and delta lobe switching associated with relative sea-level fluctuations is indicated during this time.

The presence of washed, clean silica sand, fire clay horizons, laterites, and numerous palaeosols indicate warm-humid near tropical palaeoclimate during the time of deposition of the Datta Formation.

Acknowledgements

This study benefited from the Quaid-i-Azam University 'University Research Fund' provided to Mr Shahid Iqbal. NCE in Geology is thanked for providing field support. Authors also thank Dr Muhammad Mujtaba (HDIP), Dr Muhammad Hanif (NCEG) and Prof. Michael Wagreich, Department of Geodynamics and Sedimentology (University of Vienna), for their guidance in this work.

References

- Abbasi A I, Haneef M, Obaid S, Daud F and Quraishi A W 2012 I. A Mesozoic deltaic system along the western margin of the Indian Plate: Lithofacies and depositional setting of Datta Formation, north Pakistan; *Arab. J. Geosci.* **5** 471–480.
- Arndorff L 1993 Lateral relations of deltaic paleosols from the lower Jurassic Rønne Formation on the Island of Bornholm, Denmark; *Palaeogeogr. Palaeoclimatol. Palaeoecol.* **100** 235–250.
- Arias C 2007 Pliensbachian–Toarcian ostracod biogeography in NW Europe: Evidence for watermass structure evolution; *Palaeogeogr. Palaeoclimatol. Palaeoecol.* **251** 398–421.
- Ashraf M, Ahmad M and Faruqi F A 1976 Jurassic bauxite and kaolinite deposits of Chhori area Kala Chitta Range, Punjab, Pakistan; *Geol. Bull. Punjab Univ.* **41** 41–54.
- Atkinson C D 1986 Tectonic controls on alluvial sedimentation as resolved by an ancient catena in the Capella Formation (Eocene) of northern Spain; In: *Paleosols: Their Reconstruction and Interpretation* (ed.) Wright V P, Princeton Univ. Press, Princeton, NJ, pp. 139–179.
- Baker D M, Lillie R J, Yeats R S, Johnson G D, Yousuf M and Zamin A S H 1988 Development of the Himalayan frontal thrust zone, Salt Range, Pakistan; *Geology* **16** 3–7.
- Boggs S Jr 2009 *Petrology of Sedimentary Rocks*; 2nd edn, Cambridge University Press, pp. 166–193.
- Boggs S Jr 2006 *Principles of Sedimentology and Stratigraphy*; 4th edn; Prentice Hall, Upper Saddle River, NJ.
- Blatt H 1982 *Sedimentary Petrology*; Freeman and Company, New York, pp. 105–241.
- Cadle A B and Cairncross B 1993 A sandy, bed-load dominated fluvial system deposited by lateral-accretion: Permian Karoo Sequence, South Africa; *Sedim. Geol.* **85**(1) 435–455.
- Catuneanu O 2006 *Principles of Sequence Stratigraphy*; Elsevier, pp. 21–24.
- Chandler M A 1994 Depiction of modern and Pangean deserts: Evaluation of GCM hydrological diagnostics for paleoclimate studies; In: *Pangaea: Paleoclimate, tectonics, and sedimentation during accretion, zenith, and breakup of a supercontinent* (ed.) Klein G D, Boulder, CO; *Geol. Soc. Am. Spec. Paper* **288** 117–138.
- Chandler M A, Rind D and Ruedy R 1992 Pangean climate during the Early Jurassic: GCM simulations and the sedimentary record of paleoclimate; *Geol. Soc. Am. Bull.* **104** 543–559.
- Chaudhry M N, Chuhan F A and Ghazanfar M 1994 Microfacies, diagenesis, environment of deposition and burial history of Datta Formation from Bara Oter, Dist. Abbottabad; *Kash. J. Geol.* **11** 43–59.
- Chaudhry M N, Manzoor A and Ghazanfar M 1994 Sedimentology of Datta Formation at Kalapani, Abbottabad, northwest Himalayas; *Pakistan. Geol. Bull. Punjab Univ.* **89** 11–20.
- Damborenea E S 2002 Jurassic evolution of Southern Hemisphere marine palaeobiogeographic units based on benthonic bivalves; *Geobios Mémoire Spécial n°24* 51–71.
- Dera G, Pellenard P, Neige P, Deconinck J F, Pucéat E and Dommergues J F 2009 Distribution of clay minerals in Early Jurassic Peritethyan seas: Palaeoclimatic significance inferred from multiproxy comparisons; *Palaeogeogr. Palaeoclimatol. Palaeoecol.* **271** 39–51.
- Dunham R J 1962 Classification of carbonate rocks according to depositional texture; In: *Classification of carbonate rocks* (ed.) Ham W E, *Am. Assoc. Petrol. Geol. Memoir* **1** 108–121.

- Emery D and Myers K 1996 *Sequence Stratigraphy*; Blackwell Sci. Publ. **11**–**45** 111–178.
- Fatmi A N 1974 Lithostratigraphic units of the Kohat–Potwar Province, Indus Basin, Pakistan; *Geol. Surv. Pakistan Memoir* **10** 6–65.
- Fatmi A N 1977 Mesozoic; In: *Stratigraphy of Pakistan* (ed.) Shah S M I, pp. 29–44.
- Fatmi A N, Anwar M, Hussain I, Saeed G and Latif A 1990 Stratigraphy of the Shinawari and Datta formations from the Shinawari area (western Samana Range), Kohat Dist; *Geol. Bull. Punjab Univ.* **84** 47–55.
- Fawcett P J, Barron E J, Robison V D and Katz B J 1994 The climatic evolution of India and Australia from the Late Permian to mid-Jurassic: A comparison of climate model results with the geological record; In: Pangaea: Paleoclimate, tectonics, and sedimentation during accretion, zenith, and breakup of a supercontinent (ed.) Klein G D, Boulder, CO *Geol. Soc. Am. Spec. Pap.* **288** 139–157.
- Folk R L 1974 *Petrology of Sedimentary Rocks*; Hemphill Publishing Co., Austin, TX 28.
- Freedman M G and Sanders E J 1978 *Principles of Sedimentology*; John Wiley and Sons, New York, Chichester, Brisbane, Toronto, pp. 25–54, 100–130, 190–260.
- Ferguson I R and Werritty A 1983 Bar development and channel changes in gravelly River Feshie, Scotland; *Spec. Publ. Int. Assoc. Sedim.* **6** 181–193.
- Gaetani M and Garzanti E 1991 Multicyclic history of the northern India continental margin (northwestern Himalaya); *Am. Assoc. Petrol. Geol. Bull.* **75** 1427–1446.
- Gee E R 1980 Pakistan geological Salt Range series: Directorate of overseas surveys, United Kingdom, for Govt. of Pakistan and Geol. Survey Pakistan, 6 sheets, scale 1:50,000.
- Gee E R 1989 Overview of the geology and structure of the Salt Range, with observations on related areas of north Pakistan; In: Tectonics of the western Himalayas (eds) Malinconi Co L L and Lillie R J, *Geol. Soc. Am. Spec. Publ.* **232** 95–112.
- Gowland S and Riding J B 1991 Stratigraphy, sedimentology and paleontology of the Scarborough Formation (Middle Jurassic) at Hundale Point, North Yorkshire; *Proc. Yorkshire Geol. Soc.* **48** 375–392.
- Graff-Petersen P and Bondam J 1963 Hasle Klinkerfabrik Clay Pit; Int. Clay Conf. Field Trip C, pp. 1–14.
- Hallam A 1981 A revised sea-level curve for the early Jurassic; *J. Geol. Soc. London* **138** 735–743.
- Hallam A 1984 Continental humid and arid zones during the Jurassic and Cretaceous; *Palaeogeogr. Palaeoclimatol. Palaeoecol.* **47** 195–223.
- Hallam A 1997 Estimates of the amount and rate of sea-level changes across the Rhaetian–Hettangian and Pliensbachian–Toarcian boundaries (latest Triassic to early Jurassic); *J. Geol. Soc. London* **154** 773–779.
- Hallam A 2001 A review of the broad pattern of Jurassic sea-level changes and their possible causes in the light of current knowledge; *Palaeogeogr. Palaeoclimatol. Palaeoecol.* **167** 23–37.
- Haq B U, Hardenbol J and Vail P R 1987 Chronology of fluctuating sea levels since the Triassic (250 million years ago to present); *Science* **235** 1156–1167.
- Ingram R L 1954 Terminology for the thickness of stratification and parting units in sedimentary rocks; *Geol. Soc. Am. Bull.* **65** 937.
- Iqbal S, Jan U I and Hanif M 2014 The Mianwali and Tredian formations: An example of the Triassic progradational deltaic system in the low-latitude western Salt Range, Pakistan; *Arab J. Sci. Engg.* 5489–5507, doi: 10.1007/s13369-013-0836-2.
- Iqbal S, Jan I U, Akhter G, Wagreich M and Hanif M 2014 Triassic–Jurassic boundary: Evidences from the Tethyan Salt Range, Pakistan and correlation with Europe; *J. Himalayan Earth Sci.* (Abstract volume; Earth Sciences Pakistan, 2014), pp. 36–37.
- Kadri I B 1995 *Petroleum Geology of Pakistan*; Pakistan Petroleum Limited, Graphic Publ. Karachi, pp. 40–45, 84–92.
- Kazmi A H and Jan M Q 1997 *Geology and Tectonics of Pakistan*; Graphic Publ. Karachi, pp. 25–77, 122–223.
- Kazmi A H and Abbas S G 2001 *Metallogeny and Mineral Deposits of Pakistan*; Graphic Publ. Karachi, pp. 207–215.
- Lunt I A, Smith G H S, Best J L, Ashworth P J, Lane S N and Simpson C J 2013 Deposits of the sandy braided South Saskatchewan River: Implications for the use of modern analogs in reconstructing channel dimensions in reservoir characterization; *AAPG Bull.* **97**(4) 553–576.
- McKee E D and Weir G W 1953 Terminology for stratification and cross-stratification in sedimentary rocks; *Geol. Soc. Am. Bull.* **64** 383.
- Marzo M, Nijman W and Puigdefabregas C 1988 Architecture of the Castissent fluvial sheet sandstones, Eocene, South Pyrenees, Spain; *Sedimentology* **35** 881–913.
- Massari F 1983 Tabular cross bedding in Messinian fluvial channel conglomerates, South Alps, Italy; *Int. Assoc. Sedim. Spec. Publ.* **6** 287–300.
- Miall D A 1985 Architectural-element analysis: A method of facies analysis applied to fluvial deposits; *Earth Syst. Rev.* **22**(4) 261–308.
- Moore G T, Hayashida D N, Ross C A and Jacobson S R 1992 Paleoclimate of the Kimmeridgian/Tithonian (Late Jurassic) world. I. Results using a general circulation model; *Palaeogeogr. Palaeoclimatol. Palaeoecol.* **93** 113–150.
- Mount J F 1984 Mixing of siliciclastic and carbonate sediments in shallow shelf environments; *Geology* **12**(7) 432–435.
- Mujtaba M and Memon A M 1990 Diagenesis and pore space development of Datta Sandstones in Salt Range and Surghar Range, Pakistan; *J. Hydrol. Res.* **2** 57–66.
- Mukhopadhyay G, Mukhopadhyay S K, Roychowdhury M and Parui P K 2010 Stratigraphic correlation between different Gondwana Basins of India; *J. Geol. Soc. India* **76** 251–266.
- Murris R J 1980 Middle East stratigraphic evolution and oil habitat; *Am. Assoc. Petrol. Geol. Bull.* **64** 597–618.
- Nichols G 2000 *Sedimentology and Stratigraphy*; Blackwell Science, pp. 75–145.
- Obaid S, Qureshi W A and Abbasi I A 2005 Lithofacies, sand-bodies geometry and depositional setting of the Datta Formation in Surghar Range, North Pakistan; SPE/PAPG Tech. Conf. Islamabad, pp. 113–126.
- Parrish J T, Ziegler A M and Scotese C R 1982 Rainfall patterns and the distribution of coals and evaporites in the Mesozoic and Cenozoic; *Palaeogeogr. Palaeoclimatol. Palaeoecol.* **40** 67–101.
- Pettijohn J F 1974 *Sedimentary Rocks*; Harper and Row Publ. New York, Evanston, San Francisco, and London, pp. 195–246, 316–380.
- Pettijohn F J 1975 *Sedimentary Rocks*; 3rd edn, Harper and Row, New York, NY, pp. 157–166.
- Pettijohn F J, Potter P E and Siever R 1973 Sand and sandstone, Springer-Verlag, Berlin, 617p.
- Posamentier H W and Vail P R 1988 Eustatic controls on clastic deposition. I. Sequences and systems tract models; In: Sea-level changes: An integrated approach (eds) Waliguis C K, Hastings B S, Kendall C G St. C, Posamentier H W, Ross C A and Van Wagoner J C, *Soc. Econ. Paleontol. Miner. Tulsa. USA Spec. Publ.* **42** 109–124.

- Posamentier H W, Jervey M T and Vail P R 1988 Eustatic controls on clastic deposition. II. Computational framework; In: *Sea-level changes: An integrated approach* (eds) Waliguis C K, Hastings B S, Kendall C G St. C, Posamentier H W, Ross C A and Van Wagoner J C, *Soc. Econ. Paleontol. Miner. Tulsa. USA Spec. Publ.* **42** 125–154.
- Price I 1977 Deposition and derivation of clastic carbonates on a Mesozoic continental margin, Orthis, Greece; *Sedimentology* **24** 529–546.
- Rafique M, Ashraf M, Din A and Faruqi F A 1976 Evaluation of glass sand deposits of Pezu, Bannu Dist. N.W.F.P.; *Geol. Bull. Punjab Univ.* **45** 87–92.
- Rees P M, Gibbs M T, Ziegler A M, Kutzbach J E and Behling P J 1999 Permian climates: Evaluating model predictions using global paleobotanical data; *Geology* **27** 891–894.
- Rees P M, Ziegler A M and Valdes P J 2000 Jurassic phyto-geography and climates: New data and model comparisons; In: *Warm climates in Earth history* (eds) Huber B T, Macleod K G and Wing S L, Cambridge University Press, pp. 297–318.
- Rees P M, Ziegler A M, Gibbs M T, Kutzbach J E, Behling P J and Rowley D B 2002 Permian phytogeographic patterns and climate data/model comparisons; *J. Geol.* **110** 1–31.
- Reineck E H and Singh B I 1975 *Depositional Sedimentary Environments*; Springer-Verlag, Berlin: Heidelberg, New York, pp. 21–32, 41–44, 81–147, 225–279.
- Remin Z, Krogulec T, Drela T and Surowski M 2014 The recognition of hailstone impressions in clay-rich sediment: Experimental results and relation to the neoproterozoic case; *J. Sedim. Res.* **84** 543–551.
- Reynolds A D 1994 Sequence stratigraphy and the dimensions of paralic sandstone bodies; In: *High resolution sequence stratigraphy: Innovations and applications* (ed.) Johnson S D, Univ. Liverpool, pp. 69–72.
- Richards M T 1994 Transgression of an estuarine channel-tidal flat complex: The Lower Trias of Barles, Alpes de Haute Provence; *Sedimentology* **42** 1–27.
- Ryer T A 1981 Deltaic controls on Ferron Sandstone Member of Mancos Shale: Predictive model for Cretaceous coal-bearing strata of western interior; *Am. Assoc. Petrol. Geol. Bull.* **65** 2323–2340.
- Ryseth A 1989 Correlation of depositional pattern in the Ness Formation, Oseberg area; In: *Correlation in Hydrocarbons Exploration* (ed.) Collinson J D, Graham & Trotman, London, pp. 313–326.
- Scotese C R 1994 Early and Middle Jurassic maps; In: *Pangaea: Paleoclimate, tectonics, and sedimentation during accretion, zenith, and breakup of a supercontinent* (ed.) Klein G H, *Geol. Soc. Am. Spec. Pap.* **288** 57–73.
- Shah I M S 1977 Stratigraphy of Pakistan; *Geol. Surv. Pakistan Quetta Memoir* **12** 3–4, 6–8, 23–29, 40–43.
- Shah S M I 2009 Stratigraphy of Pakistan; Govt. Pak Ministry of Petrol & Nat. Resour. Geol. Sur. Pakistan, pp. 245–273.
- Shanley K W, McCabe P J and Hettinger R D 1993 Tidal influence in Cretaceous fluvial strata from Utah, USA: A key to sequence stratigraphic interpretation; *Sedimentology* **39** 905–930.
- Smith G H S, Ashworth P J, Best J L, Lunt I A, Orfeo O and Parsons D R 2009 The sedimentology and alluvial architecture of a large braid bar, Río Paraná, Argentina; *J. Sedim. Res.* **79**(8) 629–642.
- Smith G H S, Best J L, Bristow C S and G E Petts (eds) 2006 Braided rivers: Process, deposits, ecology and management; *Int. Assoc. Sedim. Spec. Publ.* **36**.
- Tornqvist T E 1993 Holocene alteration of meandering and anastomosing fluvial systems in the Rhine–Meuse delta (central Netherlands) controlled by sea level rise.
- Tucker E M 1986 *Sedimentary Petrology: An Introduction*; Blackwell Sci. Publ., pp. 10–63, 96–158.
- Tucker E M 1988 *Techniques in Sedimentology*; Blackwell Sci. Publ., pp. 91–223.
- Tucker E M 1992 *Sedimentary petrology: An introduction to the origin of sedimentary rocks*; Blackwell Sci. Publ., pp. 10–65, 125–154.
- Van Wagoner J C, Mithcum R M Jr, Campion K M and Rahmanian V D 1990 Siliciclastic sequence stratigraphy in well logs, cores and outcrop: Concepts for high resolution correlation of time and facies; *Am. Assoc. Petrol. Geol. Methods in Exploration Series, Tulsa.* **7** 55.
- Valdes P J 1994 Atmospheric general circulation models of the Jurassic; In: *Palaeoclimates and their modeling with special reference to the Mesozoic Era* (eds) Allen J R L, Hoskins B J, Sellwood B W, Spicer R A and Valdes P J (London: Chapman and Hall), pp. 109–118.
- Valdes P J and Sellwood B W 1992 A palaeoclimate model of the Kimmeridgian; *Palaeogeogr. Palaeoclimatol. Palaeoecol.* **95** 47–72.
- Verma D, Jadhav G N, Biswal T K, Jena S K and Sharma N 2012 Characterization of hydrocarbon-bearing fluid inclusion in sandstones of Jaisalmer basin, Rajasthan: A preliminary approach; *J. Geol. Soc. India* **80** 505–514.
- Westcott W A 1993 Geomorphic thresholds and complex response of fluvial systems – some implications of sequence stratigraphy; *Am. Assoc. Petrol. Geol. Bull.* (abstract) **77** 208–218.
- Willis B J 1989 Palaeochannel reconstructions from point bar deposits: A three dimensional perspective; *Sedimentology* **36**(5) 757–766.
- Wood L J 1990 Effects of basin type on coastal plain-shelf slope systems during base level fluctuations: An experimental approach; *Am. Assoc. Petrol. Geol. Bull.* (abstract) **75** 969.
- Wood L J, Ethridge F G and Schumm S A 1991 Influence of subaqueous shelf angle on coastal plain-shelf slope deposits resulting from a rise or fall in base level; *Am. Assoc. Petrol. Geol. Bull.* (abstract) **75** 969.
- Ziegler P A 1982 *Geological atlas of western and central Europe*, 130 (Hague: Elsevier for Shell Internationale Petroleum Maatschappij).

MS received 30 April 2014; revised 3 October 2014; accepted 20 December 2014

CHAPTER 5

PALAEOCLIMATIC TRANSITION AND THE TRIASSIC–JURASSIC BOUNDARY

Multi-proxy data sets from the upper part of the Kingriali Formation and the entire Datta Formation were used to interpret the Triassic–Jurassic palaeoclimatic transition in the western Salt Range, Surghar Range and Khisor Range (Iqbal et al., 2018, under review).

A southern hemisphere record of climate change during the Triassic–Jurassic transition: evidence for hothouse climate from the Salt Range, Pakistan.

Iqbal, S., Wagreich, M., Jan, I. U., Kuerschner, W. M., and Gier, S. (2018) Global and Planetary Change (GLOPLACHA_2018_92 under review).

Shahid Iqbal wrote the article and was the corresponding author. The studied sections were selected through mutual discussion among Shahid Iqbal, Michael Wagreich, Irfanullah Jan and Michael Wolfram Kuerschner. Shahid Iqbal and Irfanullah Jan did the detailed fieldwork and sample collection. Shahid Iqbal prepared samples for thin sections, bulk mineralogy (XRD), bulk rock geochemistry and clay mineralogy. Michael Wagreich wrote parts of the introduction, driving mechanisms and global correlation and also edited the manuscript. Michael Wolfram Kuerschner revised the interpretation and correlation of the Triassic–Jurassic boundary interval. Susanne Gier helped in clay minerals preparation, processing and quantifying the clay minerals samples. Shahid Iqbal studied the thin sections, prepared all the figure and maps and interpreted all the data sets. Shahid Iqbal prepared while Michael Wagreich and Susanne Gier revised the summary figure.

A southern hemisphere record of climate change during the Triassic–Jurassic transition: evidence for hothouse climate from the Salt Range, Pakistan

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Abstract

The Late Triassic is generally considered as a time of warm semiarid to arid palaeoclimate following the Carnian Pluvial Event in the dominantly arid Triassic Period. The Triassic–Jurassic boundary interval was characterised by the change from warm, semiarid–arid to a hot and humid climate in the Tethyan domain linked to input of greenhouse gases from the Central Atlantic Magmatic Province (CAMP) activity and Pangaea breakup. This study provides the very first outcrop evidences of palaeoclimatic evolution during the Triassic–Jurassic boundary interval in the then southern hemisphere, along the eastern margin of Gondwana facing the western Tethys. In the Tethyan Salt Range of Pakistan a succession of Upper Triassic dolomites, green-black shales (Kingriali Formation) to overlying Lower Jurassic quartzose sandstones, shales, laterites and conglomerates (Datta Formation) provides information on the palaeoclimatic evolution of the area in this critical interval. Bulk and clay mineralogy of the Upper Triassic shales indicate the presence of mainly illite while kaolinite is a minor component. The kaolinite content, a reflection of the mature stage of chemical weathering and hence hot–humid conditions, increases up-section in the overlying shales and sandstone–shale succession. The following laterite–bauxite horizons lack illite

and are entirely composed of kaolinite, boehmite and haematite. The bulk rock geochemistry of the succession confirms a similar trend. The Chemical Index of Alteration (CIA_{molar}) displays an increasing trend from the Upper Triassic (CIA 68-80) to the overlying Lower Jurassic strata (CIA 90-97). The overall results for the succession reveal an increasing chemical maturity trend from Rhaetian to Hettangian thereby supporting a change from warm-arid to a hot and humid palaeoclimate, probably extreme greenhouse conditions. Similar changes in the clay mineralogy and sediment geochemistry across the Triassic–Jurassic boundary have been reported from basins across Europe. The Salt Range serves as key area to understand the southern hemisphere palaeoclimatic and palaeogeographic development across the Triassic–Jurassic boundary.

Keywords: Greenhouse; Triassic–Jurassic boundary; Pakistan; Geochemistry; Clay mineralogy

1. Introduction

The Phanerozoic palaeoclimatic fluctuations (Poulsen, 2009) had a variety of temporal scales and a wide range of climatic states, with “greenhouse” and “icehouse” phases, and hothouse and glacials as extreme end members (Kidder and Worsley, 2010). Climate proxy data (e.g. Frakes, 1979; Crowley and North, 1991; Frakes et al., 1992; Allen et al., 1994; Valdes et al., 1999; Huber et al., 2000) over the past 500 Ma, provide vast evidences suggesting that the Earth has been generally warmer than at present.

The Mesozoic Earth, in comparison to the present day, was warmer by 6 °C or more (Sellwood and Valdes, 2006). Modeling of the Late Triassic evaporation and precipitation data indicate a warm semiarid to arid palaeoworld (Simms and Ruffell, 1989; Therrien and Fastovsky, 2000; Whiteside et al., 2011). In the Tethyan domain, the Late Triassic was

dominated by warm, semiarid to arid climatic conditions (Hornung et al., 2007; Arche and López-Gómez, 2014) with two short-lived relatively wet peaks during the middle-late Carnian known as the Carnian Pluvial Event (Dal Corso et al., 2012; Ruffell et al., 2016; López-Gómez et al., 2017). Afterwards, arid conditions established and the Late Triassic has no evidence of glacial activity, making it a unique period in the Earth's Phanerozoic history (Preto et al., 2010).

In the Tethys realm the Early Jurassic marks a change from a dominantly warm-arid to a more humid palaeoclimate (Kuerschner et al., 2007; Ruckwied and Götz, 2009; Bonis and Kürschner, 2012), globally correlated with the large igneous province (LIP) volcanic activity of the Central Atlantic Magmatic Province (CAMP) at the Triassic–Jurassic boundary (Marzoli et al., 1999, 2004; Wignall, 2001; Hesselbo et al., 2002). The CAMP activity is associated with the Pangaea rifting and is considered as a favored trigger for extreme warming via greenhouse gas exhalation responsible for the Triassic/Jurassic mass extinction (Olsen, 1999; Pálffy et al., 2001; Ruhl et al., 2011).

In the then southern hemisphere, the Karoo-Ferrar Continental Flood Basalt Province (CFBP) emplacement is linked to the Early Jurassic break-up of Gondwana (Duncan et al., 1997; Pálffy et al., 1997; Pálffy and Smith, 2000), is being linked to oceanic anoxia (Hallam, 1987; Jenkyns, 1999) and global warming (Hesselbo et al., 2000). This event may have played an important role in shaping the Early Jurassic margins of the associated plates, including the northwestern margin of the Indian plate, thereby controlling the large-scale geometries of the Lower Jurassic siliciclastic depositional systems in the region.

The tectonic activities at the Triassic–Jurassic boundary and the Early Jurassic played an important role in controlling the relative sea-level fluctuation during this time interval. A short lived but pronounced sea-level fall has been reported in different parts of the world

during this interval (Haq et al., 1987; Hallam 1988, 1997, 2001; Hallam and Wignall, 1999, Haq, 2017). This sea-level fall may have exposed new areas to the greenhouse climate thereby substantially enhancing magnitude and extend of chemical weathering in these areas.

The palaeoclimatic variation and sea-level fall at the Triassic–Jurassic boundary has resulted in a pronounce regression and the deposition of quartz-rich siliciclastics (mostly Lower Jurassic) on top of the exposed, dissolved and reworked Triassic platform carbonates in many European basins (Hessilbo et al., 2004; Haas and Tardy-Filácz, 2004; Le Tourneau and Huber, 2006), Iran and adjoining Afghanistan (Klet et al., 2006; Fürsich et al., 2009; Wilmsen et al., 2009) Western Australia, Madagascar and India (Mukhopadhyay et al., 2010).

In the Salt and Trans–Indus ranges, Pakistan (Fig. 1), a thick succession of Lower Jurassic fluvio-deltaic quartz-rich siliciclastics with some shale and carbonates overly Upper Triassic platform carbonates (Iqbal et al., 2014; 2015a, b). The present paper aims to interpret the palaeoclimatic conditions in the Salt and Trans–Indus ranges (Fig. 1), Pakistan, during the Triassic–Jurassic transition and the Early Jurassic using field evidences, petrographic data, Bulk and clay mineralogy and sediment geochemistry proxies. Based on this first detailed record from the southern hemisphere, a general regional and global correlation is envisaged to establish the palaeoclimatic context of the succession. This may open a new view from the Gondawana domain in the palaeoclimatic reconstructions for a critical interval of Earth history.

2. Geological Setting and Palaeogeography

The Salt and Trans–Indus ranges in Pakistan mark the southern boundary of the Kohat–Potwar Plateau (Fig. 1). The latter forms part of the foreland zone of the

101 Himalayan Fold and Thrust Belt (HFTB) and preserves thick Precambrian to Recent
102 sediments (e.g., Shah, 2009). Stratigraphy, structural geology and tectonics have remained
103 the subjects of many studies (e.g. Molnar and Tapponnier, 1981; Hallam and Maynard,
104 1987; Baker et al., 1988; Burbank and Reynolds, 1988; Davis and Lillie, 1994; Kadri, 1995;
105 Kazmi and Jan, 1997; Jadoon et al., 1997; Shah, 2009; Mohadjer et al., 2010; Jain, 2014;
106 Iqbal et al., 2014a, b; Hanif et al., 2014).

107 The studied succession includes the Kingriali and Datta formations with a total thickness of
108 152 – 270 m. The Upper Triassic Kingriali Formation was named after the Kingriali Peak in
109 the Khisor Range (Gee, 1945). A Late Triassic, Rhaetian age is confirmed by new palynology
110 data, i.e. the presence of the dinoflagellate cysts *Rhaetogonyaulax*, *Suessia* and
111 *Beaumontella* (Kuerschner et al., 2017). The uppermost, shaly part correlates to the
112 Triassic–Jurassic mass extinction event and the base of the Hettangian (Kuerschner et al.,
113 2017). The formation conformably overlies the Chak Jabbi Limestone (Middle Triassic) in the
114 Kala Chitta area and has conformable lower contact with the Tredian Formation (Middle
115 Triassic) in all other localities. The upper contact is conformable to the Datta Formation
116 (Iqbal et al., 2017).

117 The Lower Jurassic Datta Formation was named after the Datta Nala in Surghar Range
118 (Danilchik, 1961; Danilchik and Shah, 1967). It's mainly Hettangian age was recently
119 confirmed by Kuerschner et al. (2017). The formation is predominantly of continental origin
120 (Fatmi, 1977). It is conformably overlain by Shinawari Formation (Toarcian).

121 Palaeogeographically, the Salt Range was located on the northwestern passive margin of the
122 Indian Plate in the southwestern Tethyan realm, i.e., ~20° - 25° S latitude during Late
123 Triassic–Early Jurassic (Rees et al., 1999; 2002; Damborenea, 2002; Arias, 2007; Fig. 1a). The
124 Pangaea breakup during the Triassic–Jurassic boundary interval accompanied by rifting of

the Indian Plate from Africa and Arabia resulted in a change in palaeoclimate. The associated tectonic activities controlled the west–northwestward depression in the basin area that is now the Salt Range–Kohat–Potwar Plateau (Fig. 1b, c), and the subsequent development of a large fluvial system, which resulted in the deposition of the Datta Formation on ca. 15,000 Km² area (Iqbal et al., 2015b).

3. Material and Methods

3.1 Fieldwork and Sampling

The upper part of the Kingriali Formation and the entire Datta Formation were studied in the Salt and Trans–Indus ranges (Fig. 1b, c). Detailed sections were measured at a ca. 300 km transect from Kasanwala, Kaowaali, Nammal Gorge, Zaluch Nala (western Salt Range), Chichali Nala, Gulla Khel, Baroch Nala (Surghar Range) and the Paniala anticline (Khisor Range).

Sedimentological analyses were conducted in the field (Fig. 3) including observation of the vertical and lateral lithological and facies variation, sedimentary structures and features following standard sedimentology procedures (e.g. Nichols, 2000; Reading, 2009). Marker horizons such as erosional surfaces, oxidised zones, palaeosols, rippled and cross bedded horizons and beds with plant remains were identified (Fig. 3). In total 400 archived samples were collected covering all the lithological and facies variations (Selley, 2000).

3.2 Petrographic Studies

With the exception of loose shales, 300 thin sections were prepared for all the siltstone, sandstone, conglomerates, carbonates and evaporites samples at the Department of Geodynamics and Sedimentology, University of Vienna thin section petrography lab. These thin sections were studied systematically using Leica DM2700 P

microscope with Leica MC170 HD camera. Herein, the petrographic studies were focused on sandstones thin section point counting (e.g. Dickinson, 1970; Graham et al., 1976; Appendix. 1) and its application for the interpretation of palaeoclimate and weathering trend. Petrographically, quartz was subdivided into monocrystalline (Q_m) and polycrystalline (Q_p). The latter also include chert (Ingersoll et al., 1984). The Q_m is subdivided into Q_m with unit extinction (Q_{mue}) and Q_m with undulose extinction (Q_{muu}) (Fig. 4c, e). Similarly the Q_p is subdivided into Q_p (Fig. 4f) with 2 – 3 crystals ($Q_{pq(2-3)}$) and Q_p with >3 crystals ($Q_{pq>3}$). The feldspars are subdivided into plagioclase (P) and K-feldspar (A) whereas the lithic fragments are referred to as lithic fragments (L). The data was plotted on Quartz–Feldspars–Lithic Fragments (QFL) plot (Dickinson and Suczek, 1979). To follow the practice of the existing literature (e.g. Suttner and Dutta, 1986) R is used and has the same meaning as L (Ingersoll et al., 1984). Furthermore, log/log plots for $Q_p/(F+R)$ Vs. $(Q_m+Q_p)/(F+R)$ were used to reconstruct palaeoclimate trends (Suttner and Dutta, 1986). The minimum (F+R) value was fixed as 1 to avoid out scale values (Fig. 5).

3.3 Bulk Mineralogy/XRD Analysis

Based on fieldwork observations and petrographic analysis, 100 samples were selected for XRD analysis. Powder XRD was used for the determination of the bulk mineralogical composition (Appendix. 2a). XRD analysis was conducted using PANalytical X'Pert Pro diffractometer (CuK α - radiation (40 kV, 40 mA), step size 0.0167, 5 s per step) at the Department of Geodynamics and Sedimentology, University of Vienna.

3.4 Clay mineralogy/XRD Separation

30 samples were investigated for clay mineralogy. The samples were treated with hydrogenperoxide (H₂O₂) to remove the organic matter/carbon (Van Langeveld et al., 1978; Mikutta et al., 2005). Following the organic content removal, the samples were

dispersed with a 400 W ultrasonic probe. The less than 2 μm fraction was separated using Atterberg cylinders. Oriented clay samples were prepared by pipetting 1 ml of suspension (10 mg/ml) onto a glass slide. XRD mounts were analysed, saturated with Mg and K ions and after saturated with glycerol (Mg-samples) and ethylene-glycol (K-samples) to separate smectite and vermiculite. Finally, the samples were heated to 550 °C (Moore and Reynolds, 1997). The XRD patterns (Appendix. 2b) were interpreted (Brindley and Brown, 1980; Moore and Reynolds (1997), and quantified using the correction factors (Schultz, 1964).

3.5 Bulk Geochemistry

Various geochemical proxies have reliable palaeoclimatic and palaeoenvironmental significance (e.g. Climap, 1976; Nesbitt and Young, 1982; Roser and Korsch, 1988; Wefer et al., 1999; Hofmann et al., 2001; Zachos et al., 2001; Fischer et al., 2012). In the present case 150 samples were powdered and homogenised for bulk rock geochemistry (Appendix. 3). The bulk rock geochemistry analyses were conducted in Bureau Veritas Ltd. (former Acme) Analytical Laboratories Canada using Inductively Coupled Plasma Emission Spectroscopy/Mass Spectrometry (ICP- ES/MS). The analyses include oxides, major, minor, trace and rare earth elements.

The bulk geochemical composition of the samples was measured using a combination of fusion and Aqua Regia digestion on sample splits. A split of 0.2 mg powdered and homogenised sample was fused in Lithium metaborate/tetraborate followed by Nitric acid digestion to extract Si, Nb, and Rb and measured by Inductively Coupled Plasma Optical Emission Spectrometry (ICP-OES; Spectro Ciros Vision). In addition 0.5 g sample was digested in 95 °C Aqua Regia and analysed for Al, Fe, Mg, Ca, Na, K, Ti, P, Mn, Cr, Ba, Co, Sr, Th, U, V, W, and Zr by ICP-MS (Perkin Elmer ELAN 9000, Sciex). Both splits were calibrated

against reference materials SO18, DS 10, and OREAS 45 EA. Total C (TOT/C) and total S (TOT/S) were detected by LECO analysis. 13 samples were analysed in duplicates. The average analytical error in the analysis is 0.18%.

3.5.1 Geochemical Proxies

3.5.1.1 Chemical Index of Alteration (CIA)

The Chemical Index of Alteration (CIA) is a useful proxy for palaeoclimate reconstructions (Yan et al., 2010) and is calculated (Eq. 1; Nesbitt and Young, 1982, 1989; Taylor and McLennan, 1985):

$$CIA_{(molar)} = \left[\frac{Al_2O_3(molar)}{Al_2O_3(molar) + CaO^*(molar) + Na_2O(molar) + K_2O(molar)} \right] \times 100 \text{----- Eq. 1}$$

Here the CaO* is the amount of CaO incorporated in the silicate fraction of the rock (Nesbitt and Young, 1982; McLennan et al., 1993). Further, all the quantities were used in molar proportions. To avoid errors derived from the unknown CaO content of the silicate fraction, the Chemical Proxy of Alteration (CPA) is used (Eq. 2; Buggle et al., 2011).

$$CPA_{(molar)} = \left[\frac{Al_2O_3(molar)}{Al_2O_3(molar) + Na_2O(molar)} \right] \times 100 \text{----- Eq. 2}$$

3.5.1.2 Al₂O₃ Vs. CIA plot and K₂O/ Na₂O Vs. CIA plots

Al₂O₃ Vs. CIA_(molar) plot and K₂O/Na₂O Vs. CIA_(molar) plots are used to differentiate between the arid, subtropical and tropical palaeoclimates (Goldberg and Humayun, 2010). The CIA_(molar) in these plots was modified to:

$$CIA_{(molar)} = \left[\frac{Al_2O_3(molar)}{CaO^*(molar) + Na_2O(molar) + K_2O(molar)} \right] \text{----- Eq. 3}$$

This modified CIA, a proportion between alumina and alkalis plus calcium, provides a more sensitive indicator of the degree of chemical weathering, as now the values are not restricted to less than 100 and the equation (Eq. 3) yields values of more than 100 and even up to 500 for laterites (Goldberg and Humayun, 2010).

3.5.1.3 A-CN-K Plot

Ternary plot of Al_2O_3 , $\text{CaO}^* + \text{Na}_2\text{O}$, and K_2O defines an equilateral triangle called A-CN-K diagram (Nesbitt and Young, 1984; Bahlburg and Dobrzinski, 2011). The diagram explains the linkage between the weathering trends and mineralogical compositions (Basu, 1976). Similarly the positions of smectite, illite, gibbsite, chlorite and kaolinite are indicated (Nesbitt et al., 1996).

3.5.1.4 Chemical maturity proxy plot

Compositional variations play strong role in controlling the amount of Al, Si, K and Na in sediments (Hofmann et al., 2001). Therefore, a bivariate plot of SiO_2 against total $\text{Al}_2\text{O}_3 + \text{K}_2\text{O} + \text{Na}_2\text{O}$ is generally used to represent the chemical maturity trend in the siliciclastic sediments (Suttner and Dutta, 1986; Malick and Ishiga, 2016). The chemical maturity trend is related to the palaeoclimatic conditions during the sedimentation.

4. Results

4.1 Outcrop/field data

The Kingriali Formation consists mainly of light gray, yellowish gray, thick bedded dolomite, and shows only slightly variable thicknesses from the western Salt Range (75 m – 85m) to the Surghar Range (92 m – 108 m) and the Kala Chitta Range (91 m). The

237 dolomite displays fragmental texture and is oolitic at places. Very thin beds of dark green,
 238 brownish black shale appear in the upper part. Both the occurrence frequency and thickness
 239 of the shale beds increase upsection (Fig. 3a). Further upsection, the shale is interbedded
 240 with marl and argillaceous dolomite. Dolomite with reworked clasts (Fig. 3b) and
 241 fossiliferous horizons (mostly bivalves) occurs in the uppermost part at western Salt Range.
 242 In the uppermost part of the Kingriali Formation at Nammal Gorge and Panaila anticline a
 243 conspicuous 30 – 70 cm shale occurs that includes a 2-15 cm gypsum bed (Fig. 3c, d). This
 244 unit overlies a 30 cm rusty brown, haematitic/limonitic dolomite and marks the contact of
 245 the Kingriali Formation with the overlying Datta Formation (Fig. 3c).
 246 In the western Salt Range (except Nammal Gorge) the formation is overlain by a 30 – 50 m
 247 thick succession of white, light gray, rusty brown fine to medium grained sandstones
 248 interbedded with red/maroon, orange, gray shale. This lithofacies is overlain by a 8 – 21 m
 249 thick laterite/bauxite (Fig. 3e). This laterite/bauxite is overlain by the conglomerates/pebbly
 250 sandstones of the Datta Formation in all the three sections (Fig. 3d).
 251 The thickness of the Datta Formation varies from 132 to 240 m along the transect from the
 252 western Salt Range (132 m – 200 m) to the Surghar Range (150 m – 240 m), and the Khisor
 253 Range (140 m). The lower part of the Datta Formation consists of conglomerates, pebbly
 254 and coarse sandstones of the CBFA (Iqbal et al., 2015b) in the western Salt Range (Fig. 3e).
 255 The sandstones and conglomerates are well sorted, rounded and rich in quartz. In the Gulla
 256 Khel and Paniala anticline sections the conglomerates do not exist and coarse grained, white
 257 sandstone directly overlies the Kingriali Formation. Petrified wood, thin coal patches and
 258 reworked coal/wood clasts occur (Fig. 3f).
 259 These are overlain by red-brown sandstones, siltstones and shale of the CMOBFA (Iqbal et
 260 al., 2015b). The lower bedding planes are erosive while raindrop imprints, palaeosol

horizons and plant remains occur. Mud cracks, ripples marks and bioturbation features are common.

The LFA consist of brown - black shale and associated carbonates (Iqbal et al., 2015b). The shale is carbonaceous at places while the carbonates include both pure and arenaceous dolomite. Ripple marks, bioturbation are common. Apart from some bivalves and gastropods at Zaluch Nala no fossils were observed in the succession.

4.2 Framework Composition/Petrographic studies

Of the 300 analysed samples, most are sandstones while carbonates constitute a minor proportion. Thorough dolomitisation and lack of fossil content in the carbonate samples limit their applicability for palaeoenvironmental reconstructions. In the upper part of the formation the dolomites show fragmental texture and clasts alignment. Fossiliferous horizons with bivalves and gastropods occur in the uppermost part at Kasanwala (Fig. 4a, b) and lamination resembling stromatolites occur.

Quartz is the most common component of sandstones and the total quartz (Q_{total}) averages 95.5% of the framework components. Q_{mue} predominates averaging 76.1% followed by Q_{muu} (12.5%), whereas, $Q_{pq(2-3)}$ content averages 4.3% and $Q_{pq>3}$ averages 2.6% of the bulk framework composition (Appendix. 1). High Q_{muu} and Q_p contents occur in the coarse/pebbly sandstones and conglomerates of the CBFA (Iqbal et al., 2015b). The total feldspars (F_{total}) content average only 0.15% of the bulk framework, which are highly decomposed and altered to clay minerals. Very minor potassium feldspar (K) occurs (average 0.13%) while plagioclase feldspar (F) is rare (average 0.02%). The total lithic fragments (L) constitute a minor component, averaging 1.2%. L is the combination of sedimentary lithics (L_s), metamorphic (L_m) and volcanic-hypabyssal (L_v). The QFL plot

indicates a cluster of samples near the quartz apex (Fig. 5a). The framework composition comparison, a function of the ratios of grain types and the most sensitive climatic control, is a bivariate log/log plot based on $Q_p/(F+R)$ Vs. $Q_{total}/(F+R)$ ratios and displays closely clustered sample populations in the humid zone (Fig. 5b).

4.3 Bulk mineralogy/XRD

The XRD analysis indicates the qualitative mineralogy of the selected horizons (Appendix. 2a). The upper part of the Triassic carbonates is composed of dolomite and calcite. In the shale unit, in the upper part, the carbonate content decreases and quartz and clay minerals appear. Further up section at the contact with the overlying Jurassic strata the carbonate content disappears and evaporites (gypsum) appear. In the associated shales, illite is the dominant clay mineral. These shales are overlain by laterite horizons rich in kaolinite-haematite. The kaolinite content within these laterites increases upsection and the upper part of the laterite at the Kasanwala section is entirely composed of boehmite. Upsection in all the measured sections, the coarse siliciclastics (CBFA) are quartz rich, while the associated shales with palaeosols (CMOFA) have kaolinite and quartz. The overlying shales of the LFA are composed of clay minerals, quartz and carbonates with an increasing carbonate (dolomite and ankerite) content trend upsection. Kaolinite is the dominant clay mineral but the illite content increases upsection.

4.4 Clay mineralogy

Clay mineralogy indicates the presence of a simple mixture of illite and kaolinite (Appendix. 2b). In the Kingriali Formation, illite is by far the dominant clay mineral, averaging 94.1% while the remaining 5.9% is kaolinite. The lower part of the Triassic strata indicates kaolinite as a minor component that completely disappears in the green shales.

In the overlying Datta Formation kaolinite is the dominant clay mineral, averaging 68.5% of the total clays, while the remaining 31.5% is illite. Following the very low kaolinite content during the Triassic a jump to very high kaolinite is observed at the end of the Triassic. The highest kaolinite content is present in the CMOFA (92.2%) whereas, the lowest content occurs in the LFA (37.2%). The laterite/bauxite horizons are kaolinite rich.

4.5 Bulk Geochemistry

4.5.1 Chemical Index of Alteration (CIA)

The CIA (molar) plots (Figs. 2b, 7) display low values for the Triassic strata (shales) ranging 68.5 - 80 averaging 77.7. The CIA has a mean of 92.7 for the Jurassic strata with the Kasanwala and Gulla Khel section displaying the highest average values of 94.9 and 94.6 respectively while the Nammal Gorge section has the lowest average of 88.8. Stratigraphically, the coarse grained sandstones-conglomerates (CBFA) and the associated laterites/palaeosols possess the highest CIA values generally over 95. Exceptionally high values of 99.8 are calculated for the uppermost part of the thick laterites at the Kasanwala. After the continuous high CIA values in the CBFA, a gradual drop in the CIA value is observed through the CMOFA to the LFA (Fig. 2b). However, the CIA values are still above 85 in this interval. After this interval of relatively low CIA values, the overlying sandstones (second repetition of CBFA) again display high CIA values (mostly above 90) before another drop but still more than 80 corresponding to the second repetition of the LFA. This second fluctuation corresponds to the second depositional sequence interpreted by Iqbal et al. (2015b). In order to verify the validity of the CIA, CPA (Buggle et al., 2011) was used, giving an average value of 98.8 for the Triassic strata with a minimum value of 96.2. For the Jurassic strata the CPA yields an average of 99 with a minimum value of 97.

4.5.2 Al₂O₃ Vs. CIA and K₂O/Na₂O Vs. CIA Plots

The Al₂O₃ Vs. CIA_(molar) plot for the Kingriali Formation indicates a cluster of samples in the CIA_(molar) range 1 – 2 and Al₂O₃ range of 7 – 25 (Fig. 6a). A gradual upsection increase in the Al₂O₃ content is observed. In the lower part of the succession the Al₂O₃ content increases upwards (up to 15% Al₂O₃) with no prominent change in the CIA. The boundary beds can be easily differentiated from the rest due to a distinct increase in both the Al₂O₃ and CIA (Fig. 6a).

All samples of the Datta Formation plot to the right of the Triassic samples and have 0 – 40 range for Al₂O₃ (Fig. 6a). The highest CIA and Al₂O₃ are plotted for the laterite/bauxite samples. The CBFA and sandstone-siltstone of the CMOFA display higher CIA values but the Al₂O₃ content is mostly below 15%. The lowest Al₂O₃ accompanied by relatively low CIA values correspond to the floodplain siltstone of the CMOFA while the associated palaeosols in the COMFA still plot in the high CIA and Al₂O₃ rich part. This allows their differentiation from the shales of the LFA which plot in the zone of relatively high Al₂O₃ but comparatively low CIA. The second depositional sequence displays a similar trend.

The K₂O/Na₂O vs. CIA Plot displays a wide range for the K₂O/Na₂O in the Kingriali Formation (Figs. 6b). The lowest K₂O/Na₂O ratios correspond to the lowermost parts and a gradual upsection increase is observed for all the sections. However, the Na₂O has very low values (always <0.54) hence the ratio generally fluctuates due to K₂O content (Appendix. 3). From bottom to top the K₂O/Na₂O increases from 5.67 to 89.6 in the Nammal Gorge (Fig. 2b), from 26.1 to 76 in the Gulla Khel while no significant change is observed in the Paniala section where the ratio fluctuates from 38 to 46.75. The highest K₂O/Na₂O values are recorded for the shales of the boundary bed at all the studied sections (Figs. 6b).

In the Datta Formation the K_2O/Na_2O range diminishes (0 - 44.3). The samples from the laterites, CBFA and sandstones-siltstone of the CMOFA display ratios generally below 10 whereas the shales of the CMOFA have relatively higher ratios (10 – 20). These can be easily differentiated from the shales of the LFA which possess the highest ratios (generally >20).

4.5.3 A–CN–K Plot

The A–CN–K ternary plot for the representative samples displays a cluster of samples near the A vertex close to the A–K side (Fig. 7a). Majority of the Triassic samples plot in the muscovite - illite position. The laterite/bauxite horizons of all the sections are clustered at A vertex corresponding to kaolinite position. The CBFA and associated sandstones-siltstones of CMOFA plot close to the A vertex above the Illite position. Most of the shales of the CMOFA and many of samples from the LFA plot in the illite-high illite zone. The CIA values are plotted for the samples to the left of the A–CN–K plot and correspond to the kaolinite – illite zone and are close to the A vertex of the diagram (Fig. 7a).

4.5.4 The chemical Maturity Plot (SiO_2 vs. $Al_2O_3 + Na_2O + K_2O$)

The quartz (SiO_2) vs. feldspars ($Al_2O_3 + Na_2O + K_2O$) plot roughly represents an isosceles triangle with SiO_2 line (y-axis) as its base and the two equal sides intersecting at 50% SiO_2 and roughly 30% $Al_2O_3 + Na_2O + K_2O$ (Fig. 7b). All the samples from the Kingirali Formation plot on the lower side of the diagram. The lowermost samples plot at the bottom side of the plot while, as we move upsection, the sample positions gradually move upward on the lower side of the triangle. The boundary beds from all the sections are clustered around the lower part of the vertex where the two equal sides meet (Fig. 7a).

The laterites/bauxites of the Datta Formation plot to the right of the vertex where the two equal sides meet and thus can be easily differentiated from the underlying boundary beds (Fig. 7b). The CBFA and CMOFA sands arrange close to the 100% SiO_2 vertex. The overlying

CMOFA shales gradually move from the top vertex to where the two equal sides meet. Moving further upsection in the Datta Formation the shales and associated samples of the LFA plot close to the vertex where the two equal sides meet. However, these can be easily differentiated both from the shales of the Kingriali Formation and laterites of the Datta Formation as these occupy mostly the upper, equal side of the triangle (Fig. 7b).

5. Discussion

5.1 Outcrop/field sedimentology

The thick bedded, uniform and laterally persistent dolomites of the Kingriali Formation (Fig. 3a) represent platform carbonates deposited under an overall uniform palaeoclimate (Anwar et al., 1992). The shales/marls in the upper part of the formation, and their increase both in frequency and thickness (Fig. 3a) represent the establishment of a siliciclastic feeder system (and source) in the nearby area (Mount, 1984). These are also indicative of the gradual onset of climatic change with negative impact on carbonate production (Wood, 1990). The brecciated dolomite, with its flatly aligned pebbles (Fig. 3b), and the ferruginous dolomite (Fig. 3f) indicate a sea-level drop, reworking in very shallow water and at least part-time emergence. The overlying thin bedded gypsum and gypsiferous shale (Fig. 3f) represent the final demise of carbonate production. On the other hand, the gypsum also supports the establishment of hypersaline conditions and evaporites precipitation an evidence for warm and dry conditions (Tucker, 2009).

The quartz-rich conglomerates and associated quartz arenites in the lower part of the Datta Formation (Fig. 3a, c, e) indicate a prominent regression above the marine carbonates and the onset of coarse-grained fluvial-continental conditions (Iqbal et al., 2015b). The occurrence of coal patches, plant remains and petrified wood (Fig. 3g) within the CBFA and

the overlying CMOFA support a warm and humid palaeoclimate during their deposition. The raindrop imprints provide further evidences for humid conditions. The associated laterites/bauxites and palaeosols (Fig. 3e) indicate hot and humid, waterlogged areas of a tropical to subtropical climate (e.g. Retallack, 2010; Mindszenty, 2016). Further up section (Fig. 2a, b), the LFA represents transgression and onset of marine to marginal marine condition (Iqbal et al., 2015b).

5.2 Framework Composition/petrography

The lower and middle part of the Kingriali Formation lacks identifiable fauna; however, the uniform oolitic texture, micritic nature, occurrence of bivalves and gastropods (Fig. 4a, b) coupled with persistent sediments cycles reveals a depositional setting with only minor palaeoclimatic variations, probably subtidal depths on a large carbonate platform (Anwar et al., 1992). In contrast, the fragmental texture, aligned clasts and reworked bioclasts (Fig. 3b) in the upper part of the formation advocate a significant drop in sea-level and/or accompanying tectonic activity (Sa'ad, 2017). Furthermore, the stromatolites (occurring in the overlying beds), normally flourish in the hypersaline and intertidal to supratidal conditions; on the other hand their occurrence supports the lack of dominant predators/grazing organisms and thus a possible biotic crises (e.g. Youbi et al., 2004; Luo et al., 2016). Thus their occurrence may represent the transition to extreme conditions the Triassic-Jurassic boundary (Olsen, 1999; Pálffy et al., 2010). Also, the stromatolites generally occur in the shallow to very shallow water (Preto et al., 2015). In the present case, their stratigraphic position between the underlying shallow water, reworked, ferruginous carbonates and the overlying gypsum beds indicates near surface/at least partly emergent, intertidal to supratidal conditions. The Triassic strata thus present a similar situation as

evidenced worldwide as a result of a significant and global eustatic sea-level drop at the Triassic–Jurassic boundary (Clemmensen et al., 1998; Hallam, 2001).

Quartz arenites constitute the major part of the Datta Formation (Fig. 2a, b) are texturally and compositionally supreme mature sandstones (Fig. 4c – e) and are the products of multi cyclic sediments transport and deposition (Pettijohn and Potter, 1987). Simultaneously, these are the product of intense and persistent chemical weathering, conditions that generally favour long, continued crustal stability (mostly at passive margins), warm/hot and humid palaeoclimates (Nichols, 2009; Dill, 2016). In the present case the dominance of quartz, absence or very minor occurrence of feldspars and complete absence of ferromagnesian minerals support deposition under such a hot and humid paleoclimatic setting. Furthermore, Q_{muc} constitutes around 80% of the total quartz, suggesting very strong sorting. Such conditions generally common in tropical and near tropical palaeoclimatic belts and also suggests a recycled origin of the sediments (Blatt, 1982).

The QFR triangular plot and the bivariate log/log plot of $Q_p/(F+R)$ vs. $Q_{total}/(F+R)$ indicate closely clustered sample populations in the humid part of the plot (Fig. 5a, b). These plots again support a humid palaeoclimate during the deposition of the Datta Formation. Primary absence of unstable minerals in the source areas may question the validity of the present interpretation. However, petrographic studies in the potential source areas reveal the presence of abundant feldspars and other, potentially unstable minerals (Ghazi and Mountney, 2011; Iqbal et al., 2015b; Jan et al., 2017).

5.3 Bulk Mineralogy/XRD

The presence of only carbonates in the lower and middle parts of the Kingriali Formation (Appendix. 2) indicates carbonate precipitation without (silici)clastic interruption.

447 Upsection the presence of both calcite and dolomite in the fossiliferous carbonate horizons
448 may indicate selective dolomitisation where the bioclasts may have partly or completely
449 survived dolomitisation, thereby preserving the original calcitic composition (Fig. 4a, b).
450 Such an overall vertical distribution of dolomite on the platform does not reflect long-term
451 eustasy (Balog et al., 1999). The thorough/intense dolomitisation of the lower platform part
452 reflects a semiarid, hot subtropical, seasonal setting and megamonsoonal climate. Global
453 cooling and increased humidity toward the Latest Triassic inhibited pervasive early
454 dolomitisation, leaving the upper platform largely as limestone (Balog et al., 1999). The
455 gradual appearance of clay minerals and quartz indicates a gradual onset of the siliciclastic
456 supply to the basin. The upsection disappearance of carbonates and increase in the clay
457 contents indicates a demise of carbonate precipitation. The overlying evaporites (gypsum)
458 also support semiarid to arid palaeoclimate and a probable drop in sea-level.

459 The CBFA of the Datta Formation, with dominant quartz peaks, supports the quartz-rich
460 framework composition (Fig. 5a, b). At places, the very minor kaolinite peaks within these
461 samples on one hand indicate the alteration of feldspars to clay minerals under greenhouse
462 conditions (Dill, 2016). On the other hand, clay minerals introduction into the pore spaces
463 during the comparatively low energy state of the flowing water could cause such deposition
464 (Boggs, 2009). However, the pore-space filling kaolinite is generally composed of concertina-
465 type kaolinite and occurs in kaolin arkoses whereas in the active channel system kaolinite is
466 accumulated mainly in the fine-grained overbank mudstone deposits (Dill, 2016). In the
467 present case though, field and petrographic observations favour the overbank mudstones
468 depositional scenario.

The overlying palaeosols, laterite/bauxite horizons are characteristically rich in kaolinite, boehmite and haematite. Such compositions rarely occur in warm semiarid – arid climates where surficial chemical weathering is only a minor process or is absent and also in cold and glacial climates where surficial chemical weathering proceeds very slowly and frost wedging is the dominant process (Tucker, 2009). Such deposits generally occur as residual deposits on passive continental margin and in epicontinental basins (Dill, 2016). The carbonates and shales of the LFA mark the onset of next marine/lagoonal settings (Iqbal et al., 2015b).

5.4 Clay Mineralogy

The overall dominance of illite in the Triassic shales (Appendix. 2b) supports dry semiarid to arid conditions of deposition during Rhaetian (Keller et al., 1954; Singer, 1984; Dill, 2016). The gradual upsection increase in the kaolinite content indicates a transition to warm-hot and humid palaeoclimate. The absence of kaolinite in the green shales (boundary bed) may be interpreted as representing arid and potentially cooler conditions established during volcanic winters (Schoene et al., 2010). The overlying shales of the Datta Formation are enriched in kaolinite (Fig. 7) and support hot, warm and humid palaeoclimate also indicated by other proxies.

A possible diagenetic control (Worden and Burley, 2003) over the clay minerals distribution may hinder their palaeoclimatic. Diagenetic history of kaolinite (Dill, 2016) is illustrated by a sequence of change from kaolinite through dickite into nacrite under increasing temperature and a conversion of halloysite into metahalloysite as the temperature decreases (Harvey and Murray, 1990; Polyak and Guven, 1996; Harrison and Greenberg, 1998). However, no such compositional variations were found in the present case. On the other hand it is also known that illite generally occurs in association with carbonates away

from major rivers while kaolinite generally occurs in fluvial-alluvial settings (Weaver, 1958; Whitehouse et al., 1960; Thiry, 2000).

5.5 Bulk Geochemistry

5.5.1 Chemical Index of Alteration (CIA)

CIA values mainly dependent upon chemical weathering and latter has a direct relationship with the intensity of temperature and humidity of the prevailing (palaeo)climate (Nesbitt and Young, 1984). The relatively low CIA values for the Kingriali Formation (Figs. 2b, 7) indicate a lowered intensity of chemical weathering during the Late Triassic and hence deposition under semiarid to arid palaeoclimate. However, the gradual upsection increase in the CIA values supports a gradual onset of increasing humidity in a warming, subtropical to tropical palaeogeographic setting (Figs. 6a, b). The highest CIA values correspond to the topmost beds in every section indicating the onset of comparatively warm-hot and humid palaeoclimate during the Triassic–Jurassic transition.

The high CIA values for the CBFA and CMOFA suggest their deposition in hot and humid palaeoclimate (Fig. 2b). The palaeosols of the CMOFA and the laterite/bauxite horizons possess the highest CIA values (all above 95), thus confirming their deposition under extreme greenhouse condition in tropical – near tropical palaeoclimates (Mindszenty, 2016). Further upsection, the gradual drop in the CIA values (Fig. 2b) in the LFA indicates the onset of lagoonal/marine conditions with siliciclastic input under prevailing warm and humid conditions. The second depositional sequence displays warm - humid tropical palaeoclimate.

5.5.2 Al_2O_3 Vs. CIA and $\text{K}_2\text{O}/\text{Na}_2\text{O}$ Vs. CIA Plots

Scatter plots of Al_2O_3 Vs. CIA and $\text{K}_2\text{O}/\text{Na}_2\text{O}$ Vs. CIA, calibrated with modern examples from different climates, have been used in literature to interpret climate conditions (e.g. Goldberg and Humayun, 2010 and references therein). The initial siliclastic influx in the upper part of the Kingriali Formation, marked by a vertical increase in the Al_2O_3 content (up to 15%) and no visible increase in the CIA value, indicates the onset of fine grained siliclastics under comparatively stable palaeoclimate (Figs. 2b, 6a). The overall low Al_2O_3 content of this zone also support semiarid to arid conditions globally reconstructed for the Late Triassic (Frakes, 1979; Allen et al., 1994; Balog et al., 1999; Ruckwied and Götz, 2009; Luo et al., 2016). The gradual upsection increase in both the Al_2O_3 and CIA (Fig. 6a) indicates a transition to a comparatively humid setting. The topmost samples (boundary beds) plot in the subtropical palaeoclimate. However, the associated evaporites are indicative of high evaporation under arid conditions.

The Lower Jurassic samples plot in the tropical settings (Fig. 6a). The CBFA and CMOFA display higher CIA values but the Al_2O_3 content is understandably lower than 10% as these are quartz arenites (Fig. 2b). The shales and palaeosols of the CMOFA have high CIA and Al_2O_3 . The highest CIA and Al_2O_3 occur in laterite/bauxite samples. The high kaolinite content within these horizons presents a strong case of hot and humid palaeoclimate and thus greenhouse conditions. The relatively medium kaolinite content of the shales of the LFA also explains the relatively medium Al_2O_3 values (15 – 25%) and comparatively low CIA values (Fig. 6a). The second cycle of the Datta Formation in the study area displays a similar trend.

The $\text{K}_2\text{O}/\text{Na}_2\text{O}$ Vs. CIA provides refined interpretation of the palaeoclimate as the K_2O content is a function of the illite content within the samples (Song et al., 2014). In the present case the $\text{K}_2\text{O}/\text{Na}_2\text{O}$ vs. CIA plot for the Kingriali Formation shows a population cluster in the arid setting (Fig. 6b). However, an increasing upsection trend in the $\text{K}_2\text{O}/\text{Na}_2\text{O}$

ratios supports a gradual transition from arid – semiarid to humid palaeoclimate. The widest K_2O/Na_2O range (5.6 - 89.6) in the Nammal Gorge (Fig. 2b) reflects the selective removal of Na^+ content, while K^+ (and hence illite) is highest and thus indicates weak chemical weathering in an arid setting. Westward, the decrease in K_2O/Na_2O range and so too in both the Na^+ and K^+ reveals transformation of illite to kaolinite (Wang et al., 2014). The boundary beds display subtropical palaeoclimate weathering trends (Fig. 6b).

The small K_2O/Na_2O range (0 - 44.3) for the Datta Formation coupled with high CIA values supports overall tropical conditions (Fig. 6b). The CBFA and sandstones-siltstones of the CMOFA, with ratios generally below 10, suggest the removal of K^+ and Na^+ (both K-feldspar and plagioclase) under warm humid conditions. The overlying palaeosols and laterites/bauxites also display K_2O/Na_2O ratios below 10 and their kaolinite dominance supports tropical water-logged settings (Dill, 2016). The relatively high K_2O/Na_2O ratios (10 – 20) of the overlying shales of the CMOFA make their differentiation from the LFA easier as the latter possess the highest K_2O/Na_2O ratios (generally >20) within the Datta Formation. The second depositional cycle displays a warm-humid palaeoclimate.

5.6 A-CN-K Plot

The A–CN–K ternary plot (Fig. 7a) indicates weathering trends (Selvaraj et al., 2004; Selvaraj and Chen, 2006). In the present case, the linear cluster of samples between A vertex and muscovite position along the A-K side favours a recycled origin of the sediments. Most of the Rhaetian samples plot in the muscovite – illite position thereby favouring semiarid to arid conditions. The lowermost samples in the entire sections plot in the muscovite position. As we go upsection, the progressive upward plot of the samples close to the illite zone indicates a gradual increase in the intensity of the chemical weathering.

The cluster of samples from the palaeosols, laterite/bauxite horizons, sandstones and siltstones of CBFA and CMOFA plot correspond to the kaolinite position. An east to west increasing chemical maturity trend is indicated by shales of the CMOFA and the LFA plot in the illite – kaolinite zone. In this trend, the Nammal Gorge and Zaluch Nala samples occupy positions in the illite – high illite – kaolinite zone while most of the samples from the Gulla Khel and Paniala section plot in the kaolinite zone, a strong support of recycling under a hot/warm and humid palaeoclimate. The CIA correlation with the A–CN–K plot (Bahlburg and Dobrzinski, 2011) shows a strong coherence between the two data sets (Fig. 7a).

5.7 The chemical Maturity Proxy Plot (SiO_2 Vs $\text{Al}_2\text{O}_3 + \text{Na}_2\text{O} + \text{K}_2\text{O}$)

Good correlation between the increasing SiO_2 content and $\text{Al}_2\text{O}_3 + \text{Na}_2\text{O} + \text{K}_2\text{O}$ for the Kingirali Formation (Fig. 7b) indicates gradually increasing siliciclastic influx that eventually inhabited carbonate precipitation upsection. This trend also supports increasing upsection chemical maturity. The lower samples plot in arid zone, the upper correspond to the semiarid part and the boundary beds plot between semiarid and semi-humid zones. Thus the Upper Triassic samples mark a gradual change from arid palaeoclimate to a semiarid – semi humid palaeoclimate (Fig. 7b).

The palaeosols, laterite/bauxite samples of the Datta Formation plot in the semi-humid to humid zone (Fig. 7b) indicating the onset of humid condition during the earliest Jurassic following the Rhaetian aridity. It is important to mention here that the laterite/bauxite samples, that are highly enriched in Al_2O_3 (kaolinite) content, plot with this method erroneously in the arid – semiarid part of the plot. These should not be mistakenly interpreted as the product of dry palaeoclimate, as these have lost their SiO_2 due to dissolution under extreme greenhouse conditions and thus inevitably are a product of very high intensity of chemical weathering.

The sandstones to siltstones of the CBFA and CMOFA plot close to the 100% SiO₂, thus they represent deposition under very high humidity and warm/hot temperatures, interpreted in previous sections (Figs. 5a, b). The shales of the CMOFA and LFA are positioned in the humid part of the plot favouring their deposition in hot, warm and humid palaeoclimate (Fig. 7b).

6. Prevailing Palaeoclimate

Following the deposition of the lower and middle parts of the platform carbonates of the Kingriali Formation, a siliciclastic system gradually started during the Rhaetian in the Salt and Trans–Indus ranges (Fig. 8). This siliciclastic supply gradually increased until the carbonate production completely gave way to a siliciclastic-rich system at the Triassic–Jurassic boundary. Following this initial phase, the siliciclastic system established depositing the fluvio-deltaic Datta Formation (Iqbal et al., 2015b).

Field evidences, framework mineralogy, XRD, clay mineralogy and bulk geochemistry of the Kingriali Formation indicate a semiarid to arid palaeoclimate in the source area and the depositional basin during the Rhaetian. Such conditions are generally common in warm-semiarid to arid and also in glacial regions (e.g. Singer, 1984; Boggs, 2006). In the present case glacial evidence do not occur in the area (Iqbal et al., 2015b), nor does the subtropical – near tropical palaeogeographic position of the area during this time interval (Fig. 1a) support any possibility of glaciation. Furthermore, globally there are no evidences for glaciation in the Late Triassic (Meyen, 1997; Sellwood and Valdes, 2006).

The uppermost part of the formation with increasing kaolinite content, increasing upsection trend in CIA reveals a gradual onset of the chemical sorting towards the top of the Kingriali Formation. This indicates the onset of a climate transition towards more humid conditions

during the Triassic–Jurassic boundary interval (Fig. 8). Samples from the boundary beds have higher illite content that would indicate cool and dry climate. However, the gypsum beds within the boundary interval (Figs. 3d) also support evaporation in a warm and arid palaeoclimate.

The used proxies for the Datta Formation unequivocally indicate hot/warm and humid palaeoclimate during Early Jurassic (Fig. 8). The framework composition plots indicate quartz arenites of the high humid zone (Fig. 5a, b). XRD analysis proves the presence of only quartz within these samples. The kaolinite, boehmite and haematite-rich mineralogy of the palaeosols and laterite/bauxite horizons indicates long, continued intense chemical weathering greenhouse conditions. The CIA displays the highest values (>90) for these zones within the formation confirming the hot – humid tropical near tropical palaeoclimate (Figs. 2b, 6). The Al_2O_3 Vs CIA plot of these intervals indicate extreme tropical climate that is also confirmed by the $\text{K}_2\text{O}/\text{Na}_2\text{O}$ vs. CIA plot (Fig. 6a, b). The A–CN–K plot shows a cluster of these samples in the kaolinite zone (Fig. 7a) again confirming their supreme mature stage of chemical alteration. The SiO_2 Vs $\text{Al}_2\text{O}_3 + \text{Na}_2\text{O} + \text{K}_2\text{O}$ plot for these horizons show humid conditions (Fig. 7b). The kaolinite rich mineralogy of the shales of the CMOFA and LFA shows (Appendix. 2) deposition under hot/warm palaeotemperatures and high humidity.

7. Lateral Variations

The multi-proxy palaeoweathering data provide information about east-to-west lateral variations and gradients. The shale to dolomite ratio in the upper part of the Kingriali Formation decreases from east (2.6 at Nammal Gorge) to west (1.5 at Paniala). This, on one hand indicates the proximity of the eastern part to the siliciclastic feeder source

area; on the other hand it supports the lack of an efficient transport mechanism for the
sliciclastics during Late Triassic. The thin bedded gypsum and gypsiferous shale present at
the contact with the Datta Formation in Nammal Gorge, Gulla Khel and Paniala anticline
sections support the uniform semiarid – arid conditions suitable for evaporites precipitation.
CIA values for the Kingriali Formation indicate low intensity of chemical weathering and a
general east-to-west increase (Kasanwala and Nammal Gorge sections: 73.6, to Paniala
anticline section Khisor Range: 82.9). The A–CN–K plot and the K_2O/Na_2O ratios and other
proxies confirm this east-to-west increase in the chemical maturity.
The Datta Formation has overall mature sediments, a product of clearly more hot and humid
palaeoclimate under partly extreme greenhouse tropical to subtropical conditions. Quartz-
rich conglomerates are dominant in the western Salt Range and are absent in the Trans-
Indus ranges corroborating an east-to-west clastic sediment transport. The overall
geochemical analyses do not display any major east-to-west geochemical variation.

8. Possible Drivers of Palaeoclimate Variations

8.1 Palaeogeography

The Triassic, a hothouse time with ice free poles (Preto et al., 2010), had a
dominantly warm semiarid to arid palaeoclimate with an intervening Carnian humid episode
(Ruffell et al., 2016). All the continents were locked into the supercontinent Pangaea,
surrounded by the Panthalassa Ocean (Ziegler et al., 1983). The Tethys Ocean cut into
Pangaea and was largely confined to the tropical to subtropical belt (Ziegler et al., 2003).
These were the probable reasons for hot summers, relatively cold winters and strong

monsoonal circulation in the circum-Tethyan part of Pangaea (e.g. Kutzbach and Gallimore, 1989; Dubiel et al., 1991; Jacobs and Sahagian, 1993). This mega-monsoon seasonality produced dry equatorial regions in the eastern part of Pangaea, facing the western Tethys (Parrish, 1993; Wang, 2009).

The palaeogeographic position of the study area on the southeastern, equatorial – near tropical margin of Pangaea, facing the western margin of the Tethys (Fig. 1a), and the combined geochemical results affirm a warm, semiarid to arid palaeoclimate during the Rhaetian. However, in eastern North America, a narrow equatorial humid zone and an arid belt south of 30° N, passing northward into humid temperate climates existed (Smoot and Olsen, 1988; Olsen and Kent, 1996, 2000; Kent and Olsen, 2000; Whiteside et al., 2011). The palaeogeographic position of the study area may be an example of the existence of such zonal climate north of 30° in the southern hemisphere passing southward into humid climate. However, no significant shift in palaeolatitudes have been reconstructed for the study site from Late Triassic to Early Jurassic times (e.g. Schettino and Turco, 2011), so the climate evolution cannot be explained by plate movements through latitudinal climate zones.

8.2 Tectonics and CAMP Volcanism Activity

The latest Triassic stability in climates was followed by the end-Triassic mass extinction (ETE) event (Raup and Sepkoski, 1982). This global catastrophic biotic turnover at 201.56 Ma (Blackburn et al., 2013) implies a dramatic climate turnover that coincides with the CAMP activity (Marzoli et al., 1999; Beerling and Berner, 2002; Marzoli et al., 2004; Davies et al., 2016). The CAMP activity (an area of $>2.5 \times 10^6 \text{ km}^2$) forced a severe climate change via volcanic gases emissions. Recent studies reveal a warming trend from

675 greenhouse to hothouse climate as an extinction mechanism (McElwain et al., 2007).
676 Stomatal index data suggests increase of atmospheric CO₂ to around 2000–2400 ppm, in the
677 boundary interval suggesting a 3^o – 4^oC warming (McElwain et al., 1999), whereas stable
678 isotopic evidence suggests atmospheric CO₂ up to 4400 ppm (Schaller et al., 2011).
679 Geochemical analysis of the European and Pacific basins indicates a substantial negative
680 isotope anomaly (up to -3.5‰ δ¹³C) across the Triassic–Jurassic boundary at the ETE (Galli et
681 al., 2005; Kuerschner et al., 2007; Ruhl et al., 2009; Greene et al., 2012a; Bachan et al.,
682 2014). This carbon cycle perturbation involved the massive release of greenhouse gases, i.e.
683 ~8000–9000 Gt C as CO₂ during the CAMP basaltic eruptions and ~5000 Gt C as CH₄ (e.g.
684 Pálffy et al., 2001; Beerling and Berner, 2002). The CAMP activity is considered as the major
685 reason for transition from dominantly warm, semiarid - arid to hot and humid greenhouse
686 conditions (Korte et al., 2009; Kidder and Worsley, 2010).

687 In the study area there are no direct evidences of regional basaltic flow and volcanic ash
688 layers have not been identified so far. However, the Karapa greenschist, 200 km to the
689 north of the study area, extends from Westphalian to Carnian (Pougue et al., 1992). Further
690 to the north/northeast, the Late Triassic Landai Formation (Allai Kohistan) and the Triassic–
691 Lower Jurassic Alpurai Group of Swat–Bunair area have numerous amphibolitised basaltic
692 horizons (DiPietro, 1990; Pougue et al., 1992; DiPietro et al., 1999). Unfortunately lack of
693 reliable age control restrains any direct palaeoclimatic correlation between these basaltic
694 layers and Salt and Trans–Indus ranges.

695 Seismic reflection data reveal the presence of many subsurface, north-dipping basement
696 normal faults in the Salt and Trans–Indus ranges and Kohat–Potwar Plateau, terminating at
697 the base of the Jurassic strata (Iqbal et al., 2015a). Furthermore, the Sargodha High, 50 km

the south of the Salt Range, is considered as a horst structure uplifted during the Triassic–Jurassic boundary interval (Kadri, 1995). Most recently, this high has been considered not to be a part of the foreland normal-sense high strain zones that were active in the Paleoproterozoic (Martin, 2017), thereby supporting the younger age of Sargodha High. The nature and the timing of these faults provide strong evidences for the splitting and breakup of Pangaea in the Gondwanan domain comparable to CAMP activity (Iqbal et al., 2016). On the other hand, based on the palaeogeographic position of the study area to the nearby Karoo Rift, these tectonic features may be related to the initial phase of the Karoo Rift before the onset of main basaltic flow during Toarcian (Iqbal et al., 2016).

9. Regional and Global Correlation

The existing global correlation attempts on the Triassic – Jurassic transition are mostly focused on the Carbon Isotopic Excursion (CIE) emphasizing the initial negative carbon isotopic excursion (e.g. Hillebrandt et al., 2013; Lindström et al., 2017; Fig. 9). Unfortunately there is not much literature available on the Triassic–Jurassic boundary interval outside the European basins and North America (Laurasia). This study is the first detailed attempt from the then southern hemisphere (Gondwanaland) that explains the very close correlation with the Laurasia sections. An attempt has been made to correlate the geochemical spikes observed in the present studies to the Initial Negative Isotopic Excursion (InCIE) at famous localities in the New York Canyon, England and Northern Calcareous Alps (NCA) Austria (Fig. 9). However, since the study deals with the bulk geochemistry and clay mineralogy so detail discussion in this context is provided below:

720 Following the Carnian Pluvial Event, prolonged palaeoclimatic stability established in most
 721 parts of the European basins (Preto et al., 2010). Monotonous dolomitised peritidal
 722 carbonates successions (Dolomia Principale and Hauptdolomit) were deposited in the Alps
 723 under a hot and semi-arid climate (e.g., Haas and Demény, 2002) in a large carbonate
 724 platform setting (e.g. Frisia and Wenk, 1993; Iannace and Frisia, 1994; Meister et al., 2013).
 725 In northern Italy, the Travenanzes Formation (Preto et al., 2015) indicates deposition in a
 726 wide low-relief coastal area, dry land river – muddy coastal plain and sabkha system (Breda
 727 and Preto, 2011). Seaward of this clay-rich environment, thick peritidal carbonates formed
 728 in a large-scale epicontinental platform. In the Salt and Trans–Indus ranges the Tredian
 729 Formation and the Chak Jabbi Limestone provide a similar facies and succession. The
 730 Tredian Formation is considered to have been deposited in a fluvial- deltaic setting (Iqbal et
 731 al., 2014). The Lower and middle parts of the Kingriali Formation can be considered as the
 732 stratigraphic equivalent of the Dolomia Principale and Hauptdolomit.

733 The Fleming Fjord Formation of east Greenland indicates deposition in semiarid–arid
 734 palaeoclimate with a change to a marginal warm moist temperate climate at the end of
 735 Rhaetian (Clemmensen et al., 1998). In southern Sweden, a transition from smectite
 736 dominated clays of the Maglarp B unit and the Kågeröd Formation (Norian and older) to
 737 smectite-free kaolinitic clays of the Höganäs Formation (Rhaetian – Hettangian) indicates a
 738 transition from arid to warm and humid conditions (Ahlberg et al., 2003).

739 In the Tatra Mountains, Slovakia, illite-smectite rich clays of the Carpathian Keuper and the
 740 Rhaetian Fatra Formation indicate arid environment (Michalík et al., 2013). Kaolinite
 741 appearance in the uppermost part of the Fatra Formation (~2%), reaching to a peak

abundance of 11% in the Kopienec Formation (Hettangian) indicates the gradual onset of greenhouse conditions at the Triassic–Jurassic boundary (Michalík et al., 2010).

At Kendlbachgraben, Austria, geochemistry and clay mineralogy of the Rhaetian Eiberg Member of the Kössen Formation, and the overlying boundary mudstone “Grenzmergel” of the Tiefengraben Member of Kendlbach Formation support similar interpretation to southern Sweden (Bachan et al., 2014). The mineralogical and palynological studies of the Kendlbach Formation indicate a marked shift to humid climate and freshwater runoff (Bonis et al., 2010). Furthermore, geochemistry of the lowermost “Grenzmergel” have a distinctive enrichment in the heavy REEs (Pálffy and Zajzon, 2012). High kaolinite contents are reported from the topmost Triassic *Triletes* beds in Germany (van de Schootbrugge et al., 2009).

The change in clay mineralogy in the studied sections provides a comparable palaeoclimatic transition from hot–semiarid–arid to a hot and humid setting. Geochemistry of the Triassic–Jurassic boundary interval (green shale) reveals abnormally high concentration of the total rare earths elements (REEs) comparable to that observed in the Newark Basin, Culpeper basin, Morocco (Marzoli et al., 2011) and in Kendlbachgraben section, Austria (Pálffy and Zajzon, 2012).

In Pomerania, Poland, the lower–middle Rhaetian siliciclastics (Wielichowo Beds) were deposited under semiarid palaeoclimate, whereas the Upper Rhaetian–Lower Hettangian Zagaje Formation (siliciclastics) was deposited in a fluvial-lacustrine environment under a humid period (Pieńkowski et al., 2012), similar to the change to a fluvio-deltaic to lacustrine/lagoonal setting in Pakistan.

Conclusions

In the Salt and Trans – Indus ranges of Pakistan the thick, monotonous and laterally widespread lower and middle part of the Upper Triassic Kingriali Formation represents deposition of platform carbonates under stable warm, semiarid – arid palaeoclimate. During the latest Rhaetian, fine grained siliciclastic input of illite rich shales indicates a gradual onset of a siliciclastic system under semiarid – arid settings toward the Triassic–Jurassic boundary. The mineralogy and geochemistry of the topmost clay beds provides the possible links to Pangaea splitting and CAMP activity. The overlying kaolinite rich clay beds of basal part of the Lower Jurassic Datta Formation support high input of continental detritus and deposition under conditions of extreme chemical weathering. The quartz arenite nature of the coarse silicilastics, highly mature geochemistry, presence of coal beds – coal patches and abundant plant remains strengthen the prevalence of hot and humid greenhouse palaeoclimatic conditions during the deposition of the Datta Formation. The succession provides ideal sections from the then southern hemisphere for palaeoclimatic correlation with the northern hemisphere European basins and could potentially open new location for the understanding of the palaeoclimatic fluctuation at the Triassic–Jurassic boundary.

Acknowledgments

The research is the outcome of the PhD study under the OSS-II/HEC, Pakistan. IGCP-609 support for the fieldwork and samples analyses is acknowledged. The authors are thankful to Mr. Hafiz Shahid Hussain, Mr. Mukhtiar Ghani and Mr. Syed Irfan Hashmi for

786 their help during the fieldwork. Claudia Beybel, Ilka Wünsche and Leopold Slawek are
787 thanked for their technical support in thin section preparation and Maria Meszar is thanked
788 for her help in the clay mineralogy sample preparation.

References

- Ahlberg, A., Olsson, I., and Simkevicius, P., (2003) Triassic–Jurassic weathering and clay mineral dispersal in basement areas and sedimentary basins of southern Sweden. *Sedimentary Geology* 161, 15–29. [https://doi.org/10.1016/S0037-0738\(02\)00381-0](https://doi.org/10.1016/S0037-0738(02)00381-0).
- Allen, J. R. L., Hoskins, B.J., Sellwood, B.W., Spicer, R.S., and Valdes, P.J. (Eds.), (1994) *Palaeoclimates and their Modelling: With Special Reference to the Mesozoic Era*. Chapman and Hall. 140 pp.
- Anwar, M., Fatmi, A. M., and Hyderi, I. H. (1992) Revised Nomenclature and Stratigraphy of Musakhel and Baroch Groups in Surghar Range, Pakistan. *Jour. of Geol*, 15-28.
- Arche, A., and López-Gómez, J. (2014) The Carnian Pluvial Event in Western Europe: new data from Iberia and correlation with the Western Neotethys and Eastern North America–NW Africa regions. *Earth-Science Reviews*, 128, 196-231. <http://dx.doi.org/10.1016/j.earscirev.2013.10.012>
- Arias, C. (2007) Pliensbachian–Toarcian ostracod biogeography in NW Europe: evidence for water mass structure evolution. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 251(3), 398-421.
- Bachan, A., van de Schootbrugge, B., and Payne, J. L. (2014) The end-Triassic negative $\delta^{13}\text{C}$ excursion: a lithologic test. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 412, 177-186. <http://dx.doi.org/10.1016/j.palaeo.2014.07.027>
- Bahlburg, H., and Dobrzinski, N., (2011) A review of the Chemical Index of Alteration (CIA) and its application to the study of Neoproterozoic glacial deposits and climate transitions. *Geological Society, London, Memoirs*, 36(1), 81-92. doi: 10.1144/M36.6
- Baker, D. M., Lillie, R. J., Yeats, R. S., Johnson, G. D., Yousuf, M., and Zamin, A. S. H., (1988) Development of the Himalayan frontal thrust zone: Salt Range. *Pakistan, Geol* 16:3–7
- Balog, A., Read, J. F., and Haas, J., (1999) Climate-controlled early dolomite, Late Triassic cyclic platform carbonates, Hungary. *Journal of Sedimentary Research*, 69(1) 267-282.
- Basu, A., (1976) Petrology of Holocene fluvial sand derived from plutonic source rocks: Implication to paleoclimatic interpretation: *Jour. Sed. Petrol.*, v. 46, p. 694-709.
- Beerling, D.J., and Berner, R.A., (2002) Biogeochemical constraints on the Triassic–Jurassic boundary carbon cycle event. *Glob. Biogeochem. Cycles* 16 (3), 101–113. <http://dx.doi.org/10.1029/2001GB001637>.
- Blackburn, T.J., Olsen, P.E., Bowring, S.A., McLean, N.M., Kent, D.V., Puffer, J., McHone, G., Troy Rasbury, E., Et-Touhami, M. (2013) Zircon U-Pb geochronology links the end-

- 823 Triassic extinction with the Central Atlantic Magmatic Province. *Science* 340, 941–
824 945.
- 825 Blatt, H., (1982) *Sedimentary Petrology*; Freeman and Company, New York, pp. 105–241.
- 826 Boggs, S. Jr., (2009) *Petrology of Sedimentary Rocks*; 2nd edn, Cambridge University Press,
827 pp. 166–193.
- 828 Boggs, S. Jr., (2006) *Principles of Sedimentology and stratigraphy*. Pearson Education. pp.
829 119–157.
- 830 Bonis, N.R., Ruhl, M., Kürschner, W.M. (2010) Climate change driven black shale deposition
831 during the end-Triassic in the western Tethys. *Palaeogeography, Palaeoclimatology,*
832 *Palaeoecology*, 290, 151-159. <http://dx.doi.org/10.1016/j.palaeo.2009.06.016>
- 833 Bonis, N.R. & Kürschner, W.M. (2012) Vegetation history, diversity patterns, and climate
834 change across the Triassic/Jurassic boundary. *Paleobiology*, 38, 240–264,
835 <https://doi.10.1666/09071.1>
- 836 Breda, A. and Preto, N., (2011) Anatomy of an Upper Triassic continental to marginal-marine
837 system: the mixed siliciclastic-carbonate Travenanzes Formation (Dolomites,
838 Northern Italy) *Sedimentology*, 58, 1613–1647. DOI: 10.1111/j.1365-
839 3091.2011.01227.x
- 840 Brindley, G.W., Brown, G., (1980) *Crystal Structures of Clay Minerals and Their X-ray*
841 *Identification*. Mineralogical Society, London (495 pp.).
- 842 Buggle, B., Glaser, B., Hambach, U., Gerasimenko, N., and Marković, S. (2011) An evaluation
843 of geochemical weathering indices in loess–paleosol studies. *Quaternary*
844 *International*, 240(1), 12-21.<http://doi.org/10.1016/j.quaint.2010.07.019>
- 845 Burbank, D. W., and Reynolds, R. G. H., (1988) Stratigraphic keys to the timing of thrusting in
846 the terrestrial foreland basins: applications to the northwest Himalaya. In:
847 Kleinspahn KL, Paol C (eds) *Frontiers in sedimentary geology, new perspective in*
848 *basin analysis*. Springer, Berlin, pp 331–351.
- 849 Clemmensen, L. B., Kent, D. V., and Jenkins, F. A. (1998) A Late Triassic lake system in East
850 Greenland: facies, depositional cycles and palaeoclimate. *Palaeogeography,*
851 *Palaeoclimatology, Palaeoecology*, 140(1), 135-159. PII S0031- 0 1 8 2 (9 8) 0 0 0 4 3
852 – 1.
- 853 Climap, P. M., (1976) The surface of the ice-age Earth. *Science* (New York, NY), 191(4232),
854 1131-1137.
- 855 Crowley, T. J., and North, G.R., (1991) *Paleoclimatology*. Oxford University Press. 349 pp.
- 856 Damborenea, S. E. (2002) Jurassic evolution of Southern Hemisphere marine

857 palaeobiogeographic units based on benthonic bivalves. *Geobios*, M'emoire Sp'ecial
858 35, 51-71.

859 Cullers, R. L. (2002) Implications of elemental concentrations for provenance, redox
860 conditions, and metamorphic studies of shales and limestones near Pueblo, CO, USA.
861 *Chemical Geology*, 191(4), 305-327.

862 Dal Corso, J., Mietto, P., Newton, R.J., Pancost, R.D., Preto, N., Roghi, G. and Wignall, P.B.
863 (2012) Discovery of a major negative $\delta^{13}\text{C}$ spike in the Carnian (Late Triassic) linked
864 to the eruption of Wrangellia flood basalts. *Geology*, 40, 79–82,
865 <https://doi.10.1130/G32473.1>

866 Danilchik, W., (1961) The Iron formation of the Surghar and western Salt Range, Mianwali
867 District, West Pakistan. U. S. Geol. Surv., Prof. Paper 424-D, pp. 228-231.

868 Danilchik, W., and Shah, S. M. I., (1967) Stratigraphic nomenclature of formations in Trans-
869 Indus Mountains, Mianwali District, West Pakistan: U. S. Geol. Surv., Proj. Report (IR)
870 PK-33, 45p.

871 Davis, D M., Lillie, R. J, (1994) Changing mechanical response during continental collision:
872 active examples from the foreland thrust belts of Pakistan. *J Struct Geol* 16:21–34

873 Davies, J. H. F. L., Marzoli, A., Bertrand, H., Youbi, N., Ernesto, M., and Schaltegger, U. (2016)
874 End-Triassic mass extinction started by intrusive CAMP activity. *Nature Comm.*
875 8:15596 DOI: 10.1038/ncomms15596

876 De Caritat, P., Reimann, C., Team, N. P., and Team, G. P. (2012) Comparing results from two
877 continental geochemical surveys to world soil composition and deriving Predicted
878 Empirical Global Soil (PEGS2) reference values. *Earth and Planetary Science Letters*,
879 319, 269-276.

880 Dickinson, W. R., (1970) Interpreting detrital modes of graywacke and arkose: *Jour. Sed.*
881 *Petrology*, v. 40, p. 695-707.

882 Dickinson, W. R., and Suczek, C. A., (1979) Plate tectonics and sandstone compositions: *Am.*
883 *Assoc. Petroleum Geologists Bull.*, v. 63, p. 2164-2182.

884 Dill, H. G. (2016) Kaolin: Soil, rock and ore: From the mineral to the magmatic, sedimentary
885 and metamorphic environments. *Earth-Science Reviews*, 161, 16-129.
886 <http://dx.doi.org/10.1016/j.earscirev.2016.07.003>

887 DiPietro, J. A., (1990) Stratigraphy, structure, and metamorphism near Saidu Sharif, Lower
888 Swat, Pakistan. Ph.D. thesis, Oregon State Univ., Corvallis. 182p.

889 DiPietro, J. A., Pogue, K. R., Hussain, A., and Ahmad, I. (1999) Geologic map of the Indus
890 syntaxis and surrounding area, northwest Himalaya, Pakistan. Special Papers-
891 Geological Society of America, 159-178.

892 Dubiel, R. F., Parrish, J.T., Parrish, J.M., and Good, S.C., (1991) The Pangean megamonsoon
893 — evidence from the Upper Triassic Chinle Formation, Colorado Plateau. *Palaios* 6,
894 347–370. DOI: 10.2307/3514963

895 Duncan, R. A., Hooper, P. R., Rehacek, J., Marsh, J. S., and Duncan, A. R., (1997) The timing
896 and duration of the Karoo igneous event, southern Gondwana. *J. Geophys. Res. Solid*
897 *Earth* 102(B8), 18127–18138.

898 Fatmi, A. N., (1977) Mesozoic; In: *Stratigraphy of Pakistan* (ed.) Shah S M I, GSP Mem. Vol.
899 12 pp. 29–44.

900 Fischer, G., and Wefer, G. (Eds.) (2012) *Use of proxies in paleoceanography: examples from*
901 *the South Atlantic*. Springer Science & Business Media. Pp. 1-735

902 Frakes, L. A., (1979) *Climates Throughout Geologic Time*. Elsevier, Amsterdam. 310 pp.

903 Frakes, L. A, Francis, J. E., Syktus, J. I., (1992) *Climate Modes of the Phanerozoic*. Cambridge.
904 274 pp.

905 Frisia, S. and Wenk, H. R. (1993) TEM and AEM study of pervasive, multi-step dolomitization
906 of the upper Triassic Dolomia Principale (Northern Italy) *J. Sed. Petrol.*, 63, 1049–
907 1058.

908 Fürsich, F. T., Wilmsen, M., Seyed-Emami, K., and Majidifard, M. R. (2009) Lithostratigraphy
909 of the Upper Triassic–Middle Jurassic Shemshak Group of Northern Iran. *Geological*
910 *Society, London, Special Publications*, 312(1), 129-160. DOI: 10.1144/SP312.6

911 Galli, M. T., Jadoul, F., Bernasconi, S. M., and Weissert, H. (2005) Anomalies in global carbon
912 cycling and extinction at the Triassic/Jurassic boundary: evidence from a marine C-
913 isotope record. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 216(3), 203-
914 214. doi:10.1016/j.palaeo.2004.11.009

915 Gazzi, P., (1966) Le arenarie del flysch sopracretaceo delrAppennino modenese; correlazioni
916 con il flysch di Monghidoro: *Mineralogica e Petrografica Acta*, v. 12, p. 69-97.

917 Gazzi, P., Zuffa, G. G., Gandolfi, G., and Paganelli, L., (1973) Provenienza e dispersione
918 litoranea delle sabbie delle spiagge adriatiche fra le foci delrlsonzo e del Foglia:
919 inquadramento regionale: *Memofie della Societa Geologica Italiana*, v. 12, p. 1-37.

920 Gee, E. R., (1945) The age of the saline series of the Punjab and of Kohat. *Proc Indian Acad*
921 *Sci Sect B* 14:269–310

- 922 Goldberg, K., and Humayun, M. (2010) The applicability of the Chemical Index of Alteration
923 as a paleoclimatic indicator: An example from the Permian of the Paraná Basin,
924 Brazil. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 293(1), 175-183.
925 doi:10.1016/j.palaeo.2010.05.015
- 926 Graham S. A., Ingersoll, R. V., and Dickinson, W. R., (1976) Common provenance for lithic
927 grains in Carboniferous sandstones from Ouachita Mountains and Black Warrior
928 basin: *Jour. Sed. Petrology*, v. 46, p. 620-632.
- 929 Gromet, L. P., Haskin, L. A., Korotev, R. L., and Dymek, R. F. (1984) The “North American
930 shale composite”: its compilation, major and trace element characteristics.
931 *Geochimica et Cosmochimica Acta*, 48(12), 2469-2482.
- 932 Greene, S. E., Martindale, R. C., Ritterbush, K. A., Bottjer, D. J., Corsetti, F. A., and Berelson,
933 W. M., (2012a) Recognising ocean acidification in deep time: an evaluation of the
934 evidence for acidification across the Triassic–Jurassic boundary. *Earth Sci. Rev.* 113
935 (1–2), 72–93. DOI: 10.1016/j.earscirev.2012.03.009.
- 936 Ghazi, S., and Mountney, N.P., (2011) Petrography and provenance of the Early Permian
937 fluvial Warchha Sandstone, Salt Range, Pakistan. *Sedimentary Geology*, 233: 88–110.
- 938 Haas, J., and Tardy-Filácz, E. (2004) Facies changes in the Triassic–Jurassic boundary interval
939 in an intraplate basin succession at Csővár (Transdanubian Range, Hungary)
940 *Sedimentary Geology*, 168(1), 19-48. doi:10.1016/j.sedgeo.2004.03.002
- 941 Hallam, A., (1987) Radiations and extinctions in relation to environmental change in the
942 marine Jurassic of north west Europe. *Paleobiology* 13, 152–168.
- 943 Hallam, A. (1988) A re-evaluation of the Jurassic eustasy in the light of new data and the
944 revised Exxon curve. In: Wilgus, C.K., Hastings, B.S., Kendall, G.St.C., Posamentier,
945 H.W., Ross, C.A. & van Wagoner, J.C. (eds) *Sea- Level Changes: an Integrated*
946 *Approach*. Society of Economic Paleontologists and Mineralogists, Special
947 Publications, 42, 71–108. DOI : 10.2110/sepmssp.105.
- 948 Hallam, A. (1997) Estimates of the amount and rate of sea-level change across the Rhaetian–
949 Hettangian and Pliensbachian–Toarcian boundaries (latest Triassic to early Jurassic)
950 *Journal of the Geological Society, London*, 154, 773–779.
- 951 Hallam, A., (2001) A review of the broad pattern of Jurassic sea-level changes and their
952 possible causes in the light of current knowledge. *Palaeogeography,*
953 *Palaeoclimatology, Palaeoecology*, 167(1), 23-37. PII: S0031-0182(00)00229-7
- 954 Hallam, A., and Maynard, J. B., (1987) The iron ores and associated sediments of the Chichali
955 Formation (Oxfordian to Valanginian) of the Trans-Indus Salt Range, Pakistan. *J Geol*
956 *Soc* 44(1):107–114. doi:10.1144/gsjgs.144.1.0107

- 957 Hallam, A., and Wignall, P.B. (1999) Mass extinctions and sea-level changes. *Earth-Science*
958 *Reviews*, 48, 217–250. PII: S 0 0 1 2 - 8 2 5 2 9 9 0 0 0 5 5 - 0
- 959 Hanif, M., Imraz, M., Ali, F., Haneef, M., Saboor, A., Iqbal, S., and Ahmad, S. Jr., (2014) The
960 inner ramp facies of the Thanetian Lockhart Formation, western Salt Range, Indus
961 Basin, Pakistan. *Arabian J Geosci* 7(11):4911–4926. doi:10.1007/s12517-013-1099-7
- 962 Han, W., Hong, H. L., Yin, K., Churchman, G. J., Li, Z. H., and Chen, T. (2014) Pedogenic
963 alteration of illite in subtropical China. *Clay Minerals*, 49(3), 379–390. DOI:
964 10.1180/claymin.2014.049.3.03
- 965 Haq, B. U. (2017) Jurassic Sea-Level Variations: A Reappraisal. *GSA Today*, v. 28, no. 1, doi:
966 10.1130/GSATG359A.1.
- 967 Haq, B. U., Hardenbol, J., and Vail, P. R., (1987). The chronology of fluctuating sea level since
968 the Triassic: *Science*, v. 235, p. 1156–1167,
969 <https://doi.org/10.1126/science.235.4793.1156>.
- 970 Harrison, J. L., and Greenberg, S. S., (1998) Dehydration of fully hydrated halloysite from
971 Lawrence County, Indiana. pp. 374–377.
- 972 Harvey, C., Murray, H., (1990) The geology, mineralogy and exploitation of halloysite clays of
973 Northland, New Zealand. In: Keller (Ed.), *Kaolin Symposium, Kaolin Genesis and*
974 *Utilization*, Publication Clay Mineral Society, pp. 233–248.
- 975 Hesselbo, S. P., Gröcke, D. R., Jenkyns, H. C., Bjerrum, C. J., Farrimond, P., Morgans Bell, H.
976 S., Green, O. R., (2000) Massive dissociation of gas hydrate during a Jurassic oceanic
977 anoxic event. *Nature* 406, 392–395. doi:10.1038/35019044
- 978 Hesselbo, S. P., Robinson, S. A., Surlyk, F., and Piasecki, S., (2002) Terrestrial and marine
979 extinction at the Triassic–Jurassic boundary synchronized with major carbon-cycle
980 perturbation: a link to initiation of massive volcanism? *Geology* 30, 251–254. doi:
981 10.1130/0091-7613(2002)030<0251:TAMEAT>2.0.CO;2
- 982 Hesselbo, S. P., Robinson, S. A., and Surlyk, F. (2004) Sea-level change and facies
983 development across potential Triassic–Jurassic boundary horizons, SW Britain.
984 *Journal of the Geological Society*, 161(3), 365–379.
- 985 Hillebrandt, A.V., Krystyn, L., Kürschner, W.M., Bonis, N.R., Ruhl, M., Richoz, S., Schobben,
986 M.A.N., Urlichs, M., Bown, P.R., Kment, K. and McRoberts, C.A., (2013) The global
987 stratotype sections and point (GSSP) for the base of the Jurassic System at Kuhjoch
988 (Karwendel Mountains, Northern Calcareous Alps, Tyrol, Austria). *Episodes*, 36(3),
989 pp.162–198.
- 990 Hofmann, P., Ricken, W., Schwark, L., and Leythaeuser, D. (2001) Geochemical signature and
991 related climatic-oceanographic processes for early Albian black shales: Site 417D,

- 992 North Atlantic Ocean. *Cretaceous Research*, 22(2), 243-257. DOI:
993 10.1006/cres.2001.0253
- 994 Hornung, T., Krystyn, L. & Brandner, R. (2007) A Tethys-wide mid-Carnian (Upper Triassic)
995 carbonate productivity decline: evidence for the Alpine Reingraben Event from Spiti
996 (Indian Himalaya). *Journal of Asian Earth Sciences*, 30, 285–230,
997 <https://dx.doi.org/10.1016/j.jseaes.2006.10.001>
- 998 Huber, B. T., MacLeod, K. G., and Wing, S.T. (Eds.), (2000) *Warm Climates in Earth History*.
999 Cambridge University Press. 462 pp.
- 1000 Iannace, A. and Frisia, S. (1994) Changing dolomitization styles from Norian to Rhaetian in
1001 southern Tethys realm. In: *A Volume in Honour of Dolomieu* (Eds B. Purser M. Tucker
1002 and D. Zenger), *Int. Assoc. Sedimentol. Spec. Publ.*, 21, 75–89.
- 1003 Ingersoll, R. V., Bullard, T. F., Ford, R. L., Grimm, J. P., Pickle, J. D., and Sares, S. W. (1984)
1004 The effect of grain size on detrital modes: a test of the Gazzi-Dickinson point-
1005 counting method. *Journal of Sedimentary Research*. 54(1), 103-116.
- 1006 Iqbal, S., Jan, U. I., and Hanif, M., (2014) The Mianwali and Tredian Formations: an example
1007 of the Triassic Progradational Deltaic System in the low-latitude western Salt Range,
1008 Pakistan. *Arab J Sci Eng* 39(7):5489–5507. doi:10.1007/s13369-013-0836-2
- 1009 Iqbal, S., Akhter, G., & Bibi, S. (2015a) Structural model of the Balkassar area, Potwar
1010 Plateau, Pakistan. *International Journal of Earth Sciences*, 104(8), 2253-2272. DOI:
1011 10.1007/s00531-015-1180-4.
- 1012 Iqbal, S., Jan, I. U., Akhter, M. G., and Bibi, M. (2015b) Palaeoenvironmental and sequence
1013 stratigraphic analyses of the Jurassic Datta Formation, Salt Range, Pakistan. *Journal*
1014 *of Earth System Science*, 124(4), 747-766. doi:10.1007/s12040-015-0572-y.
- 1015 Iqbal, S., and Wagreich, M. (2016) Evidences of marine regression and Pangaea breakup:
1016 Triassic-Jurassic boundary from the Tethyan Salt Range, Pakistan. 35th IGC Capetown
1017 South Africa Paper Number: 1979.
- 1018 Iqbal, S., Wagreich, M., Jan, I., Kürschner, W. M., & Gier, S. (2017) Geochemical evidences
1019 for palaeoclimatic fluctuations at the Triassic-Jurassic boundary: southwestern
1020 margin of the Neotethys in the Salt Range, Pakistan. In *EGU General Assembly*
1021 *Conference Abstracts* (Vol. 19, p. 408).
- 1022 Jacobs, D. K., Sahagian, D. L. (1993) Climate induced fluctuations in sea level during
1023 nonglacial times. *Nature* 361,710–712.
- 1024 Jadoon, I. A. K., Kemal, A., Frisch, W., and Jaswal, T. M, (1997) Thrust geometries and
1025 kinematics in the Himalayan foreland (North Potwar Deformed Zone), North
1026 Pakistan. *Geolog Rundsch* 86:120–131.

- 1027 Jain, A. K., (2014) When did India–Asia collide and make the Himalaya? *Curr Sci* 106(2):254–
1028 266.
- 1029 Jenkyns, H. C., (1988) The early Toarcian (Jurassic) anoxic event: stratigraphic, sedimentary
1030 and geochemical evidence. *Am. J. Sci.* 288, 101–151.
- 1031 Jenkyns, H. C., (1999) Mesozoic anoxic events and palaeoclimate. *Zentralbl. Geol.*
1032 *Palaeontol.*, Teil 1 7–9, 943–949.
- 1033 Kadri, I B., (1995) Petroleum geology of Pakistan. Pakistan Petroleum Limited, Karachi Pak,
1034 pp 40–61.
- 1035 Kazmi, A. H., Jan, M. Q., (1997) Geology and tectonic of Pakistan. Graphic publication,
1036 Karachi, pp 130–141.
- 1037 Keller, W. D., Westcott, J. F., and Bledsoe, A. O., (1954) The origin of Missouri fire clays. In
1038 2nd National Conference on Clays and Clay Minerals (pp. 7-46).
- 1039 Kent, D. V., and Olsen, P. E., (2000) Magnetic polarity stratigraphy and paleolatitude of the
1040 Triassic Jurassic Blomidon Formation in the Fundy basin (Canada): implications for
1041 early Mesozoic tropical climate gradients. *Earth and Planetary Science Letters* 179,
1042 311–324. [https://doi.org/10.1016/S0012-821X\(00\)00117-5](https://doi.org/10.1016/S0012-821X(00)00117-5).
- 1043 Kidder, D. L., & Worsley, T. R. (2010). Phanerozoic large igneous provinces (LIPs), HEATT
1044 (haline euxinic acidic thermal transgression) episodes, and mass extinctions.
1045 *Palaeogeography, Palaeoclimatology, Palaeoecology*, 295(1), 162-191.
1046 <https://doi.org/10.1016/j.palaeo.2010.05.036>.
- 1047 Klett, T. R., Ulmishek, G. F., Wandrey, C. J., and Agena, W. F., (2006) Assessment of
1048 undiscovered technically recoverable conventional petroleum resources of northern
1049 Afghanistan. US Geological Survey Open-File Report, 2006, 1253.
- 1050 Korte, C., Hesselbo, S. P., Jenkins, H. C., Rickaby, and R. E. M., Spotl, C., (2009)
1051 Palaeoenvironmental significance of carbon- and oxygen-isotope stratigraphy of
1052 marine Triassic–Jurassic boundary sections in SW Britain. *Journal of the Geological*
1053 *Society* 166, 431–445. doi: 10.1144/0016-76492007-177.
- 1054 Kuerschner, W. M., Bonis, N. R., and Krystyn, L. (2007) Carbon-isotope stratigraphy and
1055 palynostratigraphy of the Triassic–Jurassic transition in the Tiefengraben section—
1056 Northern Calcareous Alps (Austria) *Palaeogeography, Palaeoclimatology,*
1057 *Palaeoecology*, 244(1), 257-280. doi:10.1016/j.palaeo.2006.06.031
- 1058 Kuerschner, W. M., Iqbal, S., Wagreich, M., Jan, I. U., and Gier, S. (2017) Palynology and
1059 Carbon isotope stratigraphy of the Triassic-Jurassic transition in the Salt Range
1060 (Pakistan). In *EGU General Assembly Conference Abstracts* (Vol. 19, p. 15291).

1061 Kutzbach, J. E., and Gallimore, R. G., (1989) Pangaeen climates: megamonsoons of the
 1062 megacontinent. *Journal of Geophysical Research* 94, 3341–3358. DOI:
 1063 10.1029/JD094iD03p03341.

1064 LeTourneau, P. M., and Huber, P. (2006) Early Jurassic eolian dune field, Pomperaug basin,
 1065 Connecticut and related synrift deposits: Stratigraphic framework and paleoclimatic
 1066 context. *Sedimentary Geology*, 187(1), 63-81. doi:10.1016/j.sedgeo.2005.12.005.

1067 Lindström, S., van de Schootbrugge, B., Hansen, K.H., Pedersen, G.K., Alsen, P., Thibault, N.,
 1068 Dybkjær, K., Bjerrum, C.J. and Nielsen, L.H., (2017) A new correlation of Triassic–
 1069 Jurassic boundary successions in NW Europe, Nevada and Peru, and the Central
 1070 Atlantic Magmatic Province: A time-line for the end-Triassic mass extinction.
 1071 *Palaeogeography, Palaeoclimatology, Palaeoecology*, 478, pp.80-102.

1072 López-Gómez, J., Escudero-Mozo, M.J., Martín-Chivelet, J., Arche, A., Lago, M. and Galé, C.
 1073 (2017) Western Tethys continental-marine responses to the Carnian Humid Episode:
 1074 Palaeoclimatic and palaeogeographic implications. *Global and Planetary Change*,
 1075 148, 79–95, <https://doi.org/10.1016/j.gloplacha.2016.11.016>

1076 Luo, M., Chen, Z. Q., Shi, G. R., Fang, Y., Song, H., Jia, Z., Huang, Y. and Yang, H., (2016)
 1077 Upper Lower Triassic stromatolite from Anhui, South China: geobiologic features and
 1078 paleoenvironmental implications. *Palaeogeography, Palaeoclimatology,*
 1079 *Palaeoecology*, 452, pp.40-54. <https://doi.org/10.1016/j.palaeo.2016.04.008>

1080 Malick, B. M. L., and Ishiga, H. (2016) Geochemical Classification and Determination of
 1081 Maturity Source Weathering in Beach Sands of Eastern San'in Coast, Tango
 1082 Peninsula, and Wakasa Bay, Japan. *Earth Science Research*, 5(1), 44. DOI:
 1083 <http://dx.doi.org/10.5539/esr.v5n1p44>

1084 Martin, A. J. (2017) A review of Himalayan stratigraphy, magmatism, and structure.
 1085 *Gondwana Research*. 49 (42-80) <http://dx.doi.org/10.1016/j.gr.2017.04.031>

1086 Marzoli, A., Renne, P. R., Piccirillo, E. M., Ernesto, A., Bellieni, G., and De Min, A., (1999)
 1087 Extensive 200-million-year-old continental flood basalts of the Central Atlantic
 1088 magmatic province. *Science* 284, 616– 618.

1089 Marzoli, A., Bertrand, H., Knight, K. B., Cirilli, S., Buratti, N., Ve´rati, C., Nomade, S., Renne, P.
 1090 R., Youbi, N., Martini, R., Allenbach, K., Neuwerth, R., Rapaille, C., Zaninetti, L., and
 1091 Bellieni, G., (2004) Synchrony of the Central Atlantic magmatic province and the
 1092 Triassic-Jurassic boundary climatic and biotic crisis. *Geology* 32, 973– 976. Doi:
 1093 10.1130/G20652.1

1094 McElwain, J. C., Beerling, D. J., ad Woodward, F. I., (1999) Fossil plants and global warming
 1095 at the Triassic–Jurassic boundary. *Science* 285, 1386–1390. DOI:
 1096 10.1126/science.285.5432.1386

- 1097 McElwain, J. C., Popa, M. E., Hesselbo, S. P., Haworth, M., and Surlyk, F., (2007)
 1098 Macroecological response of terrestrial vegetation to climatic and atmospheric
 1099 change across the Triassic/Jurassic boundary in East Greenland. *Paleobiology* 33,
 1100 547–573. <https://doi.org/10.1666/06026.1>
- 1101 McLennan, S. M., Hemming, S., McDaniel, D. K., Hanson, G. M., (1993) Geochemical
 1102 approaches to sedimentation, provenance, and tectonics. In: Johnsson, M.J., Basu, A.
 1103 (Eds.), *Processes Controlling the Composition of Clastic Sediments*. Geological
 1104 Society of America, Special Papers, vol. 284, pp. 21-40. DOI: 10.1130/SPE284-p21
- 1105 Meister, P., McKenzie, J. A., Bernasconi, S. M. and Brack, P. (2013) Dolomite formation in the
 1106 shallow seas of the Alpine Triassic. *Sedimentology*, 60, 270–291. doi:
 1107 10.1111/sed.12001
- 1108 Meyen, S. V., (1997) Permian conifers of western Angaraland. Review of Palaeobotany and
 1109 Palynology 96, 351–447. [https://doi.org/10.1016/S0034-6667\(96\)00046-2](https://doi.org/10.1016/S0034-6667(96)00046-2)
- 1110 Michalík, J., Lintnerová, O., Wójcik-Tabol, P., Gaździcki, A., Grabowski, J., Golej, M., Šimo, V.
 1111 and Zahradníková, B., (2013) Paleoenvironments during the Rhaetian transgression
 1112 and the colonization history of marine biota in the Fatric Unit (Western Carpathians)
 1113 *Geologica Carpathica*, 64(1), pp.39-62. doi: 10.2478/geoca-2013-0003
- 1114 Mikutta, R., Kleber, M., Kaiser, K., and Jahn, R. (2005) Review. Soil science society of America
 1115 journal, 69(1), 120-135. DOI: 10.2136/sssaj2005.0120
- 1116 Mindszenty, A. (2016) Mindszenty, A. (2016). Bauxites: Feedbacks of System Earth at
 1117 Greenhouse times. *Geologia Croatica*, 69(1), 79-87. doi:10.4154/gc.2016.07
- 1118 Mohadjer, S., Bendick, R., Ischuk, A., Kuzikov, S., Kostuk, A., Saydullaev, U., Lodi, S., Kakar, D.
 1119 M., Wasy, A., Khan, M. A., Molnar, P., Bilham, R., and Zubovich, A. V., (2010)
 1120 Partitioning of India-Eurasia convergence in the Pamir-Hindu Kush from GPS
 1121 measurements. *Geophys Res Lett* 37:L04305. doi:10.1029/2009GL041737
- 1122 Molnar, P., and Tapponnier, P., (1981) A possible dependence of tectonic strength on the
 1123 age of the crust in Asia. *Earth Planet Sci Lett* 52:107–114
- 1124 Moore, D., and Reynolds, R.C. Jr. (1997) *X-Ray Diffraction and the Identification and*
 1125 *Analysis of Clay Minerals*, second ed.: Oxford University Press, New York, 378 p.
- 1126 Mount, J. F., (1984) Mixing of siliciclastic and carbonate sediments in shallow shelf
 1127 environments; *Geology* 12(7) 432–435.
- 1128 Mukhopadhyay, G., Mukhopadhyay, S. K., Roychowdhury, M., and Parui, P. K. (2010)
 1129 Stratigraphic correlation between different Gondwana basins of India. *Journal of the*
 1130 *Geological Society of India*, 76(3), 251-266.

- 1131 Nesbitt, H. W., and Young, G.M., (1982) Early Proterozoic climates and plate motions
1132 inferred from major element chemistry of lutites. *Nature* 299, 715-717.
- 1133 Nesbitt, H. W. and G. M. Young, (1984) Prediction of some weathering trends of plutonic
1134 and volcanic rocks based on thermodynamic and kinetic considerations. *Geochim.*
1135 *Cosmochim. Acta*, 48: 1523-1534.
- 1136 Nesbitt, H. W., and Young, G. M., (1989) Formation and diagenesis of weathering profiles.
1137 *Journal of Geology* 97, 129e-147. DOI: 10.1086/629290
- 1138 Nichols, G. (2009) *Sedimentology and stratigraphy*. John Wiley & Sons. pp. 90-133
- 1139 Nichols, G. (2000) *Sedimentology and stratigraphy*, Blackwell Science. pp. 75–145.
- 1140 Olsen, P. E., (1999) Giant lava flows, mass extinctions, and mantle plumes. *Science*, 604–
1141 605. DOI: 10.1126/science.284.5414.604
- 1142 Olsen, P. E., and Kent, D. V., (1996) Milankovitch climate forcing in the tropics of Pangea
1143 during the Late Triassic. *Palaeogeography, Palaeoclimatology, Palaeoecology* 122, 1–
1144 26.
- 1145 Olsen, P. E., and Kent, D. V., (2000) High resolution early Mesozoic Pangean climatic transect
1146 in lacustrine environments. *Zentralblatt für Geologie und Paläontologie* 11–12,
1147 1475–1496. DOI: 0340-5109/00/1998-147
- 1148 Pálffy, J., Parrish, R. R., and Smith, P. L., (1997) A U–Pb age from the Toarcian (Lower Jurassic)
1149 and its use for time scale calibration through error analysis of biochronologic dating.
1150 *Earth Planet. Sci. Lett.* 146, 659–675. PII S0012-821X 96 00241-5
- 1151 Pálffy, J., Mortensen, J. K., Carter, E. S., Smith, P. L., Friedman, R. M., and Tipper, H. W.,
1152 (2000) Timing of the end Triassic mass extinction: first on land, then in sea? *Geology*
1153 28, 39– 42.
- 1154 Pálffy, J., and Smith, P. L., (2000) Synchrony between Early Jurassic extinction, oceanic anoxic
1155 event, and the Karoo–Ferrar flood basalt volcanism. *Geology* 28, 747–750.
- 1156 Pálffy, J., Deme'ny, A., Haas, J., Hete'nyi, M., Orchard, M.J., and Vetl, I., (2001) Carbon
1157 isotope anomaly and other geochemical changes at the Triassic–Jurassic boundary
1158 from a marine section in Hungary. *Geology* 29, 1047–1050.
- 1159 Pálffy, J., and Zajzon, N., (2012) Environmental changes across the Triassic–Jurassic boundary
1160 and coeval volcanism inferred from elemental geochemistry and mineralogy in the
1161 Kendlbachgraben section (Northern Calcareous Alps, Austria) *Earth Planet. Sci. Lett.*
1162 335–336, 121–134. DOI:10.1016/j.epsl.2012.01.039

- 1163 Parrish, J. T., (1993) Climate of the supercontinent Pangea. *Journal of Geology* 101, 215–
1164 233.
- 1165 Pettijohn, F. J., and Potter, P. E. (1987) Siever, R.: Sand and sandstone. 2nd edition. Springer
1166 Verlag, p. 533.
- 1167 Pieńkowski, G., Niedźwiedzki, G., and Waksmundzka, M., (2012) Sedimentological,
1168 palynological and geochemical studies of the terrestrial Triassic–Jurassic boundary in
1169 northwestern Poland. *Geological Magazine*, 149(02), 308–332. DOI:
1170 10.1017/S0016756811000914
- 1171 Preto, N., Breda, A., Dal Corso, J., Spötl, C., Zorzi, F., and Frisia, S., (2015) Primary dolomite in
1172 the Late Triassic Travenanzes Formation, Dolomites, Northern Italy: facies control
1173 and possible bacterial influence. *Sedimentology*, 62(3), 697–716. DOI:
1174 10.1111/sed.12157
- 1175 Preto, N., Kustatscher, E., and Wignall, P., (2010) Triassic climates—state of art.
1176 *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 290, 1–10.
1177 <http://dx.doi.org/10.1016/j.palaeo.2010.03.015>.
- 1178 Polyak, V. W. C., and Guven, N., (1996) Alunite, natroalunite, and hydrated halloysite in
1179 Carlsbad Cavern. *Clay Clay Miner.* 44, 843–850.
- 1180 Poulsen, C. J. (2009) Paleoclimate modeling, pre-quatarnary. In *Encyclopedia of*
1181 *Paleoclimatology and Ancient Environments* (pp. 700–709) Springer Netherlands.
- 1182 Pogue, K. R., DiPietro, J. A., Khan, S. R., Hughes, S. S., Dilles, J. H., and Lawrence, R. D. (1992)
1183 Late Paleozoic rifting in northern Pakistan. *Tectonics*, 11(4), 871–883. DOI:
1184 10.1029/92TC00335.
- 1185 Raup, D. M., and Sepkoski Jr., J. J., (1982) Mass extinction in the marine fossil record.
1186 *Science* 215, 1501–1503.
- 1187 Reading, H. G. (Ed.) (2009) *Sedimentary environments: processes, facies and stratigraphy*.
1188 John Wiley & Sons.
- 1189 Rees, P. M., Gibbs, M. T., Ziegler, A. M., Kutzbach, J. E., and Behling, P. J., (1999) Permian
1190 climates: Evaluating model predictions using global paleobotanical data; *Geology* 27
1191 891–894.
- 1192 Rees, P. M., Ziegler, A. M., and Valdes, P. J., (2000) Jurassic phytogeography and climates:
1193 new data and model comparisons; In: *Warm climates in Earth history* (eds) Huber B
1194 T, Macleod K G and Wing S L, Cambridge University Press, pp. 297–318.

- 1195 Rees, P. M., Ziegler, A. M., Gibbs, M. T., Kutzbach, J. E., Behling, P. J, and Rowley, D. B.,
1196 (2002) Permian phytogeographic patterns and climate data/model comparisons; J.
1197 Geol. 110 1–31.
- 1198 Retallack, G.J. (2010) Lateritization and Bauxitization Events. *Econ. Geol.*, 105, 655–667. doi:
1199 10.2113/gsecongeo.105.3.655
1200 Roser, B. P., and Korsch, R. J., (1988) Provenance
1201 signatures of sandstone-mudstone suites determined using discriminant function
1202 analysis of major-element data. *Chemical geology*, 67(1-2), 119-139. DOI:
10.1016/0009-2541(88)90010-1
- 1203 Ruhl, M., W. M. Kürschner, and L. Krystyn. (2009) Triassic-Jurassic organic carbon isotope
1204 stratigraphy of key sections in the western Tethys realm (Austria). *Earth and*
1205 *Planetary Science Letters* 281:169–187. <https://doi.org/10.1016/j.epsl.2009.02.020>
- 1206 Ruhl, M., N. R. Bonis, G.-J. Reichart, J. S. Sinninghe-Damsté, and W. M. Kürschner. (2011)
1207 Atmospheric Carbon injection linked to end-Triassic mass extinction. *Science* 333,
1208 430–434. DOI: 10.1126/science.1204255
- 1209 Rudnick, R. L., and Gao, S. (2003) Composition of the continental crust. *Treatise on*
1210 *geochemistry*, 3, 659.
- 1211 Ruffell, A., Simms, M. J., and Wignall, P. B., (2016) The Carnian Humid Episode of the late
1212 Triassic: a review. *Geological Magazine*, 153(02), 271-284. DOI:
1213 <https://doi.org/10.1017/S0016756815000424>.
- 1214 Ruhl, M., Bonis, N. R., Reichart, G. J., Damsté, J. S. S., and Kuerschner, W. M., (2011)
1215 Atmospheric carbon injection linked to end-Triassic mass extinction. *Science* 333
1216 (6041), 430–434. <http://dx.doi.org/10.1126/science.1204255>.
- 1217 Ruckwied, K., and Götz, A. (2009) Climate change at the Triassic/Jurassic boundary:
1218 palynological evidence from the Furkaska section (Tatra Mountains, Slovakia)
1219 *Geologica Carpathica*, 60(2), 139-149. doi: 10.2478/v10096-009-0009-0.
- 1220 Sa’ad, Z. A. and Kedar Al-Mashaikie (2017) Tectonostratigraphic Study of Carbonate Breccias
1221 (Calciturbidites) in the Upper Triassic Baluti Formation (Northern Iraq): New Insights
1222 on Tethyan Geodynamics. In *Lithosphere Dynamics and Sedimentary Basins of the*
1223 *Arabian Plate and Surrounding Areas*. Springer International Publishing. pp. 67-86
- 1224 Schaller, M.F., Wright, J.D., Kent, D.V., (2011) Atmospheric PCO₂ perturbations associated
1225 with the Central Atlantic Magmatic Province. *Science*, 331, 1404-1409.
- 1226 Schettino, A., Turco, E., (2011) Tectonic history of the western Tethys since the Late Triassic.
1227 *Geological Society of America Bulletin* 123, 89-105.

- 1228 Schoene, B., Guex, J., Bartolini, A., Schaltegger, U., Blackburn, T.J., (2010) Correlating the
1229 end-Triassic mass extinction and flood basalt volcanism at the 100 ka level. *Geology*
1230 38, 387–390. DOI:<https://doi.org/10.1130/G30683.1>
- 1231 Schultz, L.G. (1964) Quantitative interpretation of mineralogical composition from X-ray and
1232 chemical data for the Pierre shale. US Geological Survey Prof. Paper 391-C,
1233 Washington.
- 1234 Selley, C. R., (2000) *Applied Sedimentology*: 2nd Ed. Academic Press. 182-207.
- 1235 Sellwood, B. W., and Valdes, P. J. (2006) Mesozoic climates: General circulation models and
1236 the rock record. *Sedimentary geology*, 190(1), 269-287.
1237 doi:10.1016/j.sedgeo.2006.05.013
- 1238 Selvaraj, K.; Ram Mohan, V., and Szefer, P., (2004) Evaluation of metal contamination in
1239 coastal sediments of the Bay of Bengal, India: geochemical and statistical
1240 approaches. *Mar. Pollut. Bull.* 49:174–185.
- 1241 Selvaraj, K., and Chen, C. T. A., (2006) Moderate chemical weathering of subtropical Taiwan:
1242 constraints from solid-phase geochemistry of sediments and sedimentary rocks. *The*
1243 *Journal of Geology*, 114(1), 101-116. DOI:0022-1376/2006/11401-0006
- 1244 Shah, S. M. I., (2009) *Stratigraphy of Pakistan*. Gov Pak Ministry of Petrol Nat Resour Geol
1245 Sur Pak 245–273
- 1246 Simms, M. J., and Ruffell, A. H. (1989) Synchronicity of climatic change and extinctions in the
1247 Late Triassic. *Geology*, 17(3), 265-268. doi: 10.1130/0091-7613(1989)
1248 017<0265:SOCCAE>2.3.CO;2
- 1249 Singer, A. (1984) The paleoclimatic interpretation of clay minerals in sediments—a review.
1250 *Earth-Science Reviews*, 21(4), 251-293. [https://doi.org/10.1016/0012-](https://doi.org/10.1016/0012-8252(84)90055-2)
1251 [8252\(84\)90055-2](https://doi.org/10.1016/0012-8252(84)90055-2)
- 1252 Smoot, J. P., Olsen, P. E., (1988) Massive mudstones in basin analysis and paleoclimatic
1253 interpretation of the Newark Supergroup. In: Manspeizer, W. (Ed.), *Triassic–Jurassic*
1254 *Rifting, Continental Breakup and the Origin of the Atlantic Ocean and Passive*
1255 *Margins*. Elsevier, New York, pp. 249–274.
- 1256 Song, E., Zhang, K., Chen, J., Wang, C., Jiang, G., Yin, K., Hong, H. and Churchman, J. G., (2014)
1257 *Clay Mineralogy and its Paleoclimatic Significance of the Oligocene-Miocene*
1258 *Sediments in the Gerze Basin, Tibet. Acta Geologica Sinica (English Edition)*, 2014,
1259 pp.7-23. DOI: 10.1111/1755-6724.12320
- 1260 Suttner, L. J., and Dutta, P. K. (1986) Alluvial sandstone composition and paleoclimate, I.
1261 *Framework mineralogy. Journal of Sedimentary Research*, 56(3) 329-345

- 1262 Taylor, S. R. and McLennan, S. M., (1985) The Continental Crust: Its Composition and
1263 Evolution. Blackwell, Oxford, ISBN-13: 978-0632011483, pp: 312.
- 1264 Therrien, F., and Fastovsky, D. E. (2000) Paleoenvironments of Early Theropods, Chinle
1265 Formation (Late Triassic), Petrified Forest National Park, Arizona. *Palaios*, 15(3), 194-
1266 211.
- 1267 Thiry, M. (2000) Palaeoclimatic interpretation of clay minerals in marine deposits: an
1268 outlook from the continental origin. *Earth-Science Reviews*, 49(1), 201-221.
1269 [https://doi.org/10.1016/S0012-8252\(99\)00054-9](https://doi.org/10.1016/S0012-8252(99)00054-9)
- 1270 Tucker, M. E. (Ed.) (2009) *Sedimentary petrology: an introduction to the origin of*
1271 *sedimentary rocks*. John Wiley & Sons.
- 1272 Valdes, P. J., Spicer, R. A., Sellwood, B. W., Palmer, D. C., (1999) *Understanding Past*
1273 *Climates: Modelling Ancient Weather*. Routledge. CD-ROM.
- 1274 van de Schootbrugge, B., Quan, T., Lindström, S., Püttmann, W., Heunisch, C., Pross, J.,
1275 Fiebig, J., Petschick, R., Röhling, H.-G., Richoz, S., et al., (2009) Floral changes across
1276 the Triassic/Jurassic boundary linked to flood basalt volcanism. *Nat. Geosci.* 2 (8),
1277 589–594. DOI: 10.1038/ngeo577.
- 1278 Van Langeveld, A. D., Van der Gaast, S. J., and Eisma, D. (1978) Comparison of the
1279 effectiveness of eight methods for the removal of organic matter from clay. *Clays*
1280 *and Clay Minerals*, 26(5), 361-4.
- 1281 Wang, P.X., (2009) Global monsoon in a geological perspective. *Chinese Science Bulletin* 54,
1282 1113–1136. doi: 10.1007/s11434-009-0169-4
- 1283 Weaver, C. E., (1958) A discussion on the origin of clay minerals in sedimentary rocks. *Proc.*
1284 *5th Natl. Conf., Clay and Clay Minerals*, pp. 159–173.
- 1285 Wefer, G., Berger, W. H., Bijma, J., and Fischer, G. (1999) Clues to ocean history: a brief
1286 overview of proxies. In *Use of proxies in paleoceanography* (pp. 1-68) Springer Berlin
1287 Heidelberg.
- 1288 Whitehouse, U. G., Jeffrey, L. M., Debrecht, J. D., (1960) Differential settling tendencies of
1289 clay minerals in artificial sea water. *Proc. 7th Natl. Conf., Clays and Clay Minerals*, pp.
1290 1–80.
- 1291 Whiteside, J. H., Grogana, D. S., Olsen, P. E., Kent, D. V. (2011) Climatically driven
1292 biogeographic provinces of Late Triassic tropical Pangea. *PNAS*, 108(22), 8972–8977.
1293 doi/10.1073/pnas.1102473108
- 1294 Wignall, P. B., (2001) Large igneous provinces and mass extinctions. *Earth-Sci. Rev.* 53, 1 –
1295 33. PII: S0012-8252Ž00.00037-4

- 1296 Wilmsen, M., Fürsich, F. T., Seyed-Emami, K., and Majidifard, M. R. (2009) An overview of
1297 the stratigraphy and facies development of the Jurassic System on the Tabas Block,
1298 east-central Iran. Geological Society, London, Special Publications, 312(1), 323-343.
1299 DOI: 10.1144/SP312.15
- 1300 Wood, L. J., (1990) Effects of basin type on coastal plain shelf slope systems during base
1301 level fluctuations: An experimental approach; Am. Assoc. Petrol. Geol. Bull. (abstract)
1302 75 969.
- 1303 Worden, R. H., and Burley, S. D. (2003) Sandstone diagenesis: the evolution of sand to stone.
1304 Sandstone Diagenesis: Recent and Ancient, 4, 3-44.
- 1305 Yan, D., Chen, D., Wang, Q., and Wang, J., (2010) Large-scale climatic fluctuations in the
1306 latest Ordovician on the Yangtze block, south China. *Geology* 38, 599-602. DOI:
1307 10.1130/G30961.1
- 1308 Zachos, J., Pagani, M., Sloan, L., Thomas, E., and Billups, K. (2001) Trends, rhythms, and
1309 aberrations in global climate 65 Ma to present. *Science*, 292(5517), 686-693.
- 1310 Ziegler, A. M., Eshet, G., McAllister Rees, P., Rothfus, T.A., Rowley, D.B., and Sunderlin, D.,
1311 (2003) Tracing the tropics across land and sea: Permian to present. *Lethaia* 36, 227–
1312 254. DOI: 10.1080/00241160310004657
- 1313 Ziegler, A. M., Scotese, C. R., and Barrett, S. F., (1983) Mesozoic and Cenozoic
1314 paleogeographic maps, Tidal Friction and the Earth's Rotation, 2nd edition. Springer-
1315 Verlag, Berlin, pp. 240–252.

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1317 **Figures Captions:**

1318 Figure 1. Current and palaeogeographic location of the Salt Range, Pakistan. (a) Early
1319 Jurassic palaeogeographic location of the study area. (b) Tectonic map of north
1320 Pakistan with the location of Salt Range–Kohat–Potwar Plateau. (c) Salt and Trans–
1321 Indus ranges with locations of the studied sections (modified after Rees et al. 2000;
1322 Iqbal et al., 2015b).

1323 Figure 2. (a) Representative measured stratigraphic sections in the Salt and Trans–Indus
1324 ranges. The distance between each section is displayed in kilometres. (b)

Representative section (Nammal Gorge) with various geochemical proxies plots highlighting the palaeoclimatic fluctuation in the section. CBFA = Channel Belt Facies Association, CMOFA = Channel Margin and Overbank Facies Association, LFA = Lagoonal Facies Association (see Fig. 2a for symbols and codes). The gray shaded area in the background indicates the Triassic–Jurassic transition.

Figure 3. Representative field photographs from the measured sections. (a) Field view of the Kingriali and Datta formations at Paniala anticline section (Khisor Range). (b) intraclasts in the upper part of the Kingriali Formation at Kasanwala Section. (c) Triassic–Jurassic boundary in Paniala anticline section. (d) dark shale, gypsum and rusty brown dolomite in Paniala anticline section (e) Field view of the facies and facies associations in the Datta Formation at Kasanwala Section. (f) plant remains in the Datta Formation at Nammal Gorge. (Abbreviations are explained in the text and Figure 2).

Figure 4. Representative thin section photomicrographs from the measured sections. (a, b) bivalves (Bi) and gastropods (Gp) in the upper part of Kingriali Formation at Kasanwala Section. (c, d, e) rock fragments (Rf), monocrystalline quartz with unit extinction (Qmue), monocrystalline quartz with undulose extinction (Qmuu), polycrystalline quartz (Qp), phyllosilicates (phy) in the Datta Formation. (f) dolomite with quartz pulses in Kaowaali section.

Figure 5. (a) Quartz-Feldspar-Lithic fragments (QFL) plot for the representative samples of the Datta Formation. (b) log – log plot of polycrystalline quartz (Q_p)/(F+R) vs. total quartz Q_{total} /(F+R).

Figure 6. (a) Al_2O_3 Vs. Chemical Index of Alteration ($\text{CIA}_{\text{molar}}$) plot for the representative samples. (b) $\text{K}_2\text{O}/\text{Na}_2\text{O}$ Vs. $\text{CIA}_{\text{molar}}$ plot for representative samples.

Figure 7. (a) A–CN–K plot for the representative samples, positions of reference minerals (muscovite, illite, smectite and kaolinite) are indicated on the plot (Basu, 1976; Nesbitt et al., 1996). The Triassic and Jurassic samples are separated in the lower part of the plot and the CIA range (dotted and dashed lines) and average (black dots) is plotted to lift for both the Triassic and Jurassic samples. (b) Compositional maturity proxy plot for $\text{SiO}_2\%$ vs. $(\text{Al}_2\text{O}_3+\text{K}_2\text{O}+\text{Na}_2\text{O})\%$ showing palaeoclimatic conditions during the deposition.

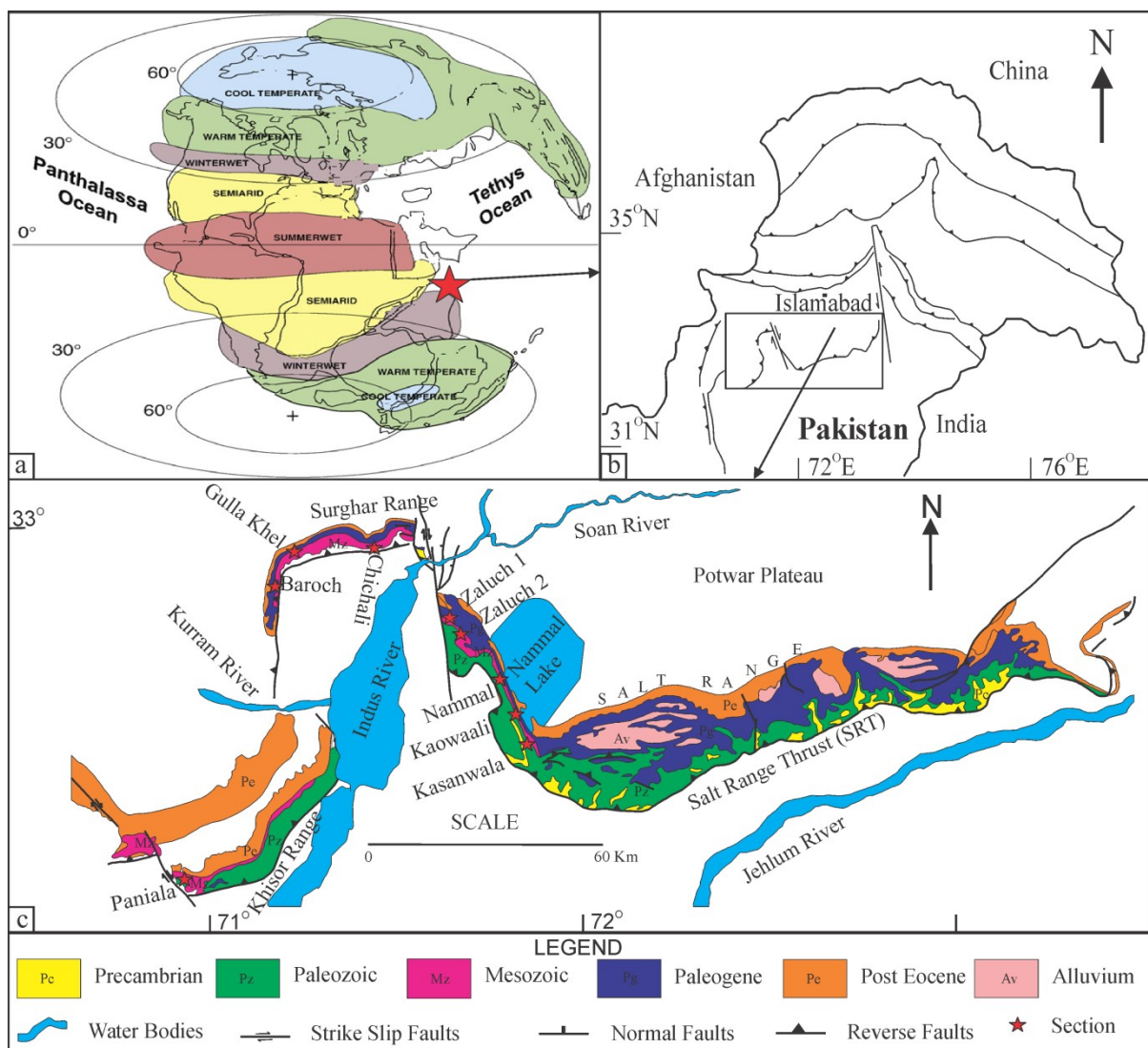
Figure 8. Weathering and palaeoclimate trends across the Triassic–Jurassic transition, reconstructed from various proxies (CIA , K^+/Na^+ , Al_2O_3 , kaolinite/illite) together with sea-level fluctuations (Iqbal et al., 2015b; Haq, 2017) for the representative Nammal Gorge section (Salt Range). Note: the dashed lines in the proxies plots are the actual values whereas the solid lines represent the general trend in each case. The increasing intensity of colour shades to the right indicates increasing values to the right side. The gray shaded area in the background indicates the Triassic–Jurassic transition.

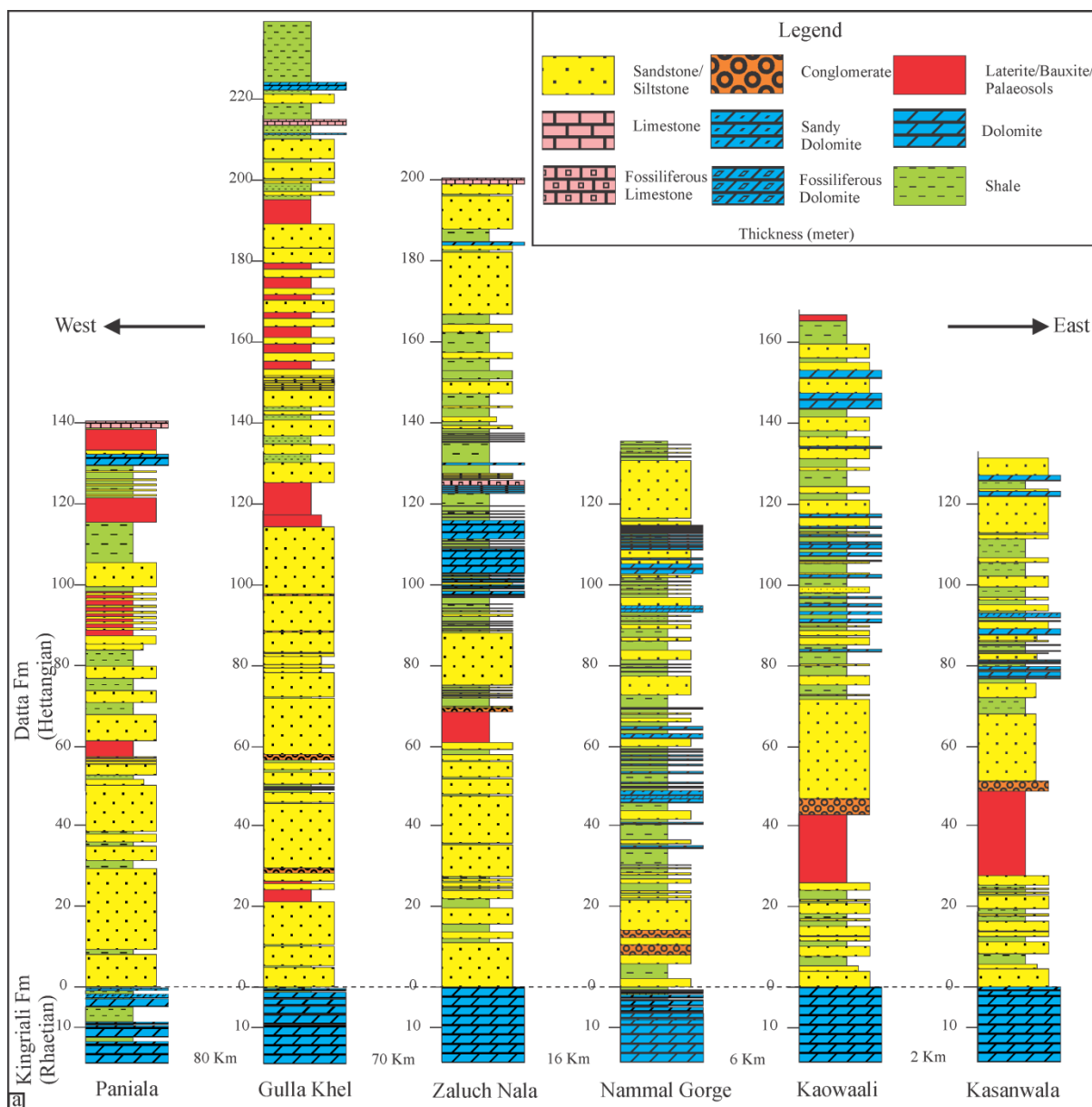
Figure 9. A general global correlation of the geochemical results (Nammal Gorge) with the base-Jurassic Global Stratotype Sections and Point (GSSP) at Kuhjoch (Austria) and other famous sections (modified from Michalík et al., 2010; Pálffy and Zajzon, 2012; Hillebrandt et al., 2013; Lindström et al., 2017). With the exception of Furkaska (Tatra Mountain) all the other sections are on the same thickness scale.

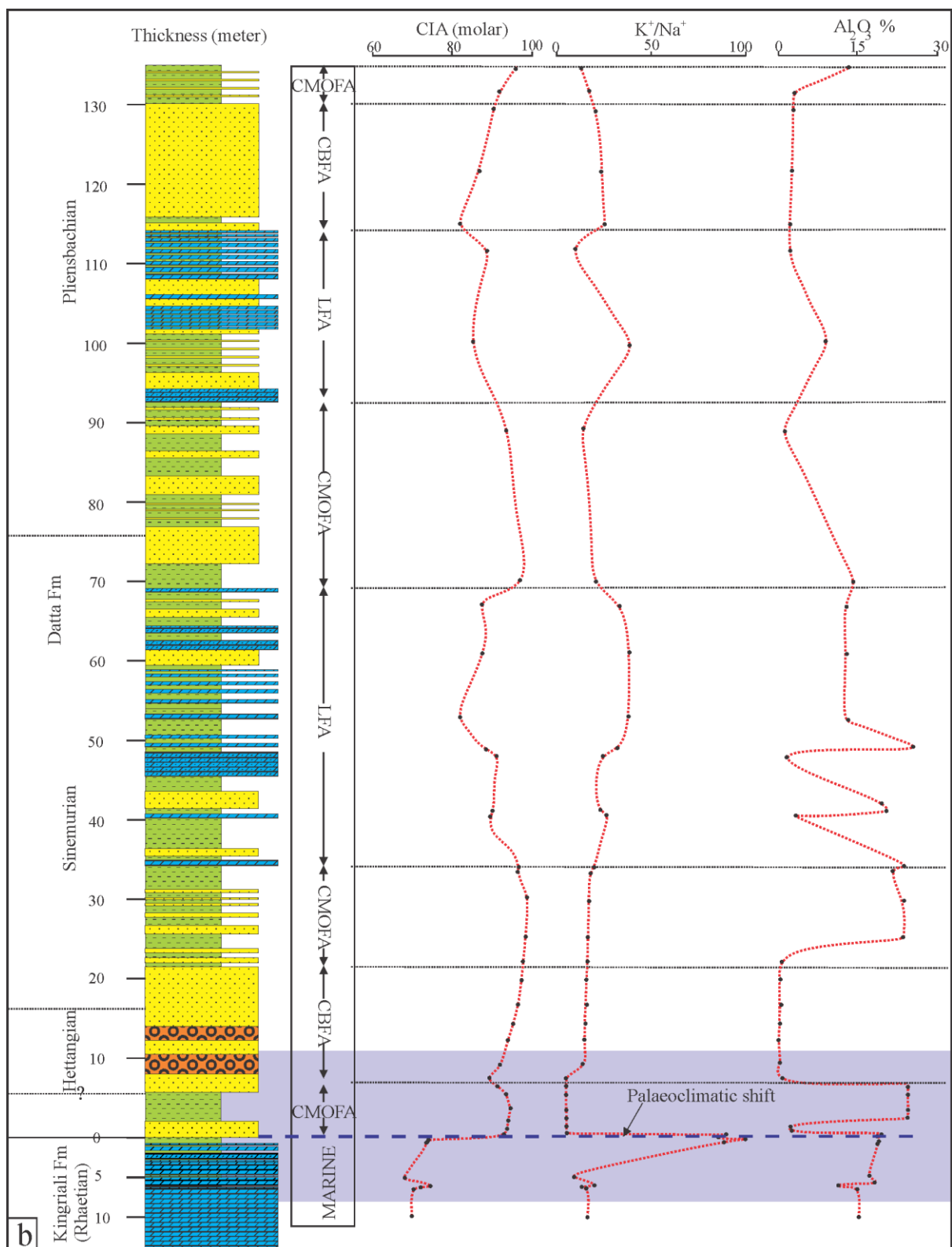
1369 Appendix 1. Framework mineralogical composition of the selected samples. Q_{mue} =
 1370 monocrystalline quartz with unit extinction, Q_{muu} = Monocrystalline quartz with
 1371 undulose extinction, Q_{pq} = polycrystalline quartz including chert, Q_{total} = total quartz,
 1372 A = k-spar, P = plagioclase, F_{total} = total feldspar, L_v = volcanic and hypabasal lithic, L_m
 1373 = metamorphic lithic, L_s = Sedimentary lithic excluding chert, L_{total} = total lithic, M =
 1374 phyllosilicates, D = heavy minerals, O = others/not recognizable, M/C =
 1375 matrix/cement (not included in total counting).

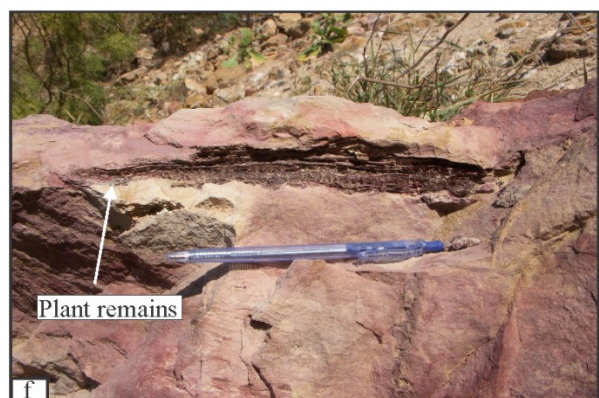
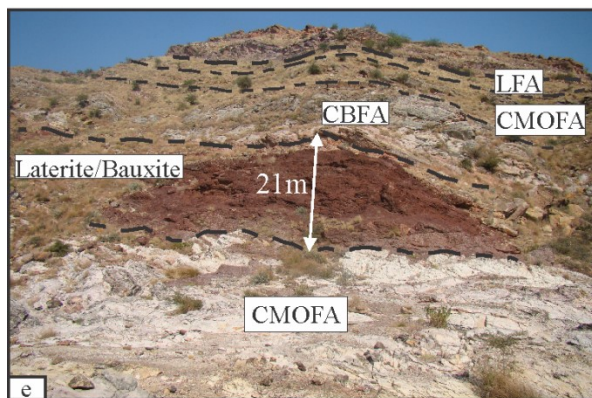
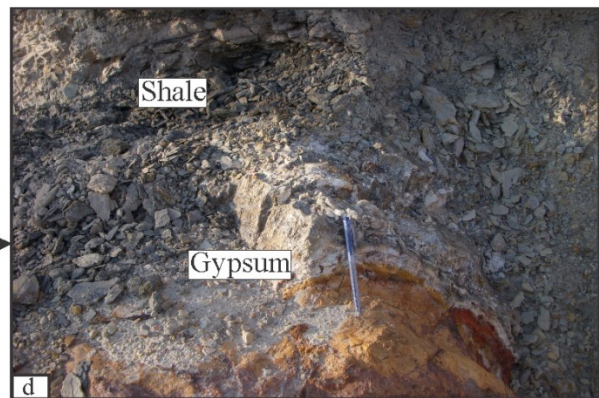
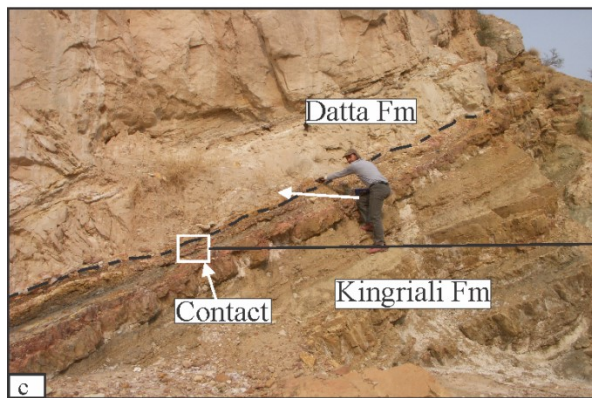
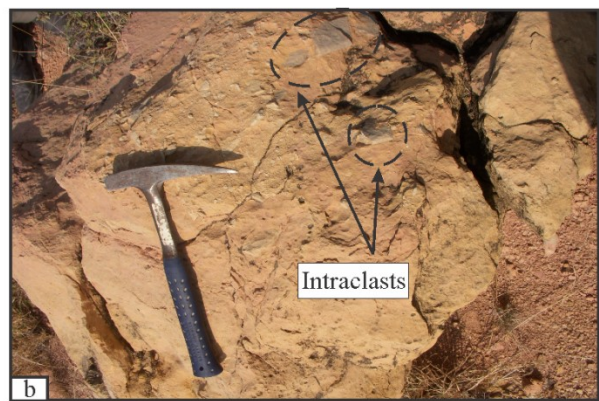
1376 Appendix 2. (a) Bulk mineralogical composition using XRD. (b) Clay mineralogy plot for
 1377 representative samples. Mg, K: Mg-, K-saturated sample; MgGly: Mg plus glycerol
 1378 saturated sample; KEG: K plus ethylene glycol saturated sample; KT: K saturated
 1379 sample heated to 550°C.

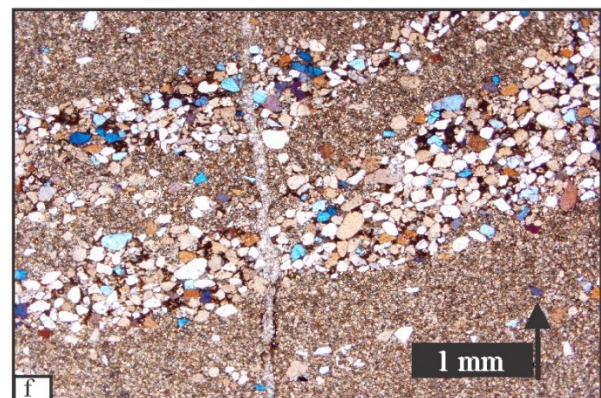
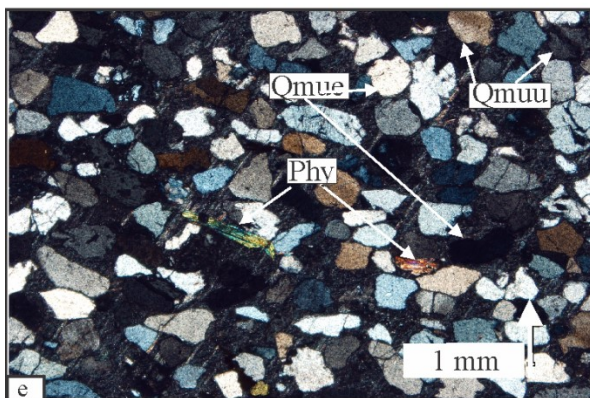
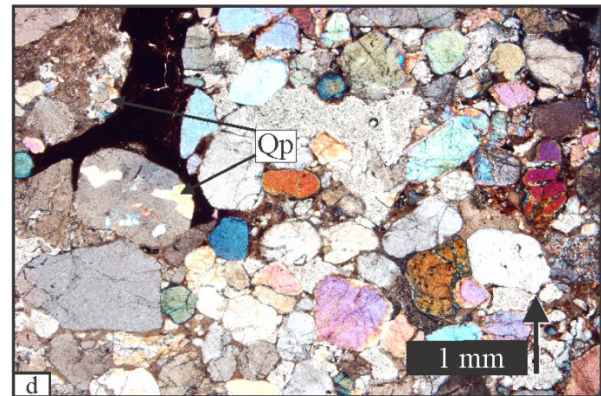
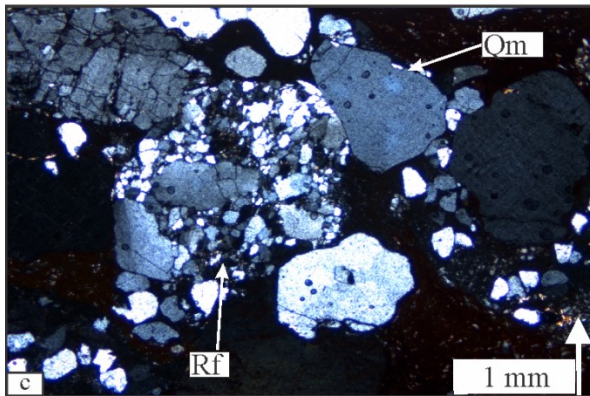
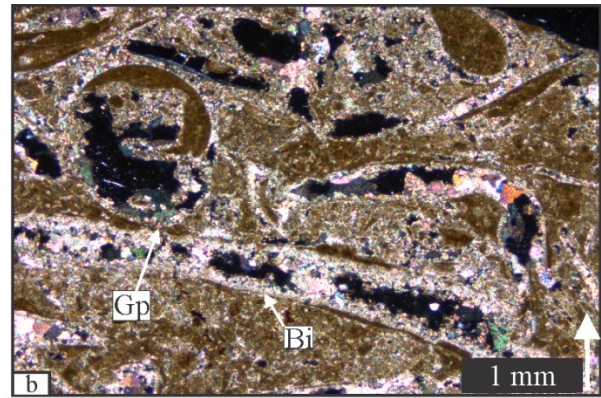
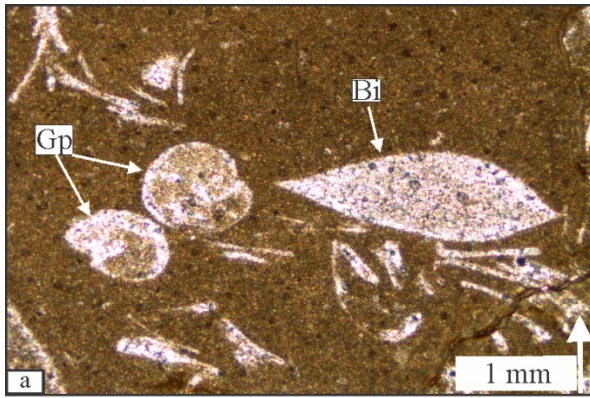
1380 Appendix 3. Bulk rock geochemistry data for the selected samples.

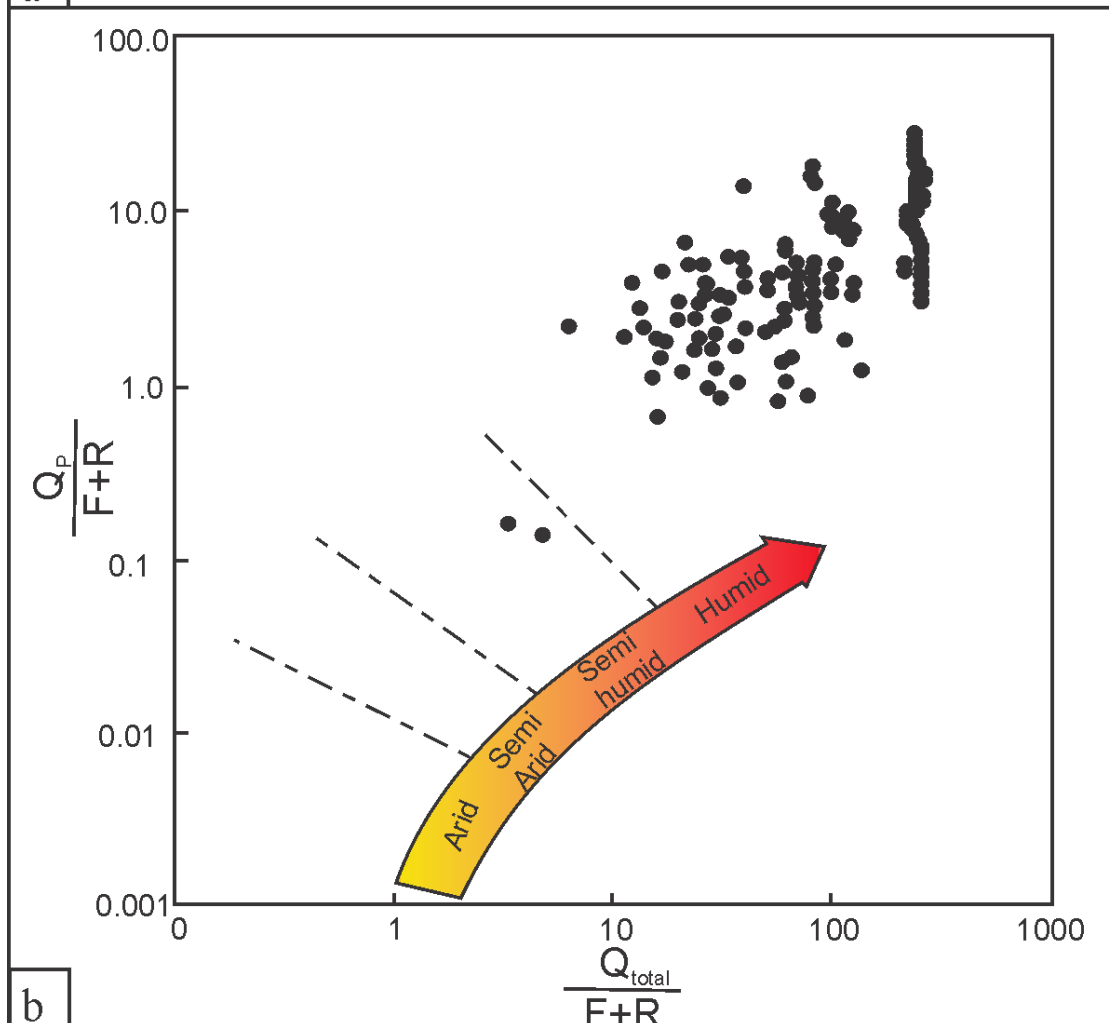
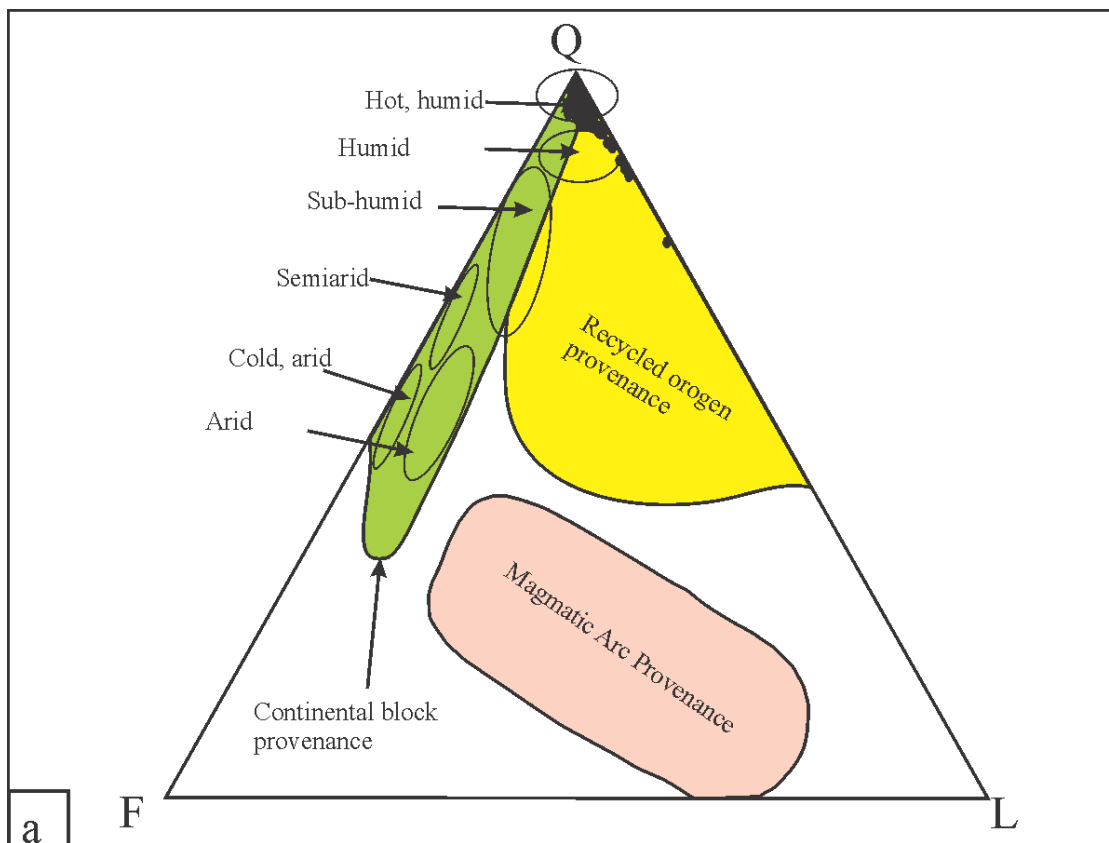


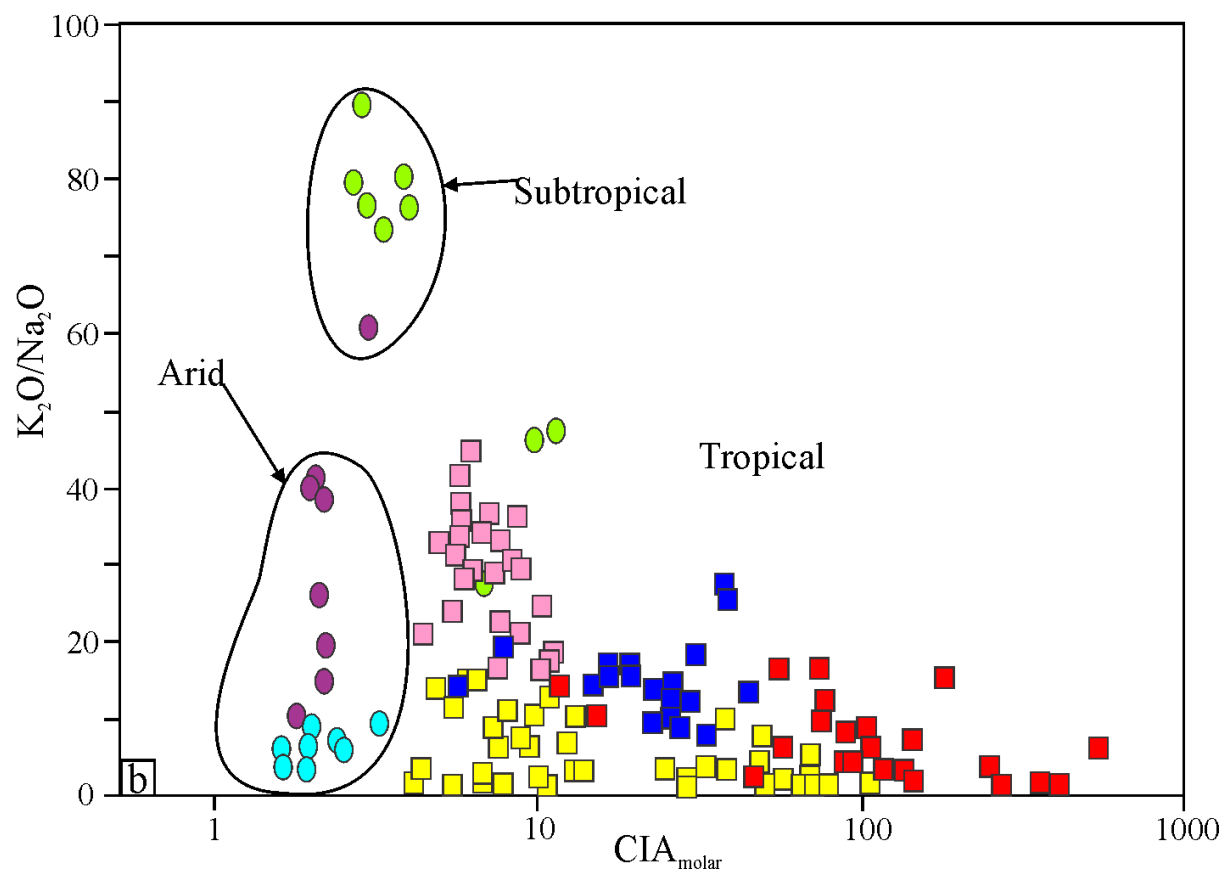
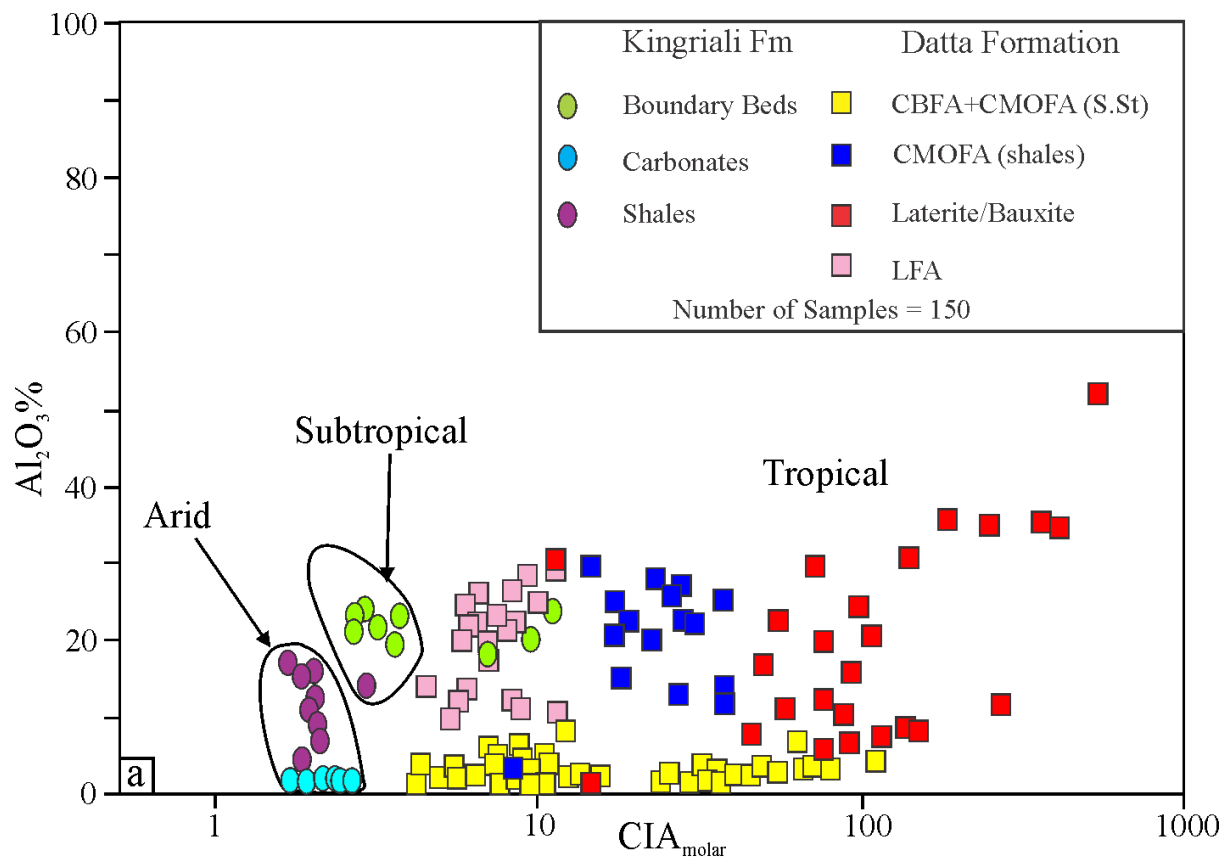


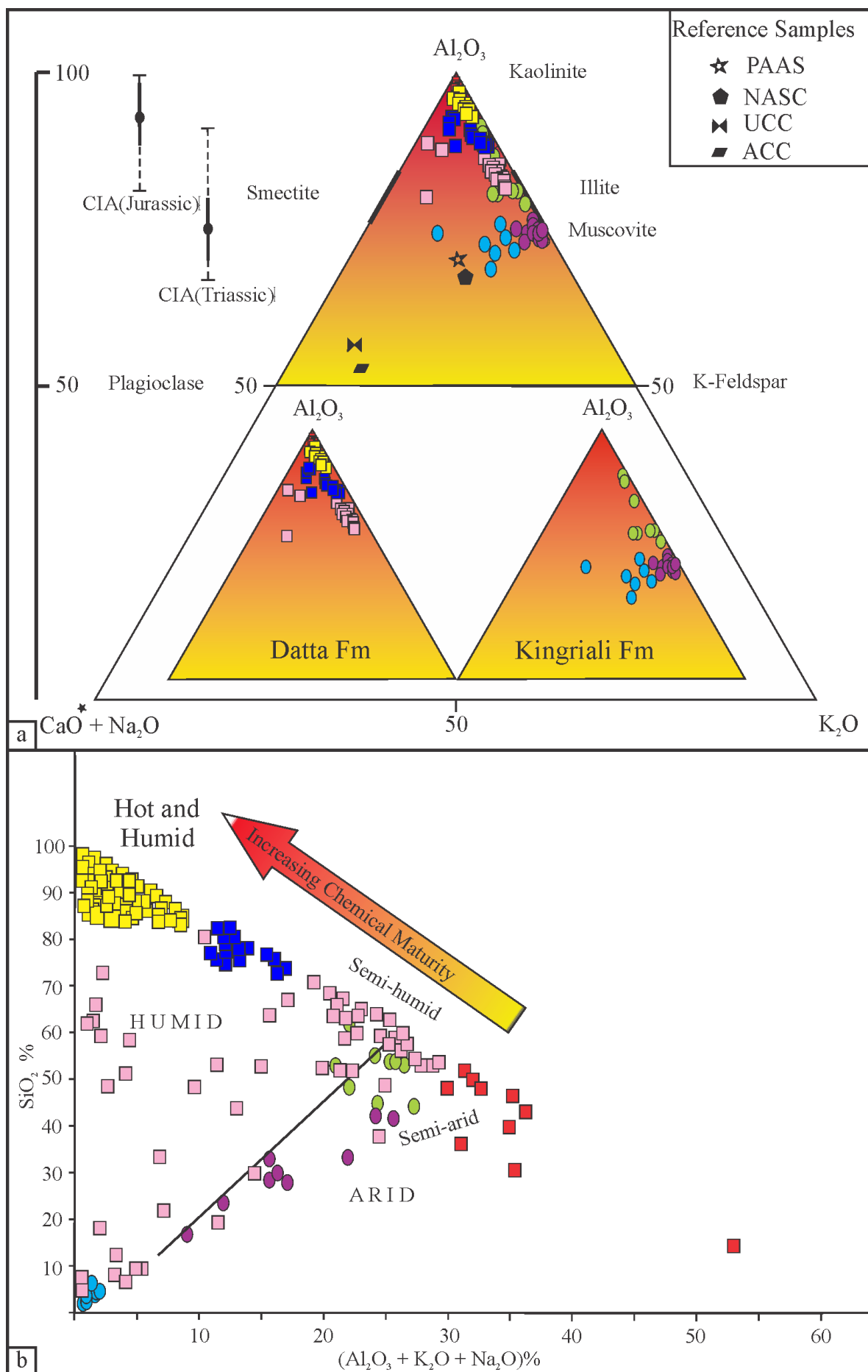


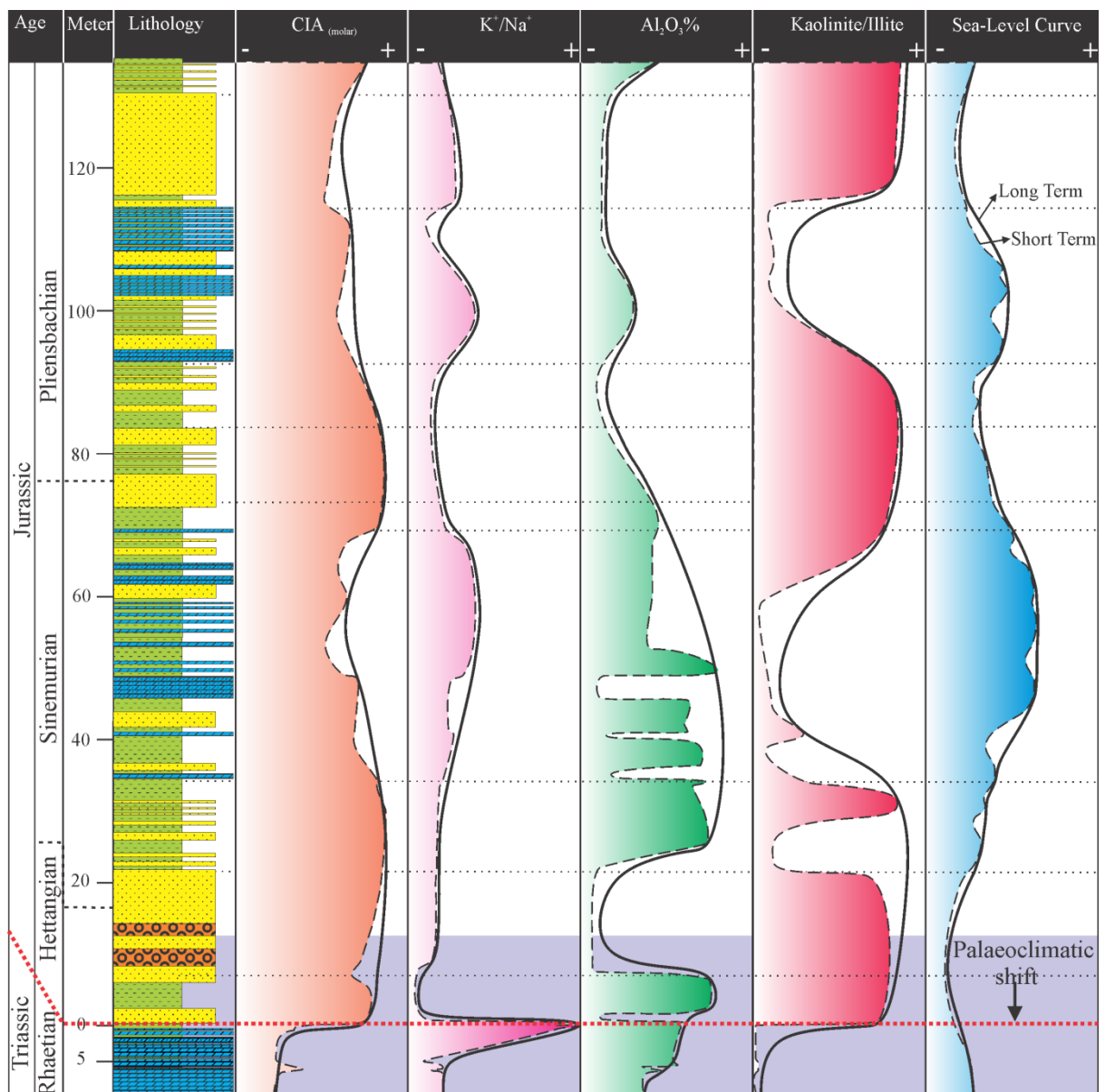


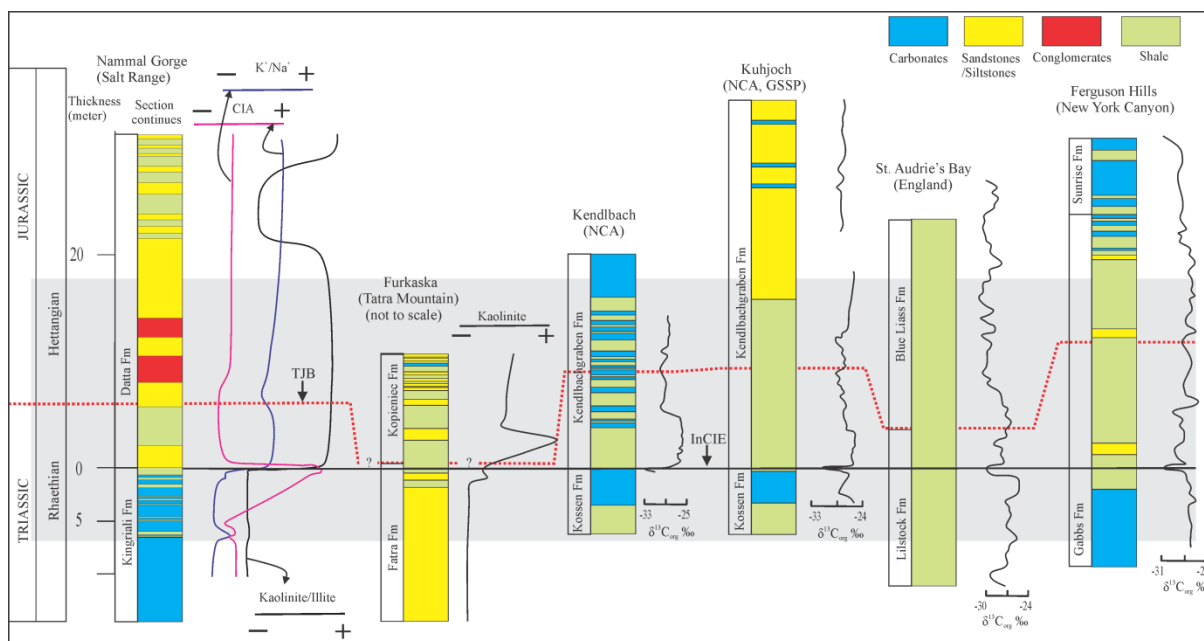












CHAPTER 6

PROVANENCE ANALYSIS

This chapter deals with the source to sink analysis of the Kingriali and Datta formations. Palaeoclimatic conditions and tectonic settings responsible for the generation of the sediments and transportation pathways have been worked out.

CHAPTER 6

PROVENANCE ANALYSIS

This chapter is an effort to reconstruct the source-to-sink history of the Upper Triassic (Kingriali Formation) to Lower Jurassic (Datta Formation) siliciclastic sediments, following the demise of the Upper Triassic carbonate platform in the Salt and Trans–Indus ranges. The provenance analysis of the siliciclastic detritus is expected to unravel the weathering, erosion and sediments generation in the source area and their transportation to the depositional sites. This will help in the reconstruction of the tectonic, palaeogeographic and palaeoclimatic history of the area during this critical time interval. The sediments provenance is based on the outcrop, petrographic data, bulk rock geochemical and heavy minerals analyses.

6.1 Framework composition/petrographic studies

Details of the petrographic studies have already been provided in chapter 4 (Iqbal et al., 2015b) and chapter 5 section 4.2 (Iqbal et al., 2018, under review). Quartz is by far the dominant constituent of the framework composition and the total quartz (Q_{total}) averages 95.5%. When subdivided based on the quartz types, Q_{mue} predominates averaging 76.1% followed by Q_{muu} (12.5%), whereas, $Q_{pq(2-3)}$ content averages 4.3% and $Q_{pq>3}$ averages 2.6% of the bulk framework composition. The total feldspars (F_{total}) content average only 0.15% of the bulk framework, which are highly decomposed and altered to clay minerals. Very minor potassium feldspar (K) occurs (average 0.13%) while plagioclase feldspar (F) is rare (average 0.02%). The total lithic fragments (L_t) constitute a minor component, averaging 1.2%. L_t is the combination of sedimentary lithics (L_s), metamorphic (L_m) and volcanic-hypabyssal (L_v).

6.1.1 Quartz-Feldspars-Lithics (QFL) plots

According to the Q_tFL_t ($L_t=RF$) all the siltstones and sandstones are compositionally quartz arenites, while the pebbly sandstones and conglomerates are lithic arenites (Pettijohn et al., 1973; Pettijohn, 1974; Fig. 6 of Iqbal et al., 2015b). However, when the L_t are subdivided into their constituent particles (Ingersoll et al., 1984), > 96% of the samples are quartz arenites and the remaining plot in the top right corner or the sublithic arenite field close to

quartz arenite (Fig. 6.1a). The Q_tFL_t plot indicates a cluster of samples near the quartz apex (Fig. 6.1b) that indicates reworking of stable continental/cratonic sediments under hot and humid palaeoclimate (Fig. 5a of Iqbal et al., 2018, under review). The Q_mFL_t plot (Fig. 6.1c) provides a high resolution image of the tectonic setting and indicates sediments derivation from stable cratonic interior (Dickinson et al., 1983). Similarly the $Q_pL_{sm}L_v$ plot (Fig. 6.1d) indicates very insignificant contribution from orogenic collisional belts, subduction zone or arc orogen.

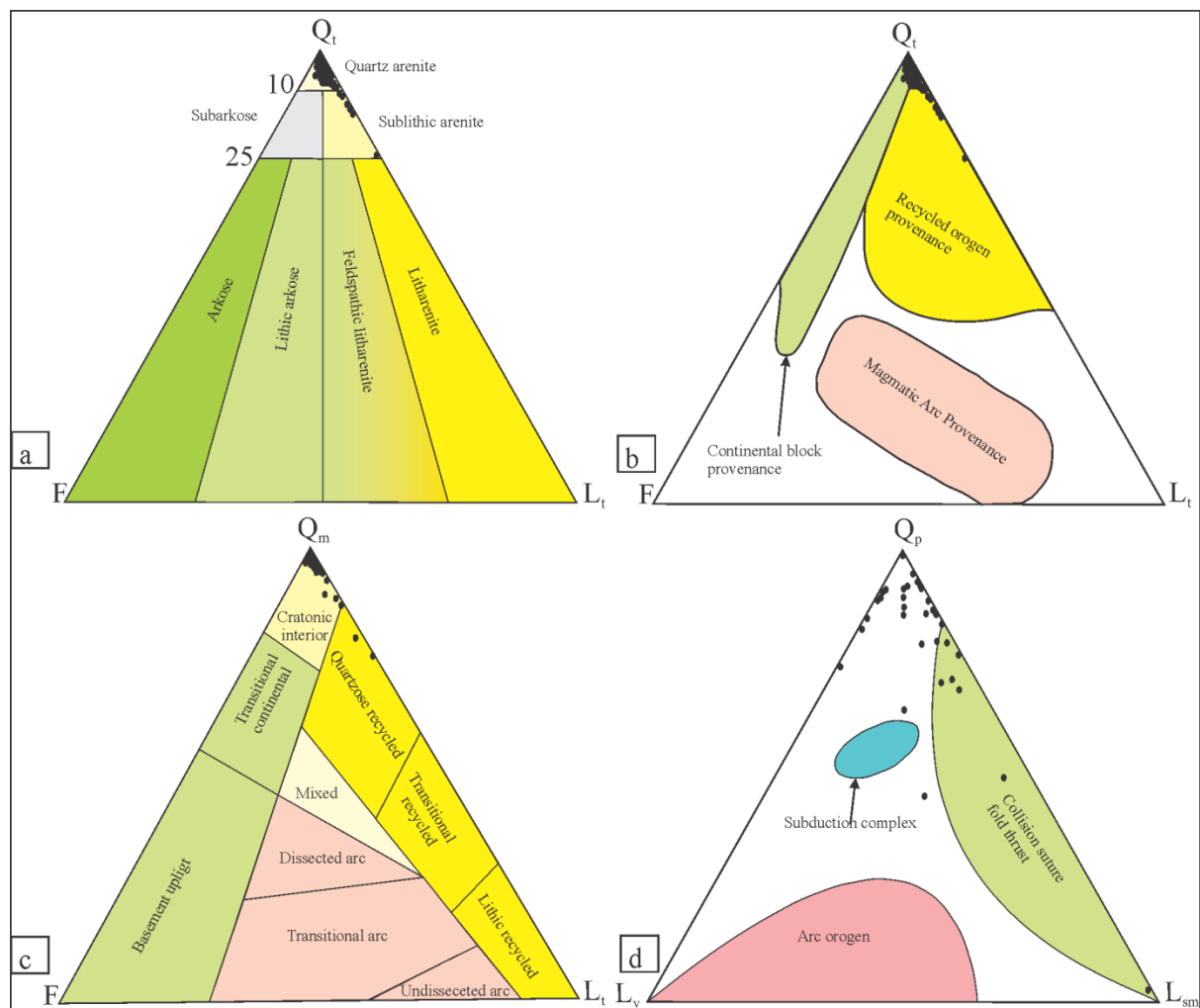


Figure 6.1 Framework mineralogical composition plots for the medium to coarse grained silicastics of the Datta Formation. a) total Quartz-Feldspars-Lithics (Q_tFL_t) plot for the sandstones classification (Ingersoll et al., 1984), b) Q_tFL_t tectonic setting discrimination plot for the sandstones (Dickinson et al., 1983), c) Q_mFL_t plot for tectonic setting discrimination (Dickinson and Suczek, 1979) and d) $Q_pL_vL_{sm}$ plot for tectonic setting discrimination (Dickinson and Suczek, 1979).

6.1.2 Quartz grains types

The four quartz types (Q_{mue} , Q_{muu} , $Q_{p(2-3)}$ and $Q_{p>3}$) can be presented in diamond-shaped provenance discrimination diagrams to interpret the source rock type (Basu et al, 1975; Tortosa et al., 1991). The dominance of Q_{mue} forces the samples to be plotted in the plutonic provenance field and very few samples in the medium to high rank metamorphic rocks (Fig. 6.2a). The composition of the parent rock was most probably granitic (Fig. 6.2b). Apparently the samples plot (Fig. 6.2b) indicates derivation of the substantial detritus from a medium to high rank metamorphic source, however when corrected for the plot (Fig. 6.2a) granitic provenance is a more justifiable interpretation in the present case.

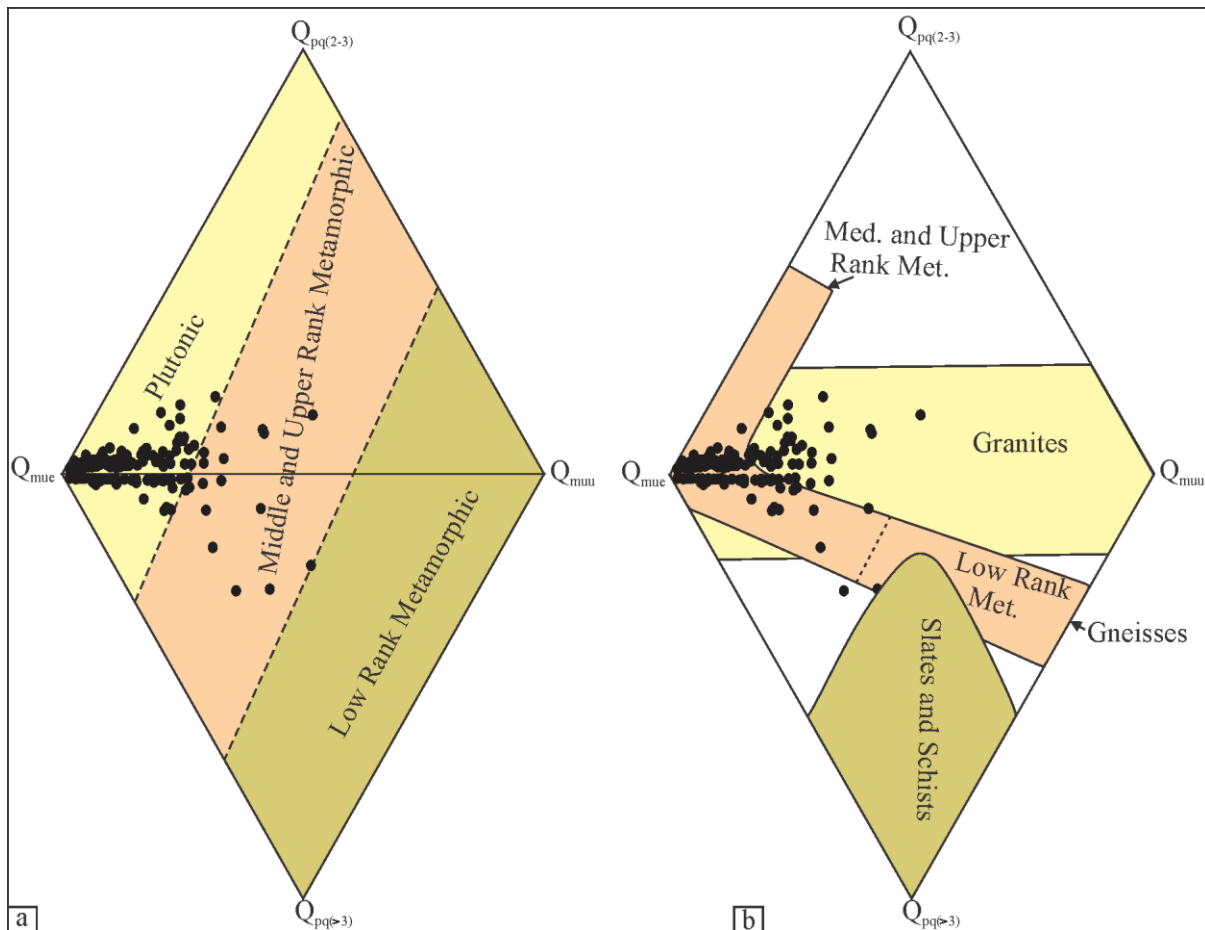


Figure 6.2 a, b) $Q_{pq<3}$ Q_{mue} $Q_{muq>3}$ pyramid plots for source rock discrimination (Basu et al., 1975; Tortosa et al., 1991).

6.2 Bulk geochemical proxies

Certain geochemical proxies are widely used to reconstruct the weathering and palaeoclimatic trend (Nesbitt and Young, 1982; 1984; 1989; Taylor and McLennan, 1985; Suttner and Dutta, 1986; Hofmann et al., 2001; Goldberg and Humayun, 2010; Abre et al., 2011; Buggle et al., 2011; Bahlburg and Dobrzinski, 2011; Malick and Ishiga, 2016), tectonic setting and source rock type (e.g. Bhatia, 1983; 1985; Bhatia and Crook, 1986; Roser and Korsch, 1986; Floyd and Leveridge, 1987; Herron, 1988; McLennan et al., 1993; McLennan, 2003; Cullers, 2002; Cingolani et al., 2003; Bracciali et al., 2007; Hegde and Chavadi, 2009; Abre et al., 2011; Stern and Wagreich, 2013; Hofmann et al., 2016). The weathering and palaeoclimatic trend proxies have already been discussed in detail in chapter 5 (Iqbal et al., 2018, under review). The tectonics and source rock identification proxies are discussed in the following.

6.2.1 Geochemical classification of sediments

Geochemical classifications of sediments include the SandClass system for geochemical classification of terrigenous sands and shales (Herron, 1988) and geochemical classification for terrigenous sands (Pettijohn et al., 1973). In the present case, the SandClass system classification [$\log(\text{Fe}_2\text{O}_3/\text{K}_2\text{O})$ vs. $\log(\text{SiO}_2/\text{Al}_2\text{O}_3)$] displays that all the siliciclastics of the Kingriali Formation (Triassic) plot into the Fe-Shale and Shale fields whereas the Datta Formation (Jurassic) sediments plot in the in the Fe-Shale and Shale and Fe-sand fields (Fig. 6.3a). None of the samples occupy the arkose and subarkose fields while very limited numbers of Jurassic samples are classified in the quartz arenite, sublithic arenite and lithic arenite fields. The geochemical classification for terrigenous sands [e.g. $\log(\text{Na}_2\text{O}/\text{K}_2\text{O})$ vs. $\log(\text{SiO}_2/\text{Al}_2\text{O}_3)$] shows that all the Triassic sediments plot in the lower left part of the diagram below the arkose field (Fig. 6.3b). A high percentage of the Jurassic sediments occupy the same region as the Triassic sediments. A substantial number of the Jurassic samples fall in the fields of quartz arenite, and sublithic arenite whereas very limited samples plot in the other fields (Fig. 6.3b).

The SandClass system classification (Fig. 6.3a) is a good proxy for distinguishing shales, Fe-rich shales, laterites and Fe-rich sandstones. The cluster of the Triassic samples confirms that these are shale and Fe-rich shales. Similarly, the lagoonal and swamp shales of the

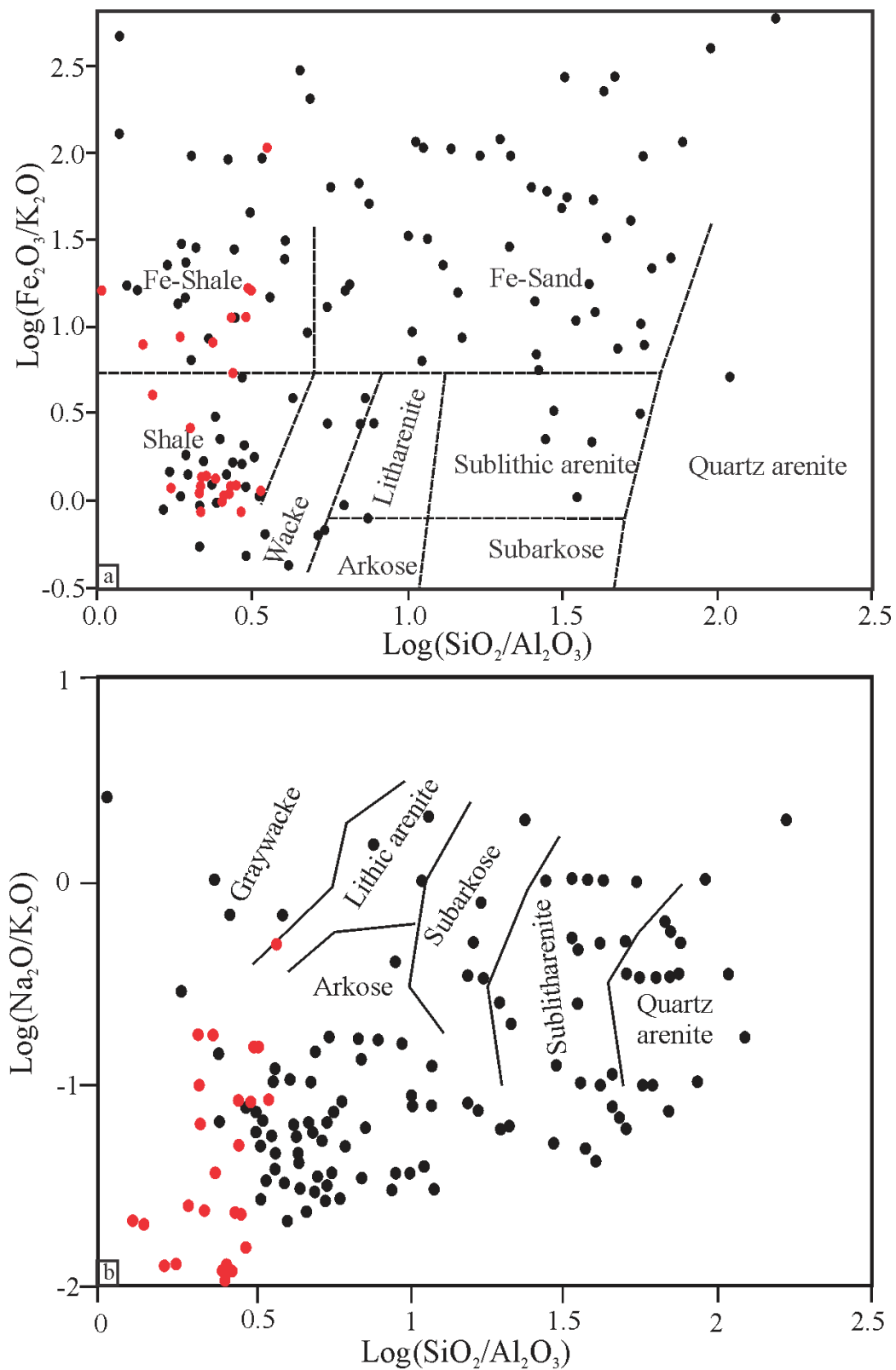


Figure 6.3 a) SandClass system for geochemical classification of the sandstones and shales (Herron, 1988), and b) geochemical classification for sandstones (Pettijohn et al., 1973); Red dots = Triassic samples, Black dots = Jurassic samples (for all figures).

Jurassic plot in the shale field whereas the paleosols clays, laterites/bauxites horizons and the flood plain clays that are rich in Fe-content plot in the Fe-rich shale field. Thus this plot provides a straightforward interpretation for the fine grain strata.

The sandstone samples, on the contrary, display no proper trend and are scattered in the Fe-rich sand field. Field observation, X-ray diffraction (XRD) and framework composition analysis of these samples provide coherent evidences that these are supreme mature quartz arenites (Fig. 6.1a; Iqbal et al., 2018, under review). However, careful observations reveal that the introduction of Fe-rich clays into the pore spaces during low energy flow stages following the deposition of very well sorted quartz sand and Fe-rich cementing material (mostly haematite) are the factors responsible for high Fe_2O_3 and Al_2O_3 content of these sediments (Iqbal et al., 2015b; 2018, under review). Thus the Fe-rich sand position of these samples is the overprint on otherwise supreme mature quartz arenite.

The geochemical classification for terrigenous sands (Fig. 6.3b) provides an easy identification for the Triassic and Jurassic shales as both occupy positions to bottom left of the arkose field (Herron, 1988). The Jurassic sandstones are dominantly clustered in the quartz arenite, sublithic arenite fields confirming their mature nature.

6.2.2 Tectonic settings and source compositional discrimination

Certain bivariate (Bhatia and Crook, 1986; Floyd and Leveridge, 1987) and ternary diagrams (Bhatia, 1985; Bhatia and Crook, 1986; Bracciali et al., 2007) of trace elements are commonly used for unravelling the tectonic settings and compositional discrimination of the siliciclastic detritus.

6.2.2a Bivariate plots

Hf(ppm) vs. La/Th plot

The bivariate plot of Hf(ppm) vs. La/Th (Fig. 6.4a) displays uniform and low La/Th values of <8 and a wide range for Hf (1 – 30 ppm). Generally, the sediments with dominantly acidic arcs provenance have uniform and low La/Th ratios. Such sediments have Hf contents of about 3-7 ppm (Floyd and Leveridge, 1987). Progressive dissection and erosion of the arc and of its plutonic roots and continental basement of ancient (meta)sedimentary rocks

concentrates the zircon contents in such sediments. Since zircon is the main host of Hf therefore the latter increases in such processes. The samples are therefore clustered roughly parallel to the Hf axis (Fig. 6.4a). The majority of the samples (more than 95%) plot

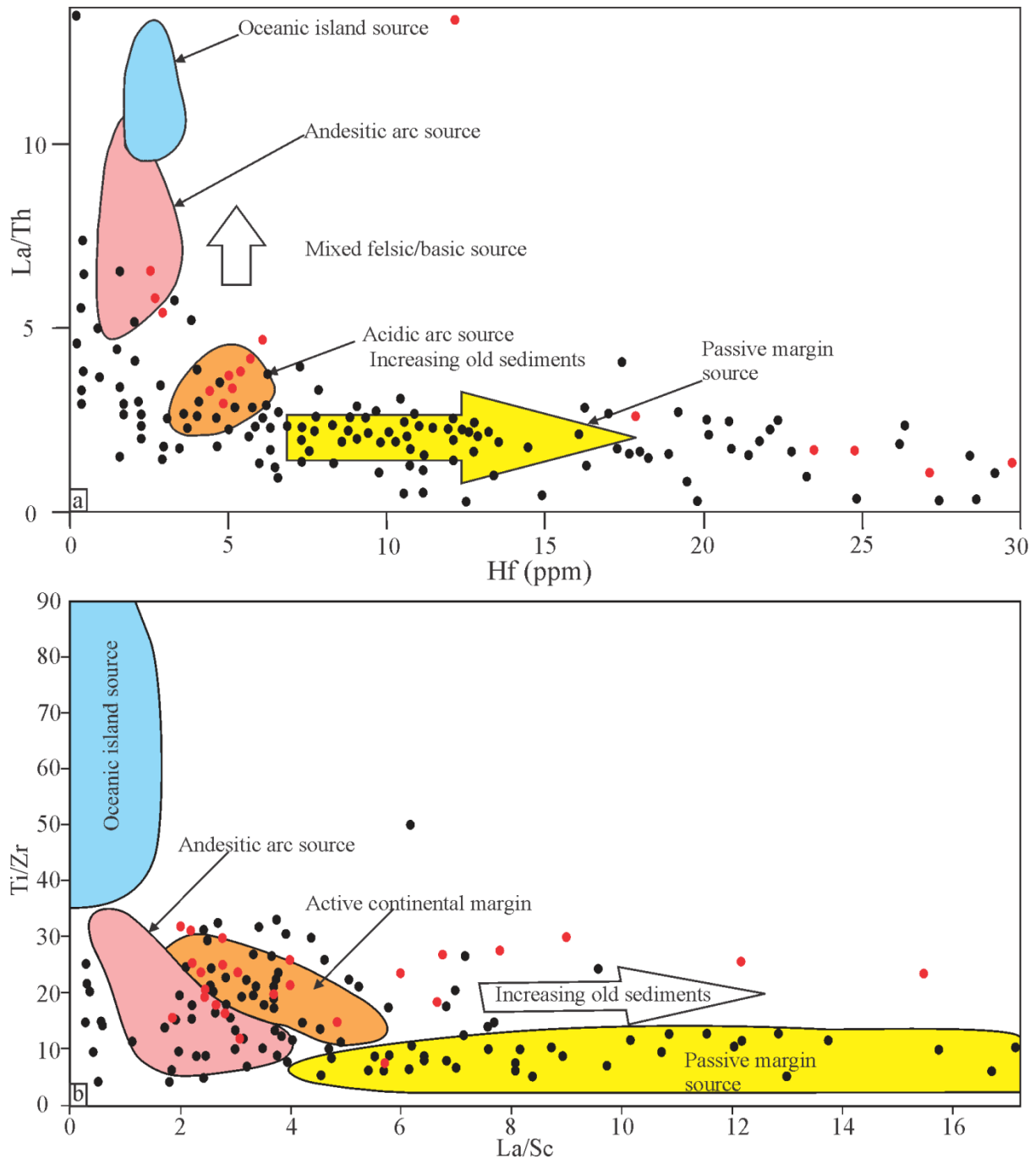


Figure 6.4 a) Hf(ppm) vs. La/Th bivariate plot (Floyd and Leveridge, 1987), and b) La/Sc vs. Ti/Zr bivariate plot (Bhatia and Crook, 1986) negating sediments derivation from oceanic island arc and acidic arc sources.

in the acidic arc source and passive margin source areas with increasing recycling component. None of the samples belong to the Oceanic arc source and only few (less than 5%) are classified into the andesitic arc source area. Thus an insignificant mixing of felsic and basic source is indicated (Fig. 6.4a).

La/Sc vs. Ti/Zr bivariate plot

The bivariate plot of La/Sc vs. Ti/Zr (Fig. 6.4b) shows that the present samples have a Ti/Zr ratio below 35 and a broad La/Sc range of 1 to > 14. The majority of the samples (around 70%) have Ti/Zr ratios below 10 accompanied by a wide range of La/Sc whereas the remaining (around 30%) samples have Ti/Zr ratio of 10 to 35 and La/Sc ratio of 1 to 4. In general, the oceanic island arc derived sediments have a higher Ti/Zr ratios (generally > 40) accompanied by very low La/Sc (< 1). Detritus derived from continental island arc displays Ti/Zr ratio between 10 and 35 whereas their La/Sc ratio is between 1 and 3. The active continental margin sediments have Ti/Zr values similar to those derived from continental island arc but are discriminated from the latter by their higher La/Sc ratio (generally ranges from 3 to 6). The passive margin siliciclastics are characterised by a very wide variation in the La/Sc ratio, and their low Ti/Zr ratios (generally < 10) discriminate them from sediments derived from other tectonic settings (Bhatia and Crook, 1986). The La/Sc vs. Ti/Zr bivariate plot attests sediments derivation from passive margin setting with a significant component from active continental margin and continental island arc settings (Fig. 6.4b). None of the samples plot in the oceanic island arc field.

Both the plots (Fig. 6.4) support derivation of the detritus from acidic source in a dominantly passive margin setting. A possible poor control on the oceanic island arc composition using Hf(ppm) vs. La/Th bivariate plot may question the validity of its application. However, the La/Sc vs. Ti/Zr bivariate plot (Bhatia and Crook, 1986) addresses this issue. The Triassic samples in both the plots favours the acidic/continental arc and active continental margin fields and only few fall into the passive margin field. Otherwise, no significant difference can be found in the positions of the Triassic and Jurassic samples.

6.2.2b Ternary plots

La–Th–Sc ternary plot

The significance of La/Th and La/Sc in the tectonic settings discrimination diagrams has already been discussed (Bhatia and Crook, 1986; Floyd and Leveridge, 1987). The Th/Sc ratio has an average of 1.6 for the shales of the Kingriali Formation while the average ratio for the siliclastics of the Datta Formation is 2.8 (Fig. 6.5a). The Th/Sc ratio is very low (average = 0.15 ± 0.08) for the oceanic island arc. Therefore, sediments derived from such settings will plot close to Sc apex of the ternary plot (Fig. 6.5a). The Th/Sc ratio averages 0.85 ± 0.13 for the continental island arc derived sediments while the ratio has an average value of 2.59 ± 0.5 for the active continental margin setting sediments. Sediments derived from passive margin settings have an average Th/Sc ratio of 3.06 ± 0.8 (Bhatia and Crook, 1986).

The La–Th–Sc ternary plot (Bhatia and Crook, 1986) reveals that most of the samples plot in the continental island arc, active continental margin and passive margin fields (Fig. 6.5a). However, with the increasing value of the Th/Sc average the error range increases. This is the limitation of the plot that does not allow a clear discrimination of the active continental margin sediment from those derived from passive margins (Fig. 6.5a). The Zr/Th ratio is helpful in such discriminations (Bhatia and Crook, 1986). Further, Co behaves concordantly with Sc (Bhatia and Crook, 1986). Therefore, the Th–Sc–Zr/10 and Th–Co–Zr/10 ternary plots (Bhatia and Crook, 1986) are valuable tools for such discriminations.

Th–Sc–Zr/10 and Th–Co–Zr/10 ternary plots

The Zr/Th ratio for the shales of the Kingriali Formation averages 12.5 and increases to 20.3 for the Datta Formation (Fig. 6.5b). The oceanic island arc sediments generally have high Zr/Th ratio (average = 48 ± 13.4) as these are depleted in Th. Additionally, such sediments are characterised by their high Th/Sc and Th/Co ratios. Therefore, sediments derived from such settings occupy positions close to the Sc and Co apices (Fig. 6.5b, c). Zr/Th ratio for the sediments of active continental margin is 9.5 ± 0.7 and coupled by low Th/Sc and Th/Co ratios. Hence such sediments plot nearest the Th pole in both the ternary plots (Fig. 6.5b, c). The passive margin siliclastics have Zr/Th ratio of 19.1 ± 5.8 and the Zr enrichment forces such sediments to plot near the Zr pole in both the diagrams (Fig. 6.5b, c). Sediments

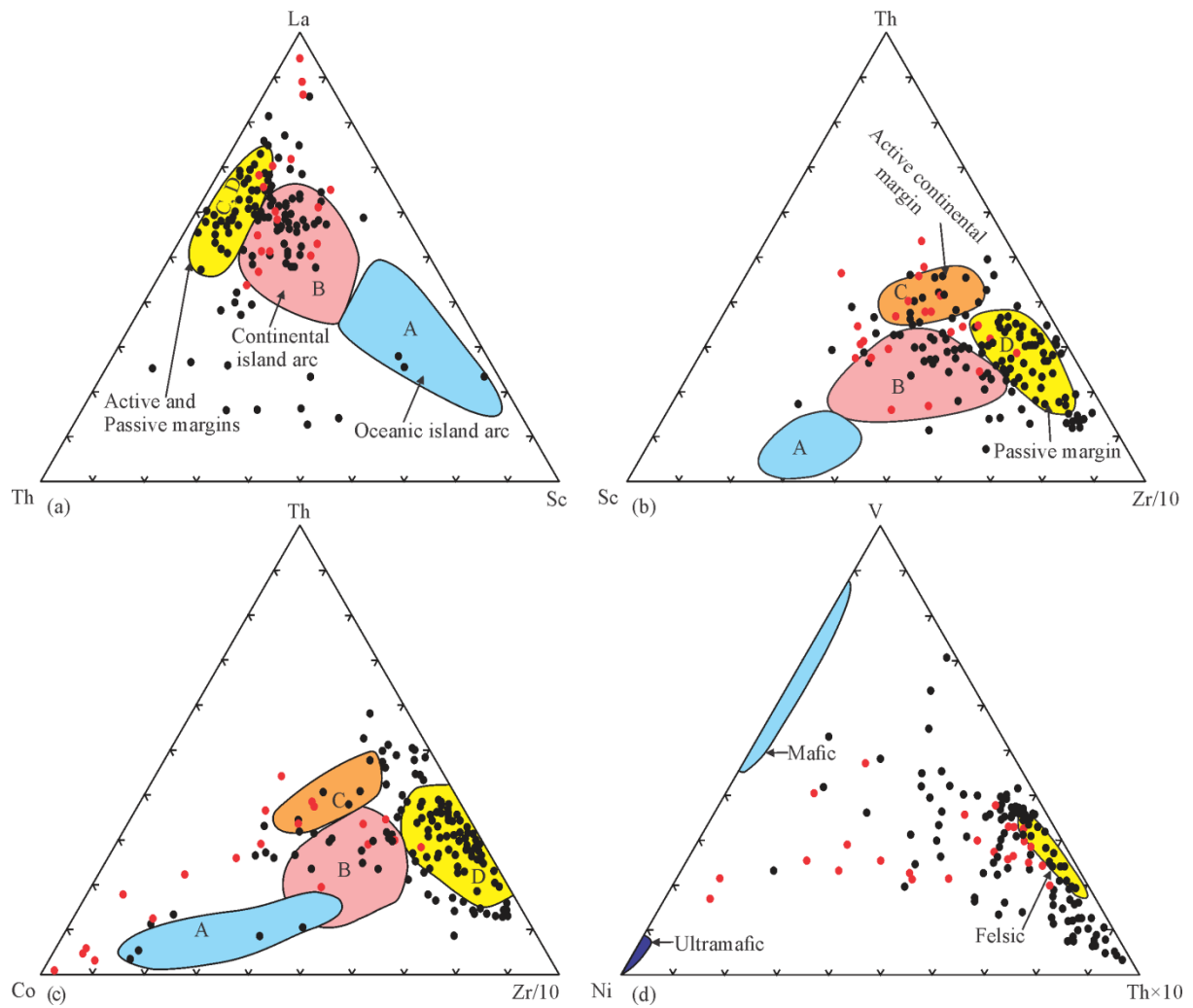


Figure 6.5 a-c) La–Th–Sc, Th–Sc–Zr/10 and Th–Co–Zr/10 ternary plots (Bhatia and Crook, 1986) for tectonic setting discrimination indicating sediments derivation from passive margin, and d) V–Ni–Th $\times$ 10 ternary plot (Bracciali et al., 2007) showing that the sediments were derived from a felsic source.

derived from continental island arc have Zr/Th average of 21.5 ± 2.4 and on both diagrams occupy an in between position (Fig. 6.5b, c). In order to bring the fields into the middle of the plots without alteration of their relative position a scaling factor has been used (Bhatia and Crook, 1986).

In the present case, the Th–Sc–Zr/10 plot (Fig. 6.5b) indicates that around 65% of the samples (almost exclusively samples of the Datta Formation) plot in the passive margin field. The remaining samples (including some samples of the Datta Formation and almost all of the Triassic Kingriali Formation samples) occupy positions of active continental margin and acidic arc source but still close to the passive margin. No sample occupies the oceanic island

arc setting. The Th–Co–Zr/10 plot (Fig. 6.5c) improves the resolution of the samples population. On this plot, > 80% of the samples occupy the passive margin field with insignificant contribution from the Triassic strata. The remaining < 20% that include all the Triassic samples are scattered over the continental island arc and active continental margin fields while the percentage of oceanic island arc samples is negligible (Fig. 6.5c). The ternary plots support sediments derivation from continental and passive margin setting and negate any incorporation from the basic source or oceanic island arc thereby supporting the interpretation of the bivariate plots (Fig. 6.4a, b).

6.2.3 Source compositions

La/Sc vs. Th/Co plot

The significance of La/Sc and Th/Co ratios has already been discussed in detail. Bivariate plot of La/Sc vs. Th/Co (Cullers, 2002) indicates a cluster of samples in the silicic rocks field and none of the samples plot in the basic rocks field (Fig. 6.6a). The V–Ni–Th $\times 10$ ternary plot (Bracciali et al., 2007) is an excellent tool for discrimination between the mafic and felsic source rocks. In general, the mafic sources are rich in V and Ni therefore sediments derived from such source rocks will occupy position along the V–Ni side of the plot. The closeness to Ni-apex supports ultramafic source derivation whereas the mafics are closer to the V-apex. As the felsic sources are rich in Th therefore such sediments plot close to Th-apex. Presently, the plot indicates a high sample population density in and around the felsic composition (Fig. 6.5d). None of the samples occupy the mafic or ultramafic composition. The two plots provide coherent evidences for the silica rich source rock of the siliciclastics with no contamination from a mafic/ultramafic source.

6.2.4 Weathering and recycling

Th/U vs. Th(ppm) bivariate plot (McLennan et al., 1993) is used to interpret the weathering trend of the siliciclastics. Presently the plot indicates Th/U values below the upper continental crust for the majority of the Triassic samples, while for the Jurassic samples (more than 90%) these values are above the upper continental crust (Fig. 6.6b). The Th values are generally more than the depleted mantle source for all the samples.

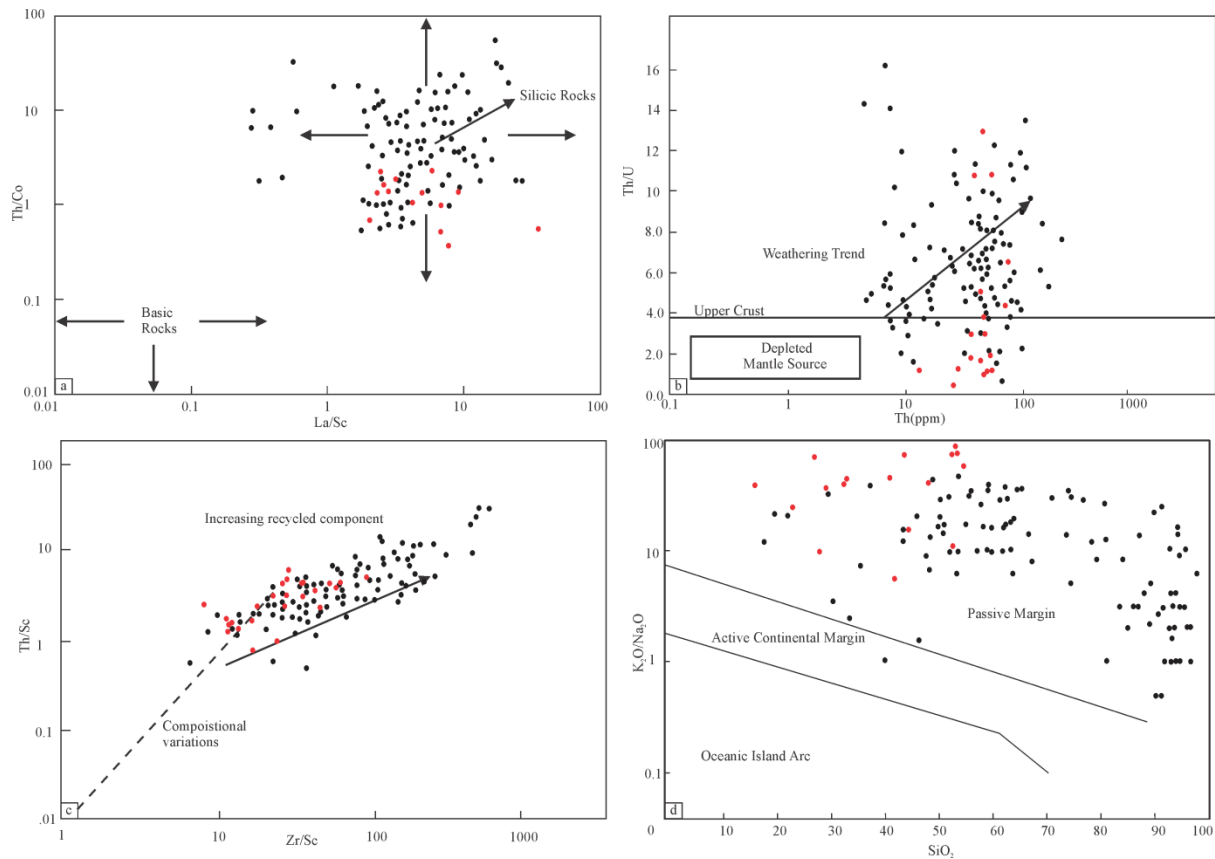


Figure 6.6 a) Bivariant plot of La/Sc vs. Th/Co (Cullers, 2002) indicating sediments derivation from a silicic source, b) Th/U vs. Th(ppm) bivariant plot (McLennan, 1993) favouring high degree of weathering for the sediments, c) Zr/Sc vs. Th/Sc bivariant plot (McLennan et al., 1993) supporting the recycled origin of the sediments, and d) bivariant plot of SiO_2 vs. $\text{K}_2\text{O}/\text{Na}_2\text{O}$ (Roser and Korsch, 1986) attesting sediments deposition on a passive margin.

The Zr/Sc vs. Th/Sc bivariant plot (McLennan et al., 1993) is widely used to identify the extent to which the sediments are recycled. In the present case the Th/Sc values are dominantly above 1 with generally higher values in the Jurassic samples in comparison to the Triassic samples (Fig. 6.6c). The Zr/Sc ratio also displays values generally above 10. The Triassic samples have low Zr/Sc values (less than 100) whereas the Jurassic samples have an overall higher range reaching up to 300.

During weathering, the Th/U value rises above upper crustal values of 3.5 to 4.0 as reflected in the Jurassic samples. Sediments weathering and recycling under oxidising conditions result in oxidation of U^{4+} to U^{6+} . The U^{6+} being more soluble and is lost and results in elevating the Th/U ratio (e.g., McLennan and Taylor, 1980; Taylor and McLennan, 1985).

Therefore, the higher position of the Jurassic samples on the Th/U axis above the upper continental crust line (Fig. 6.6b) indicates increasing intensity of chemical weathering. On the other hand, the low Th/U values in the Triassic samples indicate low chemical weathering intensity during that time. The palaeoenvironmental and palaeoclimatic interpretation for the strata (Iqbal et al., 2018, under review) supports this argument. However, the higher Th values indicate crustal differentiation in the Triassic samples.

Sediments derived from active margin tectonic settings are generally dominated by components of young undifferentiated crust (McLennan, 1993). Thus they have significantly lower Th/U values, generally below upper continental crust. Additionally, such sediments have low Th and U contents reflecting a low ratio in the source rocks (McLennan, 1989b, McLennan and Taylor, 1991) characteristic of geochemically depleted mantle source (e.g., Newman et al., 1984). Such conditions are not indicated by the studied samples.

Zr is particularly enriched in zircons while Sc generally preserves its provenance signature therefore the ratio Zr/Sc is a useful index of zircon enrichment (McLennan, 1989a). The incompatibility of Th and the compatibility of Sc in igneous systems makes Th/Sc a good indicator of igneous chemical differentiation processes. Presently, the higher Zr/Sc and Th/Sc ratios (Fig. 6.6c) of the samples indicate increasing weathering and sediments recycling from Late Triassic to Early Jurassic. The lack of low values for both the parameters negates compositional variation in the source sediments.

6.2.5 Tectonic settings

The tectonic discrimination diagram for sandstones and argillites is a bivariate plot of SiO_2 vs. $\text{K}_2\text{O}/\text{Na}_2\text{O}$ (Roser and Korsch, 1986). The $\text{K}_2\text{O}/\text{Na}_2\text{O}$ ratio displays values >1 for the entire Triassic–Jurassic strata (Fig. 6.6d). By comparison, the Triassic strata have higher ratio range (5 – 95) while the ratio range shrinks in the Jurassic strata (0.8 – 40) with the higher values in the shale and the sandstones generally display value below 15. The $\text{K}_2\text{O}/\text{Na}_2\text{O}$ ratio is discussed in detail in Iqbal et al. (2018, under review). The SiO_2 content has a 20 – 55 range (average = 45 due to incorporation of the dolomitic shales) in the Triassic shales while the Jurassic strata displays a much higher SiO_2 content (20 – 98) averaging more than 60.

The K_2O/Na_2O ratio of >1 and the overall high SiO_2 content restrict all the samples to be plotted in the passive margin setting (Fig. 6.6d). This negates the possibility of sediments derivation from a basic igneous (oceanic island arc) or an active continental margin as both of these have K_2O/Na_2O ratios < 1 and low SiO_2 content (Roser and Korsch, 1986).

6.3 Rare Earth Elements (REE)

The relatively immobile and incompatible nature of the Rare Earth Elements (REE) prevent them from being influenced by chemical alteration during transportation from source rocks to the depositional sites (Blake et al., 2017). This property makes the REE useful in tracing the source of the sediments (Garver and Scott, 1995; Cox et al., 1995). However, contradicting arguments exist suggesting that rocks exposed to intense weathering can be depleted in heavy REEs thereby impairing the REE utility as a sediment provenance indicator (Condie, 1991).

6.3.1 Chondrite and Post Archean Australian Shale (PAAS) normalised REE plots

The Post Archean Australian Shale (PAAS) normalised Rare Earth Elements + Yttrium (REE+Y) plots indicate a flat pattern, typical for continental sediments. No negative Ce anomaly which is the characteristic of marine sediments can be observed (Fig. 6.7a, b). The boundary beds samples have higher positions in the plot attributed to their higher total REE content and display slightly heavy REE (HREE) enrichment, otherwise no difference is observable between the Triassic and Jurassic strata.

The Chondrite normalised REE plots display very high enrichment of light REE (LREE) and have flat HREE pattern (Fig. 6.7c, d). A pronounced negative Eu anomaly is also observable. The boundary beds samples can be differentiated from the remaining Triassic samples by their enrichment in total REE otherwise no visible difference can be observed in the Triassic and Jurassic sediments (Fig. 6.7c, d). The chondrite-normalised REE plots pattern display distinct patterns for different source rock types (Bhatia, 1985; Bhatia and Crook, 1986; Floyd and Leveridge, 1987). Sediments of the oceanic island arcs display little enrichment of the LREE over HREE and lack visible negative Eu anomaly. Both the LREE enrichment and the negative Eu anomaly display positive correlation with the increasing silica content in the

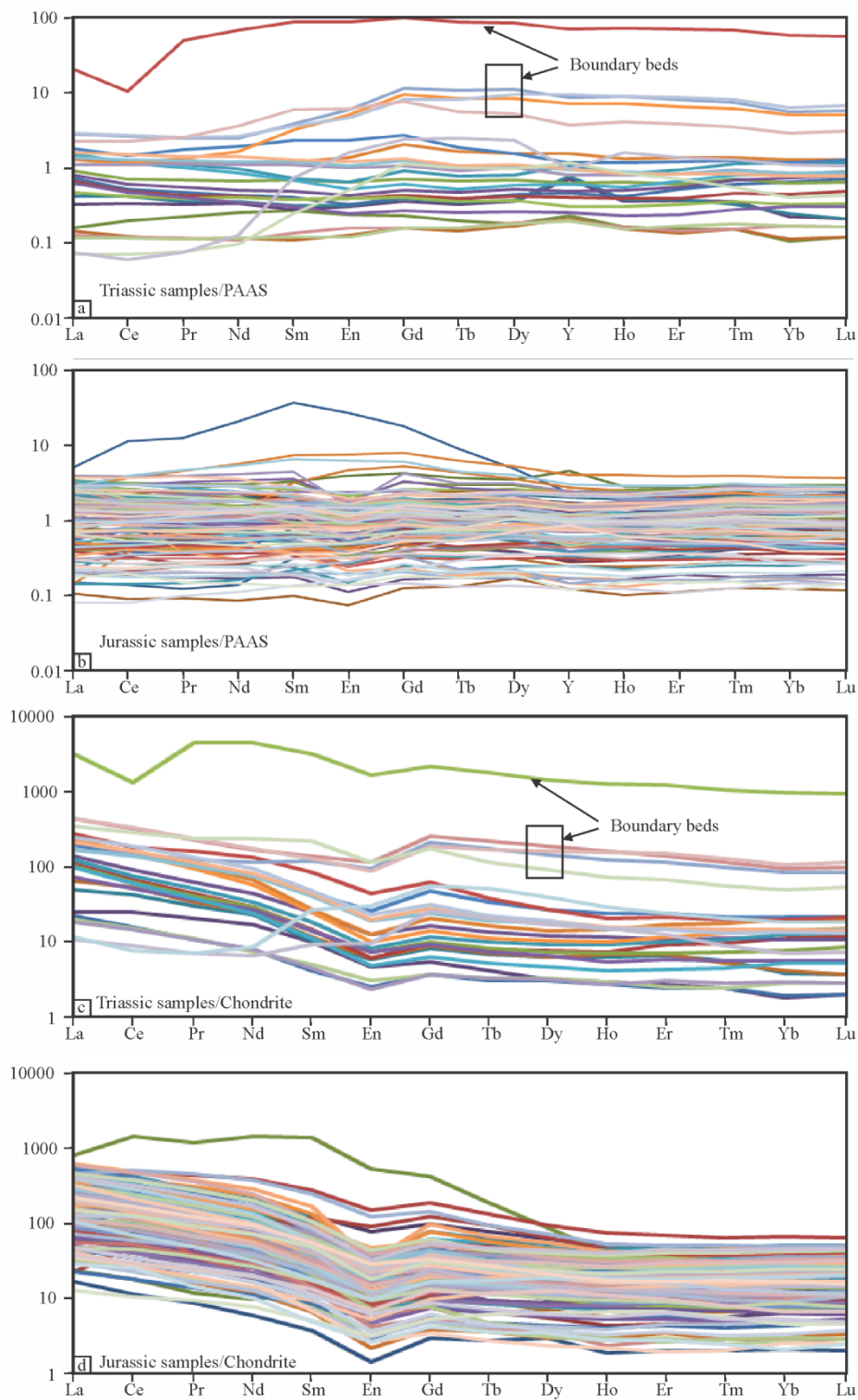


Figure 6.7 a, b) PAAS normalised REE+Y plots for the Triassic and Jurassic sediments respectively indicating reworked and homogenised sediments in study area (Cullers, 2002), c, d) Chondrite normalised REE plots for the Triassic and Jurassic strata showing a pronounced negative Eu anomaly and continental nature of the sediments (Floyd and Leveridge, 1987).

sediments. Thus the sediments of acidic source (continental island arc and active continental margin) shows considerably enriched LREE pattern over the HREE and a visible negative Eu anomaly (Taylor et al., 1968; Ewart et al., 1968; Nance and Taylor, 1976; Fleet, 1984; Cullers et al., 1987; Cullers, 1994a; 2002). The passive margins sediments are rich in quartz and derived from sedimentary and metasedimentary rocks therefore display strong LREE enrichment, possess a pronounced negative Eu anomaly and are characterised by a flat HREE pattern (Floyd and Leveridge, 1987).

6.3.2 (La/Sm)_N vs. Nb/La plot

The (La/Sm)_N vs. Nb/La plot (John et al., 2003; Boniface and Mruma, 2012) shows a good discrimination between the various tectonic settings. All the samples have Nb/La ratios below 1.4 with most of them (> 95%) having Nb/La ratios below 1.0 (Fig. 6.8). Similarly the (La/Sm)_N values are above 3 for most of the samples (> 95%). Sediments derived from oceanic island arc source have high Nb/La ratio coupled with low (La/Sm)_N values whereas the continental arc and acidic source sediments have the opposite properties and are positioned parallel to the (La/Sm)_N axis (John et al., 2003). Thus, most of the samples plot in the cratonic and passive margin field (Fig. 6.8). A very minor population of the samples are classified into continental arc and an insignificant number plot in the other fields. The tectonic settings discrimination plot indicates derivation of sediments from a dominantly felsic source and negates sediments incorporation from basic sources.

6.4 Heavy minerals analysis

Heavy minerals analyses have a wide range of applications in studying the provenance of siliciclastic rocks (e.g. Mange and Wright, 2007). Specific heavy minerals assemblages are the characteristic of specific source rocks and are thus sensitive indicators of provenance (Sevastjanova et al., 2012). Herein, heavy minerals data for the eastern most section (Kasanwala) is presented as it preserves the complete lithofacies variations in the succession. In the present case the heavy minerals assemblage is a simple mixture of the ultra-stable zircon, tourmaline and rutile with minor staurolite and rare chloritoid (and chlorite).

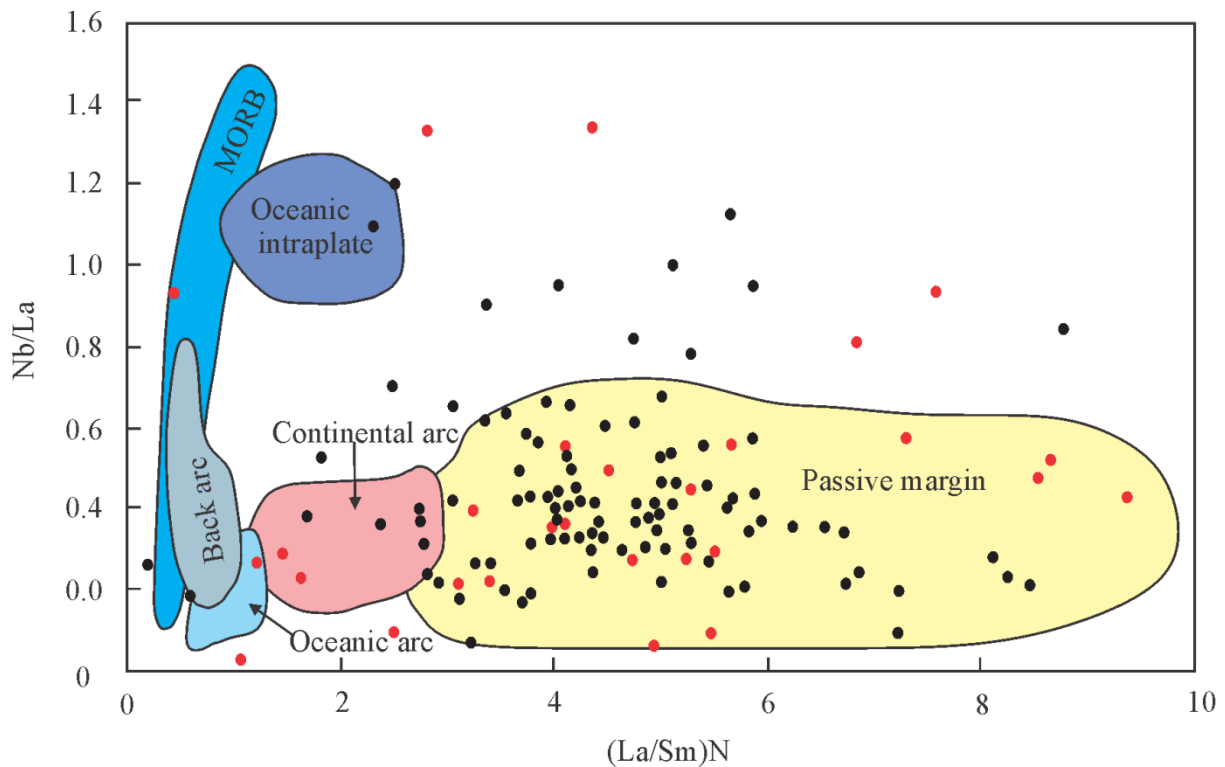


Figure 6.8 (La/Sm)N vs. Nb/La source discrimination plot (John et al., 2003) indicating sedimentation processes on a passive margin setting.

6.4.1 Tourmaline

Tourmaline is the most common transparent heavy mineral among the heavy mineral assemblage. It averages 44.3% of the whole assemblage. It averages 39.8% in the fine grained sandstone (Triassic) below the major laterite horizon. These tourmalines are mostly euhedral, angular to subangular and broken. Pink, gray, light and dark brown and green are the common colours in the plane light (PPL) (Fig. 6.9 T1. In cross light (XPL) zoning from brown, greenish brown, pink, yellowish green, light green, blue, purple to yellow rims from the centre towards the margin are observed. Most of the grains have smooth crystal faces and are devoid of inclusions and dissolution pits (Fig. 6. 10 1t – 2t). Overall these tourmalines have small grain size in comparison to those (Jurassic) overlying the laterite.

In the Jurassic strata (overlying the thick laterite) the tourmalines average 51.7% of the total assemblage and are typically coarse grained due to apparent coarse grain size of the host sandstones. The tourmalines here are typically rounded to very well rounded (Fig. 6.10 4t, 5t), broken, etched and corroded (Fig. 6.19 T4, T5) and euhedral crystals are rare. The

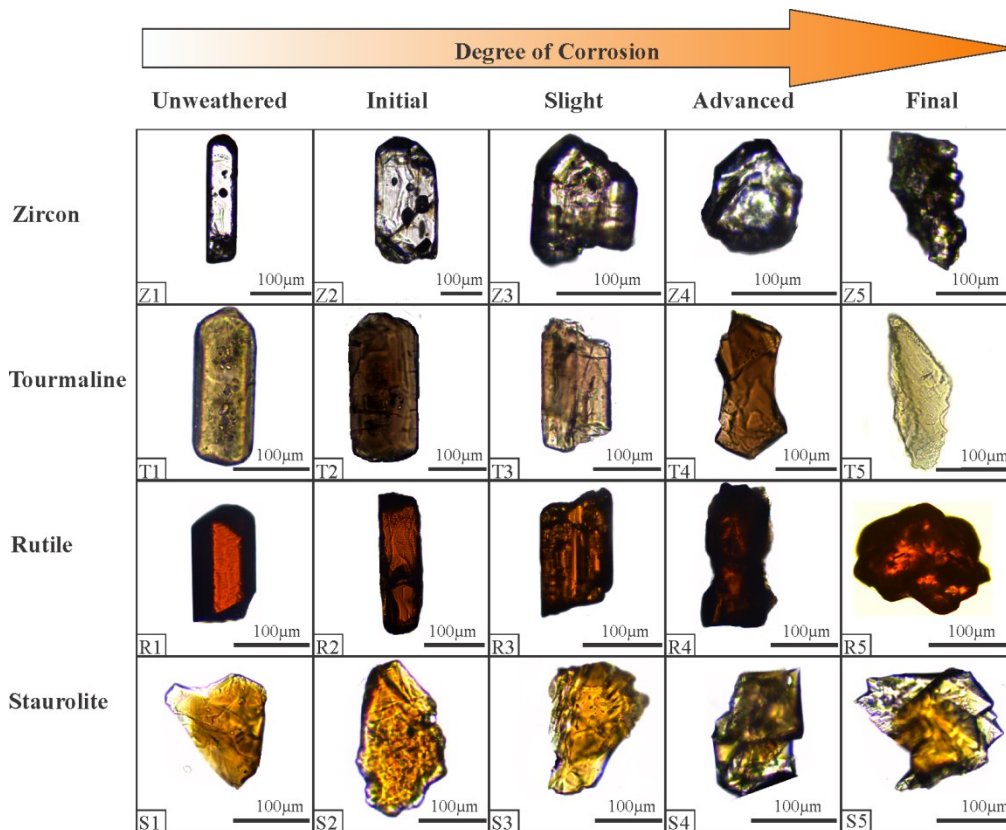


Figure 6.9 Heavy minerals indicating increasing degree of corrosion from left (unweathered stage) to right (extreme stage).

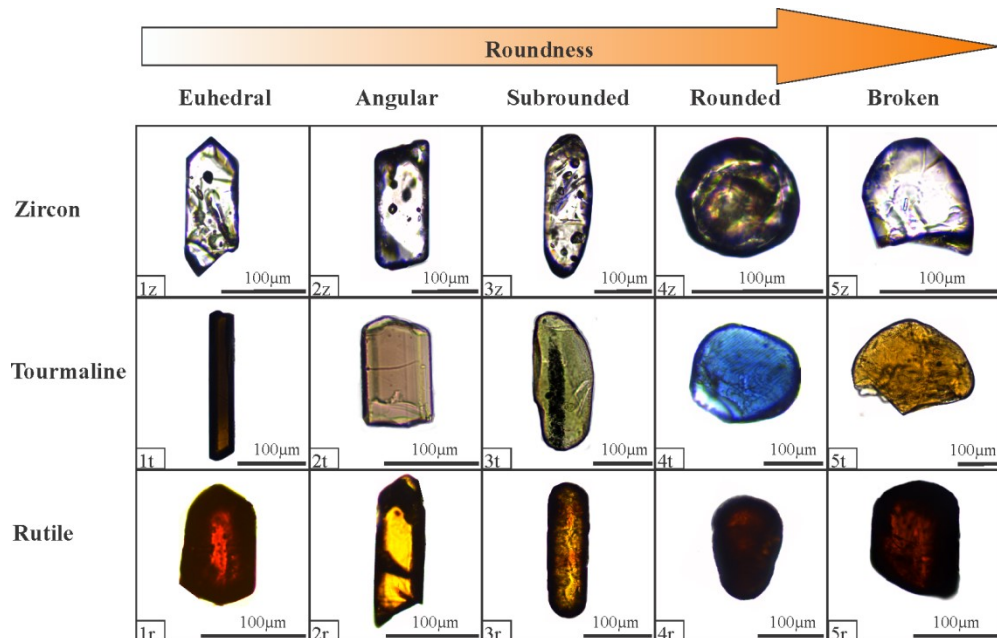


Figure 6.10 Representative heavy minerals plotted in order of increasing degree of roundness from left (euhedral) to right (broken).

tourmalines display various shades of green and greenish brown colour (Figs. 6.9 T4; Fig. 6.10 3t, 5t), pink and purple grains (Fig. 6.9 T3) occur whereas rare blue grains (Fig. 6.10 4t) also occur. With few exceptions of the sub-angular to subrounded crystals fluid inclusions are uncommon.

6.4.2 Zircon

Zircon is the second most abundant grain among the transparent heavy mineral and constitutes 32% of the Triassic heavy minerals population and 29.4% in the Jurassic. Most of the zircons are subrounded to very well rounded (Fig. 6.10 4z) and broken (Figs. 6.9 Z3; Fig. 6.10 5z), however, a substantial number of the zircons are euhedral (Fig. 6.9 Z1; Fig. 6.10 1z) and angular (Fig. 6.9 Z2; Fig. 6.10 2z). In the rounded grains the parent crystal shapes are difficult to identify, however, where possible to recognise the elongated crystals are dominant. Zoning is common and green-yellowish green to pink and yellow bands are commonly observed in XPL. Fluid inclusions are common (Fig. 6.9 Z2) while concentric fractures originating from the center of the crystal and radiating to its margin exist (Fig. 6.9 Z3). Colourless zircons predominates (Fig. 6.9 Z1; Fig. 6.10 1z) but pinkish ones are also common (Fig. 6.9 Z2). Etching, solution pits, corrosion and roundness differentiate the Jurassic zircons from the Triassic zircons which have sharp faces, edges and sharp broken surfaces.

6.4.3 Rutile

Rutile is the third most abundant transparent heavy mineral in the sediments and averages 21.5% of the assemblage (23.9% in Triassic sediments and 17.7% in the Jurassic). The rutile grains generally have thick, dark borders and are dominantly red to reddish brown in colour (Fig. 6.9). Some grains display shiny, dark yellowish brown colour. The Triassic grains generally have sharp margins and well preserved crystal faces (Fig. 6.9 R1) while roundness, corrosion and dissolution effects increase in the Jurassic grains (Fig. 6.9 R3; Fig. 6.10 4r).

6.4.4 Staurolite

Staurolite is the least abundant heavy mineral averaging nearly 3%. The only visible difference in the heavy mineral assemblage is the higher concentration of staurolite in the Triassic samples (averaging 4.1%) that diminishes to 1% in the Jurassic samples (Fig. 6.11).

Staurolite grains have marked bright yellow interference colour and straight extinction (Fig. 6.9). The subconchoidal fracture, straw yellow to yellowish brown pleochroism and saw teeth structure at the margin are the typical properties (Fig. 6.9 S1 – S5).

6.4.5 ZTR Index

The total sum of zircons, tourmalines and rutile (ZTR) is called ZTR-Index (Hubert, 1962) which displays extremely high values throughout the succession. The lowest ZTR-Index is 87 indicated for the lowest beds of the basal unit (Fig. 6.11). The ZTR-Index displays an increasing upsection trend and reaches 100 in the uppermost part of the succession (Fig. 6.11). The Triassic succession (below the thick laterite) displays an average ZTR-Index of 95.7 while it reaches to an average 98.9 in the overlying Jurassic strata. Thus the ZTR-Index reveals heavy mineral assemblage of strongly reworked and recycled sediments.

6.5. Tectonic setting using heavy minerals

6.5.1 MF–GM–MT plot

The mafic magmatic (MF), granitic and silicic metamorphic (GM), and mafic metamorphic (MT) diagram MF–GM–MT (Fig. 6.12a) of the heavy minerals assemblage provides a good discrimination between various tectonic settings (Nechaev and Isphording, 1993) and associated source rock types (Jafarzadeh et al., 2014). In the present case none of the minerals typical of MF-suite (olivine, iddingsite, all pyroxenes and green–brown hornblende) were found. Similarly the MT-suite (includes blue–green and pale-coloured hornblende, epidote and garnets) also has no representative. The heavy minerals assemblage is composed of the three ultra-stable minerals zircon, tourmaline, rutile and minor staurolite that typically belongs to the GM-suite (includes zircon, tourmaline, rutile, staurolite, sillimanite, andalusite, monazite and kyanite). Therefore, the heavy mineral assemblage is plotted as unique dot on the GM apex (Fig. 6. 12a) and hence the MF-GM-MT diagram reveals that the present sediments represent a siliciclastic detritus typical of passive margin with no contamination from mafic metamorphic or igneous source.

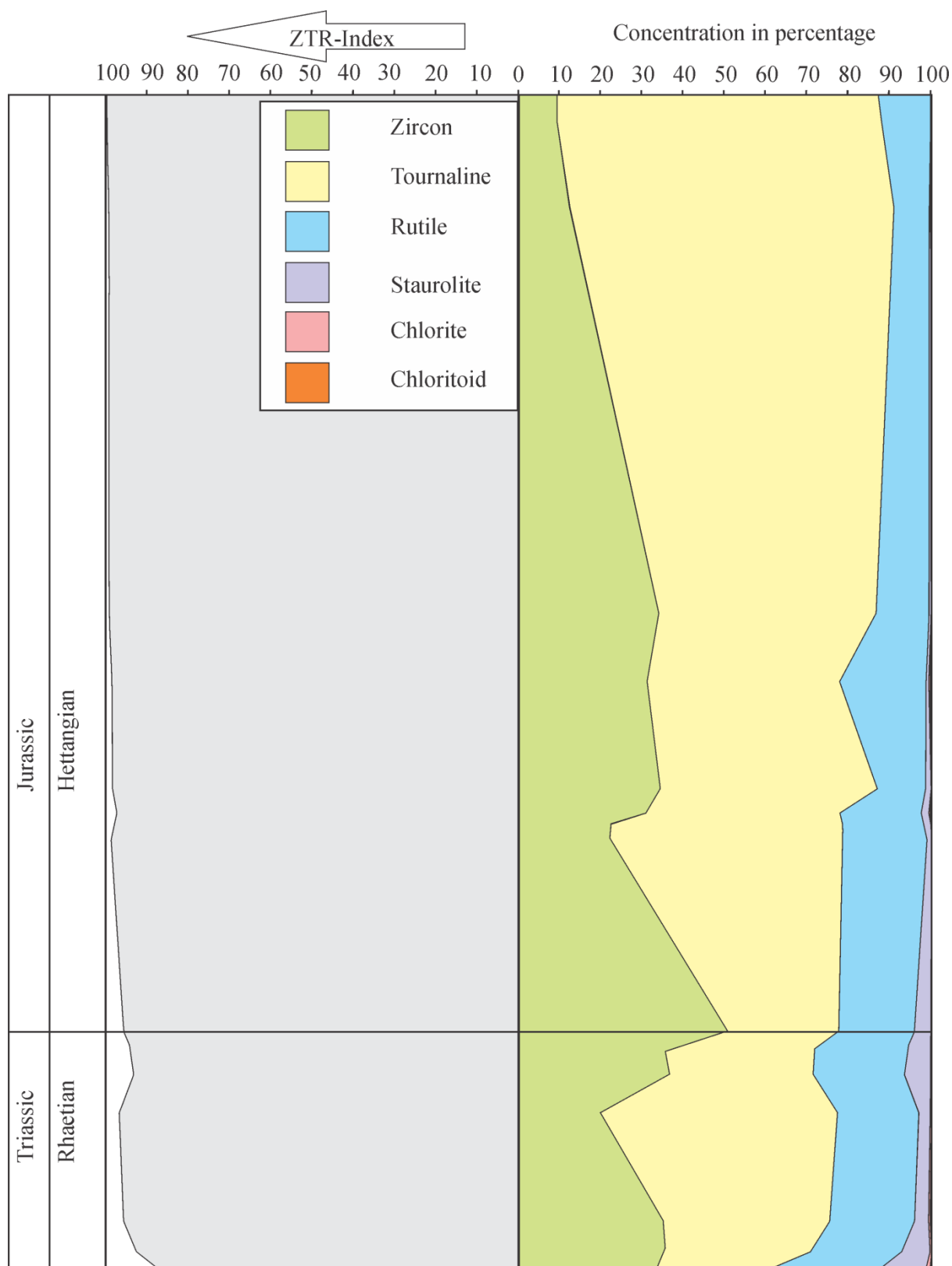


Figure 6.11 Heavy minerals distribution in the Kasanwala section. The ZTR-Index (Hubert, 1962) is displayed to the left side indicating an overall very high degree of sediments maturity.

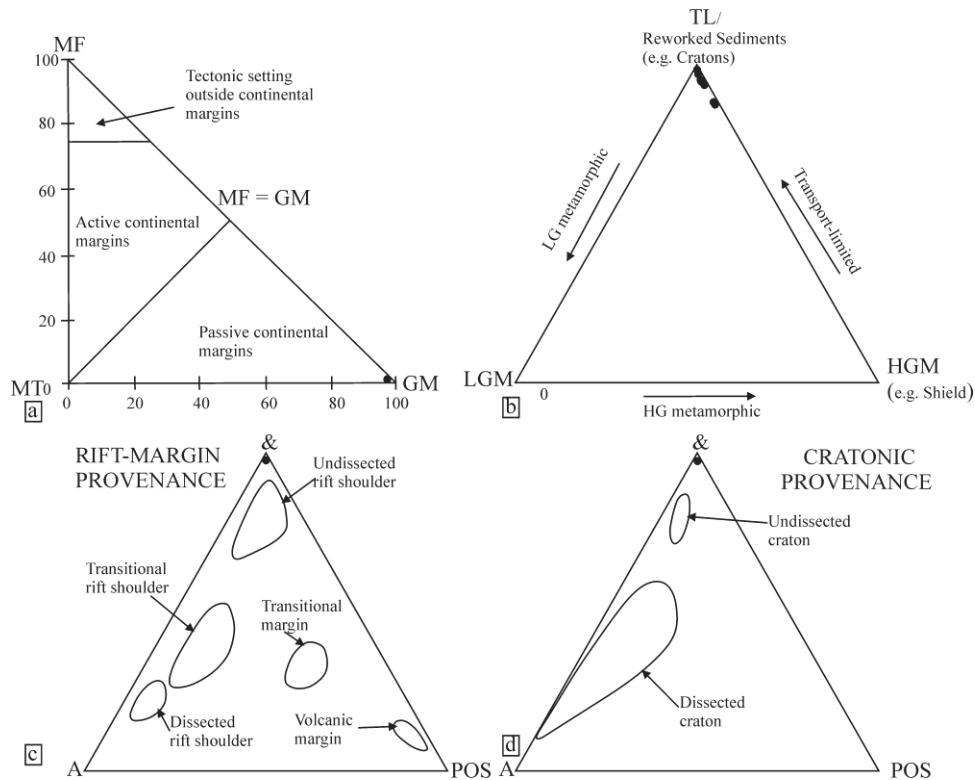


Figure 6.12 a) Tectonic setting discrimination MF-GM-MT heavy minerals plot (Garzanti and Andò, 2007) showing that the sediments are exclusively restricted to passive margin setting, b) Transport limited (TL)-Low Grade Metamorphic (LGM) – High Grade Metamorphic (HGM) plots indicating reworking of sediments in cratonic setting, c-d) &-A-POS ternary plots showing passive margin setting for the present sediments (Garzanti and Andò, 2007).

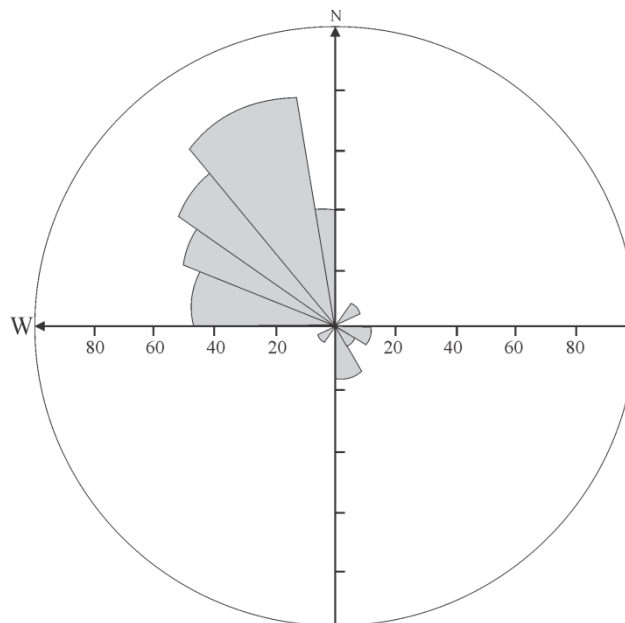


Figure 6.13 Palaeoflow data from cross-strata for the sediments favouring a north-northwestward transport of sediments. The numbers of readings are indicated on the horizontal axis.

6.5.2 Continental block provenance plots

Since the MF-GM-MT diagram (Fig. 6.12a) favours the passive margin setting for the sediments under study therefore the continental block provenance plots (Garzanti and Andò, 2007) are used to testify the hypothesis.

Continental block provenance plots are ternary diagrams for the rifted-margin and cratonic provenance (Fig. 6.12b-d). Presently, high concentration of ultra-stable minerals (zircon, tourmaline and rutile) represents a single dot at the (&) apex (Fig. 6.12c, d). This apex includes all transparent heavy minerals but excluding those not included in the other two poles. The current study indicates no contribution of total amphiboles ($A = Hb$ (hornblende amphiboles) + & A (all other amphiboles e.g. glaucophane, tremolite, actinolite) nor from pyroxenes (P) + olivine (O) + spinel (S) = POS (Garzanti and Andò, 2007; Jafarzadeh et al., 2014). Thus the supreme mature quartz arenites and the concentration of ultra-stable and well-rounded heavy mineral suites (ZTR) provide unequivocal evidences for extensive recycling of the sediments with no contemporaneous volcanic activity. Such conditions are typically common along undissected rift shoulders where volcanic activity is insignificant and along dissected rift shoulders. The pre-rift strata are exposed to prolonged weathering and recycling and extensive outcrops of the underlying basement rocks are not common (Garzanti and Andò, 2007). The quartz rich clastic sands and the low diversity of the heavy minerals favour an “Undissected Rift Shoulder” provenance. The absence of quartzofeldspathic sands and lack of hornblendes (from amphibolite-facies basement rocks) negates the “Dissected Rift Shoulder” provenance (Garzanti and Andò, 2007).

Similarly, the cratonic provenance ternary plot (Fig. 6.12d) is comparable with “Rift Shoulder” suites (Fig. 6.12c). The presence of rounded and ultra-stable heavy minerals (ZTR and staurolite) characterises areas where upper-crustal basement levels and cover strata are widely preserved (Undissected Craton subprovenance). The lack of hornblendes negates deeply eroded shield areas that expose middle-crustal levels (Dissected Craton subprovenance) that can supply abundant heavy minerals dominated by mainly blue-green to green and brown hornblende.

6.5.3 TL–LGM–HGM plot

Since the present sediments lack any of the heavy minerals that can provide evidences for a basic igneous or metamorphic provenance therefore the ternary plot of the transport limited (TL) heavy minerals (zircon, tourmaline, rutile)-low grade metamorphic (LGM) heavy minerals (epidote, chloritoid, chlorite, actinolite)-high grade metamorphic (HGM) heavy minerals (sillimanite, staurolite, garnet, andalusite) become the ideal candidate for tectonic settings and palaeoclimatic discriminations (Fig. 6.12b; Vincent et al., 2014). The cluster of samples along the TL apex indicates the presence of only ultra-stable heavy mineral suites (ZTR and staurolite). This supports sediments derivation from a mature and recycled source and the complete exclusion of unstable heavy minerals. It has already been interpreted that such evidences are indicative for cratonic basins, craton lowlands and stable basins (Garzanti and Andò, 2007). Additionally, such mature quartz rich sediments and heavy mineral suites may also indicate uplift and erosion of mature sedimentary rocks (Vincent et al., 2014). The textural maturity and palaeoclimatic and palaeoenvironmental interpretation of the sediments (Iqbal et al., 2015b; 2018, under review) support this interpretation.

6.6 Sediments transport pathways

Certain field features provide evidences for the source to sink sediments transport pathways. Among these are 1) thickness and grain size variation and 2) palaeoflow indicators:

6.6.1 Sediments thickness and grain-size variation

The lateral sediments thickness and grain-size variation trend is discussed in detail in Iqbal et al. (2018, under review). Field observations indicate that the Upper Triassic–Lower Jurassic siliciclastics display an east-southeast to west/northwest increasing thickness from western Salt Range to the Surghar Range (Iqbal et al., 2018 under review). The eastern most section in the western Salt Range (Kasanwala section) is 130 m thick that reaches to 200 m in the (Zaluch Nala) westernmost part of the western Salt Range. Further to the west the strata is thickest in the Surghar Range (235 m in the Gulla Khel section). Further to the west/southwest the thickness again reduces and the formation reduced to 140 m in the

southwestern Khisor Range (Paniala anticline section). In the north and northwest the formation is reduced to 25 – 30 m in the Kala Chitta Ranges and further northward it reduces to less than 10 m in Hazara area where it completely disappears in the northern part of the Galyat area (Hazar Kashmir syntaxis).

The Datta Formation has coarse quartz-rich conglomerates and pebbly sandstones channel belt facies in the western Salt Range. Among these the coarsest pebbles (average diameter = 5 – 7 cm) occur in the Kasanwala section (eastern part of western Salt Range) and Zaluch Nala (westernmost part of western Salt Range) has relatively fine grained pebbles (average diameter = 2 – 3 cm). Further to the west, the conglomerates disappear and in the Surghar Range medium to coarse grained sandstones are dominant. Further to the southwest the in the Khisor Range medium grained, uniform sandstones are dominant.

6.6.2 Palaeoflow indicators

The conglomerates, pebbly sandstones, sandstones and siltstones possess abundant cross beds and cross bed sets (Fig. 6.13) that can be used to work out the sediments flow pathways. Additionally, the conglomerates have some aligned pebbles with their longer axes parallel to the bedding plan and the pebble's imbrication indicative of transport direction. In the eastern part of western Salt Range (Kasanwala and Kaowaali sections) the palaeocurrent analysis from the cross-strata reveals a southeast to northwest transport of the sediments (Iqbal et al., 2015b e.g. their Fig. 5). Toward the west the in the Nammal Gorge and Zaluch Nala the palaeocurrent reveals a similar southeast to northwest palaeoflow (Fig. 6.13). Further towards the west in the Surghar Range and western Khisor Range an east-southeast to west-northwest sediments palaeoflow is indicated but with a significant east-west component. In all the section except Kasanwala and Kaowaali the palaeocurrents have some bimodal component with a northwest to southeast flow component (Fig. 6.13).

The combined analyses of the thickness variation, grains size trend and palaeocurrents data reveal the existence of a fluvial system transporting sediments from the east-southeast toward west-northwest. The lithofacies and faices association variations indicate that mostly fluvial-continental and floodplain facies exist in the western Salt Range whereas fluvio-deltaic facies are common in the Surghar and Khisor ranges (Iqbal et al., 2015b; 2018,

under review). Thus, the western Salt Range occupied a comparatively proximal position to the source region while the Surghar Range was most probably the depocenter of the deltaic system where as the Khirsor Range (at least the southwestern part) was on the distal side of the depocenter. Further to the west and northwest in the axial belt some marine strata (Shah, 2009) indicate the distal limits of the deltaic system. Similarly, subsurface data indicate that the Datta Formation pinches out in the east-southeast in the Punjab platform.

6.7 Discussion

6.7.1. Palaeogeography and detritus suppliers

The palaeogeographic position (Sorkhabi, 2014) of the study area (Fig. 6.14a, b) puts the nearby Arabian plate, Greater Somalian plate (southeastern margin of Africa) and Indian plate as the candidates for the source of the siliciclastic detritus. However, both the pinching out of the siliciclastic and decreasing grain size towards the west-northwest (towards the Arabian plate) and south-southwest (towards the Greater Somalian plate) negate siliciclastics supply from these directions. Further, the palaeoflow analysis (Fig. 6.13) arguments against sources of sediment supply from these sites. Additionally, if the sediments would have been supplied from either the Arabian plate or the Greater Somalian plate then the Axial belt (western margin of the Indian plate) which is comparatively proximal to these supposed siliciclastic feeders would have received coarse siliciclastics. By contrast, the Upper Triassic–Lower Jurassic sediments of the Axial belt are comprised of carbonates with some shale (Shah, 2009). The coherent results of the facies associations, grain size analysis and palaeocurrents data point toward an easterly source of the sediments. Thus, the Indian plate most likely served as the source for the siliciclastics (Fig. 6.15).

6.7.2 NW India passive margin

The Salt Range Basin (Kohat-Potwar Basin) represents a rim basin on the NW margin of India (Fig. 6.14c) associated with the Neo-Tethys Rift system (Stampfli et al., 1991). It possibly opened into a rift that separated Arabia and India, that extended towards Madagascar (Pillervuit, 1993) and linked southwards up with the Karoo system (Wopfner, 1993). The

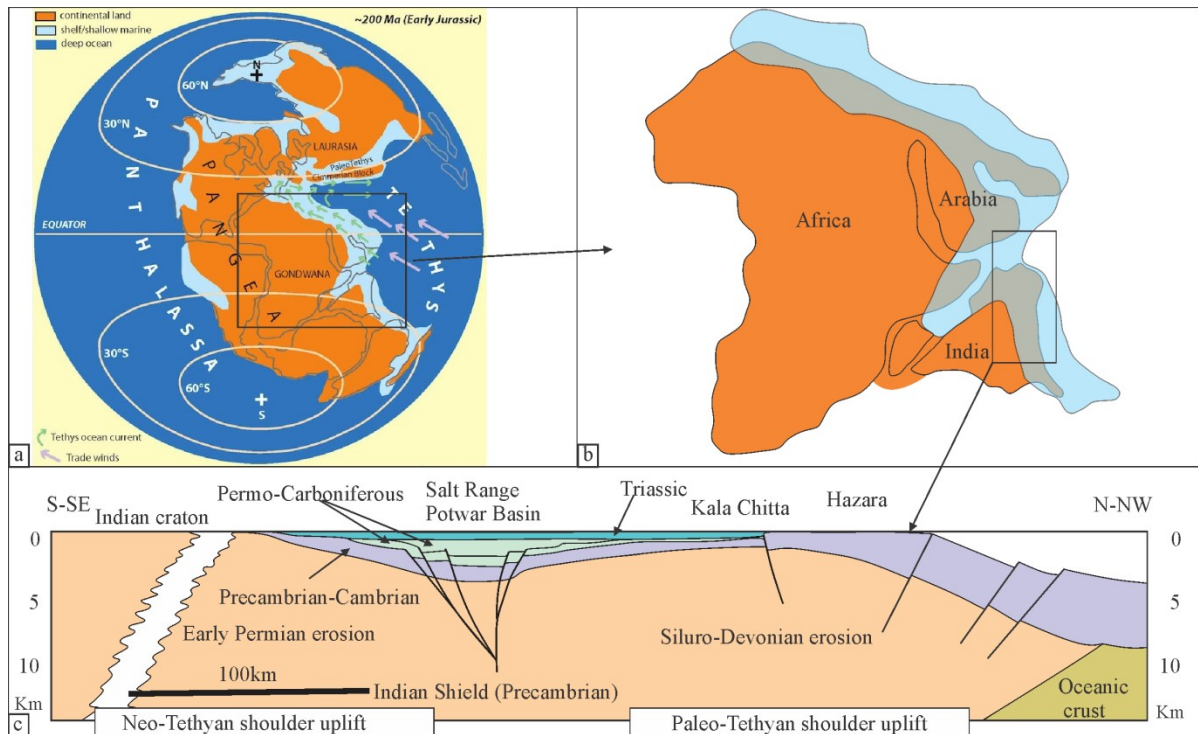


Figure 6.14 a, b) Palaeogeographic position of the Salt and Trans-Indus ranges during the Triassic–Jurassic boundary interval (Sorkhabi, 2014), and c) tentative basin cross section of the northwestern passive margin of the Indian plate during Late Triassic (modified from Stampfli et al., 2001).

ripping activity predates the Late Permian and most likely commenced during the Late Carboniferous. The initial ripping phase occurred during Early Carboniferous and was followed by middle and late ripping phases during the Early Permian and late Early Permian to Late Triassic respectively. The latter phase resulted in the opening of Neo-Tethys, development of passive margins and the thermal subsidence of the passive margin (Fig. 6.14c; Stampfli et al., 2001).

Faulting, uplift and tilting of strata produced shoulder uplift in different Indian basins during the Late Triassic (Mukhopadhyay et al., 2010; Iqbal et al., 2015a). This uplift resulted in exposure of the eastern Salt Range and adjoining areas by creating a north-northwestward tilt (Fig. 6.15a, b). Exhumation and the subsequent exposure of the sediments in the Salt Range and further south-southeast in greater India provided the main source sediments that were transported north-northwestward by the newly established fluvial system during the Early Jurassic (Fig. 6.16).

6.7.2a Possible source rocks

The roughly north-south trending, nearly 700 km long, Aravalli–Malani range (Fig. 6.16) is a part of the foreland normal-sense high strain zones that were active in the Paleoproterozoic (Martin, 2017). The range is located in the Delhi–Rajasthan area of northwest India, 600 km southeast of the Salt Range, and was a positive topographic feature by the Early Permian (Ghazi and Mountney, 2011). The Precambrian Aravalli Supergroup overlies the banded gneisses (Heron, 1953) and consists of granites, siliciclastics, carbonates, slates and phyllites (Sisodia and Chauhan, 1998). The Malani Suite comprised of porphyritic rhyolites and granites (Bhushan, 1984; Pascoe, 1959). The rhyolites have abundant pink orthoclase and some sanidine, oligoclase, quartz and rare hornblende and magnetite. The pink colour Malani Granite intrudes the rhyolites and is composed of orthoclase, quartz and sodic plagioclase, biotite, hornblende and less common occurrences of muscovite. Accessory minerals include tourmaline and fluorite.

The Lower Permian, glacial–periglacial Tobra Formation in the Salt Range contains pebbles and boulders of Aravalli–Malani origin (Jan et al., 2017). Similarly, the same ranges shed detritus for the Permian Warchha Sandstone of the Salt Range (Ghazi and Mountney, 2011). Further, the sedimentary clasts within the Warchha Sandstone are considered to have been the product of local recycling of the underlying Permian, Cambrian and Eocambrian strata in the Salt Range region itself (Ghazi and Mountney, 2011).

The exposure of those older sediments provided the potential source for the Triassic–Jurassic sediments. The equatorial to near-equatorial palaeogeographic position of the area (Fig. 6.14a) favoured the dissolution of the liable components, thereby producing supreme mature quartz arenites along the transport system (Iqbal et al., 2018 under review). The comparative low relief did not allow the derivation of sediments from diverse source rocks and the meandering fluvial system reworked and transported only the older Triassic and Permian siliciclastics (Fig. 6.15a). This is confirmed by the eastward disappearance of the Triassic and Permian strata in the present day Salt Range (Shah, 2009).

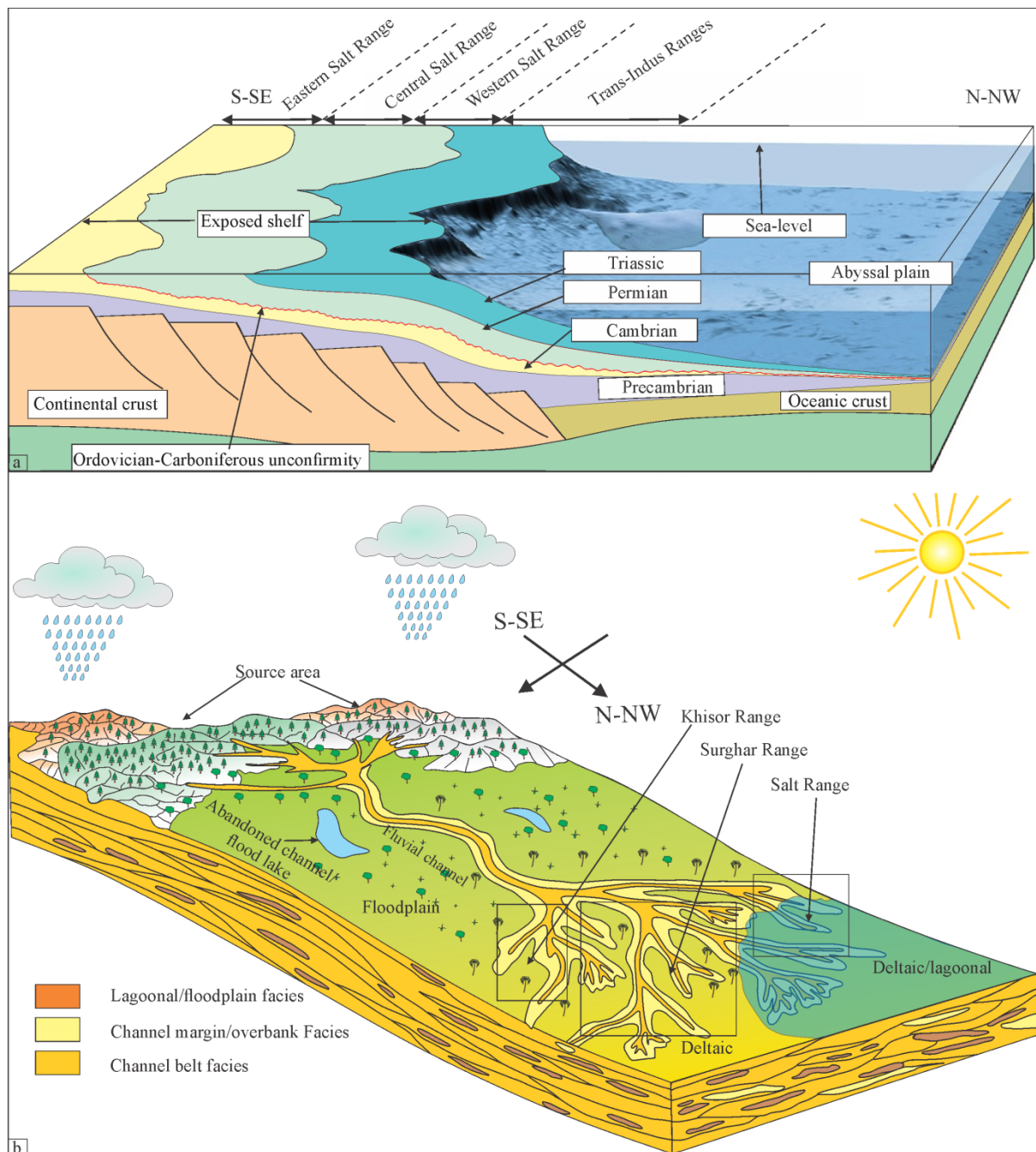


Figure 6.15 a) Cross-sectional view of the northwestern margin of the Indian plate around 200 Ma ago, b) source to sink transport model for the Late Triassic – Lower Jurassic siliciclastics in the Salt and Trans-Indus ranges. Note: both the models are not to scale.

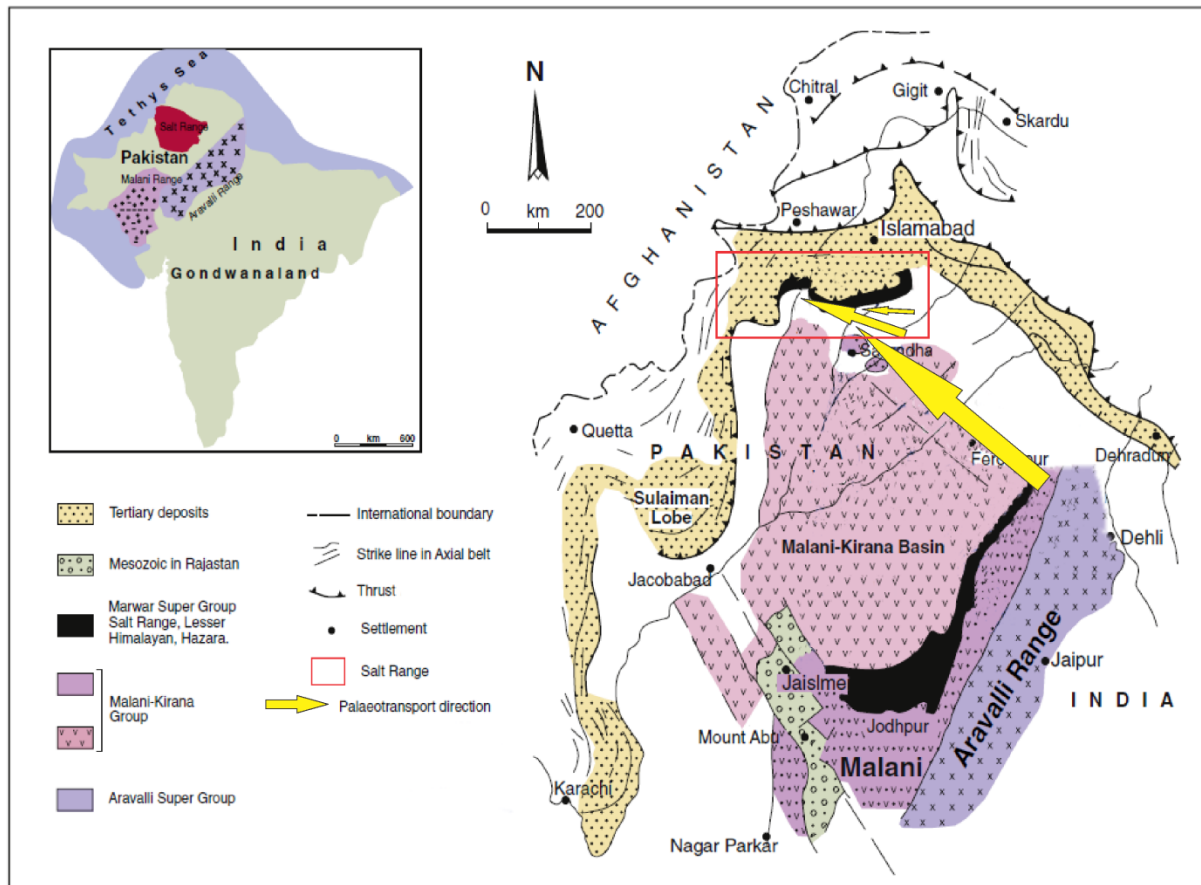


Figure 6.16. A generalised geological map of foreland Pakistan and adjoining northwest India showing the Salt Range and the possible source areas of the siliciclastics for the Upper Triassic–Lower Jurassic siliciclastics of the Kingriali and Datta formations (modified from Viridi, 1998; Ghazi and Mountney, 2011).

The extreme greenhouse conditions interpreted for the strata (Iqbal et al., 2018 under review) could have easily removed any unstable component thereby forcing supreme mature quartz arenites with very high ZTR-index, and hiding evidences for sediments incorporation from the basement or other diverse source. However, Cambrian and older strata are widely distributed throughout the Salt and Trans–Indus ranges (Fig. 6.15a) and reported in the subsurface in the Kohat–Potwar Plateau and Punjab platform (Shah, 2009) thereby indicating that the fluvial system was unable to cut into the Cambrian and older strata and the basement rocks were not exposed and not reworked.

The quartz arenite mineralogy of the sandstones (Figs. 6.1–6.2), geochemical plots (Figs. 6.3–6.8) and heavy minerals analyses (Figs. 6.8–6.12) favour sediment reworking on stable, passive margin settings. The heavy minerals analyses further attest sediment reworking

under undissected rifted-margin and undissected cratonic settings (e.g. passive margin setting) and negate the possibility of dissected rifted-margin or dissected cratonic settings (Figs. 6.13–6.15). The geochemical proxies, morphology of the individual quartz grains and heavy minerals data indicate sediments derivation from acidic source, most probably of granitic composition. The dominance of tourmaline in the heavy mineral assemblage provides evidences of detritus derivation from the Malani Granite which has tourmaline and fluorite as dominant accessory minerals (Ghazi and Mountney, 2011).

6.8 Conclusions

Following the demise of the Rhaetian platform carbonates deposition (Kingriali Formation) a fluvial system initiated during the Triassic–Jurassic boundary interval in the Salt and trans–Indus Ranges under greenhouse palaeoclimatic conditions. During the following Hettangian this fluvial system matured into a fluvio-deltaic system and was characterised by a general south-southeast to north-northwest palaeoflow. The fluvial system generally reworked the older (Carboniferous–Permian) siliciclastic strata and their detritus in the present day eastern Salt Range and adjoining western-northwestern part of the Indian craton under hot and humid palaeoclimate. These recycled sediments were deposited (as Datta Formation) in the present day western Salt Range and Trans–Indus ranges in a fluvio-deltaic depositional environment on a passive margin setting.

CHAPTER 7

PALYNOLOGY AND CARBON ISOTOPE STRATIGRAPHY

The Triassic–Jurassic transition is a major sequence boundary and in many marine basins, depending on their setting, resulted in a sedimentation break/hiatus (Haq et al., 1987; Hallam and Wignall, 1999, Hallam, 2001). Marine depositional environments became very shallow or even terrestrial (Hallam and Wignall, 1999; Haq, 2017). The lack of marine fauna, both due to extinction and sedimentary facies variation, makes the biostratigraphic resolution and correlation difficult (Marzoli et al., 1999; Olsen, 1999; Pálffy et al., 2001; Wignall, 2001; Galli et al., 2005). At the Global Stratotype Section and Point (GSSP), Kuhjoch East (Austria), the TJB corresponds to the first occurrence (FO) of the ammonite *Psiloceras spelae tiolicum* (Hillebrandt and Krystyn, 2009; Hillebrandt et al., 2013). The continental palynomorph *Cerebropollenites thiergartii* occurs 3.2 m below *P. spelae*, the closest event to the FO of *P. spelae* and allows a correlation with terrestrial sediments (Hillebrandt et al., 2013).

Palynological analysis is a useful tool to unravel vegetation patterns and thus reconstruct palaeoclimates (e.g., Barrón et al. 2006; Galfetti et al. 2007; Bonis and Kürschner, 2012). In northwest Europe, palynological records across the Triassic–Jurassic transition advocate the transient replacement of gymnosperm forests by fern and fern-associated vegetation. Such flora is generally the pioneer assemblage that occurs in disturbed ecosystems (Van de Schootbrugge et al., 2009). Similarly in the Newark Basin (North America), the transition indicates a major extinction (60%) of palynomorph taxa followed by a sharp spore spike (Fowell and Olsen, 1993; Fowell et al. 1994; Olsen et al. 2002).

Palynology and biostratigraphy are not major objectives of the present study; however, an attempt has been made to confirm the chronostratigraphic position of the Kingriali and Datta formations by using sampled fine-grained material.

7.1 Triassic–Jurassic Transition in the Salt Range, Pakistan

The lithological succession of the Triassic–Jurassic transition in the Salt Range has already been discussed in the previous chapters (Iqbal et al., 2015b, 2018 (under review). 50

Samples for palynological analysis were collected from shale beds of the succession covering all the facies variations in the Nammal Gorge (Salt Range), Gulla Khel (Surghar Range) and Paniala anticline section (Khisor Range). The present results support the presence of a rather complete record of the Triassic–Jurassic sedimentary transition in the Salt and Trans–Indus ranges, Pakistan. The palynomorphs assemblage is discussed in the following:

7.1.1 Kingriali Formation

23 shale samples from the upper part of the Kingriali Formation were processed for palynological analysis. The shales yielded pollen dominated assemblage and the palynomorphs dominantly consists of excellently preserved diverse bisaccate pollens indicating arid conditions (Fig. 7.1). The aquatic assemblage yielded the green algae *Prasinophytes* that attests sedimentation in a marine depositional environment. The age diagnostic assemblage of dinoflagellate cysts include *Rhaetogonyaulax*, *Suessia* and *Beaumontella* suggesting a Rhaetian age for the formation.

7.1.2 Datta Formation

27 samples were analysed from the continental siliciclastics of the Datta Formation that yielded spore dominated assemblages reflecting humid conditions. An abrupt change from the Rhaetian palynomorphs to the Jurassic bisaccates indicates the Triassic–Jurassic transition (Bonis et al., 2009). In the study area, the first occurrence (FO) of *Cerebropollenites thiergartii* (Fisher and Dunay, 1981; Kürschner et al., 2007) in the lower part of the formation at Gulla Khel (Fig. 7.1) indicates the approximate base of the Hettangian. Upsection the Datta Formation yielded *Classopollis turosus* dominated pollen assemblages.

7.1.3 Carbon Isotope Stratigraphy

The organic carbon isotope analysis yielded values around $-25 \delta^{13}\text{C}_{\text{org}} \text{‰}$ in the succession. However, the shales in the uppermost part of the Kingriali Formation yielded significantly more depleted values of about -33 to $-30 \delta^{13}\text{C}_{\text{org}} \text{‰}$. These distinct negative values apparently correspond to the initial C-isotopic excursion (e.g. Pálffy et al., 2001; Hesselbo et al., 2002; Guex et al., 2004; Kürschner et al., 2007; Ward et al., 2007; Ruhl et al., 2009) at

the Triassic–Jurassic transition. However, the lack of further data inhibits this potential correlation.

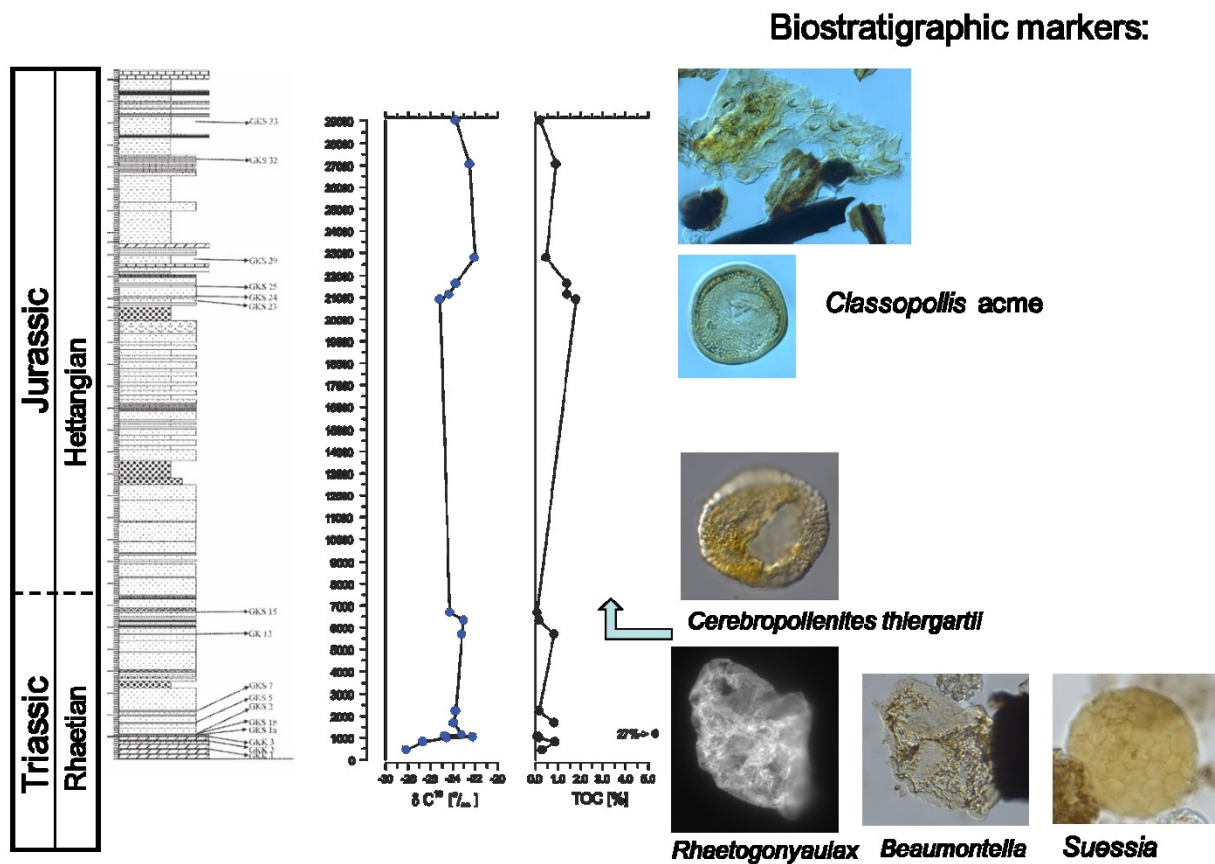


Figure 7.1 Representative section (Gulla Khel) with biostratigraphic markers, C-isotope and TOC (total organic carbon) curves. The first appearance of *Cerebropollenites thiergartii* marks the approximate base of the Hettangian and dinoflagellate cysts confirm the Rhaetian age of the Kingriali Formation (Vertical scale in cm).

CONCLUSIONS

In this section, the main conclusions of this thesis work are summarised in the order as these are presented in the thesis. In the western Salt Range and adjoining Trans–Indus ranges of Pakistan, a thick and laterally persistent dolomite of the Kingriali Formation represents the Upper Triassic platform carbonates. The overlying predominantly siliciclastic rich strata of the Datta Formation represents the Lower Jurassic fluvial-deltaic system in the area. The overall succession provides ideal sections for the Triassic–Jurassic transition from the tropical southeastern margin of the Pangaea supercontinent facing western Tethys (northwestern margin of the Indian plate).

Field evidences, petrographic studies (framework mineralogy), bulk sediments mineralogy (XRD) and bulk sediments geochemistry data indicate a prominent sea–level fall, demise of the Upper Triassic carbonate platform at the Rhaetian–Hettangian transition (Triassic–Jurassic transition) and establishment of a siliciclastic dominated fluvial-deltaic system in the area during the Hettangian. Lithofacies and facies association analyses indicate the deposition of channel belt facies, channel margin and overbank facies and lagoonal/marginal marine facies of the Datta Formation during Early Jurassic in the western Salt Range. Westward in the Surghar and Khisor ranges (Trans–Indus ranges) the Datta Formation displays facies typical of fluvial-continental setting and the lagoonal/marine facies pinch out westward. The dark grey, black Toarcian shales of the Shinawari Formation represent the next prominent marine transgression and demise of the fluvio-deltaic system of the Datta Formation in the area.

Palaeoclimatic signals from field data, framework mineralogy, XRD, oriented clay mineralogy and bulk sediments geochemical proxies indicate a change from dominantly arid-semiarid palaeoclimate during Rhaetian time to a hot and humid greenhouse palaeoclimatic condition in the early Hettangian. The Kingriali Formation with its gradual increasing upsection illite-rich shales and a parallel decrease in the carbonates thickness represents the gradual onset of a transition from a carbonate-rich depositional system to a siliciclastic-rich phase under warm/hot arid palaeoclimate during the late Rhaetian. The topmost shale and the associated evaporites (gypsum) of the Kingriali Formation further provide unequivocal evidences for arid–hothouse palaeoclimate during late Rhaetian. The topmost green shale

with its high total rare earth elements content (REE) presents a potential correlation horizon between this unit and volcanism related to the Central Atlantic Magmatic Province (CAMP) activity. The overlying supreme mature quartz arenites, kaolinite-rich clays and laterite–palaeosol horizons of the Datta Formation were deposited under hot and humid greenhouse conditions during the Hettangian.

Provenance analysis indicates that the Datta Formation consists mainly of siliciclastics of recycled origin. Evidences based on lateral sediments grain-size variation, palaeoflow analysis from cross-strata, framework composition, heavy minerals analysis and bulk geochemical proxies support sediments derivation from an east-southeast source. Shoulder uplift of the eastern Salt Range and areas further east-southeast in the form of the Indian craton during the Neothetys opening provided the source areas for sedimentary detritus. The supreme mature quartz arenite nature of the framework mineralogy and purely ultrastable heavy mineral assemblage composed of tourmaline, zircon and rutile advocates the recycling of the Permian and older siliciclastic sediments of the eastern Salt Range and further east-southeast. The acidic nature of the siliciclastics, as indicated by the geochemical proxy diagrams put rocks of the Aravalli–Malani ranges as the most probable sources of the siliciclastic detritus. Sediments reworking and recycling on a passive margin setting under the prevailing hot and humid conditions differentially removed the unstable components from the sediment load thereby producing supreme mature sediments rich in quartz, kaolinite, laterites-palaeosols and characterised by stable heavy minerals like zircon, tourmaline and rutile, i.e. a very high ZTR–Index.

Palynological analysis of the shale beds in the upper part of the Kingriali Formation yielded well preserved palynomorph assemblages dominantly consisting of diverse bisaccate pollen and the green algae *prasinophytes* indicates a marine depositional environment. The dinoflagellate cysts *Rhaetogonyaulax*, *Suessia* and *Beaumontella* confirm a Rhaetian age for the Kingriali Formation. The Datta Formation yielded a spore dominated palynological assemblage. The first occurrence of *Cerebropollenites thiergartii* corresponds to the base of the Hettangian. Further upsection, the formation yielded monotonous pollen assemblages dominantly consisting of *Classopollis turosus*. Thus, the Salt and Trans–Indus ranges of Pakistan preserve a complete record of sedimentation across the Triassic–Jurassic transition; thereby provide sections from the then southern hemisphere for this critical time

interval. Regional and global correlations based on the presented data and proxies indicates the change from a warm-arid – semiarid Rhaetian palaeoclimate to extreme warm and humid climate conditions in the Triassic–Jurassic boundary interval.

SCHLUSSFOLGERUNGEN

In diesem Abschnitt werden die wichtigsten Schlussfolgerungen dieser Dissertation in der Abfolge der Zusammenstellung zusammengefasst. In der westlichen Salt Range und den benachbarten Trans-Indus Ketten von Pakistan werden die Obertrias Plattformkarbonate durch mächtige, lateral anhaltende Dolomite der Kingriali-Formation repräsentiert. Die darüberliegenden vorwiegend siliziklastika-reichen Schichten der Datta-Formation bilden das unterjurassische fluviatile-deltaische Ablagerungssystem im Untersuchungsgebiet. Die generelle Abfolge bietet ideale Abfolgen für den Übergang von der Trias in den Jura am tropischen südöstlichen Rand des Superkontinents Pangäa am westlichen Tethysrand (nordwestlicher Rand der Indischen Platte).

Geländebefund, petrographische Untersuchungen (Sedimentmineralogie), Gesamtgesteinsmineralogie (XRD) und Gesamtgesteinschemismus zeigen einen ausgeprägten Meeresspiegelabfall, den Untergang der obertriassischen Karbonatplattform im Übergang vom Rhätium ins Hettangium (Trias-Jura-Grenze), und den Einsatz eines siliziklastika-dominierten fluviatilen-deltaischen Systems im Untersuchungsgebiet während des Hettangiums. Lithofazies und Faziesvergesellschaftungen zeigen innerhalb der Datta-Formation Ablagerungen von Rinnenfazies, Rinnenrandfazies und Überschwemmungsbereichen sowie eine lagunäre/randmarine Fazies während der Frühen Juras in der westlichen Salt Range. Westlich davon, in den Surghar und Kishor ranges (Trans-Indus Ketten) weist die Datta-Formation Faziesbereiche charakteristisch für fluviatil-kontinentale Ablagerungen und die lagunäre/randmarine Fazies keilt gegen Westen aus. Die dunkelgrauen und schwarzen Tonsteine des Toarcium der Shinawari-Formation repräsentieren die folgende deutliche Transgression und das Ende des fluviatilen-deltaischen Systems der Datta-Formation im Untersuchungsgebiet.

Paläoklimasignale aus dem Geländebefund, Sedimentmineralogie, XRD, orientierten Tonmineralproben und Gesamtgesteinsgeochemie zeigen eine Änderung von dominierend arid-semiarid Paläoklima im Rhätium zu heiß-humiden Treibhaus-Paläoklima im Frühen Hettangium. Die Kingriali-Formation weist graduell ansteigende Illit-reiche Tonsteinlagen und parallel eine Abnahme der Mächtigkeiten der Karbonatlagen auf. Dies zeigt den graduellen Einsatz eines Übergangs von einem karbonatreichen Ablagerungssystem zu einer

siliziklastika-reichen Phase unter einem warm-heißen ariden Paläoklima im Rhätium an. Die oberste Tonsteinlage und damit verbundene Evaporite (Gips) der Kingriali-Formation weist ebenfalls eindeutig auf ein arides extremes Treibhaus-Paläoklima ("hothouse") im späten Rhätium hin. Die obersten grünen Tonschiefer zeigen hohe Gehalte an Seltenen Erden (REE) und können als potentieller Korrelationshorizont zum Vulkanismus der Zentralatlantischen Magmatischen Provinz (CAMP) interpretiert werden. Die überlagernden extrem textuell reifen Quarzarenite, Kaolinit-reichen Tone und lateritische Paläobödenhorizonte der Datta-Formation wurden unter einen heißen und humiden Treibhausklima im hettangium abgelagert.

Herkunftsgebietsanalyse zeigt dass die Datta-Formation vorwiegend aus aufgearbeiteten siliziklastischen Material besteht. Auf Grund der Korngrößenvariation, der auf Schrägschichtungen beruhenden Paläoströmungsanalysen, der Mineralzusammensetzung, der Schwermineralanalyse und Gesamtgesteinschemismus kann eine ost-südöstliche Sedimentquelle angenommen werden. Die Anhebung tektonischer Horste in der östlichen Salt Range und in Gebieten ost-südöstlich davon wie etwa den Indischen Kraton während der Neotethysöffnung stellten die Quellen für das sedimentäre Material dar. Die extrem reife Quarzzusammensetzung und der hohe Anteil an ultrastabilen Schwermineralen wie Turmalin, Zirkon und Rutil spricht ebenfalls für die Aufarbeitung permischer und älterer siliziklastischer Sedimente der westlichen Salt Range und weiter süd-östlich. Der saure Gesteinscharakter wird durch geochemische Diskriminationsdiagramme weisen Gesteine der Aravalli–Malani ranges als wahrscheinlichste Herkunftsgebiete aus. Sedimentaufarbeitung und Wiederablagerung an einem passiven Kontinentalrand unter vorherrschenden heißen und humiden Bedingungen löste selektiv instabile Komponenten der Sedimentfracht und produzierte damit extrem reife Sedimente reich an Quarz, Kaolinit, Lateritische Paläoböden, charakterisiert durch Schwerminerale wie Zirkon, Turmalin und Rutil (hoher ZTR Index).

Palynologische Analysen der tonigen Lagen im oberen Teil der Kingriali-Formation zeigen gut erhaltende Palynomorphenvergesellschaftungen mit dominanten diversen bisaccaten Pollen und der ein marines Ablagerungsmilieu anzeigenden Grünalge *Prasinophytes*. Dinoflagellatenzysten wie *Rhaetogonyaulax*, *Suessia* und *Beaumontella* bestätigen ein Rhätium Alter für die Kingriali-Formation. Die Datta-Formation erbrachte eine sporen-dominierte Vergesellschaftung. Das Erstauftreten von *Cerebropollenites thiergartii* zeigt eine

ungefähre Einstufung an die Basis des Hettangiums. Im Hangenden tritt eine monotone Pollenvergesellschaftung mit vorherrschend *Classopollis turosus* auf. Aufbauen auf diesen Daten weisen die pakistanischen Salt Range und Trans-Indus Range eine komplette Sediment-Abfolge über den Übergang Trias-Jura auf und stellen damit herausragende Archive für die Südhemisphäre dar. Regionale und globale Korrelationen basierend auf den präsentierten Daten und Analysen bezeugen den Wechsel von einem warm-arid/semiariden rhätischen Paläoklima zu extrem warmen und humiden Klima im Trias-Jura Grenzübergang.

REFERENCES

Note: The references list includes only those references which have been cited in chapters 1, 2, 6 and 7. References for chapters 3, 4, and 5 are given at the end of the respective chapters (articles/manuscript).

REFERENCES

- Abre, P., Cingolani, C., Zimmermann, U., Cairncross, B., and Chemale Jr, F. (2011). Provenance of ordovician clastic sequences of the San Rafael block (central Argentina), with emphasis on the Ponón Trehué formation. *Gondwana Research*, 19(1), 275-290. <https://doi.org/10.1016/j.gr.2010.05.013>.
- Alroy, J., (2010). Fair sampling of taxonomic richness and unbiased estimation of origination and extinction rates. In: Alroy, J., Hunt, G. (Eds.), *Quantitative Methods in Paleobiology: The Paleontological Society Papers*. 16, pp. 55–80. <https://doi.org/10.1017/S1089332600001819>.
- Ashraf, M., Ahmad, M., and Faruqi, F.A., (1976). Jurassic Bauxite and kaolinite Deposits of Chhoi area Kala Chitta Range, Punjab, Pakistan. *Geol. Bull. Punjab. Univ.* 41, pp 41-54.
- Asrarullah., (1967). Geology of the Khewra dome. *Pakistan Sci. Cong.*, 18th/19th, Jamshro, Proc, PtIII ,Abs., pp. F3-F4.
- Bahlburg, H., and Dobrzinski, N., (2011). A review of the Chemical Index of Alteration (CIA) and its application to the study of Neoproterozoic glacial deposits and climate transitions. *Geological Society, London, Memoirs*, 36(1), 81-92. <https://doi.org/10.1144/M36.6>.
- Barrón, E., Gómez, J. J., Goy, A., and Pieren, A. P. (2006). The Triassic–Jurassic boundary in Asturias (northern Spain): Palynological characterisation and facies. *Review of Palaeobotany and Palynology*, 138(3-4), 187-208. <https://doi.org/10.1016/j.revpalbo.2006.01.002>.
- Basu, A., Young, S. W., Suttner, L. J., James, W. C., and Mack, G. H. (1975). Re-evaluation of the use of undulatory extinction and polycrystallinity in detrital quartz for provenance interpretation. *Journal of Sedimentary Research*, 45(4) 873-882.
- Bhatia, M. R. (1983). Plate tectonics and geochemical composition of sandstones. *The Journal of Geology*, 91(6), 611-627.
- Bhatia, M. R., (1985). Rare earth element geochemistry of Australian Paleozoic graywackes and mudrocks: provenance and tectonic control. *Sed Geol* 45 : 97-113. [https://doi.org/10.1016/0037-0738\(85\)90025-9](https://doi.org/10.1016/0037-0738(85)90025-9).

- Bhatia, M. R., and Crook, K. A. (1986). Trace element characteristics of graywackes and tectonic setting discrimination of sedimentary basins. *Contributions to mineralogy and petrology*, 92(2), 181-193. DOI <https://doi.org/10.1007/BF00375292>.
- Bhushan, S. K. (1984). Classification of Malani Igneous Suite. *Geol. Surv. India Spec. Publ.*, v.12, pp.199–205.
- Blain, S., Quéguiner, B., Armand, L., Belviso, S., Bombled, B., Bopp, L., Bowie, A., Brunet, C., Brussaard, C., Carlotti, F. and Christaki, U., (2007). Effect of natural iron fertilization on carbon sequestration in the Southern Ocean. *Nature*, 446(7139), 1070-1074. DOI: <https://doi.org/10.1038/nature05700>.
- Blake, J. M., Peters, S. C., and Johannesson, K. H. (2017). Application of REE geochemical signatures for Mesozoic sediment provenance to the Gettysburg Basin, Pennsylvania. *Sedimentary geology*, 349, 103-111. <http://dx.doi.org/10.1016/j.sedgeo.2016.12.009>.
- Blumenberg, M., Heunisch, C., Lückge, A., Scheeder, G., and Wiese, F. (2016). Photic zone euxinia in the central Rhaetian Sea prior the Triassic-Jurassic boundary. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 461, 55-64. <https://doi.org/10.1016/j.palaeo.2016.08.007>.
- Boggs, S. Jr., (2009). *Petrology of Sedimentary Rocks*; 2nd edn, Cambridge University Press, pp. 166–193.
- Boniface, N., and Mruma, A. H. (2012). Structural analysis, metamorphism, and geochemistry of the Archean granitoids-greenstones of the Sukumaland Greenstone Belt around Geita Hills, Northern Tanzania. *Natural Science*, 4(08), 526. <http://dx.doi.org/10.4236/ns.2012.48070>.
- Bonis, N. R., and Kürschner, W. M. (2012). Vegetation history, diversity patterns, and climate change across the Triassic/Jurassic boundary. *Paleobiology*, 38(2), 240-264. <https://doi.org/10.1666/09071.1>.
- Bonis, N. R., Kürschner, W. M., and Krystyn, L. (2009). A detailed palynological study of the Triassic–Jurassic transition in key sections of the Eiberg Basin (Northern Calcareous Alps, Austria). *Review of Palaeobotany and Palynology*, 156(3-4), 376-400. <https://doi.org/10.1016/j.revpalbo.2009.04.003>.
- Bracciali, L., Marroni, M., Pandolfi, L., Rocchi, S., Arribas, J., Critelli, S., and Johnsson, M. (2007). Geochemistry and petrography of Western Tethys Cretaceous sedimentary

- covers (Corsica and Northern Apennines): from source areas to configuration of margins. special papers-geological society of America, 420, 73-93. DOI: 10.1330/2006.2420(06).
- Brown, G., and Brindley, G. W. (Eds.). (1980). Crystal structures of clay minerals and their X-ray identification (Vol. 5, pp. 305-360). London: Mineralogical Society.
- Buggle, B., Glaser, B., Hambach, U., Gerasimenko, N., and Marković, S. (2011). An evaluation of geochemical weathering indices in loess–paleosol studies. Quaternary International, 240(1), 12-21. <https://doi.org/10.1016/j.quaint.2010.07.019>.
- Cingolani, C. A., Manassero, M., and Abre, P. (2003). Composition, provenance, and tectonic setting of Ordovician siliciclastic rocks in the San Rafael block: Southern extension of the Precordillera crustal fragment, Argentina. Journal of South American Earth Sciences, 16(1), 91-106. [https://doi.org/10.1016/S0895-9811\(03\)00021-X](https://doi.org/10.1016/S0895-9811(03)00021-X).
- Climap, P. M., (1976). The surface of the ice-age Earth. Science (New York, NY), 191(4232), 1131-1137.
- Chaudhry, M. N., Chuhan, F. A., and Ghazanfar, M., (1994a). Microfacies, diagenesis, environment of deposition and burial history of Datta Formation from Bara Oter, Distt. Abbottabad: Kashmir Jour. Of Geol. Vol. 11-12 pp. 43-59.
- Chaudhry, M. N., manzoor, A., and Ghazanfar, M., (1994b). Sedimentology of Datta Formation at Kalapani, Abbottabad, Northwest Himalayas, Pakistan. Geol. Bull. Punjab Univ., 89, pp.11-20.
- Colbert, E. H., Morales, M., and Minkoff, E. C. (2001). Colbert's evolution of the vertebrates: a history of the backboned animals through time. Wiley-Liss. P247.
- Condie, K.C., (1991). Another look at rare earth elements in shales. Geochim. Cosmochim. Acta 55, 2527–2531. [https://doi.org/10.1016/0016-7037\(91\)90370-K](https://doi.org/10.1016/0016-7037(91)90370-K).
- Cotter, G. D. P. (1933). The geology of the part of the Attock district west of longitude 72°45'E. Memoirs of the Geological Survey of India, 55(2), 63-161.
- Cox, L. R., Currie, E. D., Davies, L. M., Gregory, J. W., Muir-Wood, H. M., and Spath, L. F. (1930). The fossil fauna of the Samana range and some neighboring areas (Vol. 15). Government of India Central Publication Branch. Vol. 15,15p.
- Cox, R., Lowe, D.R., Cullers, R.L., (1995). The influence of sediment recycling and basement composition on evolution of mudrock chemistry in the southwestern United States.

- Geochim. Cosmochim. Acta 59 (14), 2919–2940. [https://doi.org/10.1016/0016-7037\(95\)00185-9](https://doi.org/10.1016/0016-7037(95)00185-9).
- Cullers, R. L., (1994a). The chemical signature of source rocks in size fractions of Holocene stream sediment derived from metamorphic rocks in the Wet Mountains region, USA. Chem. Geol. 113, 327–343. [https://doi.org/10.1016/0009-2541\(94\)90074-4](https://doi.org/10.1016/0009-2541(94)90074-4).
- Cullers, R. L. (2002). Implications of elemental concentrations for provenance, redox conditions, and metamorphic studies of shales and limestones near Pueblo, CO, USA. Chemical Geology, 191(4), 305-327. [https://doi.org/10.1016/S0009-2541\(02\)00133-X](https://doi.org/10.1016/S0009-2541(02)00133-X).
- Cullers, R.L., Barrett, T., Carlson, R., Robinson, B., (1987). Rare Earth element and mineralogic changes in Holocene soil and stream sediment: a case study in the Wet Mountains, Colorado, USA. Chem. Geol. 63, 275–297. [https://doi.org/10.1016/0009-2541\(87\)90167-7](https://doi.org/10.1016/0009-2541(87)90167-7).
- Danilchik, W. (1961). The iron formation of the Surghar and western Salt Range, Mianwali District, West Pakistan. US Geol Surv Prof Pap, 424, 228-231.
- Danilchik, W., and Shah, S. M. I. (1967). Stratigraphic nomenclature of formations in TransIndus Mountains Mianwali District, West Pakistan. US. Geol. Surv., Project Report, 1-45.
- Davies, R. G., and Gardezi, A. H. (1965). The presence of Bouleiceras in Hazara and its geological implications. Geol. Bull. Panj. Univ, 5, 23-30.
- Dickinson, W. R., (1970). Interpreting detrital modes of graywacke and arkose: Jour. Sed. Petrology, v. 40, p. 695-707.
- Dickinson, W. R., and Suczek, C. A., (1979). Plate tectonics and sandstone compositions: Am. Assoc. Petroleum Geologists Bull., v. 63, p. 2164-2182.
- Dickinson, W. R., Beard, L. S., Brakenridge, G. R., Erjavec, J. L., Ferguson, R. C., Inman, K. F., R. A. Knepp, F. A. Lindberg, and Ryberg, P. T. (1983). Provenance of North American Phanerozoic sandstones in relation to tectonic setting. Geological Society of America Bulletin, 94(2), 222-235.
- Ewart, A. Taylors, R., and Capp, A. C. (1968). Trace and minor element geochemistry of the rhyolitic volcanic rocks, central North Island, New Zealand. Contributions to Mineralogy and Petrology, 18, 76-104.

- Fatmi, A. N. (1974). Lithostratigraphic units of the Kohat-Potwar province, Indus basin, Pakistan. *Mem. Geol. Surv. Pakistan*, 10, 1-80.
- Fatmi, A. N., Anwar, M., Hussain, I., Saeed, G., and Latif, A., (1990). Stratigraphy of the Shinawari And Datta formations from the Shinawari area (Western Samana Range), Kohat Distt., *Geol. Bull. Punjab. Univ.* 84, pp 47-55.
- Fisher, M.J., Dunay, R.E., (1981). Palynology and the Triassic/Jurassic boundary. *Review of Palaeobotany and Palynology* 34 (1), 129–135.
- Fischer, G., and Wefer, G. (Eds.) (2012). *Use of proxies in paleoceanography: examples from the South Atlantic*. Springer Science and Business Media. Pp. 1-735
- Fleet, A. J. (1984). Aqueous and sedimentary geochemistry of the rare earth elements. In: Henderson, P., (ed.) *Rare Earth Element Geochemistry*. Elsevier, Amsterdam, 343-73. <https://doi.org/10.1016/B978-0-444-42148-7.50015-0>.
- Floyd, P. A., and Leveridge, B. E. (1987). Tectonic environment of the Devonian Gramscatho basin, south Cornwall: framework mode and geochemical evidence from turbiditic sandstones. *Journal of the Geological Society*, 144(4), 531-542. <https://doi.org/10.1144/gsjgs.144.4.0531>.
- Fowell, S. J., and Olsen, P. E. (1993). Time calibration of Triassic/Jurassic microfloral turnover, eastern North America. *Tectonophysics*, 222(3-4), 361-369. [https://doi.org/10.1016/0040-1951\(93\)90359-R](https://doi.org/10.1016/0040-1951(93)90359-R).
- Fowell, S. J., Cornet, B., and Olsen, P. E. (1994). Geologically rapid Late Triassic extinctions: palynological evidence from the Newark Supergroup. *Pangaea: Paleoclimate, Tectonics and Sedimentation During Accretion, Zenith and Break-up of a Supercontinent*. Geological Society of America Special Paper, 288, 197-206.
- Galfetti, T., Bucher, H., Ovtcharova, M., Schaltegger, U., Brayard, A., Brühwiler, T., Goudemand, N., Weissert, H., Hochuli, P.A., Cordey, F. and Guodun, K., (2007). Timing of the Early Triassic carbon cycle perturbations inferred from new U–Pb ages and ammonoid biochronozones. *Earth and Planetary Science Letters*, 258(3-4), pp.593-604. <https://doi.org/10.1016/j.epsl.2007.04.023>.
- Galli, M. T., Jadoul, F., Bernasconi, S. M., and Weissert, H. (2005). Anomalies in global carbon cycling and extinction at the Triassic/Jurassic boundary: evidence from a marine C-isotope record. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 216(3), 203-214. <https://doi.org/10.1016/j.palaeo.2004.11.009>.

- Garzanti, E., and Andò, S. (2007). Plate tectonics and heavy mineral suites of modern sands. *Developments in Sedimentology*, 58, 741-763. [https://doi.org/10.1016/S0070-4571\(07\)58029-5](https://doi.org/10.1016/S0070-4571(07)58029-5).
- Garver, J. I., Scott, T. J., (1995). Trace elements in shale as indicators of crustal provenance and terrane accretion in the southern Canadian Cordillera. *GSA Bull.* 107 (4), 440–453. DOI: [https://doi.org/10.1130/0016-7606\(1995\)107<0440:TEISAI>2.3.CO;2](https://doi.org/10.1130/0016-7606(1995)107<0440:TEISAI>2.3.CO;2).
- Gee, E.R., (1945). The age of the saline series of the Punjab and Kohat: *India Natl. Acad. Sci., Proc, Sec. B*. Vol. 14, Pt. 6, pp. 269-310.
- Gee, E. R. (1980). Salt Range series geological maps: Directorate of Overseas Surveys. United Kingdom, for the Government of Pakistan and Pakistan Geological Survey, scale, 1(50,000), 6.
- Gee, E. R., and Gee, D. G. (1989). Overview of the geology and structure of the Salt Range, with observations on related areas of northern Pakistan. *Geological Society of America Special Papers*, 232, 95-112.
- Ghazi, S., and Mountney, N. P. (2009). Facies and architectural element analysis of a meandering fluvial succession: The Permian Warchha Sandstone, Salt Range, Pakistan. *Sedimentary Geology*, 221(1), 99-126. <https://doi.org/10.1016/j.sedgeo.2009.08.002>.
- Ghazi, S., and Mountney, N. P. (2011). Petrography and provenance of the Early Permian Fluvial Warchha Sandstone, Salt Range, Pakistan. *Sedimentary Geology*, 233(1), 88-110. <https://doi.org/10.1016/j.sedgeo.2010.10.013>.
- Goldberg, K., and Humayun, M. (2010). The applicability of the Chemical Index of Alteration as a paleoclimatic indicator: An example from the Permian of the Paraná Basin, Brazil. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 293(1), 175-183. <https://doi.org/10.1016/j.palaeo.2010.05.015>.
- Graham S. A., Ingersoll, R. V., and Dickinson, W. R., (1976). Common provenance for lithic grains in Carboniferous sandstones from Ouachita Mountains and Black Warrior basin: *Jour. Sed. Petrology*, v. 46, p. 620-632.
- Guex, J., Bartolini, A., Atudorei, V., Taylor, D., (2004). High-resolution ammonite and carbon isotope stratigraphy across the Triassic–Jurassic boundary at New York Canyon (Nevada). *Earth and Planetary Science Letters* 225 (1–2), 29–41. <https://doi.org/10.1016/j.epsl.2004.06.006>.

- Guex, J., Schoene, B., Bartolini, A., Spangenberg, J., Schaltegger, U., O'Dogherty, L., Taylor, D., Bucher, H., and Atudorei, V., (2012). Geochronological constraints on post-extinction recovery of the ammonoids and carbon cycle perturbations during the Early Jurassic: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 346-347, pp. 1-11. <https://doi.org/10.1016/j.palaeo.2012.04.030>.
- Hallam, A. (1997). Estimates of the amount and rate of sea-level change across the Rhaetian–Hettangian and Pliensbachian–Toarcian boundaries (latest Triassic to early Jurassic) Journal of the Geological Society, London, 154, 773–779. <https://doi.org/10.1144/gsjgs.154.5.0773>.
- Hallam, A., (2001). A review of the broad pattern of Jurassic sea-level changes and their possible causes in the light of current knowledge. Palaeogeography, Palaeoclimatology, Palaeoecology, 167(1), 23-37. [https://doi.org/10.1016/S0031-0182\(00\)00229-7](https://doi.org/10.1016/S0031-0182(00)00229-7).
- Hallam, A., and Maynard, J. B., (1987). The iron ores and associated sediments of the Chichali formation (Oxfordian to Valanginian) of the Trans-Indus Salt Range, Pakistan. Journal of the Geological Society, 144(1), 107-114. <https://doi.org/10.1144/gsjgs.144.1.0107>.
- Hallam, A., and Wignall, P.B. (1999). Mass extinctions and sea-level changes. Earth-Science Reviews, 48, 217–250. [https://doi.org/10.1016/S0012-8252\(99\)00055-0](https://doi.org/10.1016/S0012-8252(99)00055-0).
- Haq, B. U. (2017). Jurassic Sea-Level Variations: A Reappraisal. GSA Today, v. 28, no. 1, doi: 964 10.1130/GSATG359A.1.
- Haq, B.U., Hardenbol, J., and Vail, P.R., (1987). The chronology of fluctuating sea level since the Triassic: Science, v. 235, p. 1156–1167, <https://doi.org/10.1126/science.235.4793.1156>.
- Hegde, V. S., and Chavadi, V. C. (2009). Geochemistry of late Archaean metagreywackes from the Western Dharwar Craton, South India: implications for provenance and nature of the Late Archaean crust. Gondwana Research, 15(2), 178-187. <https://doi.org/10.1016/j.gr.2008.09.006>.
- Heron, A. M. (1953). The geology of central Rajputana. Mem. Geol. Soc. Ind., 79.
- Herron, M. M. (1988). Geochemical classification of terrigenous sands and shales from core or log data. Journal of Sedimentary Research, 58(5). 820-829. <https://doi.org/10.1306/212F8E77-2B24-11D7-8648000102C1865D>.

- Hesselbo, S. P., Robinson, S. A., Surlyk, F., and Piasecki, S., (2002). Terrestrial and marine extinction at the Triassic–Jurassic boundary synchronized with major carbon-cycle perturbation: a link to initiation of massive volcanism? *Geology* 30, 251–254. [https://doi.org/10.1130/0091-7613\(2002\)030<0251:TAMEAT>2.0.CO;2](https://doi.org/10.1130/0091-7613(2002)030<0251:TAMEAT>2.0.CO;2).
- Hesselbo, S. P., Robinson, S. A., and Surlyk, F. (2004). Sea-level change and facies development across potential Triassic–Jurassic boundary horizons, SW Britain. *Journal of the Geological Society*, 161(3), 365–379. <https://doi.org/10.1144/0016-764903-033>.
- Hillebrandt, A. V., Krystyn, L., (2009). On the oldest Jurassic ammonites of Europe (Northern Calcareous Alps, Austria) and their global significance. *Neues Jb. Geol. Paläontol. Abh.* 253, 163–195. DOI: <https://doi.org/10.1127/0077-7749/2009/0253-0163>.
- Hillebrandt, A. V., Krystyn, L., Kürschner, W. M., Bonis, N.R., Ruhl, M., Richoz, S., Schobben, M.A.N., Urlichs, M., Bown, P.R., Kment, K., McRoberts, C.A., Simms, M., Tomášových, A., (2013). The Global Stratotype Sections and Point (GSSP) for the Base of the Jurassic Systemat Kuhjoch (Karwendel Mountains, Northern Calcareous Alps, Tyrol, Austria). *Episodes*. 36 pp. 162–198.
- Hofmann, P., Ricken, W., Schwark, L., and Leythaeuser, D. (2001). Geochemical signature and related climatic-oceanographic processes for early Albian black shales: Site 417D, North Atlantic Ocean. *Cretaceous Research*, 22(2), 243–257. <https://doi.org/10.1006/cres.2001.0253>.
- Hofmann, M. H., Li, X. H., Chen, J., MacKenzie, L. A., and Hinman, N. W. (2016). Provenance and temporal constraints of the early Cambrian Maotianshan shale, Yunnan Province, China. *Gondwana Research*, 37, 348–361. <https://doi.org/10.1016/j.gr.2015.08.015>.
- Hubert, J. F., (1962). A zircon–tourmaline–rutile maturity index and the interdependence of the composition of heavy mineral assemblages with the gross composition and texture of sandstones. *J. Sediment. Res.* 32, 440–450.
- Ingersoll, R. V., Bullard, T. F., Ford, R. L., Grimm, J. P., Pickle, J. D., and Sares, S. W., (1984). The effect of grain size on detrital modes: a test of the Gazzi-Dickinson point-counting method. *Journal of Sedimentary Research*. 54(1), 103–116. <https://doi.org/10.1306/212F83B9-2B24-11D7-8648000102C1865D>.

- Iqbal, S., Jan, U. I., and Hanif, M., (2014a). The Mianwali and Tredian Formations: an example of the Triassic Progradational Deltaic System in the low-latitude western Salt Range, Pakistan. *Arab J Sci Eng* 39(7):5489–5507. DOI: <https://doi.org/10.1007/s13369-013-0836-2>.
- Iqbal, S., Jan, U. I., Akhter, G., Wagreich, M., and Hanif, M., (2014b). Triassic–Jurassic boundary: evidences from the Tethyan Salt Range, Pakistan and correlation with Europe. *J Himal Earth Sci* (abstract volume), 59
- Iqbal, S., Akhter, G., and Bibi, S. (2015a). Structural model of the Balkassar area, Potwar Plateau, Pakistan. *International Journal of Earth Sciences*, 104(8), 2253–2272. DOI: <https://doi.org/10.1007/s00531-015-1180-4>.
- Iqbal, S., Jan, I. U., Akhter, M. G., and Bibi, M. (2015b). Palaeoenvironmental and sequence stratigraphic analyses of the Jurassic Datta Formation, Salt Range, Pakistan. *Journal of Earth System Science*, 124(4), 747–766. DOI: <https://doi.org/10.1007/s12040-015-0572-y>
- Iqbal, S., Wagreich, M., Jan, I. U., Kuerschner, W. M., and Gier, S., (2018 under review) A southern hemisphere record of climate change during the Triassic–Jurassic transition: evidence for hothouse climate from the Salt Range, Pakistan (GLOPLACHA_2018_92).
- Jafarzadeh, M., Harami, R. M., Friis, H., Amini, A., Mahboubi, A., and Lenaz, D. (2014). Provenance of the Oligocene–Miocene Zivah Formation, NW Iran, assessed using heavy mineral assemblage and detrital clinopyroxene and detrital apatite analyses. *Journal of African Earth Sciences*, 89, 56–71. <http://dx.doi.org/10.1016/j.jafrearsci.2013.10.005>.
- Jan, I. U., and Stephenson, M. H. (2011). Palynology and correlation of the Upper Pennsylvanian Tobra Formation from Zaluch Nala, Salt Range, Pakistan. *Palynology*, 35(2), 212–225. <https://doi.org/10.1080/gspalynol.35.2.212>.
- Jan, I. U., Stephenson, M. H., and Khan, F. R. (2009). Palynostratigraphic correlation of the Sardhai Formation (Permian) of Pakistan. *Review of Palaeobotany and Palynology*, 158(1), 72–82. <https://doi.org/10.1016/j.revpalbo.2009.08.002>.
- Jan, I. U., Shah, A., Stephenson, M. H., Iqbal, S., Hanif, M., Wagreich, M., and Hussain, H. S. (2016). The sedimentology of the Lower Permian Dandot Formation: a component of the Gondwana deglaciation sequence of the Salt Range, Pakistan. *Rivista Italiana di*

- Paleontologia e stratgrafia, 122(1), 75-90. DOI: <https://doi.org/10.13130/2039-4942/6948>.
- Jan, I. U., Iqbal, S., Davies, S. J., Zalasiewicz, J. A., Stephenson, M. H., Wagreich, Haneef, M., Hanif, M. and Ahmad, S. (2017). A Periglacial Palaeoenvironment in the Upper Carboniferous–Lower Permian Tobra Formation of the Salt Range, Pakistan. *Acta Geologica Sinica (English Edition)*, 91(3), 1063-1078. <https://doi.org/10.1111/1755-6724.13324>.
- Jadoon, I. A. K., Kemal, A., Frisch, W., and Jaswal, T. M, (1997). Thrust geometries and kinematics in the Himalayan foreland (North Potwar Deformed Zone), North Pakistan. *Geolog Rundsch* 86:120–131
- Jadoon, I. A. K, Hinderer, M., Wazir, B., Rizwan, Y. R., Bahadar, S., Hassan, M., Abbasi, Z. H., and Jadoon, S. (2015). Structural styles, hydrocarbon prospects, and potential in the Salt Range and Potwar Plateau, north Pakistan. *Arab J Geosci*. <https://doi.org/10.1007/s12517-014-1566-9>.
- Jaraula, C. M. B., Grice, K., Twitchett, R. J., Bottcher, M. E., LeMetayer, P., Dastidar, A. G., and Opazo, L. F., (2013). Elevated pCO₂ leading to Late Triassic extinction, persistent photic zone euxinia, and rising sea levels. *Geology* 41, 955 -958. <https://doi.org/10.1130/G34183.1>.
- Jehangiri, M., Hanif, M., Arif, M., Jan, I. U., and Ahmad, S., (2015). The Early Cambrian Khewra Sandstone, Salt Range, Pakistan: endorsing southern Indian provenance. *Arabian Journal of Geosciences*, 8(8), 6169-6187. DOI: <https://doi.org/10.1007/s12517-014-1649-7>.
- John, T., Schenk, V., Haase, K., Scherer, E., and Tembo, F. (2003). Evidence for a Neoproterozoic ocean in south-central Africa from mid-oceanic-ridge–type geochemical signatures and pressure-temperature estimates of Zambian eclogites. *Geology*, 31(3), 243-246. [https://doi.org/10.1130/0091-7613\(2003\)031<0243:EFANOI>2.0.CO;2](https://doi.org/10.1130/0091-7613(2003)031<0243:EFANOI>2.0.CO;2).
- Kadri, I B., (1995). Petroleum geology of Pakistan. Pakistan Petroleum Limited, Karachi Pak, pp 40–61.
- Kazmi, A. H., Jan, M. Q., (1997). Geology and tectonic of Pakistan. Graphic publication, Karachi, pp 130–141.

- Kemal, A., (1992). Geology and new trends for hydrocarbon exploration in Pakistan, in Ahmed, G., Kemal, A., Zaman, A.S.H., and Humayon, M., eds., New directions and strategies for accelerating petroleum exploration and production in Pakistan: Proceedings, international petroleum seminar, November, 22–24, 1991: Islamabad, Pakistan, Ministry of Petroleum and Natural Resources, p. 16–57.
- Khan, M. A., and Bhutta, A. A., (1983). *Concavigleichenioidites* a new Microspore genus from Datta Formation, (Jurassic), Western Salt Range, Pakistan. *Kashmir Jour. Of Geol.* Vol. 1. Pp 83-87.
- Khan, M. A., and Bhutta, A. A., (1985). On some new Early Jurassic microspores from Datta Formation Western Salt Range, Pakistan. *Kashmir Jour. Of Geol.* Vol. 5.
- Khan, M. A., and Bhutta, A. A., (1987). Comments on some new Early Jurassic Miospores from Datta Formation; Western Salt Range Pakistan. *Kashmir Jour. Of Geol.* Vol. 5. pp 153-155.
- Koroznikova, L., Klutke, C., McKnight, S., and Hall, S. (2008). The use of low-toxic heavy suspensions in mineral sands evaluation and zircon fractionation. *Journal of the Southern African Institute of Mining and Metallurgy*, 108(1), 25-33.
- Kürschner, W. M., Bonis, N. R., and Krystyn, L., (2007). Carbon-isotope stratigraphy and palynostratigraphy of the Triassic–Jurassic transition in the Tiefengraben section – Northern Calcareous Alps (Austria). *Palaeogeography, Palaeoclimatology, Palaeoecology* 244 (1–4), 257–280. <https://doi.org/10.1016/j.palaeo.2006.06.031>.
- Kuerschner, W. M., Iqbal, S., Wagreich, M., Jan, I., and Gier, S. (2017). Palynology and Carbon isotope stratigraphy of the Triassic-Jurassic transition in the Salt Range (Pakistan). In *EGU General Assembly Conference Abstracts* (Vol. 19, p. 15291).
- Latif, M. A., (1970). Micropalaeontology of the Chanali limestones. Upper Cretaceous of Hazara, West Pakistan. *Ibid.*, Sonderb. 15, pp. 25-61.
- Lindström, S., van de Schootbrugge, B., Hansen, K. H., Pedersen, G. K., Alsen, P., Thibault, N., Dybkjær, K., Bjerrum, C. J., and Nielsen, L. H., (2017). A new correlation of Triassic–Jurassic boundary successions in NW Europe, Nevada and Peru, and the Central Atlantic Magmatic Province: A time-line for the end-Triassic mass extinction. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 478, 80-102. <https://doi.org/10.1016/j.palaeo.2016.12.025>.

- Malick, B. M. L., and Ishiga, H. (2016). Geochemical Classification and Determination of Maturity Source Weathering in Beach Sands of Eastern San'in Coast, Tango Peninsula, and Wakasa Bay, Japan. *Earth Science Research*, 5(1), 44. DOI: <http://dx.doi.org/10.5539/esr.v5n1p44>.
- Mange, M. A. and Maurer, H. F. W. (1992). *Heavy minerals in colour*. London: Chapman and Hall. 1-147.
- Mange, M.A., Wright, D.T. (Eds.), (2007). *Heavy minerals in use. : Developments in Sedimentology*, 58. Elsevier, Amsterdam. 1283pp.
- Martin, A. J. (2017). A review of Himalayan stratigraphy, magmatism, and structure. *Gondwana Research*. 49 (42-80) <http://dx.doi.org/10.1016/j.gr.2017.04.031>.
- Marzoli, A., Renne, P. R., Piccirillo, E. M., Ernesto, A., Bellieni, G., and De Min, A., (1999). Extensive 200-million-year-old continental flood basalts of the Central Atlantic magmatic province. *Science* 284, 616– 618. DOI: 10.1126/science.284.5414.616.
- Marzoli, A., Jourdan, F., Puffer, J. H., Cuppone, T., Tanner, L. H., Weems, R. E., Bertrand, H., Cirilli, S., Bellieni, G., De Min, A., (2011). Timing and duration of the Central Atlantic magmatic province in the Newark and Culpeper basins, eastern U.S.A. *Lithos* 122, 175–188. <https://doi.org/10.1016/j.lithos.2010.12.013>.
- McGhee, G. R., Sheehan, P. M., Bottjer, D. J., Droser, M. L., (2004). Ecological ranking of Phanerozoic biodiversity crises: ecological and taxonomic severities are decoupled. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 211, 289–297. <https://doi.org/10.1016/j.palaeo.2004.05.010>.
- McGhee, G. R., Clapham, M. E., Sheehan, P. M., Bottjer, D. J., Droser, M. L., (2013). A new ecological- severity ranking of major Phanerozoic biodiversity crises. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 370, 260–270. <https://doi.org/10.1016/j.palaeo.2012.12.019>
- McLennan, S. M., (1989a). Rare earth elements in sedimentary rocks: Influence of provenance and sedimentary processes: *Mineralogical Society of America Reviews in Mineralogy*, v. 21, p. 169-200.
- McLennan, S. M., (1989b). Archean sedimentary rocks and the Archean mantle, in Aswal, L. D., ed., *Workshop on the Archean Mantle: LPI Technical Report 89-05*, p. 57-59. Lunar and Planetary Institute, Houston.

- McLennan, S. M. (1993). Weathering and global denudation. *The Journal of Geology*, 101(2), 295-303. <https://doi.org/10.1086/648222>.
- McLennan, S. M. (2003). Sedimentary silica on Mars. *Geology*, 31(4), 315-318. DOI: [https://doi.org/10.1130/0091-7613\(2003\)031<0315:SSOM>2.0.CO;2](https://doi.org/10.1130/0091-7613(2003)031<0315:SSOM>2.0.CO;2).
- McLennan, S. M., and Taylor, S. R. (1980). Th and U in sedimentary rocks: crustal evolution and sedimentary recycling. *Nature*, 285(5767), 621. DOI: <https://doi.org/10.1038/285621a0>.
- McLennan, S. M., and Taylor, S. R., (1991). Sedimentary rocks and crustal evolution: Tectonic setting and secular trends: *Journal of Geology*, v. 99, p. 1-21. <https://doi.org/10.1086/629470>.
- McLennan, S. M., Hemming, S., McDaniel, D. K., Hanson, G. M., (1993). Geochemical approaches to sedimentation, provenance, and tectonics. In: Johnsson, M.J., Basu, A. (Eds.), *Processes Controlling the Composition of Clastic Sediments*. Geological Society of America, Special Papers, vol. 284, pp. 21-40. DOI: 10.1130/SPE284-p21.
- Mertmann, D., and Ahmad, S., (1994). Shinawari and Samana Suk Formations of the Surghar and salt ranges, Pakistan: facies and depositional environments. *Zeitschrift der Deutschen Geologischen Gesellschaft*, 305-317.
- Mertmann, D. (2003). Evolution of the marine Permian carbonate platform in the Salt Range (Pakistan). *Palaeogeography, Palaeoclimatology, Palaeoecology*, 191(3), 373-384. [https://doi.org/10.1016/S0031-0182\(02\)00672-7](https://doi.org/10.1016/S0031-0182(02)00672-7).
- Middlemiss, C. S., (1896). The geology of Hazara and the Black Mountains. *Geol. Surv., India, Mem.*, Vol. 26, 302 p.
- Mikutta, R., Kleber, M., Kaiser, K., and Jahn, R. (2005). Review. *Soil science society of America journal*, 69(1), 120-135. DOI: 10.2136/sssaj2005.0120
- Moore, D. M. and Reynolds, R. C., Jr. (1997). *X-ray Diffraction and the Identification and Analysis of Clay Minerals*. Oxford University Press, Oxford, 378 pp.
- Morton, N., (2012). Inauguration of the GSSP for the Jurassic System. *Episodes* 35, 328–332.
- Mujtaba, M., and Memon, A. M., (1990). Diagenesis and pore space development of Datta Sandstones in Salt Range and Surghar Range. *Pak. Jour. Of Hyd. Research*. Vol. 2, Pp.57-66.
- Mukhopadhyay, G., Mukhopadhyay, S. K., Roychowdhury, M., and Parui, P. K. (2010). Stratigraphic correlation between different Gondwana basins of India. *Journal of the*

- Geological Society of India, 76(3), 251-266. DOI: <https://doi.org/10.1007/s12594-010-0097-6>.
- Nance, W. B. and Taylors, R. (1976). Rare earth patterns and crustal evolution, I: Australian post-Archean sedimentary rocks. *Geochimica et Cosmochimica Acta*, 40, 1539-51.
- Nechaev, V. P., Isphording, W. P., (1993). Heavy mineral assemblages of continental margin as indicators of plate tectonic environments. *Journal of Sedimentary Petrology* 63, 1110–1117. <https://doi.org/10.1306/D4267CB7-2B26-11D7-8648000102C1865D>.
- Nesbitt, H. W., and Young, G. M., (1982.) Early Proterozoic climates and plate motions inferred from major element chemistry of lutites. *Nature* 299, 715-717. DOI: <https://doi.org/10.1038/299715a0>.
- Nesbitt, H. W. and Young, G. M., (1984). Prediction of some weathering trends of plutonic and volcanic rocks based on thermodynamic and kinetic considerations. *Geochim. Cosmochim. Acta*, 48: 1523-1534. [https://doi.org/10.1016/0016-7037\(84\)90408-3](https://doi.org/10.1016/0016-7037(84)90408-3).
- Nesbitt, H. W., Young, G. M., (1989). Formation and diagenesis of weathering profiles. *Journal of Geology* 97, 129e-147. DOI: <https://doi.org/10.1086/629290>.
- Newman, S, Macdougall, J. D and Finkel, R. C., (1984). ^{230}Th - ^{238}U disequilibrium in island arcs: Evidence from the Aleutians and the Marianas: *Nature*, v. 308, p. 268-270.
- Obaid, S., Qureshi, W. A., and Abbasi, I. A., (2005). Lithofacies, sand-bodies geometry and depositional setting of the Datta Formation in Surghar Range, North Pakistan. SPE/PAPG Technical Conference Islamabad. Pp 113-126.
- Olsen, P. E., (1999). Giant lava flows, mass extinctions, and mantle plumes. *Science*, 604–605. DOI: 10.1126/science.284.5414.604.
- Olsen, P. E., Koeberl, C., Huber, H., Montanari, A., Fowell, S. J., Et-Touhami, M., and Kent, D. V. (2002). Continental Triassic-Jurassic boundary in central Pangea: Recent progress and discussion of an Ir anomaly. *special papers-geological society of America*, 505-522.
- Pálffy, J., Demény, A., Haas, J., Hetényi, M., Orchard, M.J., and Vetl, I., (2001). Carbon isotope anomaly and other geochemical changes at the Triassic–Jurassic boundary from a marine section in Hungary. *Geology* 29, 1047–1050. [https://doi.org/10.1130/0091-7613\(2001\)029<1047:CIAAOG>2.0.CO;2](https://doi.org/10.1130/0091-7613(2001)029<1047:CIAAOG>2.0.CO;2).
- Pascoe, E. H., (1959). *Manual of Geol. India and Burma Vol. II* India Govt, press Calcutta. 484-1338.

- Pettijohn, F. J., (1974). *Sedimentary Rocks*; Harper and Row Publ. New York, Evanston, San Francisco, and London, pp. 195–246, 316–380.
- Pettijohn, F. J., Potter, P. E., and Siever, R., (1973). *Sand and sandstone*, Springer-Verlag, Berlin, 617p.
- Pettijohn, F. J., Potter, P. E., and Siever, R. (1987). *Sand and Sandstone*. 553 pp.
- Pillecuit, A., (1993). Les Blocs exotiques du Sultanat d'Oman, évolution paléogéographique d'une marge passive flexurale. *Mémoires de Géologie*, Lausanne, 17: 1-249.
- Raup, D. M., and Sepkoski, J. J. (1982). Mass extinctions in the marine fossil record. *Science*, 215(4539), 1501-1503.
- Richoz, S., Van De Schootbrugge, B., Pross, J., Püttmann, W., Quan, T. M., Lindström, S., Heunisch, C., Fiebig, J., Maquil, R., Schouten, S. and Hauzenberger, C. A., (2012). Hydrogen sulphide poisoning of shallow seas following the end-Triassic extinction. *Nature Geoscience*, 5(9), 662-667. <https://doi.org/10.1038/ngeo1539>.
- Roser, B. P., and Korsch, R. J. (1986). Determination of tectonic setting of sandstone-mudstone suites using content and ratio. *The Journal of Geology*, 94(5), 635-650. <https://doi.org/10.1086/629071>.
- Roser, B. P., and Korsch, R. J., (1988). Provenance signatures of sandstone-mudstone suites determined using discriminant function analysis of major-element data. *Chemical geology*, 67(1-2), 119-139. DOI: [https://doi.org/10.1016/0009-2541\(88\)90010-1](https://doi.org/10.1016/0009-2541(88)90010-1).
- Rafique, M., Ashraf, M., Din, A., and Faruqi, F. A., (1976). Evaluation of glass sand deposits of Pzeu, Bannu Distt. N.W.F.P. *Geol. Bull. Punjab. Univ.* 45, pp 87-92.
- Ruhl, M., Bonis, N. R., Reichart, G. J., Damsté, J. S. S., and Kürschner, W. M., (2011). Atmospheric carbon injection linked to end-Triassic mass extinction. *Science* 333 (6041), 430–434. <http://dx.doi.org/10.1126/science.1204255>.
- Ruhl, M., Kürschner, W. M., Krystyn, L., (2009). Triassic–Jurassic organic carbon isotope stratigraphy of key sections in the western Tethys realm (Austria). *Earth Planet. Sci. Lett.* 281, 169–187. doi:10.1016/j.epsl.2009.02.020
- Schoene, B., Guex, J., Bartolini, A., Schaltegger, U., Blackburn, T. J., (2010). Correlating the end-Triassic mass extinction and flood basalt volcanism at the 100 ka level. *Geology* 38, 387–390. <https://doi.org/10.1130/G30683.1>.
- Schultz, L. G. (1964). Quantitative interpretation of mineralogical composition from X-ray and chemical data for the Pierre Shale (No. 391-C). 1-31

- Selley, C. R., (2000). *Applied Sedimentology*: 2nd Ed. Academic Press. 182-207.
- Sepkoski, J. J., (1996). Patterns of Phanerozoic extinction: a perspective from global data bases. In: Walliser, O.H. (Ed.), *Global Events and Event Stratigraphy*. Springer-Verlag, Berlin, pp. 35–51. https://doi.org/10.1007/978-3-642-79634-0_4.
- Sevastjanova, I., Hall, R., and Alderton, D. (2012). A detrital heavy mineral viewpoint on sediment provenance and tropical weathering in SE Asia. *Sedimentary Geology*, 280, 179-194. <https://doi.org/10.1016/j.sedgeo.2012.03.007>.
- Shah, S. M. I. (Ed.). (1977). *Stratigraphy of Pakistan* (Vol. 12). Director General, Geological Survey of Pakistan. 41-56.
- Shah, S. M. I., (1980). *Stratigraphy and economic geology of central Salt Range*. Geol. Surv. Pakistan, Rec., Vol. 52, 104 p.
- Shah, S. M. I., (2009). *Stratigraphy of Pakistan*. Gov Pak Ministry of Petrol Nat Resour Geol Sur Pak 245–273.
- Sisodia, M. S., and Chauhan, D. S. (1998). Indian Precambrian, ed. Paliwal, BS. 171-182.
- Sorkhabi, R., (2014). How Much Oil in the Middle East?.
- Stampfli, G. M., Marcoux, J. and Baud, A., (1991). Tethyan margins in space and time. In: J.E.T. Channel, E.L. Winterer and L.F. Jansa (eds), *Paleogeography and Paleooceanography of Tethys*. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 87: 373-410. DOI: [https://doi.org/10.1016/0031-0182\(91\)90142-E](https://doi.org/10.1016/0031-0182(91)90142-E).
- Stampfli, G. M., Mosar, J., Favre, P., Pillecuit, A. and Vannay, J-C. (2001). Permo-Mesozoic evolution of the western Tethys realm: the Neo-Tethys East Mediterranean Basin connection. In: P.A. Ziegler, P. A., Cavalza, W., Robertson, A. H. F., and Crasquin-Soleau, S., (eds), *Peri-Tethys Memoir 6: Peri-Tethyan Rift/Wrench Basins and Passive Margins*. *Mém. Mus. natn. Hist. nat.*, 186 : 51-108. Paris ISBN: 2-85653-528-3.
- Stern, G., and Wagreich, M. (2013). Provenance of the Upper Cretaceous to Eocene Gosau Group around and beneath the Vienna Basin (Austria and Slovakia). *Swiss Journal of Geosciences*, 106(3), 505-527. DOI: <https://doi.org/10.1007/s00015-013-0150-8>.
- Suttner, L. J., and Dutta, P. K. (1986). Alluvial sandstone composition and paleoclimate, I. Framework mineralogy. *Journal of Sedimentary Research*, 56(3) 329-345. <https://doi.org/10.1306/212F8909-2B24-11D7-8648000102C1865D>.
- Taylor, R., Ewart, A. and Capp, A. C. (1968). Leucogranites and rhyolites: trace element evidence for fractional crystallization and partial melting. *Lithos*, 1, 179-86.

- Taylor, S. R. and McLennan, S. M., (1985). *The Continental Crust: Its Composition and Evolution*. Blackwell, Oxford, ISBN-13: 978-0632011483, pp: 312.
- Tortosa, A., Palomares, M., and Arribas, J. (1991). Quartz grain types in Holocene deposits from the Spanish Central System: some problems in provenance analysis. *Geological Society, London, Special Publications*, 57(1), 47-54. <https://doi.org/10.1144/GSL.SP.1991.057.01.05>.
- Van de Schootbrugge, B., Quan, T.M., Lindström, S., Püttmann, W., Heunisch, C., Pross, J., Fiebig, J., Petschick, R., Röhling, H.G., Richoz, S. and Rosenthal, Y., (2009). Floral changes across the Triassic/Jurassic boundary linked to flood basalt volcanism. *Nature Geoscience*, 2(8), 589-594. DOI: <https://doi.org/10.1038/ngeo577>.
- Van Langeveld, A. D., Van der Gaast, S. J., and Eisma, D. (1978). Comparison of the effectiveness of eight methods for the removal of organic matter from clay. *Clays and Clay Minerals*, 26(5), 361-4.
- Vincent, H., Wach, G., and Ketannah, Y. (2014). Heavy mineral record of Andean uplift and changing sediment sources across the NE margin of South America: a case study from Trinidad and Barbados. *Geological Society, London, Special Publications*, 386(1), 217-241. <http://dx.doi.org/10.1144/SP386.19>.
- Virdi, N. S., (1998). Coexisting Late Proterozoic glaciogene sediments and evaporites in the Lesser Himalaya and western Indian Shield-Expression of contemporaneity of low latitude glaciation and tropical desiccation. In: Paliwal, B.S. (Ed.), *The Indian Precambrian*: Sci. Pub. Ind. Jodhpur, pp. 502–511.
- Wadia, D. N. (1966). *Geology of India (rev.)*. MacMillan Company Ltd., London, *Geological Magazine*, 1(1939), 76.
- Ward, P.D., Garrison, G.H., Williford, K.H., Kring, D.A., Goodwin, D., Beattie, M.J., McRoberts, C.A., (2007). The organic carbon isotopic and paleontological record across the Triassic–Jurassic boundary at the candidate GSSP section at Ferguson Hill, Muller Canyon, Nevada, USA. *Palaeogeography, Palaeoclimatology, Palaeoecology* 244 (1–4), 281–289. <https://doi.org/10.1016/j.palaeo.2006.06.042>.
- Wefer, G., Berger, W. H., Bijma, J., and Fischer, G. (1999). Clues to ocean history: a brief overview of proxies. In *Use of proxies in paleoceanography* (pp. 1-68) Springer Berlin Heidelberg. DOI: https://doi.org/10.1007/978-3-642-58646-0_1.

- Wignall, P. B., (2001). Large igneous provinces and mass extinctions. *Earth-Sci. Rev.* 53, 1 – 33. [https://doi.org/10.1016/S0012-8252\(00\)00037-4](https://doi.org/10.1016/S0012-8252(00)00037-4).
- Wopfner, H., (1993). Pange an depositional sequences between Karoo and Lut. In: U. Thorweihe and H. Schandelmeier (eds), *Geoscientific Research in northeast Africa*. A.A. Balkema, Rotterdam; Brookfield: 189-195.
- Wotzlaw, J.-F., Guex, J., Bartolini, A., Gallet, Y., Krystyn, L., McRoberts, C.A., Taylor, D., Schoene, B., Schaltegger, U., (2014). Towards accurate numerical calibration of the late Triassic: high-precision U-Pb geochronology constraints on the duration of the Rhaetian. *Geology* 42, 571–574. <https://doi.org/10.1130/G35612.1>.
- Wynne, A. B., (1978). On the geology of the Salt Range in the Punjab. *Ibid., Mem., Vol. 14*, 313p.
- Wynne, A. B., (1879). Further notes on the geology of the upper Punjab. *Ibid., Rec., Vol. 12, Pt.2*, pp. 114-133.
- Zachos, J., Pagani, M., Sloan, L., Thomas, E., and Billups, K. (2001). Trends, rhythms, and aberrations in global climate 65 Ma to present. *Science*, 292(5517), 686-693. DOI: 10.1126/science.1059412.

APPENDICES

Appendix 1. Point counting data of the framework composition of the selected sandstones samples. Note: some samples have been averaged to 500 counts to keep the data uniform. monocrystalline = Q_m ; polycrystalline = Q_p ; monocrystalline quartz with unit extinction = Q_{mue} ; monocrystalline quartz with undulose extinction = Q_{muu} ; polycrystalline quartz with 2 – 3 crystals = $Q_{pq(2-3)}$; polycrystalline quartz with >3 crystals = $Q_{pq>3}$; total quartz = Q_{total} ; plagioclase = P; K-feldspar = A; total feldspar = F_{total} ; volcanic and hypabyssal lithics = L_v ; metamorphic lithics = L_m ; sedimentary lithics excluding chert = L_s ; total lithics = L_{total} . Phyllosilicates = M, ferromagnesian minerals = D, others/unidentifiable = O matrix/cement = M/C.

Appendix 2(a, b). a) Bulk mineralogical composition for the representative samples using XRD. The common mineral peaks have been labeled. b) XRD mineralogical composition for the oriented clay samples with major peaks being labeled.

Appendix 3. Bulk sediments geochemistry of the representative samples of the Kingriali and Datta formations.

Appendix 4. Point counting data of the heavy minerals of the selected samples.

Appendix 1.

Sample	Q _{mue}	Q _{muu}	Q _{pq (2-3)}	Q _{pq>3}	Q _{total}	A	P	F _{total}	L _v	L _m	L _s	L _{total}	M	D	O	M/C	Total
DU1	378	42	23	7	450	0	0	0	0	0	0	0	8	42	0	22	500
DU2	327	58	67	25	477	0	0	0	0	0	0	0	2	21	0	30	500
DU3	332	113	32	17	494	0	0	0	5	0	0	5	0	1	0	10	500
DU4	407	75	13	2	497	0	0	0	0	0	0	0	0	3	0	3	500
DU5	347	96	26	17	486	0	2	2	2	3	5	10	0	2	0	45	500
DU6	310	78	60	38	486	0	2	2	0	4	0	4	2	6	0	84	500
DU7	310	74	74	36	494	0	0	0	0	0	0	0	0	2	4	26	500
DU8	324	74	52	40	490	0	0	0	0	2	4	6	0	0	4	44	500
KF 1	444	32	16	4	496	0	0	0	0	0	0	0	0	0	4	100	500
KF 2	416	52	22	6	496	0	0	0	0	0	0	0	0	2	2	114	500
KF 3	424	64	8	2	498	0	0	0	0	0	0	0	0	2	0	182	500
KF 4	428	28	24	14	494	0	0	0	0	0	0	0	0	6	0	116	500
KF 5	386	46	52	12	496	0	0	0	0	0	0	0	2	2	0	206	500
KF 7	388	82	20	8	498	0	0	0	0	0	2	2	0	0	0	196	500
KF 10	444	34	14	6	498	0	0	0	0	0	0	0	0	0	2	160	500
KF 12	452	30	10	4	496	0	0	0	0	0	0	0	0	4	0	80	500
KF 13	446	28	18	0	492	0	0	0	0	0	0	0	2	6	0	166	500
KF 14	314	20	8	8	350	0	0	0	0	0	104	104	24	14	8	464	500
DK-a	196	86	36	110	428	4	0	4	10	36	18	64	0	4	0	254	500
DK-b	394	72	28	4	498	0	0	0	0	0	2	2	0	0	0	138	500
DK-c	394	68	20	6	488	2	0	2	2	0	2	4	0	4	2	160	500
DK-d	430	36	12	2	480	0	0	0	0	0	0	0	4	10	6	134	500
DK-17	238	90	74	70	472	0	0	0	4	14	4	22	0	0	6	180	500
DK-19	378	90	18	6	492	0	0	0	2	2	2	6	0	0	2	120	500
DK-21	402	70	8	14	494	2	0	2	0	4	0	4	0	0	0	110	500
DK-22	450	30	12	4	496	0	0	0	0	0	0	0	0	4	0	78	500
DK-24	452	20	4	0	476	0	0	0	0	0	0	0	2	18	4	146	500
DK-25	460	18	8	0	486	0	0	0	0	0	0	0	0	14	0	198	500
DK-27	430	48	8	6	492	4	0	4	0	2	0	2	0	2	0	126	500
DK-28	438	44	12	2	496	0	0	0	0	0	0	0	0	4	0	84	500
DK-37	434	30	10	4	478	0	0	0	0	0	0	0	4	16	2	34	500
DK-42	384	100	6	4	494	2	0	2	0	0	0	0	0	4	0	288	500
DK-54	458	18	8	6	490	0	0	0	0	0	0	0	2	8	0	100	500
DK-55	428	66	4	0	498	0	0	0	0	0	0	0	0	2	0	102	500
DK-56	438	42	14	0	494	2	0	2	0	2	0	2	0	2	0	84	500
DK-57	422	62	12	0	496	2	0	2	0	0	0	0	0	2	0	192	500
DK-58	458	32	8	2	500	0	0	0	0	0	0	0	0	0	0	238	500
DK-62	478	2	0	0	480	0	0	0	0	0	0	0	0	16	4	114	500
DN-1	152	165	53	90	460	3	0	3	2	20	12	34	0	3	0	340	500
DN-2	400	74	22	2	498	0	0	0	0	0	0	0	0	2	0	196	500
DN-3	330	92	42	8	472	2	0	2	0	0	0	0	0	24	2	70	500
DN-5	225	152	43	35	455	10	0	10	5	10	15	30	0	3	2	530	500
DN-6	302	118	22	43	485	3	0	3	0	2	8	10	2	0	0	75	500
DN-7	195	132	36	123	486	0	0	0	5	2	5	12	0	2	0	408	500
DN-9	424	34	10	6	474	0	0	0	0	0	0	0	6	16	4	288	500
DN-13	338	100	40	5	483	0	0	0	3	0	2	5	0	12	0	164	500

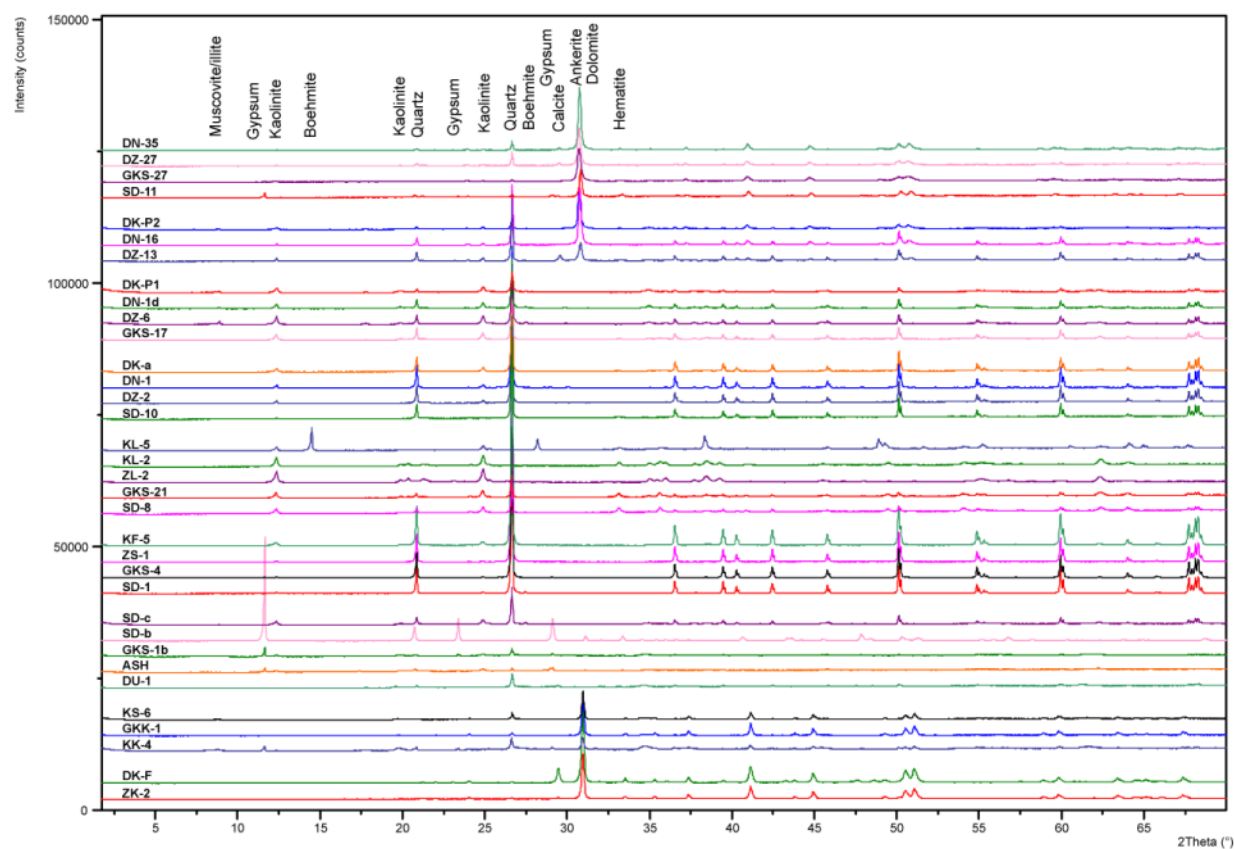
Appendix 1.

Sample	Q _{mue}	Q _{muu}	Q _{pq (2-3)}	Q _{pq>3}	Q _{total}	A	P	F _{total}	L _v	L _m	L _s	L _{total}	M	D	O	M/C	Total
DN-14	406	48	20	4	478	0	0	0	0	0	0	0	8	12	2	440	500
DN-16	375	102	13	5	495	0	0	0	0	5	0	5	0	0	0	715	500
DN-19	458	12	8	0	478	0	0	0	0	0	0	0	7	10	5	250	500
DN-21	400	30	28	6	464	0	0	0	4	0	0	4	14	18	0	74	500
DN-25	366	63	24	6	459	4	0	4	8	4	4	16	0	14	7	101	500
DN-26	428	44	18	2	492	2	0	2	4	0	0	4	0	2	0	52	500
DN-27	426	52	17	2	497	0	0	0	0	0	3	3	0	0	0	339	500
DN-28	336	98	42	6	482	4	0	4	2	0	2	4	0	2	8	596	500
DN-29	347	123	12	2	484	0	0	0	7	0	9	16	0	0	0	612	500
DN-31	476	16	4	2	498	0	0	0	0	0	0	0	0	2	0	102	500
DN-32	318	154	18	8	498	0	0	0	0	2	0	2	0	0	0	386	500
DN-34	334	82	12	6	434	0	0	0	2	0	0	2	4	0	60	478	500
DN-38	460	24	8	4	496	0	0	0	0	0	0	0	0	4	0	120	500
DN-39	376	89	27	4	496	0	0	0	2	0	2	4	0	0	0	182	500
DN-40	440	24	16	4	484	0	0	0	2	0	6	8	0	0	8	82	500
DN-41	366	120	10	2	498	0	0	0	2	0	0	2	0	0	0	200	500
DN-42	384	88	14	10	496	0	0	0	2	0	0	2	0	0	2	84	500
DN-43	432	54	6	0	492	0	0	0	2	0	0	2	0	4	2	262	500
ZS-1	454	28	16	2	500	0	0	0	0	0	0	0	0	0	0	142	500
ZS-2	459	14	20	2	495	0	0	0	2	0	0	2	0	3	0	136	500
ZS-3	458	26	10	4	498	0	0	0	0	0	0	0	0	2	0	46	500
ZS-4	454	34	2	2	492	0	0	0	0	0	0	0	0	2	6	28	500
ZS-5	460	32	4	4	500	0	0	0	0	0	0	0	0	0	0	16	500
ZS-6	482	12	4	2	500	0	0	0	0	0	0	0	0	0	0	72	500
ZS-8	486	4	0	2	492	0	0	0	0	0	0	0	0	6	2	34	500
ZS-9	486	4	0	0	490	0	0	0	0	0	0	0	0	2	8	36	500
ZS-11	478	12	10	0	500	0	0	0	0	0	0	0	0	0	0	24	500
ZS-12	482	8	4	2	496	0	0	0	0	0	0	0	0	4	0	22	500
ZS-13	456	34	4	2	496	0	0	0	0	0	0	0	0	0	4	26	500
DZ-2	472	22	2	0	496	0	0	0	0	0	0	0	0	0	4	62	500
DZ-12	445	30	20	0	495	0	0	0	0	0	0	0	0	0	5	180	500
DZ-13	442	14	12	4	472	0	0	0	0	0	0	0	0	0	28	240	500
DZ-15	465	25	5	0	495	0	0	0	0	0	0	0	5	0	0	190	500
DZ-18a	480	5	15	0	500	0	0	0	0	0	0	0	0	0	0	10	500
DZ-19	445	30	10	5	490	10	0	10	0	0	0	0	0	0	0	425	500
DZ-21	412	28	8	10	458	0	0	0	2	0	16	18	0	6	18	378	500
DZ-23	460	30	6	0	496	2	0	2	0	0	0	0	0	2	0	52	500
DZ-26	440	52	2	4	498	0	0	0	0	0	0	0	0	2	0	50	500
GKS-3	476	18	2	2	498	0	0	0	0	0	0	0	2	0	0	40	500
GKS-4	422	52	16	6	496	0	0	0	0	4	0	4	0	0	0	88	500
GKS-6	444	46	6	0	496	0	0	0	0	0	0	0	4	0	0	32	500
GKS-8	444	24	2	2	472	0	0	0	0	0	0	0	28	0	0	66	500
GKS-10	480	10	4	0	494	0	0	0	0	0	0	0	6	0	0	18	500
GKS-12	332	108	28	20	488	2	0	2	2	2	0	4	0	0	6	118	500
GKS-18	485	15	0	0	500	0	0	0	0	0	0	0	0	0	0	110	500
GKS-19	460	20	5	0	485	0	0	0	0	0	0	0	10	5	0	45	500

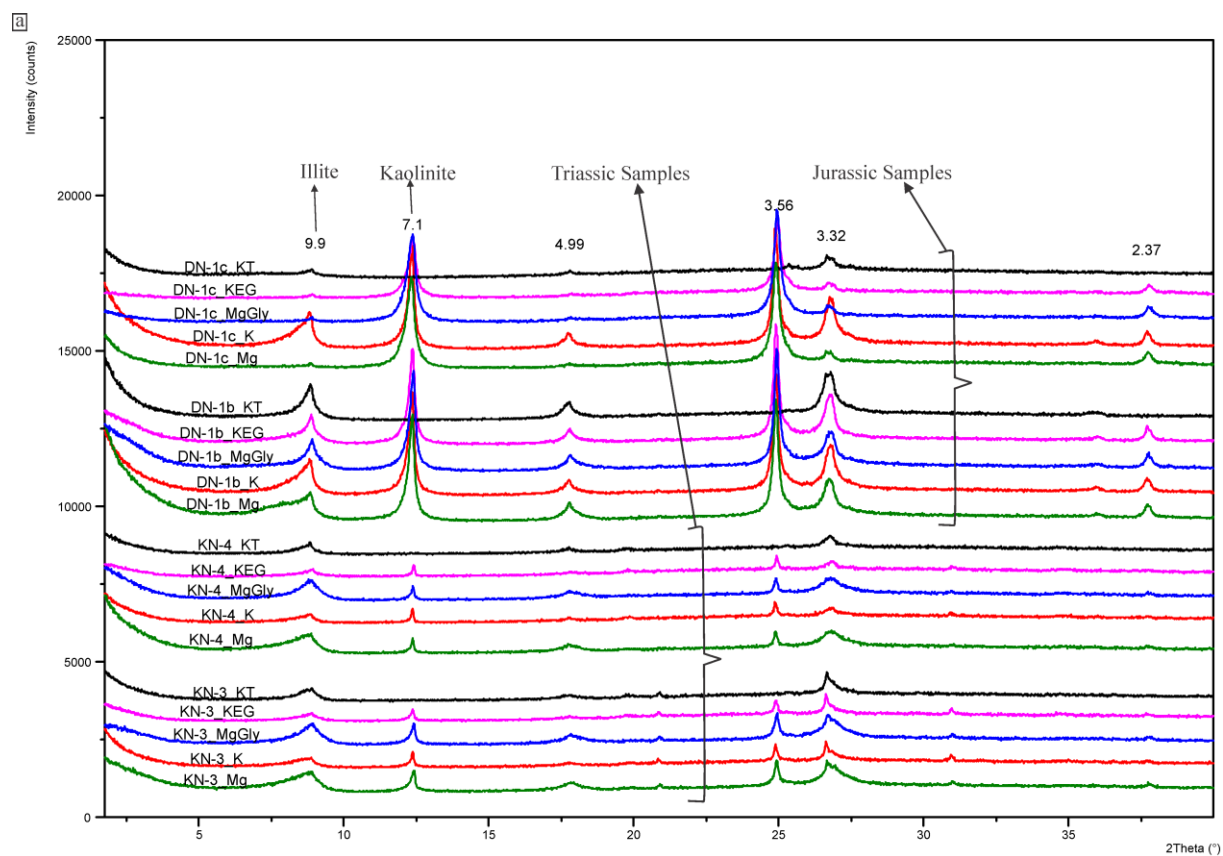
Appendix 1.

Sample	Q _{mue}	Q _{muu}	Q _{pq (2-3)}	Q _{pq>3}	Q _{total}	A	P	F _{total}	L _v	L _m	L _s	L _{total}	M	D	O	M/C	Total
GKS-20	444	22	12	2	480	0	0	0	0	0	0	0	0	20	0	382	500
GKS-22	474	12	0	0	486	0	0	0	0	0	0	0	6	4	4	66	500
GKS-26	458	34	4	2	498	0	0	0	0	0	0	0	0	2	0	48	500
GKS-31	490	0	0	0	490	0	0	0	0	0	0	0	10	0	0	25	500
SD-1	450	45	0	0	495	0	0	0	0	0	0	0	0	5	0	85	500
SD-4	348	98	36	10	492	2	0	2	0	0	0	0	0	4	2	186	500
SD-5	482	8	2	2	494	0	0	0	0	0	0	0	0	4	2	70	500
SD-6	427	60	13	0	500	0	0	0	0	0	0	0	0	0	0	413	500
SD-9	448	40	10	2	500	0	0	0	0	0	0	0	0	0	0	102	500
SD-10	372	98	20	8	498	0	0	0	0	2	0	2	0	0	0	204	500

Appendix 2a.



Appendix 2b.



Appendix 3.

	Method	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200
	Analyte	SiO2	Al2O3	Fe2O3	MgO	CaO	Na2O	K2O	TiO2	P2O5	MnO	Cr2O3
	Unit	%	%	%	%	%	%	%	%	%	%	%
	MDL	0.01	0.01	0.04	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.002
Section	Kasanwala											
Sample No.	Sample type											
DK-T	Outcrop	1.25	0.5	0.9	19.27	31.17	0.02	0.11	0.02	0.1	0.06	<0.002
DK-F	Outcrop	1.64	0.52	2.13	16.8	33.29	0.02	0.13	0.03	0.69	0.13	<0.002
DK-F	Outcrop	1.6	0.52	2.21	17.14	33.05	0.02	0.13	0.03	0.73	0.13	0.003
DK-K	Outcrop	4.31	1.52	2.13	17.11	30.69	0.02	0.39	0.07	0.07	0.07	0.003
KF-1	Outcrop	93.84	2.95	0.55	0.06	0.42	<0.01	<0.01	0.21	0.03	<0.01	0.031
KF-2	Outcrop	93.88	3.05	1.03	0.02	0.14	<0.01	0.02	0.41	0.03	<0.01	0.027
KF-3	Outcrop	86.76	7.59	0.98	0.04	0.07	0.01	0.03	0.78	0.05	<0.01	0.019
KF-4	Outcrop	92.81	2.19	2.39	0.04	0.06	<0.01	<0.01	0.76	0.02	0.02	0.03
KF-5	Outcrop	90.13	5.41	1.01	0.03	0.12	0.02	<0.01	0.44	0.05	<0.01	0.023
KF-6	Outcrop	85.12	8.1	2.32	0.03	0.06	0.01	0.02	0.98	0.02	<0.01	0.018
KF-7	Outcrop	92.72	3.77	1.31	0.02	0.08	<0.01	0.02	0.35	0.02	<0.01	0.021
KF-8	Outcrop	88.01	6.77	0.95	0.03	0.08	<0.01	0.04	1	0.04	<0.01	0.024
KF-10	Outcrop	91.62	4.66	1.23	<0.01	0.03	<0.01	<0.01	0.32	0.03	<0.01	0.029
KF-11	Outcrop	74.42	16.05	0.93	0.04	0.06	0.02	0.1	1.34	0.17	<0.01	0.026
KF-13	Outcrop	93.04	3.68	0.58	0.02	0.1	<0.01	0.04	0.71	0.03	<0.01	0.021
KF-14	Outcrop	53.53	11.32	25.84	0.07	0.17	0.02	0.12	2.22	0.06	0.03	0.022
KL-1	Outcrop	30.72	35.02	18.4	0.03	0.04	0.02	0.07	2.46	0.04	<0.01	0.032
KL-2	Outcrop	40.15	34.83	9.82	0.02	0.04	0.02	0.02	1.91	<0.01	<0.01	0.017
KL-3	Outcrop	35.93	30.74	19	0.04	0.07	0.02	0.14	2	0.07	<0.01	0.021
KL-5	Outcrop	14.68	53.09	13.9	0.09	0.15	0.01	0.06	3.41	0.04	<0.01	0.029
KL-7	Outcrop	46.38	35.25	0.51	0.07	0.11	0.02	0.03	3.4	0.03	<0.01	0.034
DK-a	Outcrop	79.21	10.63	4.19	0.07	0.16	<0.01	0.08	0.55	0.07	<0.01	0.047
DK-b	Outcrop	95.29	2.24	0.99	0.04	0.05	<0.01	0.03	0.12	0.02	<0.01	0.035
DK-c	Outcrop	93.93	1.69	0.82	0.07	0.9	0.02	0.27	0.24	0.02	<0.01	0.044
DK-d	Outcrop	91.09	3.13	1.48	0.14	0.31	0.02	0.47	0.49	0.02	<0.01	0.033
DK-E	Outcrop	43.5	12.95	33.91	0.1	1.31	0.03	0.36	0.98	0.68	0.01	0.024
KW-1	Outcrop	48.34	8.62	29.21	0.2	0.51	0.07	0.45	0.76	0.34	<0.01	0.035
DK-P1	Outcrop	53.65	24.84	3.13	1.25	0.64	0.08	3.55	1.19	0.07	<0.01	0.025
DK-P2	Outcrop	29.71	12.18	4.22	8.23	16.12	0.06	1.84	0.64	0.07	0.09	0.009
DK-P3	Outcrop	49.18	19.13	4.03	3.25	5.21	0.07	2.86	1.09	0.06	0.04	0.02
DK-P2a	Outcrop	29.37	12.13	4.19	8.45	16.01	0.06	1.84	0.64	0.05	0.09	0.011
Section	Kaowaali											
Sample No.	Sample type											
KK-1a	Outcrop	87.17	1.29	2.1	0.09	4.27	0.03	0.41	0.42	2.92	<0.01	0.051
KK-2a	Outcrop	88.77	4.24	2.74	0.17	0.17	0.02	0.81	0.4	0.17	0.03	0.047
KK-3a	Outcrop	98.87	0.14	0.64	0.01	0.03	<0.01	0.02	0.02	0.02	<0.01	0.065
KK-4a	Outcrop	94.92	2.35	0.53	0.09	0.13	0.01	0.5	0.44	0.02	<0.01	0.037
KK-6a	Outcrop	83.66	2.93	3.88	0.78	2.62	<0.01	0.6	0.38	0.08	0.04	0.042
KK-7a	Outcrop	95.4	2.12	0.6	0.06	0.06	0.02	0.36	0.49	<0.01	<0.01	0.038
DU1	Outcrop	53.45	9.78	13.88	2.03	4.93	0.04	5.15	0.22	1.96	<0.01	0.051
DZ4	Outcrop	70.96	17.54	0.8	0.36	0.09	0.07	1.99	1.29	0.05	<0.01	0.015
DZ6	Outcrop	55.8	26.61	1.38	0.56	0.06	0.09	2.69	1.92	0.06	<0.01	0.021

Appendix 3.

	Method	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200
	Analyte	SiO2	Al2O3	Fe2O3	MgO	CaO	Na2O	K2O	TiO2	P2O5	MnO	Cr2O3
	Unit	%	%	%	%	%	%	%	%	%	%	%
	MDL	0.01	0.01	0.04	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.002
DZ7	Outcrop	4.04	0.56	0.49	0.71	0.66	0.27	0.13	0.13	<0.01	<0.01	<0.002
DZ9a	Outcrop	59.38	21.36	4.43	0.78	0.09	0.09	2.97	1.38	0.06	0.05	0.017
DZ11	Outcrop	57.37	25.54	4.14	0.27	0.11	0.05	0.48	1.13	0.07	<0.01	0.013
DZ12	Outcrop	59.19	19.96	6.06	0.88	0.11	0.08	2.99	1.25	0.08	0.13	0.014
DZ12a	Outcrop	65.43	19.28	1.73	0.61	0.17	0.08	2.81	1.27	0.06	<0.01	0.013
Section	Nammal											
Sample No.	Sample type											
KK-1	Outcrop	52.74	15.9	5.17	2.46	1.44	0.4	4.6	1.12	0.08	0.01	0.016
KK-2	Outcrop	28.42	12.21	4.13	11.12	13.38	0.31	3.1	0.62	0.17	0.03	0.007
KK-3	Outcrop	44.62	19.18	5.99	4.85	3.48	0.31	4.87	0.85	0.32	0.02	0.015
KK-4	Outcrop	42.09	18.24	5.69	5.43	4.83	0.89	5.05	1.05	0.14	0.02	0.016
DKA-10	Outcrop	53.6	19.59	7	2.65	0.44	0.08	6.31	1.07	0.24	0.01	0.016
DKA-11	Outcrop	53.2	20.04	6.69	2.56	0.38	0.07	6.27	1.04	0.34	0.01	0.016
Dka - 12	Outcrop	52.68	19.75	6.5	2.57	0.31	0.08	6.12	1.03	0.49	0.01	0.013
DKA-13	Outcrop	53.78	20.51	4.82	1.82	0.33	0.06	4.79	0.99	1.88	<0.01	0.026
DN-1a	Outcrop	6.31	0.58	4.55	0.18	9.57	0.03	0.04	0.05	0.59	<0.01	0.003
DN-1d	Outcrop	62.09	20.99	1.08	0.31	0.03	0.06	0.94	3.27	0.07	0.01	0.02
DN-1j	Outcrop	58.9	1.87	28.18	0.77	1.54	<0.01	0.1	0.38	0.06	0.57	0.01
DN-1	Outcrop	89.22	3.45	0.75	0.03	0.08	0.06	0.13	0.38	0.03	<0.01	0.043
DN-4	Outcrop	93.18	1.7	0.83	0.17	0.03	0.05	0.08	1.16	0.03	<0.01	0.047
DN-5	Outcrop	63.28	1.39	28.42	0.07	0.35	0.01	0.1	0.08	0.06	0.32	0.032
DN8	Outcrop	58.23	22.67	0.84	0.2	0.04	0.04	0.62	2.26	0.07	<0.01	0.017
DN8a	Outcrop	54.95	24.92	1.91	0.38	0.05	0.07	1.16	3.52	0.05	0.02	0.02
DN-11	Outcrop	7.86	4.15	10.39	12.04	25.55	0.02	0.44	0.24	0.06	0.29	0.005
DN12	Outcrop	50.69	21.51	6.54	0.58	0.08	0.11	2.14	1.38	0.13	<0.01	0.017
DN-16	Outcrop	46.24	2.21	5.75	7.24	15.82	0.01	0.2	0.21	0.05	0.11	0.018
DN16first	Outcrop	50.62	26.4	4.55	0.86	0.27	0.12	3.21	1.38	0.2	0.01	0.018
DN17	Outcrop	73.95	14.21	1.55	0.43	0.12	0.08	2.58	1.18	0.06	<0.01	0.011
DN-19	Outcrop	73.62	2.14	2.96	2.99	6.78	0.02	0.27	0.63	0.02	0.06	0.023
DN22	Outcrop	74.69	14.02	1.31	0.33	0.05	0.07	1.94	1.35	0.04	<0.01	0.011
DN24	Outcrop	63.46	15.22	2.46	0.3	0.02	0.04	0.66	1.57	0.04	0.01	0.012
DN30	Outcrop	74.44	9.92	4.27	0.35	0.05	0.05	1.59	1.41	0.05	<0.01	0.012
DN-35	Outcrop	12.01	3.02	6.04	12.29	26.48	0.04	0.24	0.22	0.03	0.1	0.005
DN-37	Outcrop	89.56	3.25	1.34	0.28	0.05	0.03	0.61	0.62	0.02	<0.01	0.013
Section	Zaluch											
Sample No.	Sample type											
ZK-1	Outcrop	4.33	1.43	2.77	18.13	28.99	0.02	0.24	0.06	0.02	0.11	0.002
ZK-2	Outcrop	3.88	1.36	2.75	18.59	28.77	0.02	0.24	0.06	<0.01	0.11	<0.002
ZS-1	Outcrop	96.29	1.6	0.43	<0.01	0.32	<0.01	0.02	0.18	0.03	<0.01	0.043
ZS-3	Outcrop	93.45	3.65	0.28	<0.01	0.04	<0.01	0.04	0.62	0.05	<0.01	0.027
ZS-4	Outcrop	86.99	7.95	0.82	0.02	0.05	<0.01	0.13	0.7	0.03	<0.01	0.015
ZS-6	Outcrop	76.88	12.45	4.29	0.05	0.06	0.01	0.27	0.67	0.03	<0.01	0.013
ZS-8	Outcrop	91.51	1.63	3.03	0.01	0.49	<0.01	0.03	1.5	0.04	<0.01	0.025
ZS-9	Outcrop	85.93	8.64	0.97	<0.01	0.01	<0.01	0.03	0.84	0.02	<0.01	0.016

Appendix 3.

	Method	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200
	Analyte	SiO2	Al2O3	Fe2O3	MgO	CaO	Na2O	K2O	TiO2	P2O5	MnO	Cr2O3
	Unit	%	%	%	%	%	%	%	%	%	%	%
	MDL	0.01	0.01	0.04	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.002
ZS-11	Outcrop	95.74	2.55	0.35	<0.01	<0.01	<0.01	0.02	0.1	0.02	<0.01	0.015
ZS-12	Outcrop	94.24	3.38	0.62	<0.01	<0.01	<0.01	0.01	0.2	0.02	<0.01	0.049
ZS-14	Outcrop	62.78	22.85	3.6	0.06	0.05	0.02	0.32	1.25	0.05	<0.01	0.012
ZL-1	Outcrop	49.84	29.77	7.23	0.06	0.07	0.02	0.32	0.94	0.06	<0.01	0.016
ZL-2	Outcrop	43.74	35.83	2.62	0.17	0.2	0.01	0.15	2.51	0.06	<0.01	0.043
DZ-1	Outcrop	48.73	30.5	1.75	0.56	0.77	0.15	2	1.7	0.17	<0.01	0.025
DZ-2	Outcrop	83.94	3.96	8.08	0.06	0.16	0.01	0.08	0.46	0.12	<0.01	0.029
DZ-3	Outcrop	95.62	2.02	0.75	0.02	0.04	<0.01	0.1	0.14	0.01	<0.01	0.013
DZ-4	Outcrop	62.91	22.34	1.3	0.61	0.25	0.1	2.86	1.36	0.07	<0.01	0.018
DZ-6	Outcrop	64.64	21.19	1.22	0.49	0.11	0.08	2.7	1.33	0.06	<0.01	0.016
DZ-8	Outcrop	59.81	25.55	0.92	0.26	0.19	0.05	0.75	2.03	0.05	<0.01	0.019
DZ-9	Outcrop	66.62	20.35	0.66	0.25	0.06	0.05	0.66	3.11	0.07	<0.01	0.018
DZ-10	Outcrop	20.06	11.1	13.7	7.82	17.2	0.05	1	0.55	0.05	0.2	0.007
DZ-11	Outcrop	52.06	28.4	2.7	0.92	0.16	0.09	2.61	1.53	0.08	0.01	0.021
DZ-13	Outcrop	58.3	4.13	4.9	4.09	11.5	0.02	0.32	0.5	0.02	0.14	0.012
DZ-14	Outcrop	56.45	23.51	2.51	0.99	1.42	0.08	2.64	1.51	0.08	<0.01	0.02
DZ-15	Outcrop	80.98	11.02	0.55	0.16	0.11	0.06	0.73	1.23	0.05	<0.01	0.013
DZ-17	Outcrop	18.26	1.8	2.26	0.56	42.01	0.02	0.24	0.16	0.03	0.12	0.004
DZ-18	Outcrop	63.91	22.09	2.86	0.13	0.11	0.03	0.56	1.35	0.08	<0.01	0.015
DZ-18a	Outcrop	66.5	1.18	1.08	0.15	16.81	<0.01	0.14	0.28	<0.01	0.03	0.013
DZ-20	Outcrop	62.19	20.99	2.96	0.9	0.16	0.07	2.52	1.27	0.04	0.01	0.018
DZ-23	Outcrop	94.07	2.45	0.65	0.1	0.32	0.02	0.31	0.53	0.03	<0.01	0.014
DZ-25	Outcrop	90.55	2.31	4.34	0.08	0.37	0.03	0.08	0.78	0.01	0.01	0.012
DZ-26	Outcrop	97.59	0.91	0.3	0.02	0.14	<0.01	0.06	0.31	0.01	<0.01	0.02
DZ-27	Outcrop	22.58	6.39	8.74	9.12	20.9	0.03	0.6	0.48	0.02	0.12	0.009
Section	Gulla Khel											
Sample No.	Sample type											
GKK-1	Outcrop	16.61	6.96	2.79	15.27	20.86	0.05	1.99	0.35	0.09	0.05	0.005
GKK-2	Outcrop	23.34	9.26	3.45	12.55	18.1	0.1	2.61	0.54	0.06	0.03	0.009
GKK-3	Outcrop	54.88	18.56	4.78	3.08	0.99	0.09	5.49	1.12	0.11	0.01	0.015
GKS-1A	Outcrop	27.52	13.4	32.37	2.21	1.31	0.05	3.66	0.58	0.82	0.05	0.013
GKS-1B	Outcrop	44.1	22.66	5.48	2.06	1.4	0.06	4.56	1.07	0.69	0.02	0.021
GKS-2	Outcrop	33.42	6.22	4.71	0.14	1.19	0.15	0.37	0.86	0.05	<0.01	0.009
GKS-4	Outcrop	94.44	2.41	0.36	0.02	0.49	<0.01	0.03	0.21	0.02	<0.01	0.031
GKS-5	Outcrop	53.74	28.25	1.56	0.28	0.3	0.09	0.87	1.48	0.07	<0.01	0.021
GKS-6	Outcrop	92.68	3.56	0.55	0.03	0.04	0.01	0.1	1.2	0.05	<0.01	0.024
GKS-7	Outcrop	53.46	27.32	4.21	0.26	0.49	0.08	0.68	1.38	0.12	<0.01	0.019
GKS-8	Outcrop	80.69	11.66	0.67	0.07	0.06	0.01	0.25	1.51	0.06	<0.01	0.015
GKS-12	Outcrop	96.48	1.27	1.17	0.02	0.03	<0.01	0.01	0.32	0.01	<0.01	0.026
GKS-16	Outcrop	81.29	11.82	0.67	0.02	0.03	<0.01	<0.01	1.05	0.05	<0.01	0.02
GKS-17	Outcrop	67.05	16.94	7.19	0.06	0.06	0.03	0.23	1.44	0.07	<0.01	0.02
GKS-18	Outcrop	63.63	20.62	5.6	0.06	0.08	0.02	0.12	1.1	0.03	<0.01	0.021
GKS-21	Outcrop	47.96	24.38	16.62	0.04	0.08	0.02	0.17	1.13	0.07	<0.01	0.031
GKS-22	Outcrop	78.45	12.51	2.04	0.04	0.04	<0.01	0.12	1.3	0.05	<0.01	0.017

Appendix 3.

	Method	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200
	Analyte	SiO2	Al2O3	Fe2O3	MgO	CaO	Na2O	K2O	TiO2	P2O5	MnO	Cr2O3
	Unit	%	%	%	%	%	%	%	%	%	%	%
	MDL	0.01	0.01	0.04	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.002
GKS-23	Outcrop	59.81	22.24	1.48	0.22	0.02	0.06	0.92	2.1	0.06	<0.01	0.019
GKS-26	Outcrop	93.06	1.81	2.52	0.02	0.03	0.02	0.06	0.6	0.07	<0.01	0.024
GKS-27	Outcrop	8.25	4.47	10.49	11.48	26.52	0.02	0.34	0.24	0.03	0.3	0.009
GKS-29	Outcrop	57.86	24.75	1.86	0.54	0.72	0.08	1.96	1.33	0.05	<0.01	0.021
GKS-30	Outcrop	8.97	4.81	7.12	11.79	27.55	0.03	0.51	0.24	0.01	0.18	0.006
Section	Paniala											
Sample No.	Sample type											
KS-4	Outcrop	48.34	16.85	6.18	4.44	2.88	0.12	5.08	0.89	0.28	0.01	0.014
KS-6	Outcrop	32.81	11.8	4.61	9.74	11.8	0.09	3.74	0.63	0.49	0.04	0.011
KS-10	Outcrop	29.61	13.54	6.95	9.23	12.27	0.07	2.66	0.69	0.31	0.12	0.01
KS-11a	Outcrop	41.35	23.71	7.66	1.36	0.13	0.04	1.87	1.04	0.17	0.04	0.014
ASH	Outcrop	33.21	20.07	14.54	1.7	3.14	0.04	1.83	0.81	0.42	0.04	0.013
SD-a	Outcrop	1.66	0.48	6.53	18.11	28.77	0.03	0.06	0.03	<0.01	0.42	<0.002
SD-b	Outcrop	9.08	7.2	1.45	0.09	26.24	0.04	0.09	0.04	<0.01	<0.01	0.003
SD-c	Outcrop	61.69	19.55	4.1	0.72	0.09	0.09	2.44	1.49	0.14	<0.01	0.015
SD-1	Outcrop	94.28	2.7	0.26	0.08	0.01	0.03	0.26	0.78	0.03	<0.01	0.013
SD-2	Outcrop	51.04	29.7	2.43	0.57	0.2	0.11	1.56	1.33	0.06	<0.01	0.018
SD-3	Outcrop	50.36	29.57	3.44	0.69	0.06	0.11	2.08	1.25	0.06	<0.01	0.02
SD-4	Outcrop	62.06	0.89	4.8	5.85	9.84	0.02	0.19	0.56	0.27	0.15	0.027
SD-5	Outcrop	89.22	6.13	0.43	0.01	<0.01	<0.01	0.05	1.42	0.03	<0.01	0.019
SD-6	Outcrop	51.7	3.8	35.32	0.36	0.12	0.02	0.33	0.74	0.02	0.71	0.012
SD-7	Outcrop	78.16	12.69	1.05	0.25	0.1	0.1	1.13	1.58	0.06	<0.01	0.017
SD-8	Outcrop	52.27	20.01	18.13	0.03	<0.01	0.02	0.19	1.2	0.03	<0.01	0.023
SD-9	Outcrop	90.74	0.6	6.55	0.03	0.08	0.02	0.01	0.19	0.02	<0.01	0.024
SD-10	Outcrop	83.53	0.9	12.64	0.01	0.02	<0.01	0.03	0.2	0.02	<0.01	0.023
SD-11	Outcrop	5.25	1.18	18.55	11.5	27.59	0.09	0.06	0.1	0.03	0.72	0.004
SD-12	Outcrop	2.14	1.04	0.88	0.16	31.95	0.02	0.03	0.06	<0.01	0.06	<0.002
SD-13	Outcrop	8.14	2.97	5.91	15.77	26.83	0.03	0.21	0.21	0.01	0.17	0.004
SD-14	Outcrop	2.23	4.96	0.68	0.05	28.79	0.54	0.21	0.06	0.02	<0.01	<0.002
SD-15	Outcrop	37.52	22.21	3.04	1.32	7.73	0.06	2.17	1	0.06	0.01	0.017

Appendix 3.

	LF200 Ba PPM 1	LF200 Ni PPM 20	LF200 Sc PPM 1	LF200 LOI % -5.1	LF200 Sum % 0.01	LF200 Be PPM 1	LF200 Co PPM 0.2	LF200 Cs PPM 0.1	LF200 Ga PPM 0.5	LF200 Hf PPM 0.1	LF200 Nb PPM 0.1	LF200 Rb PPM 0.1	LF200 Sn PPM 1
Section Sample No.	Kasanwala												
DK-T	21	<20	1	46.3	99.66	<1	5.6	0.4	0.8	0.1	0.6	5.5	<1
DK-F	33	<20	<1	44.3	99.7	<1	10.4	0.4	0.6	0.2	0.8	5.8	<1
DK-F	35	<20	<1	44.4	99.95	<1	12.5	0.2	<0.5	0.2	1.5	6.4	<1
DK-K	41	<20	2	43.6	99.95	<1	8.1	1	0.8	0.4	1.6	19.9	<1
KF-1	34	<20	3	1.8	99.9	<1	3.3	<0.1	2.6	3.7	6.5	0.6	<1
KF-2	34	<20	5	1.3	99.93	<1	4.2	<0.1	3.6	10.2	9.3	1.2	<1
KF-3	72	<20	9	3.4	99.77	<1	2.4	<0.1	8.3	21.8	18.7	1.4	2
KF-4	66	<20	12	1.5	99.79	1	9.4	<0.1	2.8	28.7	19.1	0.8	3
KF-5	38	<20	4	2.6	99.87	<1	3	<0.1	6.5	6.2	10.7	0.6	2
KF-6	40	<20	10	3.1	99.81	2	2.7	<0.1	10.1	30.5	22.5	0.8	2
KF-7	24	<20	3	1.6	99.95	<1	2.7	<0.1	4.4	7.1	8.3	0.8	<1
KF-8	59	<20	8	2.9	99.81	<1	3.3	0.1	8.7	18.3	21.1	2.5	2
KF-10	29	<20	3	2	99.9	<1	2.7	<0.1	4.9	3.9	7.8	0.6	1
KF-11	101	<20	11	6.4	99.59	<1	2.5	0.3	18.8	20.8	33	4	2
KF-13	23	<20	4	1.7	99.93	<1	1.5	<0.1	7.2	10.8	16.1	1.6	4
KF-14	62	50	40	6.1	99.56	3	48.5	0.1	16.1	51.4	40.7	4.7	4
KL-1	39	53	43	12.9	99.79	<1	6.6	0.2	48	19.7	46	2.2	9
KL-2	11	65	30	13	99.87	<1	4.2	0.2	44.4	12.4	34.4	0.9	4
KL-3	66	<20	33	11.7	99.74	1	5.7	0.4	34.6	20.9	37	5.5	6
KL-5	32	55	73	14.2	99.71	3	5.5	0.3	56.9	24.8	62.5	2.8	11
KL-7	17	<20	21	13.9	99.72	<1	1.3	<0.1	53	38.2	65.3	1.2	10
DK-a	59	<20	10	4.9	99.88	2	2	0.1	30.7	6.5	12.3	2.4	3
DK-b	45	<20	2	1.1	99.93	<1	2.1	<0.1	4.3	2.2	3	1.8	3
DK-c	63	<20	2	1.9	99.9	<1	2.8	0.1	2	5.8	4.8	7.5	2
DK-d	101	<20	4	2.6	99.85	<1	1.9	0.4	4.3	13.6	8.5	15.2	7
DK-E	116	<20	28	5.9	99.76	<1	2.7	2.5	20.8	11	17.6	17.3	4
KW-1	316	<20	20	11.2	99.78	1	2.5	0.6	48.7	10.5	14.7	24.2	11
DK-P1	343	25	20	11.3	99.81	3	6.4	13.4	31.2	7.7	26.9	168.2	7
DK-P2	161	30	15	26.7	99.88	<1	13.4	6	13.5	3.7	13	84.3	3
DK-P3	256	24	18	14.9	99.84	3	13	9.8	24.2	6.7	23.6	135.4	5
DK-P2a	139	27	13	26.9	99.78	2	12	5.3	13.1	3.6	12.5	78.1	3
Section Sample No.	Kaowaali												
KK-1a	51	22	8	1.1	99.86	<1	3.4	0.6	1.6	12.7	8.6	14.2	1
KK-2a	97	<20	6	2.4	99.95	<1	7.8	1.2	5.3	6.3	7.8	29.6	1
KK-3a	9	<20	<1	0.2	100.02	<1	3	<0.1	2.5	0.9	2.3	0.4	<1
KK-4a	22	<20	4	0.9	99.94	<1	1.6	0.4	4.4	9.1	8.6	14.2	1
KK-6a	36	<20	7	4.8	99.86	<1	3.7	0.8	4.3	7.6	7.2	19.4	1
KK-7a	19	<20	3	0.8	99.95	<1	1.9	0.4	3.8	8.6	9.1	8.9	1
DU1	51	<20	10	8.4	99.85	6	1.7	13.9	14	2.1	6.1	213.5	1
DZ4	254	<20	14	6.7	99.82	3	2.4	8.9	27.3	16.1	22.8	91.3	5
DZ6	314	<20	21	10.5	99.72	5	3.1	17	50.1	22.4	38.3	142.3	7

Appendix 3.

	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200
	Ba	Ni	Sc	LOI	Sum	Be	Co	Cs	Ga	Hf	Nb	Rb	Sn
	PPM	PPM	PPM	%	%	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM
	1	20	1	-5.1	0.01	1	0.2	0.1	0.5	0.1	0.1	0.1	1
DZ7	31	<20	2	92.9	99.94	2	1.9	0.3	0.9	2.1	1.2	3.2	4
DZ9a	469	28	19	9.2	99.78	8	19.7	13.9	28.3	12.7	27.5	140.2	6
DZ11	182	<20	16	10.7	99.85	2	6.3	4.7	30.8	7.1	25.4	29.2	4
DZ12	264	32	21	9.1	99.83	3	21.9	11.9	23.6	8.6	22.3	131.4	5
DZ12a	307	<20	16	8.4	99.83	3	5	10.4	23.7	11.6	22.4	117.4	4
Section	Nammal												
Sample No.													
KK-1	193	27	11	15.9	99.85	8	11.5	19.3	23.5	9.7	25.3	221.1	4
KK-2	84	<20	9	26.4	99.89	5	10.7	12.1	15.1	3.5	14.2	149.5	2
KK-3	148	29	15	15.3	99.87	5	14	20.5	24.3	4.7	19.5	243.8	3
KK-4	136	23	12	16.4	99.83	10	14.1	19.5	26.9	8.2	24.4	250.3	4
DKA-10	105	175	15	8.7	99.76	14	238	26.4	27	8.9	26.5	303	3
DKA-11	101	149	16	9.1	99.76	18	271.9	25.9	27.4	7.5	23.8	314.5	3
Dka - 12	121	138	17	10.2	99.73	13	246.5	26.6	30.3	7.8	26.8	312.7	3
DKA-13	651	68	23	9.5	98.62	10	51.6	25.7	40.1	16.5	23.8	238.7	4
DN-1a	77	<20	3	67.7	89.59	<1	1.5	0.1	0.6	1.5	1.4	1.7	2
DN-1d	163	24	24	10.8	99.71	<1	9.1	8.4	30.7	26.2	60.9	43.7	9
DN-1j	28	<20	17	7.5	99.88	<1	34.3	0.5	2	6.3	6	5.2	1
DN-1	57	<20	3	5.8	99.96	<1	2.4	0.2	6.4	5.9	8.3	3.2	2
DN-4	39	<20	4	2.5	99.84	<1	2.8	<0.1	3.4	28.4	20.8	2.1	2
DN-5	24	<20	12	5.8	99.91	<1	23.9	<0.1	2	1.4	2.1	3.6	<1
DN8	113	33	21	14.8	99.75	1	9.7	6.2	28.1	17.4	41.4	29.4	7
DN8a	180	28	27	12.7	99.76	<1	8.1	9.9	32.8	22.7	63.8	55.9	9
DN-11	37	<20	6	38.9	99.94	1	7.4	1.9	5	1.3	4.9	21	1
DN12	187	<20	19	16.6	99.82	<1	7.2	11.1	26.2	10.8	24.7	95.9	6
DN-16	19	<20	5	22.1	99.91	<1	7.1	0.6	1.7	1.6	3.8	9.8	1
DN16first	242	33	24	12.1	99.79	5	26.4	14.7	31.9	7.8	27.1	146.8	6
DN17	267	<20	12	5.6	99.82	1	7.2	5.2	12.7	16.2	21.8	87.1	3
DN-19	46	<20	5	10.4	99.89	<1	5.3	0.6	2.1	8.6	10.3	11.4	2
DN22	223	<20	13	6	99.85	<1	3.1	4.9	13.3	13	23.8	70.4	4
DN24	89	61	19	16	99.84	8	22.8	4.9	18	14.3	26.7	27.3	6
DN30	175	<20	12	7.8	99.87	1	2.4	4.6	14.2	13.3	23.2	57.4	4
DN-35	22	<20	4	39.5	99.93	<1	3.9	1.1	3.1	2.1	3.9	12.5	<1
DN-37	61	<20	4	4.2	99.94	<1	3.4	1.7	4.8	10.5	10	28.9	2
Section	Zaluch												
Sample No.													
ZK-1	11	<20	2	43.6	99.68	<1	18.5	0.7	1.3	0.4	1.7	9.8	<1
ZK-2	10	20	2	43.8	99.65	2	24.4	0.7	1	0.3	1.6	10.7	<1
ZS-1	15	<20	<1	1.1	100	1	2.8	<0.1	1.5	2.3	5.5	1.3	<1
ZS-3	40	<20	4	1.7	99.87	<1	1.9	0.2	4.1	19.2	13.7	1.9	1
ZS-4	63	<20	6	3.2	99.89	<1	2.1	0.9	10.3	17.3	16	6.5	1
ZS-6	133	<20	8	5.2	99.93	4	4.8	1.2	14.3	6.2	14	12.6	2
ZS-8	30	<20	7	1.3	99.59	<1	5.9	<0.1	1.8	63.9	27.2	1.5	2
ZS-9	61	<20	8	3.4	99.91	<1	7.4	0.2	9.5	16.2	23.1	1.8	2

Appendix 3.

	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200
	Ba	Ni	Sc	LOI	Sum	Be	Co	Cs	Ga	Hf	Nb	Rb	Sn
	PPM	PPM	PPM	%	%	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM
	1	20	1	-5.1	0.01	1	0.2	0.1	0.5	0.1	0.1	0.1	1
ZS-11	17	<20	1	1.2	100	<1	2.4	<0.1	2.6	2	2.3	1.1	2
ZS-12	25	<20	2	1.5	99.99	<1	2.8	<0.1	3.7	2.8	5.5	1.3	<1
ZS-14	134	<20	17	8.9	99.87	4	3.7	1.4	28	12.3	26.1	12.9	4
ZL-1	171	<20	18	11.6	99.91	2	2.4	1.1	34.2	3.2	20.8	13.9	4
ZL-2	48	<20	14	14.4	99.76	3	15.4	0.7	35.2	27.5	53.3	4.6	9
DZ-1	309	23	26	13.3	99.74	2	4.5	8.6	38.8	22.3	33.5	82.9	6
DZ-2	42	<20	4	3	99.91	2	2.3	0.3	3.2	10.9	8.8	3	<1
DZ-3	30	<20	1	1.3	100.01	<1	1.2	0.4	1.6	1.6	2.4	4.4	<1
DZ-4	305	24	19	8	99.82	<1	2.9	12.3	29.9	10.8	26	131.5	<1
DZ-6	324	<20	18	8	99.86	5	3.4	10.7	25.6	12	24.6	125.2	5
DZ-8	128	<20	19	10.2	99.82	<1	3.8	7.1	32.5	18.9	39.1	35.4	2
DZ-9	140	<20	23	7.9	99.77	<1	2.5	5.7	23.9	20	55.4	32.8	7
DZ-10	145	<20	12	28.1	99.79	<1	15.5	4.5	11.5	3	11.2	46.6	2
DZ-11	248	34	19	11.2	99.81	3	13	11.2	38.9	8.1	30.7	114.4	6
DZ-13	51	<20	6	15.9	99.85	1	8.2	1.2	8.1	8.7	9.6	14.9	<1
DZ-14	282	28	21	10.5	99.8	2	6.5	10	34.5	10.9	30.9	120.9	5
DZ-15	117	<20	17	5	99.89	<1	1.6	3.7	13.4	13	20.7	30.2	3
DZ-17	38	<20	4	34.5	99.94	<1	5.8	0.5	3.2	1.7	3.1	9.4	<1
DZ-18	205	<20	13	8.6	99.79	3	6.9	2	26.6	16.9	29.5	24.4	4
DZ-18a	23	<20	2	13.8	99.96	1	3.5	0.2	<0.5	5.2	4.1	4.8	<1
DZ-20	203	40	20	8.7	99.85	4	23.5	8.8	26.1	9.7	23.7	111.9	5
DZ-23	46	<20	4	1.4	99.92	<1	5.5	0.9	4.1	10.4	9.8	12.7	<1
DZ-25	56	<20	4	1.3	99.92	<1	5.2	0.4	3.7	13.5	13	3.8	<1
DZ-26	16	<20	2	0.6	99.99	<1	1.3	0.2	<0.5	4.7	3.7	2.9	<1
DZ-27	56	<20	7	30.8	99.79	<1	4.9	2.7	7.9	4	8.1	27.7	1
Section	Gulla Khel												
Sample No.													
GKK-1	37	23	9	34.9	99.92	1	10.3	4.7	7.8	1.9	7.9	98.7	1
GKK-2	62	22	6	29.8	99.89	3	12.7	7.6	12.2	3.5	12.4	120.3	2
GKK-3	138	32	12	10.6	99.79	12	20.7	13.1	23.3	12.2	25	232.2	4
GKS-1A	61	188	22	17.6	99.59	32	279.2	9	16	5.7	12.1	164.6	2
GKS-1B	114	112	16	17.5	99.66	13	69.4	14	22.2	9.1	23.7	199.3	4
GKS-2	70	<20	8	52.7	99.8	2	3.4	0.8	6.6	16.4	16.6	12.1	3
GKS-4	18	<20	1	2	99.98	<1	2.1	<0.1	2.1	3.6	4.9	1.4	<1
GKS-5	229	52	16	13.1	99.78	<1	51.7	6.3	36.9	16.8	33.2	51.6	1
GKS-6	30	<20	6	1.5	99.69	<1	2.7	0.2	3	44.5	22.6	3.2	<1
GKS-7	254	<20	18	11.7	99.76	3	9.3	5	33.7	11	29.7	42.6	5
GKS-8	102	<20	10	4.7	99.76	<1	2.7	1.9	13.7	34.1	29.6	15.1	3
GKS-12	18	<20	1	0.6	99.95	<1	9	<0.1	0.6	7.5	6.8	0.4	<1
GKS-16	48	<20	8	4.9	99.87	<1	3.5	<0.1	14.1	16.3	21.3	0.8	3
GKS-17	105	<20	17	6.7	99.79	2	3.4	0.5	21.8	21.9	29.9	9	<1
GKS-18	86	21	14	8.6	99.88	2	6.4	0.8	27.7	11.1	23.2	4.4	<1
GKS-21	55	<20	24	9.3	99.83	3	2.6	0.5	26.5	8.6	23.7	7.1	4
GKS-22	41	<20	13	5.2	99.79	1	2.1	<0.1	14.7	23.2	25.2	4.8	3

Appendix 3.

	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200
	Ba	Ni	Sc	LOI	Sum	Be	Co	Cs	Ga	Hf	Nb	Rb	Sn
	PPM	PPM	PPM	%	%	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM
	1	20	1	-5.1	0.01	1	0.2	0.1	0.5	0.1	0.1	0.1	1
GKS-23	114	38	23	12.8	99.78	3	10.6	6.2	28.4	17.8	36.9	43.3	6
GKS-26	17	<20	3	1.7	99.93	2	1.9	<0.1	2.1	12.1	10.5	1.9	1
GKS-27	32	<20	10	37.8	99.91	2	10.7	1	4.2	2	4.7	13.4	1
GKS-29	190	28	23	10.6	99.83	3	9.1	10.8	30.2	9.4	26.6	99.8	1
GKS-30	44	22	6	38.5	99.75	<1	9.8	1.5	5.8	1.7	5	18.3	<1
Section	Paniala												
Sample No.													
KS-4	110	47	13	14.7	99.79	4	14.3	10.8	21.7	12.7	22.5	231.6	3
KS-6	96	23	9	23.9	99.7	4	10.6	7.3	13.7	13.2	14.1	156.1	2
KS-10	112	31	9	24.3	99.74	10	23.4	7.3	17.3	4.3	16.8	107	2
KS-11a	143	39	12	22.4	99.76	7	26.9	12.8	29.7	5.7	24.6	108.7	3
ASH	108	50	11	23.9	99.78	8	66	6.6	24.1	4.7	19.4	76.2	3
SD-a	28	34	<1	43.6	99.67	5	51	0.2	<0.5	0.3	0.7	4	<1
SD-b	4532	42	1	25.2	69.95	<1	14.9	0.7	<0.5	0.3	2.6	6.9	<1
SD-c	360	<20	15	9.4	99.75	2	4.6	6	22.7	20.2	32.3	99.8	3
SD-1	64	<20	3	1.4	99.88	<1	0.6	0.4	3.7	21.5	15.3	9.6	1
SD-2	232	21	18	12.8	99.85	1	6.3	5.5	37.6	9.7	30.5	66.2	4
SD-3	204	27	18	12.2	99.86	3	7.9	8.9	38.6	7	27.6	99.8	2
SD-4	142	<20	3	15.2	99.84	<1	6.9	0.2	0.7	7.3	11	5.9	<1
SD-5	51	<20	7	2.5	99.83	<1	1.9	<0.1	6	29.3	27.9	2	2
SD-6	107	27	19	6.7	99.84	3	35.9	1	5.5	19.4	14.6	14.5	2
SD-7	191	<20	9	4.6	99.77	2	2.6	2.8	14.5	26.4	35.3	48.1	<1
SD-8	53	<20	15	8	99.85	<1	3.2	0.8	22.2	15	23.9	9.3	4
SD-9	10	<20	1	1.7	99.97	<1	1.3	<0.1	6.1	3	5.5	1.2	6
SD-10	10	<20	1	2.6	99.97	<1	1.8	0.1	2.5	2.9	4.3	2	15
SD-11	8	24	3	34.6	99.7	2	28.8	0.4	1.2	1.4	1.9	3.1	<1
SD-12	18	<20	<1	21.3	57.64	<1	14.2	0.4	2.8	0.7	3.6	3.5	2
SD-13	29	29	3	39.4	99.69	<1	21	0.8	3.4	1.6	4.5	10.2	<1
SD-14	5	<20	21	25.1	62.6	3	0.5	<0.1	<0.5	0.4	1.3	11	<1
SD-15	152	36	14	16.5	91.68	6	13.4	10.3	29.8	5	23.3	115.6	5

Appendix 3.

	LF200 Sr PPM 0.5	LF200 Ta PPM 0.1	LF200 Th PPM 0.2	LF200 U PPM 0.1	LF200 V PPM 8	LF200 W PPM 0.5	LF200 Zr PPM 0.1	LF200 Y PPM 0.1	LF200 La PPM 0.1	LF200 Ce PPM 0.1	LF200 Pr PPM 0.02	LF200 Nd PPM 0.3
Section Sample No.	Kasanwala											
DK-T	78.3	<0.1	1.3	1	10	<0.5	5.1	6.1	5.9	15.5	1.95	7.9
DK-F	76.5	<0.1	0.9	2.4	10	<0.5	7	19.1	12.1	26.5	2.85	10.5
DK-F	93.2	<0.1	0.8	2.3	19	<0.5	7.6	20.9	15.4	32.8	3.13	11.4
DK-K	98.4	0.2	1.4	1.7	14	<0.5	16.8	5.6	5.4	9.8	0.99	3.7
KF-1	41.2	0.3	5.1	1	15	2.3	140.6	21.4	26.7	48.5	8.84	41
KF-2	43.1	0.6	13.8	2.1	49	2.1	395.6	12.3	28.4	77.8	6.67	23.2
KF-3	92.3	1.3	41.6	3.5	23	2.4	828.5	56.4	74.8	158.4	15.77	57.5
KF-4	11.5	1.3	18	5.1	17	3.1	1051.2	28.6	5.6	10.8	1.09	4.4
KF-5	56.1	0.7	14.2	1.3	24	2.2	230.4	7.5	54.8	116.4	13.08	39.3
KF-6	32.3	1.5	29.3	3.8	64	2.7	1158.6	28.6	23.9	45.1	4.76	15.9
KF-7	27.4	0.5	13.6	1.2	20	1.7	277	7.9	24.1	56.3	5.4	16.6
KF-8	39.3	1.5	33.9	4	30	2.6	692.7	31.2	51.5	100.6	10.92	37.5
KF-10	25	0.5	8.8	0.9	18	2.1	145.8	6.5	34.5	68.5	7	22.7
KF-11	140.1	2.1	75.2	11.5	37	3.1	778.8	61.5	188.2	888.7	111.35	661.2
KF-13	12.1	1.1	22.9	3.2	28	2.1	420.8	15.6	30.2	57	6.44	21.3
KF-14	48.4	2.9	88.1	15.3	78	3.9	1909	65.6	127.5	251.7	26.68	94.5
KL-1	53.8	3.5	42.9	8.4	266	3.8	715.6	33.7	16.2	62.4	4.39	16.5
KL-2	80.6	2.4	28.4	5	193	2.6	457.8	29.3	8.3	33.6	4.42	21.8
KL-3	120.1	2.5	37.5	6.3	282	3.1	770.8	40.3	63.9	146.7	13.73	57.7
KL-5	125.1	4.2	52	19.5	343	4.3	933.9	42.8	20.3	63.1	7.1	29.3
KL-7	45	4.7	42.5	7	168	5.6	1386.6	45.1	11.6	26	2.32	8
DK-a	77.4	1.1	18.9	2.1	141	2.4	228.1	15.2	18.7	51.1	5.01	17.4
DK-b	12.5	0.3	3.5	0.6	40	1.9	83.2	4.5	7.6	14.4	1.5	5.3
DK-c	24.6	0.4	4	1	30	2.9	227.9	6.5	10.7	20.7	2.24	8.1
DK-d	28.2	0.7	9.1	1.6	35	2.6	525.6	13.8	18.1	33.6	3.62	13.3
DK-E	65.9	1.4	25.5	5.5	664	2.9	410.1	44.6	16.1	30	3.13	12.4
KW-1	91.4	1.1	44.1	6.8	256	1.5	408.1	23.7	22.1	59.1	5.78	20.5
DK-P1	236.5	2	28	2.7	184	3	265.8	24.4	72.7	142.5	14.76	50.4
DK-P2	164.1	0.9	13.8	1.1	93	1.5	131.1	23.6	37.4	74.9	8.23	30.7
DK-P3	181	1.7	22.9	2.6	141	2.7	243.3	30.7	60.4	122.2	12.86	45.3
DK-P2a	135.9	0.9	13.6	1	88	1.3	124.1	22	32.6	68.8	7.89	30.5
Section Sample No.	Kaowaali											
KK-1a	181.3	0.5	34.2	4.9	25	3.8	504.3	77	69.8	173.2	20.95	80.7
KK-2a	90.1	0.5	8.8	1.7	49	3.7	252.2	12	21.3	45.1	5.29	20.9
KK-3a	4.8	0.1	0.7	<0.1	9	3.7	30.5	1.3	2.6	4.8	0.59	1.8
KK-4a	34.6	0.6	5.7	1.3	55	2.2	360.4	7.4	12.6	23.9	2.48	7.8
KK-6a	112.4	0.4	5.6	1.7	130	2.9	302.4	10.5	11.4	25	3.06	12.2
KK-7a	12.7	0.7	5.3	1.3	24	2.9	339.8	12.5	11.5	22	2.57	9.1
DU1	84.9	0.3	6.5	2.2	148	<0.5	54	120.2	95.6	172.4	23.41	93.7
DZ4	98.9	1.9	22.9	6.6	124	3.1	567.8	40.2	52.1	97	10.62	36.8
DZ6	181.9	2.9	38.3	10.1	199	4.3	829.1	53.1	94.8	168.1	18.03	65

Appendix 3.

	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200
	Sr	Ta	Th	U	V	W	Zr	Y	La	Ce	Pr	Nd
	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM
	0.5	0.1	0.2	0.1	8	0.5	0.1	0.1	0.1	0.1	0.02	0.3
DZ7	125.8	0.1	2	1.4	37	<0.5	82.8	3.3	4	7	0.81	2.7
DZ9a	124.1	2	27.2	4.2	155	3.3	447.5	45.9	62	120.2	13.63	49.8
DZ11	126.4	1.7	28.9	3.6	83	2.4	255.4	34.8	114.2	205.5	21.31	71.4
DZ12	91.1	1.7	22.3	3.3	159	3.2	307.6	38.4	53.7	100.5	11.57	41.8
DZ12a	119.9	1.8	23.9	3.9	136	2.7	407.1	36.6	53.9	96.9	10.98	39.2
Section	Nammal											
Sample No.												
KK-1	129.3	1.8	24.2	18.5	108	2.4	345.7	12.1	27.1	41.6	4.14	13.6
KK-2	115	0.9	14.5	9.1	81	1.8	124.8	10.7	24.7	39	3.81	13.3
KK-3	159.9	1.4	18.4	8.7	105	1.9	165.1	16.7	34.4	56.8	6.01	20.8
KK-4	346.6	1.6	22.4	11.2	121	2.7	306	13	30.1	48	4.86	16
DKA-10	224.6	1.9	23.7	1.8	113	2.2	320.7	28.2	55.3	96.5	9.66	31.1
DKA-11	202	1.5	19.9	1.8	102	2	264.7	40.9	48	91.7	9.85	36.4
Dka - 12	250.1	1.9	27.7	2.5	113	2.1	287.7	30.7	67.8	114.3	15.36	61.6
DKA-13	1744.9	1.6	27.5	18.9	131	2.3	633.1	1898.5	790.9	832.6	442.88	2161.1
DN-1a	132.5	<0.1	0.8	4.4	32	<0.5	70.8	72.4	5.3	29.8	6.11	39
DN-1d	134.8	4.5	30.9	15	189	5.2	954.6	60.4	60.9	124.3	12.8	42.7
DN-1j	98	0.5	5.4	1.3	54	1.4	241.8	15.5	8.5	17	2.1	8.4
DN-1	84.5	0.6	11.4	1.5	62	2.5	207.7	10.4	14.6	33.1	3.51	12.8
DN-4	67.3	1.6	20.3	3.7	39	4.1	1051	27.1	32.3	71.7	7.61	28.3
DN-5	21.5	0.1	3.6	0.4	46	1.8	50.4	10.1	5.3	11.1	1.27	5
DN8	68.3	3.1	26.9	10.8	127	3.6	638.4	107.5	108.8	286.7	40.01	179.1
DN8a	143.2	4.8	33.5	13.3	185	5.5	839.5	60.3	56.3	104.6	10.66	35.2
DN-11	121.8	0.4	4.3	0.4	36	0.6	45.9	13.5	14.3	27.5	3.01	11.4
DN12	187.8	2	27.3	3.9	174	3.1	366.4	30.4	69.5	136.9	14.71	49.1
DN-16	100.6	0.2	5	0.4	58	1.2	67.1	14.9	17.4	39.1	4.71	18.4
DN16first	365.8	2	24.4	2.3	183	3.8	260.6	30	80.8	160.5	17.45	62.6
DN17	136.5	1.6	19.3	3.3	94	2.7	592.2	38.7	48.2	98.2	11.58	40.1
DN-19	55.4	0.8	7.5	1.8	32	2.8	311.5	19.7	15.7	37.9	4.11	15.5
DN22	92.1	1.9	20.6	3.1	102	2.9	461.6	39.5	47.5	102.4	10.99	40.7
DN24	117.1	2.2	23	4.4	85	3.3	521.9	44.5	42	76.3	8.63	34.2
DN30	63.3	1.8	17	3.4	86	3.1	457	30.6	37.6	71.9	7.9	29.8
DN-35	195.3	0.3	3.8	0.6	38	0.9	76.2	8.9	10.8	20.7	2.33	9
DN-37	32.4	0.7	9.3	1.5	49	1.7	378.6	19.5	18.8	36.2	3.83	13.7
Section	Zaluch											
Sample No.												
ZK-1	90	0.1	1.6	5.4	32	<0.5	15.2	5.2	4.8	9.3	1.03	3.4
ZK-2	96.1	<0.1	1.3	6.4	50	<0.5	14.3	5.3	4.4	9.2	0.99	3.7
ZS-1	30.3	0.3	5	0.6	14	3	84.2	4.6	12.8	32.2	3.43	13.2
ZS-3	100.5	1	18.8	2.7	20	3	719.2	63.1	51.8	140.2	14.62	57.5
ZS-4	47.3	1.3	22.1	2.4	33	2.1	639.8	25.1	37.1	82.4	9.29	35.6
ZS-6	67	0.9	18.3	1.8	44	1.8	228.4	20.7	54.3	107	11.99	40.7
ZS-8	13.8	2.1	112.1	13.6	44	3.3	2357.7	64.5	146.5	301.2	34.88	130.6
ZS-9	28.4	1.5	29.3	2.8	23	2.1	597.5	21.8	38	75.4	8.57	30.8

Appendix 3.

	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200
	Sr	Ta	Th	U	V	W	Zr	Y	La	Ce	Pr	Nd
	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM
	0.5	0.1	0.2	0.1	8	0.5	0.1	0.1	0.1	0.1	0.02	0.3
ZS-11	21	0.2	4.4	0.3	<8	1.2	70.9	3.9	23.2	49.2	5.66	17.8
ZS-12	24.8	0.3	7.2	0.5	18	3.5	104.5	3.9	24.3	65.5	6.11	23.3
ZS-14	120.8	2	37	6.4	94	2.3	427.8	29.8	97.7	137.7	13.93	35.2
ZL-1	163.1	1.6	19	2.6	99	2.6	112.9	12	110.1	150.1	16.46	43.1
ZL-2	267.4	4	105	4.6	129	3.9	953.9	42.9	39.6	133.9	10.18	39.5
DZ-1	422.8	2.5	43.5	6.9	145	4.1	799.9	40.8	96.5	237.1	25.98	89.8
DZ-2	81.6	0.7	16.6	7	34	2.3	407.2	13.3	28.1	82.2	8.92	33.2
DZ-3	28.2	0.3	2.5	0.5	11	1.1	58.9	4.6	7.6	15.4	1.73	6.3
DZ-4	160.3	2.2	24.9	3.9	155	2.5	414.7	40.4	62	130.1	15.32	56.7
DZ-6	135.3	1.9	25.1	3.9	136	3	417	38.7	56.6	117.4	12.74	47.8
DZ-8	115.4	2.8	30.9	6	129	5	733	39.3	50.2	82.7	9.45	33.1
DZ-9	81.5	4.1	41.1	9.8	116	5	721	80.4	105.5	311.6	41.51	171.5
DZ-10	186.3	0.7	12.7	1.8	79	2.5	101.6	20.4	32.1	68.5	7.35	25.5
DZ-11	266.1	2.1	29.9	3.5	189	3.9	303.2	32.2	74.3	144.8	16.17	56.8
DZ-13	125.4	0.6	8.9	1.5	60	1.6	331.8	14.8	22.5	48.8	5.17	18.3
DZ-14	327	2.3	25.1	3.3	185	3.3	406.1	31.6	66.8	132.6	13.75	46.7
DZ-15	168	1.7	16.9	3	70	2.8	472.6	28.4	36.9	74.2	7.5	25.3
DZ-17	206.7	0.2	3.8	0.8	25	0.7	65.3	16.1	16.6	36.7	4.44	16.9
DZ-18	133.6	2.1	56	4.8	86	2.6	628.3	44	141.1	292.6	33.69	112.7
DZ-18a	175.6	0.2	3.8	0.6	21	1.1	193.4	14.4	11	27.6	3.27	12.2
DZ-20	127.8	1.9	24.7	2.9	171	3.5	335.1	28.8	56.6	119.6	12.01	39
DZ-23	37.8	0.7	8.9	1.9	44	1.8	397.6	19.5	27.2	57	6.18	22.1
DZ-25	24.5	1.1	9.7	2.5	51	2.1	514.9	19.3	9.7	16	1.81	6.2
DZ-26	12.6	0.3	2.8	0.5	16	2	178.9	6.5	6.9	14	1.53	5.8
DZ-27	120.8	0.5	8.4	1.1	69	1.3	151.3	19.5	25.9	57.9	6.35	24.4
Section	Gulla Khel											
Sample No.												
GKK-1	142	0.6	6.9	4.7	51	0.9	66.3	8.2	17.8	32.6	3.45	12.7
GKK-2	166.5	0.6	13.2	17.1	94	2.1	124.5	6.8	23.5	37	3.55	11
GKK-3	189.2	1.8	27	11.4	129	2.5	450.5	16.1	57.5	93.3	8.92	27.4
GKS-1A	162.1	0.8	18.8	5.8	129	1.3	223.7	191.9	41.4	92.6	11.76	52.7
GKS-1B	587.6	1.7	36.1	7.7	201	2.3	345.8	228.3	105.6	205.7	21.78	78.5
GKS-2	203.6	1.1	35	34.4	44	1.1	574.7	23.4	45.6	91.4	10.21	39.7
GKS-4	40.3	0.3	6.4	0.9	13	2.1	124.6	8.3	15.7	34.3	4.03	14.2
GKS-5	145.3	2.3	50.6	10.8	137	2.9	598	63.8	122.1	243.5	26.26	89.8
GKS-6	28.5	1.6	76.8	8.6	44	3.8	1665.6	55.4	111.4	230.8	28.4	106.8
GKS-7	280.9	2.1	50	4	165	4.2	409.5	40.2	124.2	241.6	23.31	76.5
GKS-8	88.3	2.2	61.5	6	55	6.5	1235.9	66.9	96.9	193.1	20.42	71.7
GKS-12	9.8	0.5	16.2	2.1	15	1.9	292.8	10.4	25.4	55.2	6.69	25.2
GKS-16	74.6	1.4	32.8	6.9	52	2.7	582.5	22.1	96.1	198.3	20.16	62.2
GKS-17	116.6	2	51.2	5.4	131	2.2	808.5	37.2	104.6	206.2	21.54	65.1
GKS-18	74	1.8	29.5	2.3	106	1.8	426	18.5	40.5	83.5	8.28	27.8
GKS-21	91.2	1.7	31.3	3.6	179	2.4	316.5	15.4	61.7	113.9	8.27	19.3
GKS-22	70.5	1.9	32.3	3.2	65	1.9	859.3	22.4	29.8	78.9	3.36	9.8

Appendix 3.

	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200
	Sr	Ta	Th	U	V	W	Zr	Y	La	Ce	Pr	Nd
	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM
	0.5	0.1	0.2	0.1	8	0.5	0.1	0.1	0.1	0.1	0.02	0.3
GKS-23	162.5	2.8	26.8	6.4	138	4	644.4	42.8	45.2	88.1	9.96	34.8
GKS-26	27.2	0.7	8.1	1.5	37	2	455.4	11.7	11.6	24.3	2.75	9.8
GKS-27	147.1	0.4	6.6	0.4	58	0.9	81.3	21.8	27.4	60.4	6.57	24.6
GKS-29	89.3	2	31.1	3.4	180	3.4	334.1	42.5	86.1	171.9	19.36	71.6
GKS-30	143	0.4	6.2	3.1	38	<0.5	67.9	13	20.2	41.8	5.31	20.5
Section	Paniala											
Sample No.												
KS-4	130	1.7	25.4	18	108	2	454.3	21.1	40.7	87.2	9.87	34.7
KS-6	173.1	0.9	23.5	7.1	84	2	492.6	24.8	51	101.8	10.86	37.9
KS-10	242.9	1	22.4	4.2	86	1.6	154.1	25.7	60.2	119	12.57	44.2
KS-11a	395.8	1.7	37.7	5.5	105	2.6	208.1	251.7	107.5	213.8	22.71	83.1
ASH	333	1.3	24.2	5.8	102	1.8	175.6	97.9	84.7	180.8	22.68	113.1
SD-a	89.5	<0.1	0.4	2.5	<8	<0.5	10.3	31.2	2.6	5.5	0.67	3
SD-b	65.8	0.2	0.5	3.3	13	<0.5	14.7	26	2.8	4.8	0.65	4
SD-c	383.8	2.5	48.8	10.3	121	2.8	734.3	56.6	106.9	231.6	25.29	91.7
SD-1	48.3	1.1	31.6	3.1	24	1.5	763.5	21.4	50.1	102.4	11.33	39.6
SD-2	173.8	2.3	44.6	4	114	2.7	359.4	26.6	89.9	174.4	18	60.5
SD-3	158.5	2	36.7	4.7	126	2.8	250.7	22.6	77	127.9	13.31	44.4
SD-4	51.4	0.6	20.2	1.7	31	1.7	284.5	27.3	30.3	75.5	9.82	39.7
SD-5	59.1	1.9	44	3.9	47	2.1	1055.3	36.5	45.2	89.4	10.66	41.1
SD-6	34.1	1	40.8	5.2	64	1.6	712	27.6	34.8	76.6	8.3	32.4
SD-7	101.9	2.4	39.9	11	81	2.7	992.5	61	95.8	196.2	21.79	75.4
SD-8	84.9	1.8	55.2	3.9	269	2.4	515.9	25.6	25.1	42	4.47	14.5
SD-9	23.9	0.3	4.9	2	84	1.6	112.6	3.3	8.1	17.5	2.15	7.4
SD-10	35.7	0.3	6.2	0.7	24	1.9	113.3	5.3	8.7	17.8	2.1	7.7
SD-11	170.7	<0.1	2	0.9	25	<0.5	59.2	22.8	9	18.3	2.27	8.7
SD-12	91	0.2	0.6	0.4	<8	<0.5	26.4	9.5	3	6.3	0.87	3.5
SD-13	141.4	0.4	4.1	1.1	32	0.6	58.9	9.5	11.1	20.9	2.47	9.1
SD-14	191.4	0.1	0.9	0.6	21	<0.5	24.2	3.1	6.6	15.3	1.31	4.6
SD-15	231.3	1.6	21.7	2.4	134	2.2	181.5	23.9	52	100	11.41	42.7

Appendix 3.

	LF200 Sm PPM 0.05	LF200 Eu PPM 0.02	LF200 Gd PPM 0.05	LF200 Tb PPM 0.01	LF200 Dy PPM 0.05	LF200 Ho PPM 0.02	LF200 Er PPM 0.03	LF200 Tm PPM 0.01	LF200 Yb PPM 0.05	LF200 Lu PPM 0.01	TC000 TOT/C %	TC000 TOT/S %
Section Sample No.	Kasanwal											
DK-T	1.48	0.26	1.07	0.15	0.77	0.16	0.44	0.06	0.29	0.05	13.12	<0.02
DK-F	1.52	0.33	1.67	0.25	1.55	0.35	1.05	0.14	0.62	0.09	12.71	<0.02
DK-F	1.75	0.34	1.79	0.26	1.59	0.38	1.1	0.13	0.68	0.09	12.45	<0.02
DK-K	0.61	0.14	0.74	0.11	0.74	0.15	0.39	0.06	0.31	0.05	12.26	<0.02
KF-1	5.94	1.01	4.57	0.71	4.09	0.81	2.43	0.34	2.14	0.33	0.12	<0.02
KF-2	4.4	0.62	3.45	0.48	2.34	0.47	1.51	0.24	1.74	0.27	0.06	<0.02
KF-3	10.76	1.89	10.78	1.85	11.31	2.19	6.31	0.87	5.87	0.86	0.03	<0.02
KF-4	1.73	0.49	3.48	0.71	4.88	1.05	3.41	0.53	3.7	0.6	0.03	<0.02
KF-5	4.73	0.63	2.7	0.33	1.7	0.29	0.9	0.14	1.01	0.15	0.03	<0.02
KF-6	3.69	0.72	4.11	0.77	4.86	1.14	3.49	0.55	3.94	0.64	0.03	<0.02
KF-7	3.05	0.4	2.26	0.32	1.66	0.31	1	0.16	1.18	0.18	0.03	<0.02
KF-8	6.5	0.91	5.82	0.92	5.48	1.13	3.41	0.53	3.83	0.59	0.04	<0.02
KF-10	2.62	0.4	1.74	0.24	1.33	0.24	0.72	0.11	0.78	0.12	<0.02	<0.02
KF-11	206.4	29.44	84.11	6.82	21.51	2.15	5.93	1.02	7.24	1.01	0.05	<0.02
KF-13	3.78	0.47	3.26	0.52	2.91	0.56	1.95	0.3	2.08	0.31	0.04	<0.02
KF-14	17.9	2.01	15.17	2.24	12.95	2.54	7.53	1.17	8.21	1.26	0.07	<0.02
KL-1	3.23	0.63	3.72	0.76	5.5	1.25	4.06	0.68	4.7	0.74	0.05	0.06
KL-2	5.19	0.9	4.94	0.8	5.08	1.13	3.52	0.55	3.71	0.58	0.04	<0.02
KL-3	10.75	1.74	8.02	1.15	7.05	1.44	4.28	0.71	4.59	0.73	0.02	<0.02
KL-5	6.15	1.11	5.97	1.08	7.46	1.6	5.29	0.85	5.96	0.95	0.1	0.02
KL-7	2.08	0.5	3.48	0.81	6.33	1.47	4.94	0.81	5.6	0.86	0.03	<0.02
DK-a	2.83	0.48	2.3	0.41	2.69	0.57	1.66	0.26	1.7	0.27	0.05	0.07
DK-b	0.97	0.12	0.76	0.13	0.73	0.16	0.54	0.07	0.51	0.08	0.02	<0.02
DK-c	1.3	0.16	1.15	0.17	1.06	0.22	0.69	0.1	0.69	0.12	0.21	0.15
DK-d	2.26	0.28	2.13	0.34	2.15	0.46	1.45	0.23	1.59	0.24	0.06	0.24
DK-E	4.32	1.12	7.43	1.23	7.21	1.42	3.85	0.49	3.19	0.48	0.1	<0.02
KW-1	3.52	0.72	3.69	0.66	4.29	0.88	2.68	0.41	2.69	0.42	0.17	0.82
DK-P1	7.62	1.25	4.99	0.76	4.67	0.94	2.86	0.46	3	0.46	0.31	0.03
DK-P2	5.91	1.13	5.21	0.78	4.62	0.9	2.52	0.4	2.48	0.38	6.22	0.03
DK-P3	7.65	1.39	6.18	0.96	5.8	1.15	3.34	0.51	3.36	0.52	1.99	<0.02
DK-P2a	5.77	1.1	5.22	0.76	4.42	0.92	2.59	0.36	2.38	0.34	6.39	0.03
Section Sample No.	Kaowaali											
KK-1a	17.51	3.63	16.65	2.45	13.62	2.52	6.88	0.95	5.85	0.9	0.14	0.03
KK-2a	3.91	0.72	3.19	0.42	2.33	0.44	1.24	0.2	1.46	0.22	0.1	0.1
KK-3a	0.2	0.03	0.18	0.03	0.19	0.02	0.12	0.02	0.13	0.02	0.04	<0.02
KK-4a	1.33	0.2	1.05	0.17	1.25	0.26	0.83	0.14	0.98	0.18	0.06	0.03
KK-6a	2.7	0.51	2.26	0.34	2.02	0.39	1.16	0.19	1.35	0.19	1.01	0.09
KK-7a	1.45	0.24	1.53	0.28	1.87	0.4	1.38	0.22	1.35	0.23	0.09	<0.02
DU1	18.58	4.26	19.55	2.76	15.06	2.81	7.79	0.98	5.43	0.82	0.55	<0.02
DZ4	5.54	1.12	5.57	1.02	6.8	1.38	4.41	0.67	4.39	0.68	0.3	0.06
DZ6	10.55	1.72	8.69	1.47	9.67	1.9	5.71	0.85	5.91	0.91	0.38	0.08

Appendix 3.

	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	TC000	TC000
	Sm	Eu	Gd	Tb	Dy	Ho	Er	Tm	Yb	Lu	TOT/C	TOT/S
	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	%	%
	0.05	0.02	0.05	0.01	0.05	0.02	0.03	0.01	0.05	0.01	0.02	0.02
DZ7	0.54	0.08	0.58	0.1	0.73	0.1	0.32	0.05	0.34	0.05	62.97	0.97
DZ9a	9.18	1.79	8.14	1.38	8.29	1.52	4.83	0.67	4.75	0.71	0.85	0.08
DZ11	10.57	1.78	7.64	1.23	7.41	1.44	4.2	0.65	4.51	0.66	0.07	0.02
DZ12	7.68	1.52	7.13	1.14	6.76	1.4	3.97	0.62	4.07	0.61	0.41	0.02
DZ12a	7.06	1.33	6.11	1.03	6.09	1.24	3.66	0.59	3.92	0.6	0.39	0.06
Section	Nammal											
Sample No.												
KK-1	2.21	0.35	1.89	0.29	1.9	0.41	1.5	0.24	1.79	0.3	4.94	0.78
KK-2	2.1	0.42	1.97	0.3	1.82	0.39	1.13	0.18	1.23	0.21	6.03	0.49
KK-3	3.77	0.7	3.31	0.5	2.95	0.63	1.73	0.28	1.73	0.27	1.26	0.71
KK-4	2.73	0.47	2.32	0.36	2.26	0.5	1.58	0.27	2.01	0.33	1.74	1
DKA-10	4.01	0.7	4.18	0.59	3.45	0.82	2.71	0.45	3.25	0.5	<0.02	<0.02
DKA-11	6.61	1.48	9.51	1.24	6.67	1.32	3.93	0.54	3.55	0.54	<0.02	<0.02
Dka - 12	12.98	2.56	12.43	1.4	6.63	1.16	3.43	0.49	3.26	0.54	0.04	0.03
DKA-13	487.19	96.04	458.24	66.43	365.23	71.5	201.03	27.13	162.11	23.73	0.05	0.08
DN-1a	17.59	5.1	24.71	3.32	15.37	2.45	5.1	0.53	2.57	0.37	34	5.32
DN-1d	7.43	1.46	7.63	1.43	9.93	2.17	6.72	1.1	7.15	1.15	2.04	0.09
DN-1j	2.14	0.47	2.66	0.45	2.83	0.63	1.79	0.27	1.83	0.28	0.23	0.61
DN-1	2.38	0.37	1.91	0.3	1.91	0.38	1.21	0.18	1.26	0.2	2.3	0.25
DN-4	4.85	0.53	4.31	0.65	4.32	0.96	2.99	0.48	3.39	0.56	0.88	0.06
DN-5	1.21	0.29	1.81	0.33	1.97	0.41	1.09	0.17	1.15	0.18	0.43	0.65
DN8	40.9	8.23	36.83	4.77	23.1	3.98	10.97	1.56	10.44	1.54	3.78	0.14
DN8a	6.19	1.44	7	1.36	9.55	2.01	6.3	1.01	6.64	1.09	1.28	0.14
DN-11	2.04	0.45	2.35	0.34	2.11	0.42	1.16	0.17	1.04	0.15	11.6	0.17
DN12	6.98	1.23	5.44	0.94	5.51	1.15	3.56	0.56	3.89	0.58	0.91	1.73
DN-16	3.69	0.73	3.4	0.51	2.82	0.56	1.49	0.21	1.32	0.2	6.49	0.63
DN16first	8.69	1.48	5.77	0.91	5.26	1.1	3.48	0.55	3.65	0.58	0.52	0.48
DN17	7.13	1.31	6.62	1.08	6.64	1.43	4.42	0.69	4.85	0.73	0.42	0.13
DN-19	3.24	0.67	3.2	0.53	3.66	0.68	2.16	0.32	2.16	0.34	2.95	0.13
DN22	7.18	1.24	6.32	1.07	6.83	1.39	4.3	0.63	4.59	0.69	0.21	0.12
DN24	7.39	1.54	7.76	1.35	8.5	1.67	5.26	0.76	4.93	0.78	1.19	1.98
DN30	4.94	0.98	4.74	0.86	5.45	1.25	3.41	0.56	3.79	0.56	0.57	0.9
DN-35	1.69	0.35	1.6	0.25	1.55	0.29	0.93	0.13	0.8	0.12	12.16	0.06
DN-37	2.31	0.46	2.69	0.46	3.06	0.71	2.13	0.32	2.26	0.37	1.73	0.06
Section	Zaluch											
Sample No.												
ZK-1	0.75	0.17	0.73	0.12	0.8	0.16	0.41	0.06	0.46	0.07	11.96	<0.02
ZK-2	0.67	0.13	0.74	0.12	0.77	0.15	0.48	0.07	0.47	0.07	12	<0.02
ZS-1	2	0.34	1.55	0.16	0.86	0.16	0.43	0.07	0.46	0.07	0.11	<0.02
ZS-3	9.62	1.9	10.48	1.55	9.45	1.91	5.54	0.73	4.3	0.7	0.04	<0.02
ZS-4	6.13	1.09	5.27	0.79	5.16	1.01	3.07	0.51	3.55	0.58	0.02	<0.02
ZS-6	7.07	1.16	5.57	0.78	4.68	0.87	2.42	0.39	2.54	0.41	0.08	<0.02
ZS-8	24.37	1.73	19.49	2.35	13.36	2.29	6.73	1.13	7.62	1.18	0.12	<0.02
ZS-9	5.32	0.67	4.45	0.71	4.42	0.85	2.71	0.44	3.07	0.49	0.02	<0.02

Appendix 3.

	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	TC000	TC000
	Sm	Eu	Gd	Tb	Dy	Ho	Er	Tm	Yb	Lu	TOT/C	TOT/S
	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	%	%
	0.05	0.02	0.05	0.01	0.05	0.02	0.03	0.01	0.05	0.01	0.02	0.02
ZS-11	2.01	0.23	1.07	0.14	0.87	0.13	0.44	0.07	0.49	0.07	<0.02	<0.02
ZS-12	3.03	0.4	1.52	0.17	0.97	0.16	0.42	0.07	0.49	0.07	0.02	<0.02
ZS-14	4.88	0.9	4.73	0.81	5.68	1.16	3.41	0.54	3.76	0.56	0.04	<0.02
ZL-1	5.13	1.01	3.89	0.55	3.32	0.57	1.83	0.28	1.89	0.3	0.02	<0.02
ZL-2	8.75	1.49	7.74	1.14	7.43	1.63	5.14	0.8	5.5	0.88	0.08	<0.02
DZ-1	11.47	1.71	7.89	1.15	7.62	1.51	4.81	0.8	5.76	0.86	0.11	0.35
DZ-2	6.34	1.11	6.46	0.89	3.5	0.61	1.76	0.23	1.57	0.27	0.08	0.05
DZ-3	1.11	0.21	0.95	0.13	0.74	0.15	0.47	0.06	0.44	0.06	0.04	0.03
DZ-4	10.58	2.02	8.72	1.34	7.87	1.51	4.65	0.7	4.47	0.71	0.3	<0.02
DZ-6	8.99	1.75	8.51	1.17	7.16	1.41	4.16	0.67	4.25	0.67	0.3	0.06
DZ-8	5.94	1.08	5.11	0.93	6.25	1.45	4.4	0.64	4.46	0.72	0.21	0.03
DZ-9	36.17	6.83	28.14	3.41	17.41	2.84	8.23	1.23	7.95	1.2	0.1	<0.02
DZ-10	4.81	0.88	4.12	0.61	3.86	0.77	2.17	0.36	2.25	0.36	6.11	0.02
DZ-11	9.14	1.69	6.74	1.09	6.46	1.39	4.09	0.64	4.17	0.64	0.26	<0.02
DZ-13	3.35	0.53	2.83	0.48	2.78	0.58	1.8	0.26	1.76	0.29	4.04	<0.02
DZ-14	7.71	1.32	5.35	0.93	5.51	1.2	3.77	0.63	4.16	0.63	0.47	0.04
DZ-15	4.28	0.71	3.87	0.67	4.62	1	3.14	0.49	3.46	0.52	0.06	0.15
DZ-17	3.35	0.65	3.19	0.52	2.78	0.59	1.78	0.23	1.47	0.26	9.84	<0.02
DZ-18	15.34	2.15	10.12	1.56	8.44	1.69	5.02	0.77	5.26	0.81	0.08	<0.02
DZ-18a	2.52	0.53	2.61	0.41	2.52	0.5	1.41	0.19	1.38	0.19	3.74	<0.02
DZ-20	6.23	1.12	5.28	0.82	5.51	1.1	3.39	0.54	3.67	0.58	0.15	<0.02
DZ-23	3.84	0.66	3.34	0.56	3.26	0.71	2.18	0.36	2.44	0.38	0.08	<0.02
DZ-25	1.37	0.3	1.92	0.43	3.05	0.73	2.39	0.37	2.57	0.41	0.12	<0.02
DZ-26	1.05	0.17	0.96	0.15	1.03	0.21	0.71	0.11	0.77	0.11	0.05	<0.02
DZ-27	4.27	0.82	4.13	0.59	3.6	0.7	2.03	0.29	1.82	0.27	7.55	0.03
Section	Gulla Khel											
Sample No.												
GKK-1	2.09	0.42	1.88	0.27	1.61	0.3	0.94	0.14	0.9	0.14	9.13	0.28
GKK-2	1.68	0.27	1.25	0.19	1.15	0.23	0.69	0.11	0.83	0.13	7.35	0.74
GKK-3	3.8	0.57	2.82	0.4	2.54	0.54	1.89	0.32	2.29	0.38	0.25	0.25
GKS-1A	17.81	5.44	43.98	6.45	36.25	6.98	18.62	2.46	13.91	2.15	0.06	2.02
GKS-1B	21.24	6.55	53.2	8.29	47.89	8.9	23.09	2.92	15.5	2.48	0.08	2.67
GKS-2	7.02	0.91	5.48	0.72	4.2	0.81	2.26	0.35	2.39	0.38	29.73	1.67
GKS-4	1.87	0.26	1.49	0.23	1.38	0.29	0.85	0.12	0.79	0.13	0.03	0.27
GKS-5	14.01	2.28	12.21	1.94	11.5	2.39	6.94	1.03	6.77	1.09	0.95	0.4
GKS-6	19.81	1.92	15.57	2.05	10.96	2.13	6.07	0.91	5.9	0.93	0.03	<0.02
GKS-7	11.35	1.81	9.23	1.23	6.99	1.23	3.53	0.51	3.32	0.48	0.11	0.23
GKS-8	12.05	1.53	11.36	1.7	10.42	2.1	6.26	0.92	5.87	0.91	0.04	<0.02
GKS-12	4.81	0.58	3.88	0.5	2.47	0.45	1.28	0.18	1.17	0.19	0.02	<0.02
GKS-16	7.13	0.93	4.71	0.74	4.44	0.91	2.81	0.45	3.07	0.5	0.03	<0.02
GKS-17	8.05	1.28	6.71	1.17	7.22	1.6	5.23	0.81	5.61	0.87	0.07	<0.02
GKS-18	4.33	0.69	3.11	0.53	3.21	0.68	2.18	0.36	2.4	0.38	0.05	<0.02
GKS-21	2.73	0.52	2.38	0.45	3.33	0.65	2.13	0.36	2.49	0.37	0.04	<0.02
GKS-22	2.13	0.41	2.36	0.49	3.94	0.88	3.05	0.5	3.53	0.57	0.03	<0.02

Appendix 3.

	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	LF200	TC000	TC000
	Sm	Eu	Gd	Tb	Dy	Ho	Er	Tm	Yb	Lu	TOT/C	TOT/S
	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	%	%
	0.05	0.02	0.05	0.01	0.05	0.02	0.03	0.01	0.05	0.01	0.02	0.02
GKS-23	5.96	1.18	6.07	1.1	7.7	1.52	4.82	0.77	4.95	0.78	1.76	0.95
GKS-26	2.14	0.26	1.86	0.31	1.96	0.43	1.37	0.21	1.37	0.21	0.04	0.17
GKS-27	4.66	0.93	4.28	0.69	4.46	0.85	2.53	0.4	2.56	0.37	10.62	0.23
GKS-29	13.25	2.6	11.18	1.61	9.35	1.79	5.21	0.78	5.07	0.76	0.44	0.39
GKS-30	3.73	0.69	3.22	0.45	2.66	0.47	1.4	0.19	1.22	0.19	10.9	0.05
Section	Paniala											
Sample No.												
KS-4	6.16	1.07	4.91	0.7	4.33	0.8	2.35	0.37	2.3	0.36	1.61	0.04
KS-6	6.72	1.14	5.72	0.76	4.69	0.87	2.41	0.36	2.34	0.35	4.78	0.03
KS-10	7.11	1.31	6.22	0.8	4.84	0.83	2.34	0.33	2.18	0.33	4.79	0.95
KS-11a	19.77	5.04	37.98	6.18	41.42	8.74	24.96	3.2	17.68	2.84	0.45	4.11
ASH	32.81	6.68	35.7	4.26	22.77	3.99	10.86	1.39	8.13	1.32	0.27	3.5
SD-a	1.36	0.54	5.3	0.76	4.18	0.84	2.01	0.22	1.13	0.18	11.97	0.16
SD-b	4.05	1.74	11.29	1.86	9.97	1.59	3.9	0.51	3	0.45	0.04	13.44
SD-c	15.31	2.48	12.28	1.59	10.03	2.02	6.41	1	6.81	1.08	0.23	0.08
SD-1	6.44	0.61	4.91	0.62	3.54	0.7	1.96	0.31	2.27	0.35	<0.02	<0.02
SD-2	8.37	1.38	6	0.81	5.23	0.94	2.9	0.44	3.06	0.49	0.1	0.02
SD-3	7.35	1.28	5.38	0.79	4.51	0.88	2.54	0.39	2.73	0.4	0.12	0.08
SD-4	8	1.32	7.78	1.07	5.64	0.97	2.55	0.32	1.96	0.28	4.11	0.11
SD-5	8.48	1.22	7.39	1.11	7.23	1.43	4.75	0.72	4.77	0.79	0.02	0.03
SD-6	7.15	0.99	6.03	0.9	5.8	1.14	3.61	0.56	3.88	0.6	0.09	0.08
SD-7	12.58	1.79	10.86	1.7	10.51	2.19	7.08	1.05	7.31	1.1	0.14	<0.02
SD-8	2.68	0.54	3.26	0.61	4.51	0.97	2.82	0.46	3.07	0.49	0.04	<0.02
SD-9	1.01	0.15	0.68	0.1	0.58	0.12	0.31	0.05	0.4	0.06	0.06	0.06
SD-10	1.48	0.2	1.22	0.16	0.9	0.18	0.56	0.08	0.54	0.09	0.06	0.11
SD-11	1.94	0.39	2.5	0.41	3.12	0.6	1.55	0.2	1.08	0.16	8.44	2.3
SD-12	0.75	0.18	1.03	0.22	1.55	0.36	1.04	0.14	0.72	0.11	0.12	17.33
SD-13	1.68	0.35	1.69	0.26	1.5	0.31	1.03	0.13	0.85	0.12	10.76	0.6
SD-14	0.73	0.15	0.84	0.13	0.83	0.15	0.43	0.06	0.34	0.06	0.04	15.96
SD-15	8.08	1.59	6.7	0.97	5.17	0.93	2.66	0.42	2.69	0.38	0.18	3.88

Appendix 3.

	AQ200	AQ200	AQ200	AQ200	AQ200	AQ200	AQ200	AQ200	AQ200	AQ200	AQ200
	Mo	Cu	Pb	Zn	Ni	As	Cd	Sb	Bi	Ag	Au
	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPB
	0.1	0.1	0.1	1	0.1	0.5	0.1	0.1	0.1	0.1	0.5
Section	Kasanwala										
Sample No.											
DK-T	0.1	1.2	0.9	18	4.9	<0.5	0.2	<0.1	<0.1	<0.1	<0.5
DK-F	0.3	1.6	1.2	41	10.9	1.1	0.6	<0.1	<0.1	<0.1	<0.5
DK-F	0.3	1.4	1.4	40	11.4	0.7	0.5	<0.1	<0.1	<0.1	1.1
DK-K	<0.1	0.5	1.2	32	9.2	<0.5	0.3	<0.1	<0.1	<0.1	1.2
KF-1	2.4	5.6	1.9	6	8.5	<0.5	<0.1	<0.1	<0.1	<0.1	<0.5
KF-2	1.6	9.9	2.1	8	8.1	<0.5	<0.1	<0.1	<0.1	<0.1	0.7
KF-3	0.6	3.3	4.3	3	5.4	<0.5	<0.1	<0.1	<0.1	<0.1	<0.5
KF-4	1.6	11.8	7.8	19	11.6	<0.5	<0.1	<0.1	<0.1	<0.1	<0.5
KF-5	1.3	9.1	3	5	7.1	<0.5	<0.1	<0.1	<0.1	<0.1	<0.5
KF-6	0.3	4.3	4.4	4	5.4	<0.5	<0.1	<0.1	<0.1	<0.1	<0.5
KF-7	1.1	13.3	1.7	6	8.2	<0.5	<0.1	<0.1	<0.1	<0.1	<0.5
KF-8	0.8	11.3	5.5	4	6.1	<0.5	<0.1	<0.1	<0.1	<0.1	<0.5
KF-10	1.8	7.6	1.6	4	8.2	<0.5	<0.1	<0.1	<0.1	<0.1	<0.5
KF-11	0.6	2.8	7	2	5	<0.5	<0.1	<0.1	<0.1	<0.1	1.2
KF-13	0.8	8.1	3.6	2	3.8	<0.5	<0.1	<0.1	<0.1	<0.1	<0.5
KF-14	1.4	94.1	13.7	97	62.8	<0.5	0.4	<0.1	<0.1	<0.1	0.7
KL-1	1.4	22.4	25.3	14	15.4	<0.5	<0.1	0.2	0.5	<0.1	<0.5
KL-2	0.7	21.8	11.3	5	15.3	<0.5	<0.1	<0.1	0.3	<0.1	0.6
KL-3	1.5	23.5	21.3	9	7.9	1.9	<0.1	0.4	0.4	<0.1	<0.5
KL-5	1.3	266	7.8	38	29	<0.5	0.2	<0.1	0.8	<0.1	0.9
KL-7	<0.1	2.3	3.3	<1	1.2	<0.5	<0.1	0.1	0.6	<0.1	<0.5
DK-a	1.7	7.6	13.2	2	5.8	5	<0.1	0.1	0.3	<0.1	<0.5
DK-b	2.5	14.4	4.7	4	8.4	0.8	<0.1	<0.1	<0.1	<0.1	<0.5
DK-c	4.3	14.7	11.4	6	9.6	3.4	<0.1	0.1	<0.1	<0.1	<0.5
DK-d	2.2	11.8	13.2	3	6	<0.5	<0.1	<0.1	<0.1	<0.1	<0.5
DK-E	2.4	10	59.7	16	7.9	15.6	<0.1	1.2	1.3	<0.1	1.4
KW-1	17.2	85.9	97.9	37	2.8	297.1	0.3	3.5	0.3	0.3	1.8
DK-P1	0.1	14.4	13.3	6	6.7	1.8	<0.1	<0.1	0.6	<0.1	<0.5
DK-P2	0.1	16.6	10.9	11	13.3	5.4	<0.1	<0.1	0.3	<0.1	<0.5
DK-P3	0.1	12.5	14.3	10	15.8	6.1	<0.1	<0.1	0.6	<0.1	0.7
DK-P2a	<0.1	16.3	10.4	9	13.4	5	<0.1	<0.1	0.4	<0.1	<0.5
Section	Kaowaali										
Sample No.											
KK-1a	5.9	7.7	8.1	45	22.1	12.7	0.2	0.2	0.1	<0.1	<0.5
KK-2a	4.2	9.1	8.2	29	18.4	3.3	0.1	<0.1	0.1	<0.1	0.6
KK-3a	6.1	5.1	1.9	6	12.3	0.9	<0.1	<0.1	<0.1	<0.1	0.5
KK-4a	3.2	3.5	2.5	2	6.5	0.9	<0.1	<0.1	<0.1	<0.1	1.3
KK-6a	4.1	9.6	7.3	6	9.9	6.2	<0.1	<0.1	<0.1	<0.1	1
KK-7a	2.9	7	2.7	5	7.2	<0.5	<0.1	<0.1	<0.1	<0.1	0.6
DU1	0.3	3.2	3.4	19	9	11.6	<0.1	<0.1	<0.1	<0.1	<0.5
DZ4	<0.1	7.8	2.4	5	0.8	<0.5	<0.1	<0.1	0.4	<0.1	1
DZ6	<0.1	5.5	2.7	6	1.2	<0.5	<0.1	<0.1	0.7	<0.1	0.6

Appendix 3.

	AQ200	AQ200	AQ200	AQ200	AQ200	AQ200	AQ200	AQ200	AQ200	AQ200	AQ200
	Mo	Cu	Pb	Zn	Ni	As	Cd	Sb	Bi	Ag	Au
	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPB
	0.1	0.1	0.1	1	0.1	0.5	0.1	0.1	0.1	0.1	0.5
DZ7	0.4	15.5	20.4	6	8.2	0.6	<0.1	0.2	0.1	<0.1	0.9
DZ9a	0.1	30.7	16.4	27	23.9	<0.5	0.4	0.1	0.7	<0.1	<0.5
DZ11	0.1	12.8	10.7	4	4.4	<0.5	<0.1	<0.1	<0.1	<0.1	1.1
DZ12	<0.1	55.4	30.4	34	21.4	2	0.2	0.1	0.6	<0.1	<0.5
DZ12a	0.1	26.1	10.7	10	5.7	0.8	<0.1	0.1	0.6	<0.1	<0.5
Section	Nammal										
Sample No.											
KK-1	8.6	42.9	27.6	13	22.6	15.5	<0.1	0.3	0.4	<0.1	<0.5
KK-2	2.7	21.6	14.4	10	18.3	10	<0.1	<0.1	0.2	<0.1	0.6
KK-3	2.6	19.2	16.1	11	22.6	11	<0.1	<0.1	0.3	<0.1	<0.5
KK-4	4.4	19.1	22.3	11	17.6	16.6	<0.1	<0.1	0.3	<0.1	<0.5
DKA-10	0.3	6.3	11.6	39	100	<0.5	0.2	<0.1	0.3	<0.1	<0.5
DKA-11	<0.1	11.3	21.5	54	84.2	<0.5	0.1	<0.1	0.3	<0.1	<0.5
Dka - 12	0.2	37.7	13.9	188	58	<0.5	<0.1	<0.1	0.3	<0.1	2.8
DKA-13	0.6	80.8	11.1	172	26.4	<0.5	<0.1	<0.1	0.3	<0.1	1.7
DN-1a	8.9	4.1	216.5	9	3.4	5.3	0.1	1.5	<0.1	0.2	1.5
DN-1d	0.1	52.5	21.1	6	9	1.7	<0.1	0.1	0.7	0.3	0.7
DN-1j	0.9	8.1	4.3	19	21.7	1.7	<0.1	<0.1	<0.1	<0.1	<0.5
DN-1	2.8	6.7	14.6	3	7.4	0.8	<0.1	0.1	0.2	<0.1	<0.5
DN-4	3.1	6.6	7.6	7	9.9	<0.5	0.2	<0.1	<0.1	<0.1	1.6
DN-5	2.6	3.4	2.7	44	24.9	1.6	<0.1	<0.1	<0.1	<0.1	<0.5
DN8	0.6	83	13.5	40	20.4	<0.5	0.5	0.2	0.7	0.1	1.2
DN8a	0.1	45.4	18.2	8	7.3	2.6	<0.1	0.1	0.9	0.2	2.1
DN-11	0.2	4.4	5.3	35	6.2	3.2	<0.1	<0.1	0.1	<0.1	<0.5
DN12	0.2	8.2	10.6	34	6.2	5.7	<0.1	<0.1	0.7	<0.1	<0.5
DN-16	1.1	4.1	11.2	11	15.8	15.9	<0.1	<0.1	0.1	<0.1	<0.5
DN16first	<0.1	29.1	11	19	10.8	4.8	<0.1	<0.1	0.7	<0.1	0.7
DN17	<0.1	12.9	7.8	31	3.3	5.2	<0.1	<0.1	0.4	<0.1	0.5
DN-19	1.5	5.3	3.2	50	8.3	<0.5	<0.1	<0.1	<0.1	<0.1	<0.5
DN22	<0.1	9	4.3	13	1.8	2.7	<0.1	<0.1	0.4	<0.1	<0.5
DN24	0.4	16.9	11.8	117	52.3	0.7	0.2	<0.1	0.3	<0.1	<0.5
DN30	0.2	7.5	13.8	10	1.7	1.2	<0.1	<0.1	0.3	<0.1	<0.5
DN-35	0.4	3.6	5.1	18	5.5	1.4	<0.1	<0.1	<0.1	<0.1	<0.5
DN-37	0.7	5.4	1.4	17	5.2	<0.5	0.2	<0.1	0.1	<0.1	<0.5
Section	Zaluch										
Sample No.											
ZK-1	0.4	18.7	1.3	23	14.4	<0.5	0.6	<0.1	<0.1	<0.1	1.7
ZK-2	0.4	104	1.3	22	19.4	<0.5	0.8	<0.1	<0.1	<0.1	0.6
ZS-1	4.3	6.5	1.3	3	10.8	<0.5	<0.1	<0.1	<0.1	<0.1	1.7
ZS-3	1.4	3.9	3.4	2	6.2	<0.5	<0.1	<0.1	<0.1	<0.1	1.1
ZS-4	0.5	4.4	2.6	1	4.4	<0.5	<0.1	<0.1	<0.1	<0.1	1.4
ZS-6	0.5	3.7	6.2	2	6.2	<0.5	<0.1	<0.1	<0.1	<0.1	0.8
ZS-8	1	7.8	7.1	14	13.4	<0.5	<0.1	<0.1	0.3	<0.1	0.7
ZS-9	0.6	9.2	5	3	8.8	<0.5	<0.1	<0.1	<0.1	<0.1	<0.5

Appendix 3.

	AQ200	AQ200	AQ200	AQ200	AQ200	AQ200	AQ200	AQ200	AQ200	AQ200	AQ200
	Mo	Cu	Pb	Zn	Ni	As	Cd	Sb	Bi	Ag	Au
	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPB
	0.1	0.1	0.1	1	0.1	0.5	0.1	0.1	0.1	0.1	0.5
ZS-11	1.4	4.7	0.9	2	5.2	<0.5	<0.1	<0.1	<0.1	<0.1	1.1
ZS-12	4.1	8.3	2.4	3	11.4	1	<0.1	<0.1	<0.1	<0.1	0.8
ZS-14	<0.1	23.2	5.2	2	5.8	<0.5	<0.1	<0.1	<0.1	<0.1	1.2
ZL-1	0.2	8.7	7.4	2	1.6	<0.5	<0.1	<0.1	0.1	<0.1	0.8
ZL-2	<0.1	43.5	30	3	8.5	12.1	<0.1	0.3	1.4	<0.1	1.1
DZ-1	0.2	4.4	32.1	4	2	0.7	<0.1	<0.1	0.7	<0.1	1
DZ-2	3	10.7	7.5	43	6.6	14.6	<0.1	<0.1	0.2	<0.1	<0.5
DZ-3	1.2	12.9	4.5	5	5.9	1.7	<0.1	<0.1	<0.1	<0.1	1.6
DZ-4	<0.1	25.5	10.2	2	2.3	<0.5	<0.1	<0.1	0.4	<0.1	0.8
DZ-6	<0.1	22.8	11.7	9	2.4	<0.5	<0.1	<0.1	0.5	<0.1	2.2
DZ-8	0.2	33.3	10.4	<1	1.2	<0.5	<0.1	<0.1	0.4	<0.1	3.7
DZ-9	<0.1	11.9	8.8	<1	1	<0.5	<0.1	<0.1	0.6	<0.1	4.9
DZ-10	0.4	5.2	7.2	21	11.3	4.1	<0.1	<0.1	0.3	<0.1	<0.5
DZ-11	<0.1	18.5	16.7	29	9.7	5.3	<0.1	<0.1	0.7	<0.1	0.7
DZ-13	0.5	4.2	2.7	18	7.4	1.9	<0.1	<0.1	0.1	<0.1	1.2
DZ-14	<0.1	19.1	9	12	7.6	1.7	<0.1	<0.1	0.6	<0.1	<0.5
DZ-15	0.2	4.2	5	2	1.3	<0.5	<0.1	<0.1	0.1	<0.1	<0.5
DZ-17	0.6	4.9	8.2	5	5.9	10.3	<0.1	<0.1	<0.1	<0.1	<0.5
DZ-18	<0.1	7	5	6	10.3	<0.5	<0.1	<0.1	<0.1	<0.1	<0.5
DZ-18a	1	2.8	4.9	4	5.2	1	<0.1	<0.1	<0.1	<0.1	<0.5
DZ-20	0.2	17	21.6	25	16.7	6.1	<0.1	<0.1	0.5	<0.1	0.6
DZ-23	0.8	2.6	10.3	8	4.5	1.7	<0.1	<0.1	<0.1	<0.1	<0.5
DZ-25	0.9	6.5	2.4	11	7.4	<0.5	<0.1	<0.1	<0.1	<0.1	<0.5
DZ-26	1.7	8.9	1.6	3	6.3	<0.5	<0.1	<0.1	<0.1	<0.1	<0.5
DZ-27	0.2	6.9	4.4	13	3.5	4	<0.1	<0.1	0.2	<0.1	0.5
Section	Gulla Khel										
Sample No.											
GKK-1	8.9	7.2	22.8	9	19.4	11	<0.1	<0.1	<0.1	<0.1	<0.5
GKK-2	15.8	24.1	33.5	13	17.9	7.6	<0.1	0.3	0.2	0.5	0.5
GKK-3	0.7	26.6	28.5	9	25.2	2.5	<0.1	<0.1	0.3	0.2	2.1
GKS-1A	2.5	15.4	14.7	1322	202.3	3	1.5	0.1	0.3	<0.1	<0.5
GKS-1B	3.5	36.7	20	165	102.5	3.8	3.8	0.2	0.3	<0.1	<0.5
GKS-2	8.5	22.8	122.4	5	6.1	6.4	0.1	1	<0.1	0.1	0.8
GKS-4	2.4	7.7	2.2	5	6.5	<0.5	<0.1	<0.1	<0.1	<0.1	<0.5
GKS-5	1.4	29.5	18.9	27	38.3	<0.5	0.2	<0.1	0.2	0.2	0.9
GKS-6	1	6.1	8.5	5	5.3	<0.5	<0.1	<0.1	<0.1	<0.1	0.5
GKS-7	0.1	7.9	7.8	6	8.9	<0.5	<0.1	<0.1	0.1	<0.1	<0.5
GKS-8	0.2	4.8	4.4	3	4.7	<0.5	<0.1	<0.1	<0.1	<0.1	<0.5
GKS-12	2.5	13	6.1	26	14.9	<0.5	<0.1	<0.1	<0.1	<0.1	<0.5
GKS-16	0.4	13.2	5.6	1	3	<0.5	<0.1	<0.1	<0.1	<0.1	<0.5
GKS-17	0.5	1.7	6	5	4.1	<0.5	<0.1	<0.1	<0.1	<0.1	<0.5
GKS-18	0.3	47.8	4.5	2	11.4	<0.5	<0.1	<0.1	<0.1	<0.1	0.6
GKS-21	0.9	8.7	31.5	2	2	3.1	<0.1	0.2	0.3	<0.1	<0.5
GKS-22	0.4	1.8	2.7	3	3.6	<0.5	<0.1	<0.1	<0.1	<0.1	<0.5

Appendix 3.

	AQ200	AQ200	AQ200	AQ200	AQ200	AQ200	AQ200	AQ200	AQ200	AQ200	AQ200
	Mo	Cu	Pb	Zn	Ni	As	Cd	Sb	Bi	Ag	Au
	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPM	PPB
	0.1	0.1	0.1	1	0.1	0.5	0.1	0.1	0.1	0.1	0.5
GKS-23	0.6	9	8.7	19	11.9	0.8	<0.1	<0.1	0.4	<0.1	<0.5
GKS-26	2.5	5.6	15.5	6	4.1	8.9	<0.1	<0.1	<0.1	<0.1	<0.5
GKS-27	0.2	2.8	8.8	29	8.5	8.5	<0.1	<0.1	0.1	<0.1	<0.5
GKS-29	<0.1	15.9	10.6	29	5.8	2.7	<0.1	<0.1	0.4	<0.1	<0.5
GKS-30	0.1	5	12.3	13	10.4	8.1	<0.1	<0.1	0.2	<0.1	<0.5
Section	Paniala										
Sample No.											
KS-4	12.1	28.9	37.6	18	35.5	14.9	0.1	0.2	0.3	0.5	0.9
KS-6	1.6	26.1	14.3	17	13.3	4.7	<0.1	<0.1	0.3	0.3	<0.5
KS-10	1.6	6.5	22.2	38	22.2	9	<0.1	<0.1	0.2	<0.1	<0.5
KS-11a	0.5	11.8	34.6	38	23	9.5	<0.1	<0.1	0.2	<0.1	<0.5
ASH	1	10	20.5	193	39.1	8.6	<0.1	<0.1	0.2	<0.1	0.7
SD-a	1.5	0.8	1.9	69	30.8	0.6	1.2	<0.1	<0.1	0.2	1
SD-b	2.2	4.2	1.4	71	34.3	2.2	0.1	<0.1	<0.1	0.1	<0.5
SD-c	0.5	13.9	14.9	13	5.2	9.4	<0.1	<0.1	0.2	<0.1	<0.5
SD-1	0.6	5.2	3.2	2	2.8	<0.5	<0.1	<0.1	<0.1	<0.1	0.7
SD-2	0.1	22.6	8.4	4	2.3	<0.5	<0.1	<0.1	<0.1	<0.1	1.5
SD-3	<0.1	26.1	3.9	3	4.5	<0.5	<0.1	<0.1	0.1	<0.1	1
SD-4	2	3.8	3.5	8	8.2	1.7	<0.1	<0.1	<0.1	<0.1	<0.5
SD-5	0.3	3.6	4.4	2	3.6	<0.5	<0.1	<0.1	0.1	<0.1	0.9
SD-6	1.3	14.2	36.1	101	25.8	1.3	<0.1	<0.1	<0.1	<0.1	<0.5
SD-7	0.3	9.3	9	2	2.9	<0.5	<0.1	<0.1	0.2	<0.1	0.5
SD-8	0.7	2.3	14.7	2	1	1.8	<0.1	0.1	0.3	<0.1	<0.5
SD-9	4	26.7	24.3	13	4	110.5	<0.1	<0.1	<0.1	<0.1	<0.5
SD-10	1.8	22.7	1.1	48	4.4	1.1	<0.1	<0.1	<0.1	<0.1	<0.5
SD-11	0.7	1.4	2.8	277	22.1	1.4	4.5	<0.1	<0.1	<0.1	<0.5
SD-12	0.2	0.9	1.4	130	16.5	<0.5	0.5	<0.1	<0.1	<0.1	<0.5
SD-13	0.4	3.9	6.1	100	22.6	1.2	<0.1	<0.1	<0.1	<0.1	<0.5
SD-14	0.2	1.7	0.5	16	2.8	1.1	<0.1	<0.1	<0.1	<0.1	<0.5
SD-15	<0.1	13.6	7.9	11	15.6	5.8	<0.1	<0.1	0.4	<0.1	<0.5

Appendix 3.

	AQ200	AQ200	AQ200
	Hg	Tl	Se
	PPM	PPM	PPM
	0.01	0.1	0.5
Section	Kasanwala		
Sample No.			
DK-T	<0.01	<0.1	<0.5
DK-F	<0.01	<0.1	<0.5
DK-F	<0.01	<0.1	<0.5
DK-K	<0.01	<0.1	<0.5
KF-1	<0.01	<0.1	<0.5
KF-2	<0.01	<0.1	<0.5
KF-3	<0.01	<0.1	<0.5
KF-4	<0.01	<0.1	<0.5
KF-5	<0.01	<0.1	<0.5
KF-6	<0.01	<0.1	<0.5
KF-7	<0.01	<0.1	<0.5
KF-8	<0.01	<0.1	<0.5
KF-10	<0.01	<0.1	<0.5
KF-11	<0.01	<0.1	<0.5
KF-13	<0.01	<0.1	<0.5
KF-14	<0.01	<0.1	<0.5
KL-1	<0.01	<0.1	<0.5
KL-2	<0.01	<0.1	<0.5
KL-3	<0.01	<0.1	<0.5
KL-5	<0.01	<0.1	<0.5
KL-7	<0.01	<0.1	<0.5
DK-a	<0.01	<0.1	0.6
DK-b	<0.01	<0.1	<0.5
DK-c	<0.01	0.1	3.2
DK-d	<0.01	<0.1	<0.5
DK-E	0.01	0.1	3
KW-1	0.04	10.9	3.4
DK-P1	<0.01	0.2	<0.5
DK-P2	<0.01	0.1	<0.5
DK-P3	<0.01	0.2	<0.5
DK-P2a	<0.01	0.1	<0.5
Section	Kaowaali		
Sample No.			
KK-1a	0.03	0.2	<0.5
KK-2a	<0.01	0.2	<0.5
KK-3a	<0.01	<0.1	<0.5
KK-4a	<0.01	<0.1	<0.5
KK-6a	0.09	<0.1	<0.5
KK-7a	0.03	<0.1	<0.5
DU1	0.01	0.3	0.7
DZ4	<0.01	0.2	<0.5
DZ6	0.02	0.4	<0.5

Appendix 3.

	AQ200	AQ200	AQ200
	Hg	Tl	Se
	PPM	PPM	PPM
	0.01	0.1	0.5
DZ7	0.1	<0.1	<0.5
DZ9a	0.02	<0.1	<0.5
DZ11	<0.01	<0.1	<0.5
DZ12	0.1	0.2	<0.5
DZ12a	<0.01	0.1	<0.5
Section	Nammal		
Sample No.			
KK-1	0.13	1	<0.5
KK-2	0.02	1.3	<0.5
KK-3	0.02	1.3	<0.5
KK-4	0.03	1.5	<0.5
DKA-10	<0.01	0.4	<0.5
DKA-11	0.04	0.4	<0.5
Dka - 12	0.03	0.5	<0.5
DKA-13	0.02	0.9	<0.5
DN-1a	0.12	0.2	<0.5
DN-1d	0.05	0.3	<0.5
DN-1j	<0.01	<0.1	<0.5
DN-1	<0.01	<0.1	<0.5
DN-4	<0.01	<0.1	<0.5
DN-5	<0.01	<0.1	<0.5
DN8	0.02	<0.1	0.8
DN8a	0.06	0.1	<0.5
DN-11	<0.01	<0.1	<0.5
DN12	<0.01	0.3	<0.5
DN-16	<0.01	<0.1	<0.5
DN16first	0.02	0.2	<0.5
DN17	<0.01	0.1	<0.5
DN-19	<0.01	<0.1	<0.5
DN22	0.02	<0.1	<0.5
DN24	0.01	0.3	<0.5
DN30	0.02	0.1	<0.5
DN-35	<0.01	<0.1	<0.5
DN-37	<0.01	<0.1	<0.5
Section	Zaluch		
Sample No.			
ZK-1	<0.01	<0.1	<0.5
ZK-2	<0.01	<0.1	<0.5
ZS-1	<0.01	<0.1	<0.5
ZS-3	<0.01	<0.1	<0.5
ZS-4	<0.01	<0.1	<0.5
ZS-6	<0.01	<0.1	<0.5
ZS-8	<0.01	<0.1	<0.5
ZS-9	<0.01	<0.1	<0.5

Appendix 3.

	AQ200	AQ200	AQ200
	Hg	Tl	Se
	PPM	PPM	PPM
	0.01	0.1	0.5
ZS-11	<0.01	<0.1	<0.5
ZS-12	<0.01	<0.1	<0.5
ZS-14	<0.01	<0.1	<0.5
ZL-1	<0.01	<0.1	<0.5
ZL-2	0.06	<0.1	<0.5
DZ-1	<0.01	0.2	<0.5
DZ-2	<0.01	0.1	<0.5
DZ-3	<0.01	<0.1	<0.5
DZ-4	<0.01	<0.1	<0.5
DZ-6	<0.01	0.2	<0.5
DZ-8	0.01	0.1	<0.5
DZ-9	0.01	<0.1	<0.5
DZ-10	<0.01	<0.1	<0.5
DZ-11	0.01	<0.1	<0.5
DZ-13	<0.01	<0.1	<0.5
DZ-14	<0.01	<0.1	<0.5
DZ-15	<0.01	<0.1	<0.5
DZ-17	<0.01	<0.1	<0.5
DZ-18	<0.01	<0.1	<0.5
DZ-18a	<0.01	<0.1	<0.5
DZ-20	<0.01	<0.1	<0.5
DZ-23	<0.01	<0.1	<0.5
DZ-25	<0.01	<0.1	<0.5
DZ-26	<0.01	<0.1	<0.5
DZ-27	<0.01	<0.1	<0.5
Section	Gulla Khel		
Sample No.			
GKK-1	0.01	1	<0.5
GKK-2	0.07	1.1	6.1
GKK-3	0.1	1	1
GKS-1A	0.01	0.5	<0.5
GKS-1B	0.02	0.8	<0.5
GKS-2	0.08	0.9	1.4
GKS-4	<0.01	<0.1	<0.5
GKS-5	0.02	0.7	0.5
GKS-6	<0.01	<0.1	<0.5
GKS-7	<0.01	<0.1	<0.5
GKS-8	<0.01	<0.1	<0.5
GKS-12	<0.01	<0.1	<0.5
GKS-16	<0.01	<0.1	<0.5
GKS-17	<0.01	<0.1	<0.5
GKS-18	<0.01	<0.1	<0.5
GKS-21	<0.01	<0.1	<0.5
GKS-22	<0.01	<0.1	<0.5

Appendix 3.

	AQ200	AQ200	AQ200
	Hg	Tl	Se
	PPM	PPM	PPM
	0.01	0.1	0.5
GKS-23	0.02	<0.1	<0.5
GKS-26	<0.01	0.1	<0.5
GKS-27	<0.01	<0.1	<0.5
GKS-29	<0.01	<0.1	<0.5
GKS-30	<0.01	<0.1	0.8
Section	Paniala		
Sample No.			
KS-4	0.02	1.4	0.5
KS-6	<0.01	0.6	<0.5
KS-10	<0.01	0.7	<0.5
KS-11a	0.02	0.5	1.5
ASH	0.01	1	0.8
SD-a	<0.01	0.8	<0.5
SD-b	<0.01	<0.1	1.5
SD-c	0.01	0.2	1
SD-1	<0.01	<0.1	<0.5
SD-2	<0.01	0.3	<0.5
SD-3	<0.01	0.1	<0.5
SD-4	<0.01	<0.1	<0.5
SD-5	<0.01	<0.1	<0.5
SD-6	<0.01	2.6	0.5
SD-7	0.01	<0.1	<0.5
SD-8	<0.01	0.1	<0.5
SD-9	<0.01	0.1	<0.5
SD-10	<0.01	<0.1	<0.5
SD-11	<0.01	0.1	<0.5
SD-12	<0.01	<0.1	<0.5
SD-13	<0.01	0.1	<0.5
SD-14	0.01	<0.1	<0.5
SD-15	<0.01	0.1	<0.5

Appendix 4.

Sample	Zircon	Tourmaline	Rutlie	Starulite	Chlorite	Chloritoid	Others	Total
KF-1	84	67	67	29	2	1	0	250
KF-2	141	36	56	16	1	0	0	250
KF-4	88	101	51	8	1	1	0	250
KF-8	50	143	49	7	1	0	0	250
KF-10	92	86	55	17	0	0	0	250
KF-13	43	99	85	23	0	0	0	250
KF-14	130	65	45	10	0	0	0	250
DZ-2	58	113	77	2	0	0	0	250
ZS-5	111	71	65	3	0	0	0	250
ZS-7	80	101	59	9	1	0	0	250
ZS-9	87	112	49	2	0	0	0	250
SZ-11	61	136	50	3	0	0	0	250
ZS-13	18	160	68	4	0	0	0	250
GKS-3	113	86	49	1	0	0	1	250
GKS-4	114	60	73	3	0	0	0	250
GKS-6	87	105	56	2	0	0	0	250
DK-b	56	141	47	6	0	0	0	250
DK-c	78	117	49	4	1	1	0	250
DK-d	85	132	30	3	0	0	0	250
DK-7	110	88	50	1	0	0	1	250
DK-10	112	62	72	4	0	0	0	250
DK-11	88	103	57	2	0	0	0	250
DK-18	57	137	49	7	0	0	0	250
DK-26	78	116	50	4	1	1	0	250
DK-30	84	134	29	3	0	0	0	250
DK-41	33	193	23	1	0	0	0	250
DK-43	28	193	29	0	0	0	0	250
DZ-23	31	197	21	1	0	0	0	250
DZ-25	24	196	30	0	0	0	0	250