

Foraminiferal biostratigraphy, facies and sequence stratigraphy analysis across the K-Pg Boundary in Hazara, Lesser Himalayas (Dhudial Section)

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ABSTRACT

The sedimentary strata were sampled in the lesser Himalayas to probe paleoenvironmental changes across the Cretaceous-Paleogene (K-Pg) boundary in the eastern Tethys. The study provides an integrated lithologic and bio-sequence stratigraphic analysis, leading to paleoecology and paleoenvironmental interpretations. The planktic foraminiferal limestone of the late Cretaceous is overlain by lateritic sandstones and sandy foraminiferal limestones, the latter being of Paleocene age. Though the deposition of Cretaceous strata mainly occurred in transgressive and high stand system tracts, the top of Cretaceous is marked by type-I sequence boundary and low stand system tract, corresponding to the Paleocene Hangu Formation. Deposits below the K-Pg boundary zone interval have been correlated to the late Cenomanian *Rotalipora reichel* biozone to early Campanian *Globotruncana ventricosa* zone, with absence of Maastrichtian fauna. A marked change in fauna above the K-Pg boundary zone interval has been observed and manifested by presence of larger benthic foraminifera such as *Lockhartia* Davies, 1932 and *Globanomalina* Haque, 1956 genera. The boundary occurs at the contrasting inter-facial contact of the two rock units and advocates an early lowered sea-levels or dead ocean model. An organic bed of late Turonian-early Coniacian corresponds to the probable presence of the OAE3 and could represent a missing link in the late Cretaceous of lesser Himalayas in the Pakistani domain. Prior to the K-Pg event and Indo-Eurasian collision, an influx of siliciclastics suggests a major episode of uplift and shortening caused by ophiolite obduction or magmatic

KEY WORDS
Cretaceous-Paleogene,
Tethys,
Lesser Himalayas,
Kawagarh Formation,
Foraminifera,
OAE3.

upwelling during the Campanian. The subsequent erosion and its re-deposition shaped the platform, evolving it from relatively steeper ramp geometry in the Campanian to gentler epeiric ramp in the Selandian and Thanetian, and triggered deposition of shallow ramp larger benthic foraminiferal facies. The boundary is similar in nature with erosional phase in the whole region but its duration was prolonged in the study section and its upper limit has some regional changes. As finding of this study, the late Cretaceous “Nara Sandstone Member” of the Kawagarh Formation in Hazara area of earlier workers could be revised as Paleocene Hangu Formation.

RÉSUMÉ

Biostratigraphie (foraminifères), évolution des faciès et analyse séquentielle autour de la limite K-Pg dans la zone d’Hazara, bas Himalaya (coupe de Dhudial).

Des couches sédimentaires ont été échantillonnées dans le bas Himalaya en vue d’étudier les changements paléoenvironnementaux autour de la limite Crétacé-Paléogène (K-Pg) dans l’est de la Téthys. L’étude fournit une analyse stratigraphique intégrée, lithologique et bio-séquentielle, conduisant à des interprétations paléoécologiques et paléoenvironnementales. Le calcaire à foraminifères planctoniques du Crétacé supérieur est recouvert par des grès latéritiques et des calcaires sableux à foraminifères datés du Paléocène. Bien que les sédiments crétacés soient transgressifs et principalement associés à des cortèges de haut niveau, le sommet est marqué par une limite de séquence de type I et la zone du cortège de bas niveau correspondant à la formation paléocène de Hangu. Les dépôts situés sous la limite K-Pg ont été attribués à l’intervalle compris entre les biozones à *Rotalipora reichel* du Cénomanien supérieur et celles à *Globotruncana ventricosa* du Campanien inférieur, la faune maastrichtienne étant absente. Un changement paléoécologique marqué au-dessus de la limite a été observé et se manifeste par la présence de grands foraminifères benthiques tels que les genres *Lockhartia* Davies, 1932 et *Globanomalina* Haque, 1956. La zone de la limite K-Pg montre ainsi un contact faciologique et écologique net suggérant une chute eustatique brutale et/ou un modèle « d’océan mort ». Un niveau riche en matière organique, datant du Turonien supérieur-Coniacien inférieur, pourrait correspondre à l’OAE3. Si cette identification était confirmée, cela correspondrait à une découverte inédite de l’OAE3 dans le domaine pakistanais de l’Himalaya. Avant la limite K-Pg et la collision indo-eurasienne, un afflux de sédiments silicoclastiques suggère un épisode majeur de soulèvement et de raccourcissement causé par l’obduction des ophiolites ou un *upwelling* magmatique pendant le Campanien. L’érosion associée et le dépôts suivants de sédiments fluvio-deltaïques et marins ont façonné la plate-forme, la faisant passer d’une géométrie de rampe à pente raide au Campanien à une rampe à pente plus douce au Selandien et au Thanétien, ceci étant associé au dépôt de rampes peu profondes à faciès de grands foraminifères benthiques. La limite K-Pg montre donc une érosion importante dans toute la région de l’Himalaya, mais la durée de cette érosion et son extension dans le temps varient localement, comme en témoigne l’étude de la coupe de Dhudial. Comme le montre le résultat de cette étude, le « *Nara Sandstone Member* » de la Formation Kawagarh, dans la région de Hazara, attribué au Crétacé supérieur dans des travaux précédents, doit maintenant être considéré comme appartenant à la Formation de Hangu datant du Paléocène.

MOTS CLÉS
Crétacé-Paléogène,
Téthys,
bas Himalaya,
Formation Kawagarh,
Foraminifera,
OAE3.

INTRODUCTION

The Cretaceous-Paleogene (K-Pg) boundary is widely studied to know the causes of the extinction event that occurred during the time period and its effect on the flora and fauna. At its extreme, the end Cretaceous mass extinction vanished about 75% of marine and almost 50% of land species (Ryder *et al.* 1996; Hallam & Wignall 1997; Alroy 2008). However, the event was biologically selective as the extinction had greater intensity in the Northern Hemisphere (Keller *et al.* 1993; Nichols & Johnson 2002; Jiang *et al.* 2010; Donovan *et al.* 2016). The extinction event records paleoceanographic, paleobiological and climatic changes evolving from much warmer to cool greenhouse and are responsible for wiping out the planktonic foraminifera in majority of sections around the world (Molina *et al.* 1996; Hay & Floegel 2012; Robertson

et al. 2013; Kaiho *et al.* 2016; Bardeen *et al.* 2017; Arenillas *et al.* 2018). Immediately after the event, the marine ecosystem recovered under the “Strangelove Ocean” conditions and the benthic foraminifera re-organized after their drop due to biologic productivity (Rhodes & Thayer 1991; Alegret *et al.* 2001, 2002). Prior to the main extinction event, the Oceanic Anoxic Events (OAEs) developed as consequence of short-lived periods of organic carbon burial. These events are distinguished by their pervasive distribution of distinct black shale beds with marked carbon isotopic excursions. The relationship of plankton turnover and carbon isotopic excursions with the major OAEs points to widespread changes in the ocean-climate system. The mid-late Cretaceous was a time of a switch in the character of the ocean-climate system subjecting to increased rates of tectonic activity and shifting paleogeography (Jones & Jenkyns 2001). The OAEs and resultant black shales during

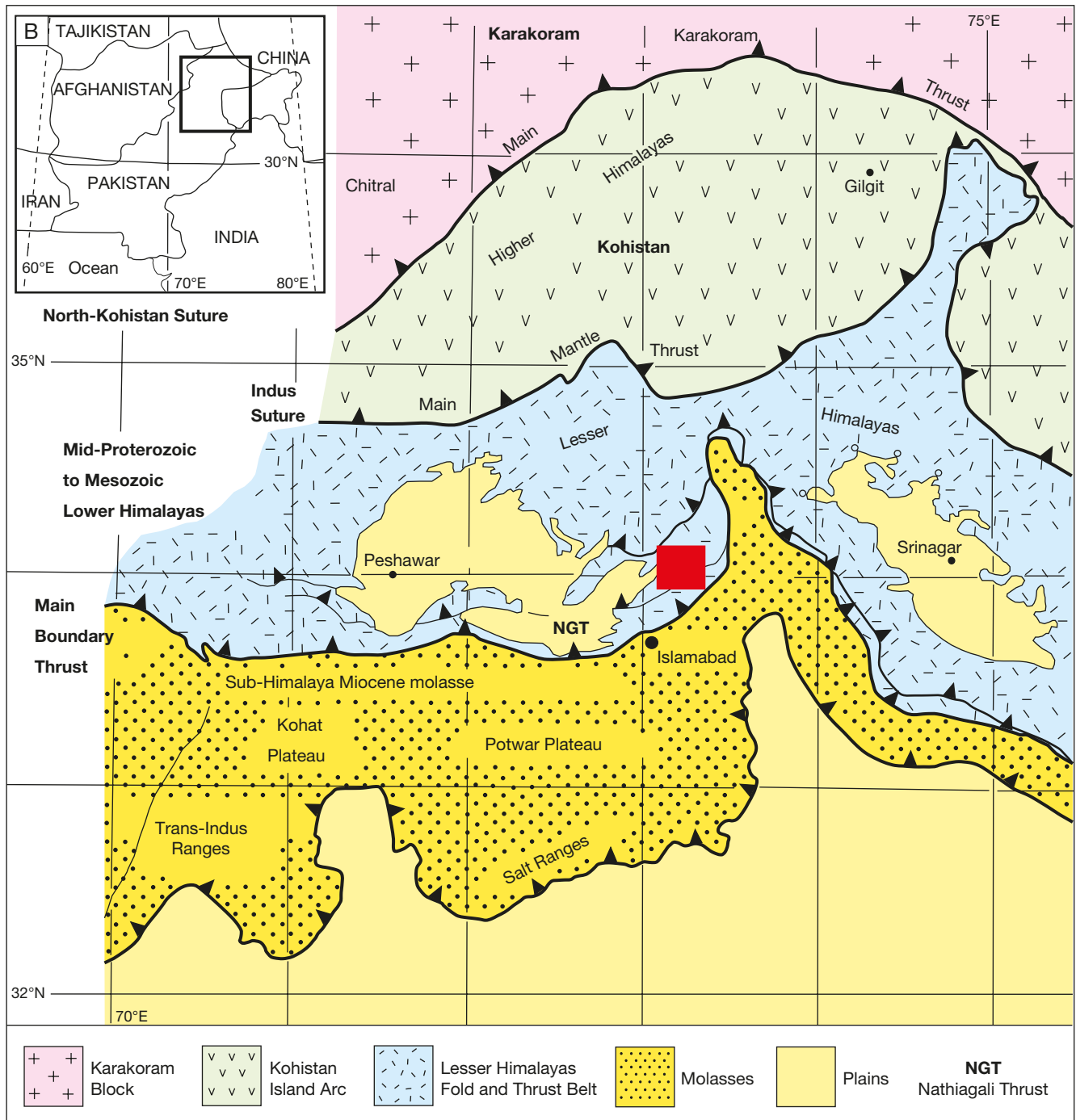


Fig. 1. — Regional geological map after Burg (2011). The red square corresponds to studied location at Dhudial.

this time period are characterized by periodic and contemporaneous deposition in a broad range of oceanic settings. Those that occurred in the early Aptian (*c.* 120 Ma) and surrounding the Cenomanian-Turonian boundary (*c.* 93.5 Ma) have global distribution and comparatively high organic carbon content (Jenkyns 2003; Kuroda & Ohkouchi 2006). The stratigraphic and micropaleontological evidences of these black shales in the Atlantic, Indian and Pacific oceans have provided good opportunities regarding the origin and understanding of OAEs over the last few decades. The enhanced burial of organic matter can

be explained by a boost in primary productivity or improved preservation scenarios (of OAE2) but the actual triggering mechanism of increased carbon sequestration are not evidently identified. In addition to the enhanced burial of organic matter, Turgeon & Creaser (2008) explain triggering of the OAE2 by a massive magmatic episode as shown by marine osmium isotope records. While most of these studies were focused on the western Tethys (e.g., Gambacorta *et al.* 2016), work related with OAEs from the eastern Tethys is minimal and notably include Li *et al.* (2006), Wendler *et al.* (2009), Zhang *et al.* (2016).

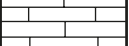
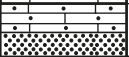
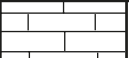
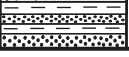
Paleocene	Lockhart		Monotonous limestone, slightly nodular
	Hangu		Sandstone in lower part, sandy limestone in upper part
Cretaceous	Kawagarh		Medium bedded limestone and subordinate marls
	Lumshiwal		Sandstone, limestone and subordinate shales/marls
	Chichali		Belemnites bearing greenish shale beds
Jurassic (Lower-Mid)	Samana Suk		Limestone with occasional intercalations of shale/marls
	Shinawari		Sandstone, limestone and shales
	Datta		Calcareous sandstone and shales

FIG. 2. — Generalized stratigraphy of the study area (Shah 2009).

Recent studies of mid-Cretaceous calcareous nannofossils and planktic foraminifera from deep-sea sections have established an integrated calcareous plankton biostratigraphy and improved geochronology. The record of K-Pg boundary is preserved in the sediments of the eastern Tethys Ocean and spread across various sedimentary basins of Pakistan. This contact is to date the only obvious exposure in the Upper Indus Basin of Pakistan. A valid demarcation and changes across this contact is a missing link and the objectives of present study focus on: 1) analyzing facies distribution; 2) studying the various planktonic bio-events of the upper Cretaceous Dhudial Section; 3) looking for presence of OAEs in the Dhudial Section; and 4) establishing sequence stratigraphic framework by integration of facies and foraminiferal biostratigraphic analysis across the Cretaceous-Paleogene boundary in the eastern Tethys Ocean in the lesser Himalayas.

GEOLOGICAL SETTINGS

The late Cretaceous strata in the Indus Basin of Pakistan was part of the temperate carbonate platform system along the southern margin of the eastern Tethys Ocean. The study area is part of the lesser Himalayas (Fig. 1) bounded on the north and south by the Main Mantle Thrust (MMT) and Main Boundary Thrust (MBT), respectively (Burg 2011; Kazmi & Jan 1997). The sedimentary deposits comprise pelagic and hemi-pelagic limestones of outer ramp to slope environments (Shah 2009). The area has general exposure of rock units ranging in age from Pre-Cambrian to Miocene, including the Cretaceous-Paleogene contact between the Upper Cretaceous Kawagarh Formation and Paleocene Hangu Formation (Warwick *et al.* 1995; Shah 2009). The recognition of the late Cretaceous stage boundaries in the Indus Basin have been hampered by gaps and incomplete sections. However, the Upper Indus Basin of Pakistan provides more continuous pelagic-hemi pelagic sedimentation, represented by late Cretaceous Kawagarh Formation (Fig. 2). The measured interval of the late Cenomanian-early Campanian succession (Latif 1970; Shah 2009) is represented by 70 m thickness of Kawagarh Formation and 05 m of early Paleogene

Hangu Formation at Dhudial (34°10'01"N, 73°39'86"E), north of the city Abbottabad (34°09'21"N, 73°13'10"E) and of the capital Islamabad in northern Pakistan. The Cretaceous unit is composed of medium to thick bedded limestones, which are dark grey on fresh surfaces and bluish to pale yellow on weathered surface and bear some intercalation of marls. The lower part of Paleogene (Hangu Formation) is comprised of lateritic sandstones with ferruginous cements, interpreted as exposed deltaic plains (Warwick *et al.* 1995; Munir *et al.* 2005; Shah 2009). The Hangu Formation in adjacent areas like Balakot and Azad Kashmir, as per Munir *et al.* (2005), with 1-26 meters thickness variation, has records of laterites, pisolitic bauxites, carbonaceous shales and coal seams, conforming with the transitional marine/deltaic environment. whereas its upper part is represented by larger benthic foraminiferal sandy limestones. Since the Dhudial Section is located in the lesser Himalayas of the intensely deformed Himalayan fold and thrust belt, earlier researchers were reluctant to carry out noteworthy analysis. Very less literature review is available by researchers and their work is restricted to pioneer values. The noteworthy ones include Latif (1970), Ahsan *et al.* (1993) and Shah (2009), who worked on facies variations in the early Paleocene, lithostratigraphy of the southern Hazara and the occurrence of oolitic hematite, marking the K-T boundary.

MATERIAL AND METHODS

The Dhudial Section (Fig. 3) is located 5 km north of Abbottabad City on the Abbottabad-Thandiani road, where it is excellently exposed along a quarry comprising of late Cretaceous-Paleocene rock units (Fig. 4A-C). A total of 40 samples were collected from the Dhudial Section at regular intervals of 2 meters. Each sample from limestones/marly limestones was cut into two pieces, one of which was retained while other was processed using the freeze-thaw method of Slipper (1996) in which the trampled samples were immersed in a supersaturated solution of Na₂SO₄. The process was repeated 2-3 times, washed and passed through and 63 µm sieve till foraminiferal

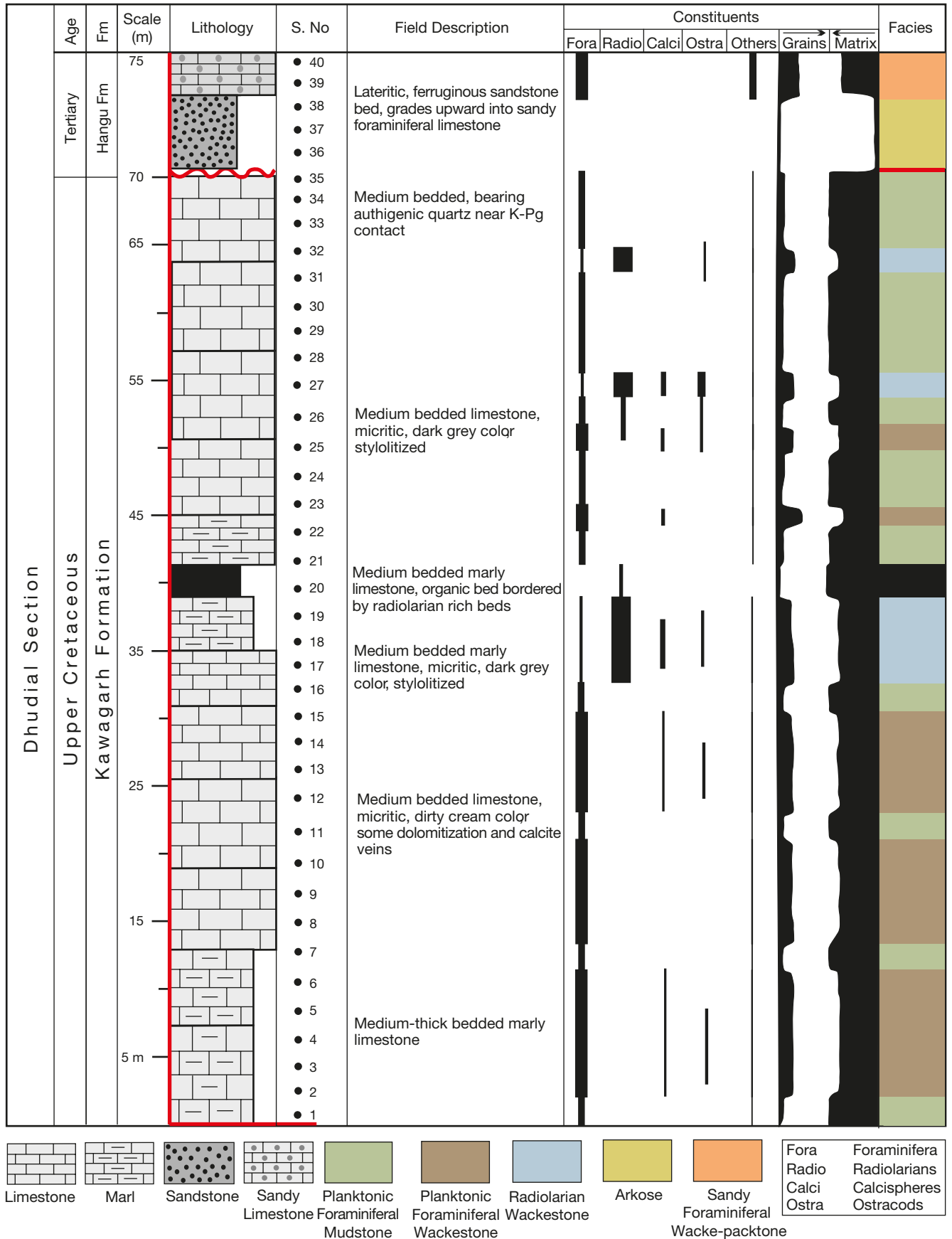


FIG. 3. — Lithostratigraphic column showing the lithology, constituents and facies of the Dhudial Section.



FIG. 4. — Field photograph of the studied Dhudial Section: **A**, stratigraphic contacts of different ages; **B**, authigenic quartz near the boundary; **C**, Outcrop view of probable OAE3.

residues were obtained. However, the foraminiferal preservation was poor and because of carbonate adhesion, the specimens were difficult to free. The retained samples were mounted on slides to prepare thin sections for petrographic analysis under standard procedures. Dunham (1962) and Flügel (2004) were used for classification and interpretation of facies combined with paleoenvironmental estimates by using the agglutinated foraminiferal morphogroup method (Kaminski & Gradstein 2005; Cetean *et al.* 2011; Murray *et al.* 2011). The method presumes that species with different test shapes have different

locales and nourishment strategies, and disparity in the comparative abundance of the morphogroups can reflect environmental changes. A sea level curve has been drawn to mention the system tracts and sequences of the formation by integrating microfacial interpretation and associated fauna. The sea level curves of Haq (2014); Kominz *et al.* (2008) and Müller *et al.* (2008) were used for interpretation of sequence stratigraphic analysis. A total of 28 planktonic foraminiferal species were identified following methods of Robaszynsky & Caron (1979) and Caron (1985).

RESULTS

KAWAGARH FORMATION

The lower part of the Kawagarh Formation is dominated by planktonic foraminiferal wackestones facies representing medium to thick bedded limestones, dark grey on fresh and yellowish on weathered surface. The facies is comprised of grains (25%) among them 15% are planktonic foraminifera as shown in Fig. 5A and includes the genera of *Hedbergella* Brönnimann & Brown, 1958 (Brönnimann & Brown 1958), *Globigerinelloides* Cushman & Ten Dam, 1948 (Cushman & Ten Dam 1948), *Rotalipora* Brotzen, 1942 (Brotzen 1942), *Whiteinella* Pessagno, 1967 (Pessagno 1967), *Heterohelix* Ehrenberg, 1843 (Ehrenberg 1843), *Helvetoglobotruncana* Reiss, 1957 (Reiss 1957), *Dicarinella* Porthault, 1970 (Porthault *et al.* 1970) and *Praeglobotruncana* Bermúdez, 1952 (Bermúdez 1952). The non-foraminiferal content includes ostracods (5%), sponge spicules (3%) and calcispheres (2%). Ferroan dolomitic crystals and calcitic veins are common throughout the facies; the former one has commonly developed along stylolitic boundaries (Fig. 5B). The microfacies, at some places, also have packstone textures (Fig. 5C). The radiolarian rich wackestones facies (Fig. 5D) occur in middle part of the formation and also recur as thin beds in the upper portion. The facies contains 25% of grains which include radiolarians (12%) as obvious in Fig. 5E, calcispheres (5%), planktonic foraminifera (4%), ostracods (2%) and sponge spicules (2%). The planktonic foraminifera include genera of *Heterohelix*, *Whiteinella*, *Marginotruncana* Hofker, 1956 (Hofker 1956) and *Globotruncana* Cushman, 1927 (Cushman 1927) – in the uppermost beds. Though interbeds of the planktonic foraminiferal lime mudstone facies are present in the lower part but it prevails in the upper part of the formation. The facies is comprised of <10% grains in the form of planktonic foraminifera (5%) including genera of *Hedbergella*, *Globigerinelloides*, *Rotalipora*, *Whiteinella*, *Heterohelix*, *Dicarinella*, *Marginotruncana*, *Praeglobotruncana*, *Globotruncana*, some ostracods (2%), radiolarians (1%) and sponge spicules (1%). The facies is in contact with wackestone facies having a sharp dolomitized boundary (Fig. 5F). Stylolitization and often cross cutting calcite veins (Fig. 5G) are not uncommon, particularly in the upper half of Kawagarh Formation. The frequently zoned ferroan dolomite crystals (Fig. 5H) are considered as secondary. The facies shows development of authigenic quartz (Fig. 4B) and an organic rich unit (Fig. 4B) below the contact with the Paleocene.

HANGU FORMATION

The 3-meter thick sandstone lithofacies of the Paleogene is hard, compact and characterized by a light brown color. In addition to quartz and feldspar, chert rock fragments and accessory minerals such as mica (muscovite) are obvious. Quartz (Fig. 5I) is most dominant, majority of which is monocrystalline and sub-angular to sub-rounded. The feldspar grains belong to the K-feldspar group, medium to coarse sand size with well-rounded boundaries (Fig. 5J). Lithics of chert (Fig. 5K) are present whereas accessory minerals include

muscovite mica and some ore minerals. They are sub-spherical to sub-rounded boundaries. The constituents are clustered by a combination of quartz overgrowths, interstitial calcite cements, and ferruginous matrix (Fig. 5L). This moderately sorted, fine grained sandstone is texturally and chemically sub-mature as indicated by the abundance of grains/matrix and Q/F ratios rock term the rock as sub-arkose. Dominance of monocrystalline quartz, some chert fragments and highly rounded grains indicate a dominantly igneous and sedimentary source. The overlying 2 m bed of sandy foraminiferal limestone facies yield angular to sub-angular siliciclasts and skeletal allochems embedded in a mix of matrix and sparry calcite cement. Skeletal allochems are partially micritized and include *Lockhartia* (Davies 1932), *Globanomalina* (Haque 1956) and miliolids. Upwards gradation of facies into pure foraminiferal limestones has been recorded in the Thanetian Lockhart Formation of Hanif *et al.* 2013).

DISCUSSION

FORAMINIFERAL BIOSTRATIGRAPHY

The late Cretaceous biostratigraphic correlation on planktonic foraminifera has received attention due to presence of significant microfossils in deep marine sequences. Early studies conducted by Eicher (1969, 1972) formed a critical framework for the planktic foraminiferal biostratigraphy and paleoecology whereas Leckie (1989) considerably developed its paleoclimatic and paleo-oceanographic applications. Regarding the preliminary biostratigraphic framework in the study area, Latif (1970) reported Coniacian-Campanian foraminifera in the Changla Gali Section of Hazara area this being confirmed by Butt (1992). Due to the structural complexities, limited work has been conducted and the current Dhudial Section provides the best exposure to date of late Cretaceous-Paleogene strata in the region, not described before. This study describes the presence of Cenomanian-Early Campanian planktic foraminiferal species ranges and applies it on the standard planktic foraminiferal biozones of Robaszynsky & Caron (1995) to achieve long range correlations. The range chart based on abundance, diversity, FADs and LADs of different foraminifera (Fig. 6) recognize eight different biozones. These biozones include: 1) *Rotalipora reichel* zone; 2) *Rotalipora cushmani* zone; 3) *Whiteinella archaeocretacea* zone; 4) *Helvetoglobotruncana helvetica* zone; 5) *Marginotruncana sigali* zone; 6) *Dicarinella concavata* zone; 7) *Globotruncana elevata* zone; and 8) *Globotruncana ventricosa* zone; they have been marked in Figure 8.

The base is occupied by *Rotalipora reichel* zone of lower Cenomanian (Robaszynsky *et al.* 1984) and includes assemblages of *Hedbergella planispira* (Tappan 1940), *Heterohelix globulosa* (Ehrenberg 1840), *Globigerinelloides bollii* (Pessagno 1967), *G. bentonensis* (Morrow 1934) and *Rotalipora appenninica* (Renz 1936). The *Rotalipora cushmani* zone (13-24 m) bears some age diagnostic foraminifera of which the *Rotalipora deekei* (Franke 1925) and *R. cushmani* (Morrow 1934) are main elements. Since the species of *Praeglobotruncana prae-helvetica*

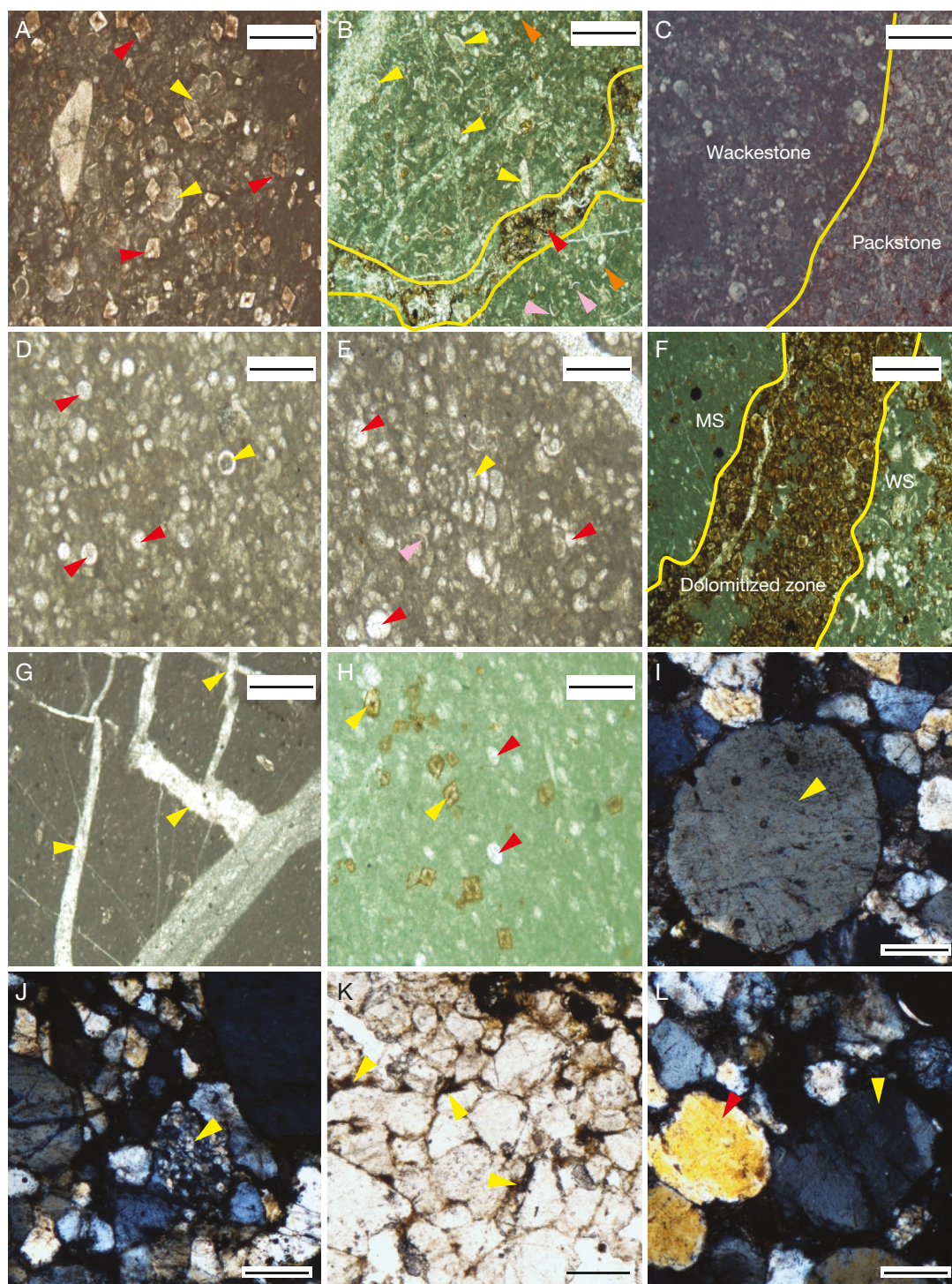


FIG. 5. — Photomicrographs of thin slides: **A**, Planktonic foraminifera (yellow arrowheads) and ferroan euhedral dolomite crystals (red arrowheads) (S# 09); **B**, dolomitized zone (red arrowheads) with planktonic foraminifera (yellow arrowheads), radiolarians (orange arrowheads) and ostracods (pink arrowheads) (S# 25); **C**, contact of the foraminiferal wackestone and packstone patch (S# 22); **D**, radiolarian rich facies showing radiolarians (red arrowheads) and calcispheres (yellow arrowheads) (S# 18); **E**, *Heterohelix* sp. (yellow arrowheads) along with radiolarians (red arrowheads) and ostracods (pink arrowheads) (S# 17); **F**, dolomitized zone between the lime mudstone (MS) and wackestone (WS) (S# 11); **G**, tectonic induced fracturing (S# 07); **H**, radiolarians (red arrowheads) and zoned ferroan dolomite crystals in the mudstone facies (S# 11); **I**, well rounded feldspar grains (S# 38); **J**, chert lithic (S# 36); **K**, ferruginous matrix (S# 37); **L**, quartz (red arrowheads) and feldspar (yellow arrowheads) grains in Hangu Formation (S# 38). Scale bars: A-C, F, G, 200 µm; D, E, H-L, 100 µm.

(Trujillo 1960), *Anaticinella multiloculata* (Morrow 1934) are absent, the sub-zones of Keller *et al.* (2001) cannot be established and this zone is therefore, marked only by presence of *Rotalipora* extinction sub-zone. This sub-zone is defined by

the extinction of *R. deekei* to the extinction of *R. cushmani* as a consequence of OAE2 (Keller & Pardo 2004) near the Cenomanian-Turonian boundary (93.90 Ma). Three divisions of *Whiteinella archaeocretacea* zone (24-28 m), dominated

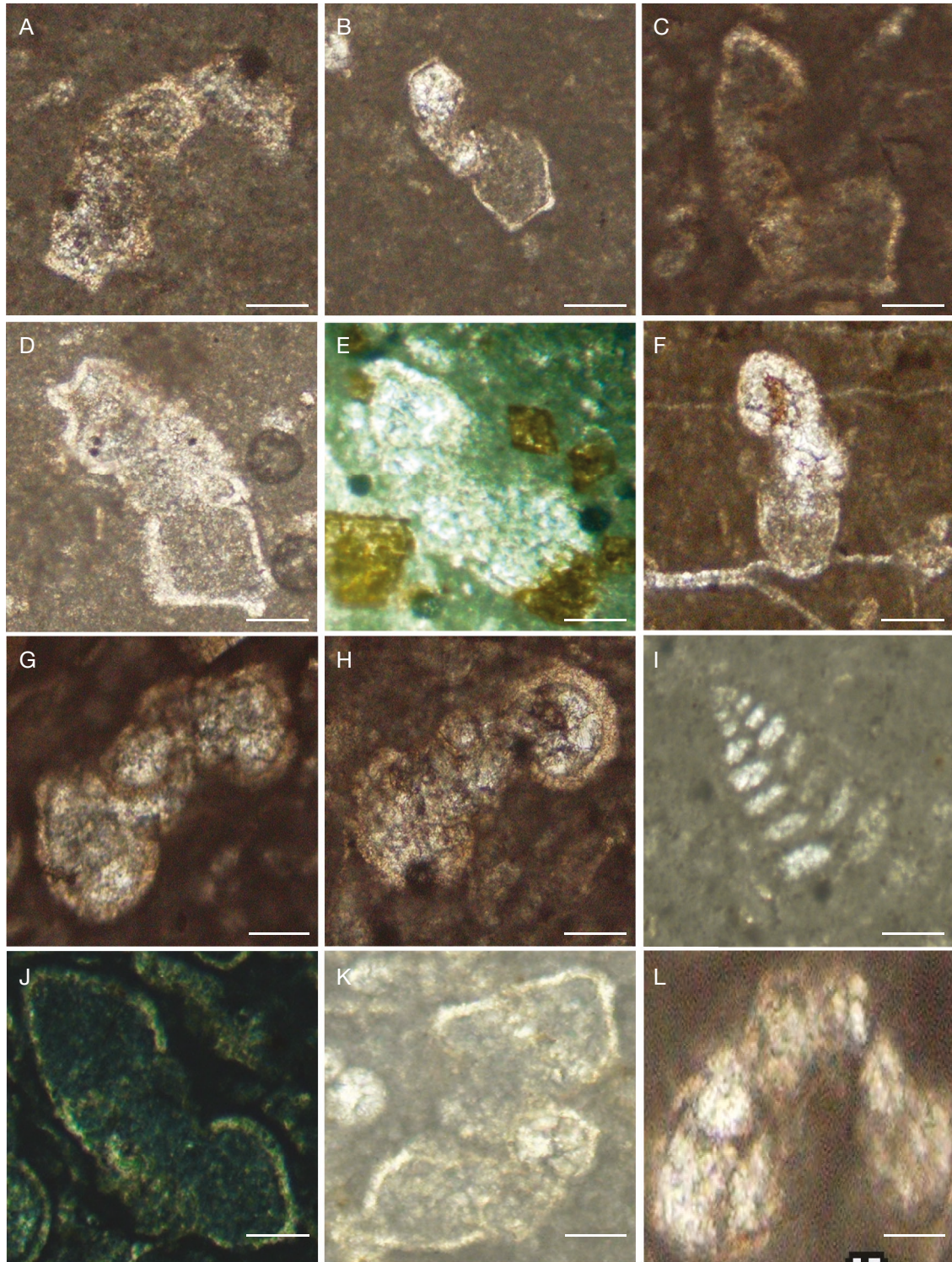


FIG. 6. — A, *G. ventricosa* (S# 33); B, *G. linneiana* (S# 31); C, *D. concavata* (S# 24); D, *Dicarinella hagni* (S# 12); E, *M. sigali* (S# 20); F, *H. planispira* (S# 23); G, *W. baltica* (S# 14); H, *H. helvetica* (S# 14); I, *Heterohelix moremani* (S# 13); J, *Rotalipora appenninica* (S# 08); K, *R. cushmani* (S# 06); L, *W. archaeocretacea* (S# 15). Scale bars: 80 μ m.

largely by the heterohelids, are identified (Keller & Pardo 2004). The lowermost *G. bentonensis* sub-zone is defined by the interval from disappearance of *R. cushmani* to extinction of *G. bentonensis* (93.86 Ma). *Dicarinella hagni* sub-zone, a replacement of *A. multiloculata* sub-zone of Keller *et al.*

(2001), occurs between the demise of *G. bentonensis* and the onset of *Heterohelix* species such as *H. Moremani* (Cushman 1938) and *H. reussi* (Cushman 1938) in the Dhudial Section. The uppermost *Heterohelix moremani* sub-zone marks FAD of heterohelix shift to the first appearance of *Helvetoglo-*

botruncana helvetica (Bolli 1945). The *W. archaeocretacea* assemblage shows the massive presence of the species of the genus *Whiteinella*, *W. archaeocretacea*, *W. baltica* (Douglas & Rankin 1969) and *H. globulosa*. The early Turonian marker *H. helvetica* (93.29 Ma) initiates the *H. helvetica* zone (28–35 m) and is also considered as a foraminiferal biomarker near the Cenomanian-Turonian boundary (93.50 Ma). The *H. helvetica* zone, like *W. archaeocretacea* zone, is marked by assemblages of genera of the *Whiteinella*, *Heterohelix*, *Dicarinella algeriana* (Caron 1966) and *Praeglobotruncana gibba* (Klaus 1960). The lower part of upper Turonian marks the initiation of *Marginotruncana sigali* zone (35–41 m) and yielded *M. sigali* (Reichel 1950), *M. pseudolinneiana* (Pessagno 1967), some *Whiteinella* and *Heterohelix* species, *D. algeriana* and *H. planispira*. The *Heterohelix* species especially *H. reussi* is abundant and *H. globulosa* shows its appearance in the upper part of *M. sigali* zone whereas majority of the whiteinellids die out.

The zones of *Dicarinella primitiva*, at the onset of Santonian (85.8 Ma), and *Dicarinella asymmetrica* of upper Santonian, have not been recorded in the present study. However, they have been included in the *Dicarinella concavata* zone (41–56 m) of Caron (1985) and represents a time interval from the early Coniacian to late Santonian. Along with the *Dicarinella concavata* (Brotzen 1934) The zone also has presence of *Heterohelix* species, in particular the *Heterohelix reussi*, *H. globulosa*, *H. planispira*, *M. sigali* and *M. pseudolinneiana*. The globotruncinids make its appearance towards the top and marks the start of *Globotruncana elevata* (56–61 m) and *Globotruncana ventricosa* (61–70 m) zones. The *G. elevata* zone (upper Santonian to lower Campanian) of Caron (1985) like *D. primitiva* zone is missing from the study; however, the overlying *G. ventricosa* zone (Robaszynsky & Caron 1995) of lower Campanian is widely present and marked by presence of *Globotruncana ventricosa* (White 1928), *Globotruncana arca* (Cushman 1926), *G. linneiana* (d'Orbigny 1839), *H. reussi* and *H. globulosa*.

The present study depicts that deposits above the early Campanian have experienced an abrupt change of fauna from late Cretaceous planktonic to Paleocene benthic. The Paleocene genera of Lockhartia (larger benthic) and Globanomalina (smaller benthic) in the sandy foraminiferal limestone bed, above the sandstone unit, manifest a faunal change. The lateritic? ferruginous sandstone unit, located between 70 and 73 m (Fig. 3), and subsequent sandy foraminiferal limestone bed overlying the late Cretaceous beds and underlying the late Paleocene foraminiferal limestones remains undated. Khan & Ahmad (1966) and Shah (2009) have called the upper sandstones of the Cretaceous Kawagarh Formation in Hazara area as “Nara Sandstone Member” and correlated them with the Maastrichtian Pab Sandstone in the Sulaiman Ranges. However, this correlation was merely on facies basis, lacked any sort of evidence and appears to have an age assignment problem. The absence of Maastrichtian fauna in the Kawagarh Formation of the studied Dhudial Section and overall, in the Hazara area makes the correlation of this sandstone unit with the Maastrichtian Pab Sandstone dubious. The gradational

contact of this sandstone with middle Paleocene foraminifera bearing sandy limestone bed and conformable presence of Thanetian foraminiferal limestones (Lockhart Formation) of Hanif *et al.* (2013) positions the unit in Selandian. Using nannofossils, dinoflagellates and pollens samples from the Salt Ranges, Köthe (1988) and Warwick *et al.* (1995) have confirmed middle Paleocene (Selandian) age for the Hangu Formation. Thus, the late Cretaceous “Nara Sandstone Member” of Kawagarh Formation in Hazara area of Khan & Ahmad (1966) and Shah (2009) could be revised as middle Paleocene Hangu Formation. Though the main extinction event started well before the onset of Paleocene, the absence of Maastrichtian and Danian fauna in the Upper Indus Basin hinders the establishment of the K-Pg boundary on faunal record. Due to these reasons, the boundary zone is marked on basis of facies change, and correspond to the sandy ferruginous bed interval between 70 and 73 m at the base of the Hangu Formation, this being supported by the appearance of middle Paleocene benthic fauna immediately above.

PALEOECOLOGY

The spatial distribution and abundance of planktonic foraminiferal assemblages are associated with sea water stratification, vertical changes in temperatures and density gradients (Keller 2002; Price & Hart 2002). Based on the foraminiferal assemblages, the upper euphotic zone, lower euphotic zone, disphotic deeper water zones have been established (Fig. 7). The keeled species have generally been assigned deeper water paleoecologic requirements (Norris & Wilson 1998; Leckie *et al.* 2002).

The keeled genera of rotaliporids, globotruncanids and marginotruncanids, representing deeper oligotrophic waters (Corfield *et al.* 1990; Price & Hart 2002; Norris & Wilson 1998; Mikadze 2010; Abramovich *et al.* 2002), are present in lower and upper parts of the late Cretaceous unit in this study. The rotaliporids thrived in lower part of the Dhudial Section (0–24 m) in disphotic zones (>200 m) and were extinct largely due to well-developed oxygen minimum zones coincident with OAE2. The middle portion (24–38 m) correspond to *W. archaeocretacea*, *H. helvetica* zones and lower part of *M. sigali* zone associated with upper euphotic zone and are explained by more oxygenated waters in between the reduced OMZs below and above. The genera *heterohelix*, *hedbergella*, *globigerina*, *whiteinella* and *dicarinella* symbolize shallow (<100 m), normal saline water (Price & Hart 2002). The *H. globulosa*, *W. baltica* often occur in nature in symbiotic relationship with dinoflagellates in the upper euphotic zone (<100 m). However, the virtual absence of genus *Hedbergella*, occupying near-surface waters, excluding *Hedbergella planispira*, indicates lower part of upper euphotic zone fauna in the current study. Occurrence of *H. planispira* allied with low oxygen tolerant dwellers- the heterohelicids- particularly the *H. moremani* and *H. reussi* (Keller & Pardo 2004), could suggest a partially developed oxygen minimum zone (OMZ) during deposition. This interpretation is also supported by the presence of species of *D. hagni* (Scheibnerová 1962) and *D. algeriana*; both of which descended from the keeled glo-

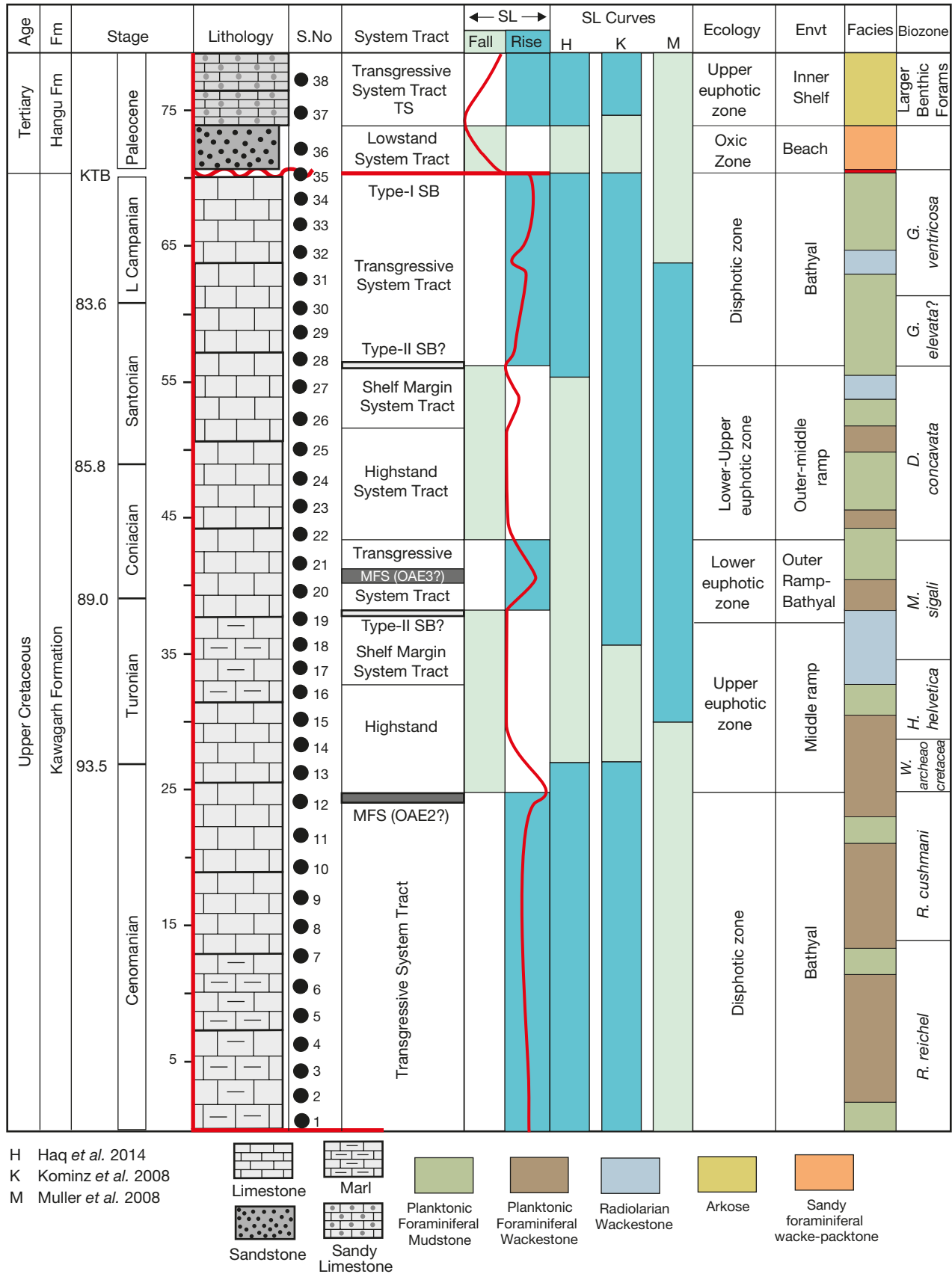


FIG. 7. — Chart showing the sequence stratigraphic framework, sea level changes and paleoecology of the Dhudial Section.

botruncanids and inhabitants of low oxygenated waters. The overlying OMZ is represented by *M. sigali* zone (38–44 m) corresponding to the lower euphotic zone (>100 m). Following the OMZ, a brief period of lower-upper euphotic waters (44–51 m) occur as revealed by *D. concavata* zone. The rise in sea levels and oligotrophic waters of the disphotic zones led the globotruncanids to flourish in the upper part of the Dhudial Section (51–70 m) before evolving to lowered sea levels and/or ‘dead ocean model’ in the early Campanian. The conditions changed abruptly and led to development of oxic zones depositing the Paleocene laterites and sandy foraminiferal limestones of oxic and upper euphotic zones, respectively.

FACIES ANALYSIS

The sediments in most part of the Dhudial Section are composed of fine grained limestones below the K-Pg boundary zone interval, and sandstones and sandy limestones above. The lower part of the Dhudial Section is occupied by late Cretaceous planktonic foraminiferal wackestone, spanning the Cenomanian and lower Turonian stages, and contains relatively deep water (near thermocline) fauna such as *Rotalipora*, *Heterohelix*, *Hedbergella*, *Dicarinella*, as well as *Marginotruncana* (in middle part of the formation). The facies is attributed to have been deposited in middle bathyal depths, the breaking of thermocline effectively eliminating the deeper dwelling species. Following the demise of deep dwelling fauna like *Rotalipora* on maximum sea levels (Leckie 1989), the intervening zone with upper euphotic zone fauna could suggest lowered sea levels. This can be inferred from the extinction of deep-sea benthic foraminifera across the Cenomanian-Turonian Boundary (Kaiho 1998). The relatively high diversity of the biota in the intervening zone (Fig. 8) suggests loss of water stratification and thorough mixing of the nutrients, reaching to various parts in the ocean depth profile (Hays *et al.* 2005). Patches of foraminiferal packstone texture support the circulation of the waters in middle-distal outer ramp environment. The radiolarian rich wackestones occur as transitional bed (31–38 m) between the lower foraminiferal wackestones and overlying foraminiferal lime mudstones during middle to late Turonian. Two environments of deposition have been suggested for this radiolarian-rich facies with different causes: 1. Middle to outer ramp due to the absence of deep waters planktonic foraminiferal fauna and presence of low salinity tolerant *Hedbergella planispira*. The high levels of dissolved silica may be due to the growth of the Kohistan Island Arc within Tethys during late Cretaceous could therefore probably explain the bloom of radiolarians 2. Open marine/bathyal (lower euphotic zone) associated with the possible presence of OAE3 in the Coniacian-Santonian time periods (Arthur *et al.* 1990) or latest Turonian Hitchwood event of Jarvis *et al.* (2006) in sections including Tethys when primary plankton resource was on a high (Bomou *et al.* 2013; Leckie *et al.* 2002). The later one is suggested because of possible presence of organic rich deposits (Fig. 4C) but absence of deep dwelling planktonic foraminifera makes the interpretation ambiguous. Because of the long-ranges and ocean-wide distribution, the radiolarites apparently characterize anachronistic facies, but

stepwise extinctions and constant radiations precede and follow the events (Erbacher & Thurow 1997). Leaching of nutrients during drowning events and development of reactive Fe limitation led to an increased productivity, an expanded oxygen minimum zone (OMZ), and amplified organic matter preservation at the inception of OAE3. Whereas this loss of deep habitat reduced the deeper dwelling forms, it allowed the shallower radiolarians to thrive (Erbacher & Thurow 1997). Likewise, the expanded oxygen minimum zone at the OAE3 also destroy the deeper habitats of the planktonic foraminifera, and though this assumption has been well established for the OAE2 by Erbacher *et al.* (1996), the same can also be considered for the probable OAE3 facies. Considering the radiolarians as shallower coupled with general absence of deep dwelling foraminifera, the current facies can be explained according to Erbacher & Thurow (1997) and Erbacher *et al.* (1996). Conversely, rapid submarine eruptions offer a viable alternative for traditional climatic scenarios for high productivity of silica that are irrelevant to greenhouse-to-icehouse climatic change (Racki & Cordey 2000). Therefore, the possible volcanic activity in the Kohistan Island Arc within the eastern Tethys, explaining for nurturing of radiolarians, can advocate for the first interpretation as well. Radiolarian's occurrence and partial disappearance of foraminifera has also been reported from the Coniacian Niobrara Formation, Canada (Diaz & Velez 2018), comparable with the Coniacian radiolarians of Pugh *et al.* (2014) in the Sverdrup Basin, Arctic Canada. They believed that the high supplement of silica due to volcanic events favored the paleoecological conditions for radiolarians to flourish. Regarding the OAE3, which is the least known amongst all events of the Cretaceous, its development and paleoenvironments were dependent on regional or local conditions (Bomou *et al.* 2013). Even though Wagreich (2012) has reported local organic rich layers in Pakistan (location not mentioned) but he generally considered that the OAE3 organic deposits are largely absent from the Tethys. He assumed the OAE3 does not define a single distinct time period, spanned over a longer period and occurred in different basins in different times. The organic rich horizon of present study could either represent the Hitchwood Event of Jarvis *et al.* (2006) or possibly linked with OAE3 of Wagreich (2012) because of the longer duration of the event. However, further studies are required to better support the assumption. The upper part of the late Cretaceous is marked by foraminiferal lime mudstones representing relatively shallower deposition during Coniacian to middle Santonian, and deeper till early Campanian; according to the occurrence of globotruncanids at this level. The facies is in contact with planktonic foraminiferal wackestones with an intervening ferroan dolomitized boundary, attributed to migration of fluids along bedding planes and stylolitized sutures.

Facies over the K-Pg boundary are markedly different i.e. fine-medium grained sandstones with ferruginous matter, calcite and silica cements. Unfortunately, the boundary zone does not bear any characteristic sedimentary structures, probably due to the extreme thin nature of the zone, it does have a yellowish to brownish lateritic horizon with developed

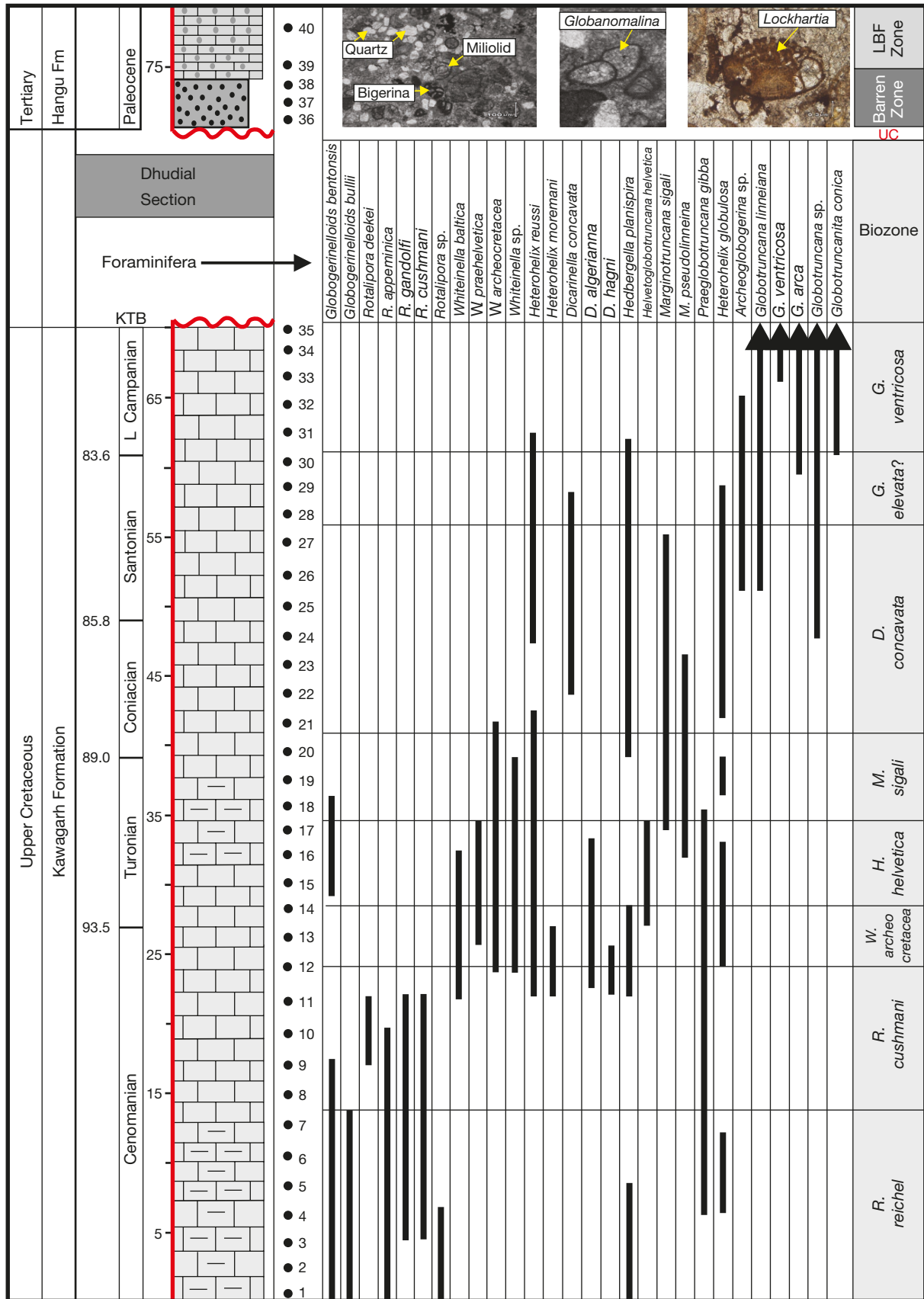


FIG. 8. — Chart showing the planktonic foraminiferal distribution and biozones of the Dhudial Section.

crystals of authigenic quartz. The texturally and chemically sub-mature sub-arkose, and moderate sorting suggest transitional marine (deltaic) environment shaping the grains into sub-angular to rounded outlines. The fine-grained, sub-arkosic texture and ferruginous nature, is supported by the laterites and bauxites of Munir *et al.* (2005), in the lower part, in continental-transitional marine environment. The deposits have been considered as residual by Umar *et al.* (2015), following intense chemical weathering of the exposed Kawagarh Formation. Elsewhere, in sections located outside the domain of the studied area, the basal parts of Hangu Formation have presence of paleosol horizons, red beds and pisolitic bauxites of Warwick *et al.* (1995), Danilchik & Shah (1987), Whitney *et al.* (1990). The widespread presence of exposed facies in the region rationalizes the continental-marginal marine paleo deposition for the facies of Hangu Formation. Dominance of monocrystalline quartz, presence of feldspars, some chert fragments and shapes of quartz grains indicate a dominantly igneous and sedimentary source rock. The upward gradation into sandy limestones, bearing Paleocene larger benthic foraminifera (LBF) such as *Lockhartia*, *Globanomalina* and miliolids, portray an inner shelf (lagoonal) depositional environment. The facies and fauna depict a sudden shift in the paleogeography from relatively steeper ramp type geometry in the early Campanian to shallow epeiric ramp (Fig. 9) in Selandian and Thanetian as manifested by the siliciclastics (Hangu Formation of this study) and widespread carbonates of the Lockhart Formation of Hanif *et al.* (2013). The morphometric change in ramp profile is believed to be because of deposition of the erosional load, during Campanian-Selandian, in deeper parts of the basin, making angle of the ramp gentler. Continued deposition led to a much gentler ramp to almost an epeiric platform by the Thanetian; subsequently depositing the Lockhart Formation. The intervening lower lateritic sandstone bed of the Hangu Formation, represents an erosional phase and consequent shallower deposition.

The facies below the K-Pg boundary zone interval in other regional sections have been reported by different workers. According to Shah (2009), the late Cretaceous facies, towards the west and southwest in the Kohat Plateau and Trans-Indus Ranges (Fig. 2), shows a deepening trend with marls and calcareous shales as its elements. Latif (1970) in Hazara area (Dhudial Section located in), and Butt (1992) in western sections, have reported Coniacian to Campanian planktonic foraminifera. Further deep towards the southwest, in the Sulaiman Range, the late Cretaceous unit of creamy white micritic and porcelaneous Parh Limestone is linked with the Kawagarh Formation (Butt 1992; Shah 2009; Khan 2013). Butt (1992) suggests a major transgression during the Coniacian-Campanian followed by a shallowing trend in late Campanian. This late Campanian regression (early Maastrichtian of Khan 2013) in Sulaiman Range is marked by carbonate breccia and presence of thick mixed siliciclastics of Mughal Kot Formation which have an overlying unconformable contact with the Paleocene strata with an intervening laterite (Shah 2009). Thus, the overall absence of late Campanian and Maastrichtian strata are conjunct with the findings of the present work. Converse to the

west and southwest, the whole of late Cretaceous is absent in the southern section's such as the Salt Ranges (Fig. 2) and the Tertiary sediments lie directly on the early Cretaceous or older rocks along a major (late Jurassic-Paleogene) hiatus. Whether or not this hiatus has any relationship with the Campanian-Selandian hiatus of present study but it is likely that uplift due to hot spot development of Garzanti & Hu (2014) during the Campanian triggered erosion of the older rocks and developed a major unconformity. In the Ladakh Himalaya (northwestern India), the facies below the K-Pg are represented by the typical basinal Kan Ji Formation characterizing Campanian- Maastrichtian terrigenous detritus as dark marls, mudstones and Zoophycus bearing sandstones (Green *et al.* 2008). This basinal unit, compared with the late Cretaceous of Pakistan, has a late Maastrichtian top part comprising shallow marine mixed-siliciclastic sediments (Marpo Limestone), cross bedded Danian sandstone (Stumpata Quartz Arenite) and late Paleocene Dibbling Limestone with abundant smaller benthos (Green *et al.* 2008). This shallowing upward facies at the Maastrichtian-Danian hiatus could be correlated with the Campanian-Selandian hiatus of the current study, however, in Hazara area, the hiatus duration was a bit longer. The late Paleocene Dibbling Limestone conforms with the transgression during Thanetian that also deposited the Lockhart Formation of Hanif *et al.* (2013). Further east in the Garhwal Himalayas, Utrarrrakhand, India, the shallow marine oomicritic limestones of late Cretaceous Nilkanth Formation (Najman *et al.* 1993) has contact with late Paleocene-Early Eocene Subathu Formation. The contact is marked by a thin (1.8 m) paleosol horizon with plant rootlets and is almost similar in duration to the present study. Wan *et al.* (2007) has described late Cretaceous (Maastrichtian) Qubeiya Formation, a conglomeratic sandstone (Quxia Formation) along a hiatus in the southern Tibet, China. In the absence of age diagnostic fauna, Wan *et al.* (2007) have 'tentatively' placed the Quxia Formation in Danian, above which the lies Thanetian Jialazi Formation with interbedded conglomerates and limestones at its base. The pause in sedimentation of current study conforms with the Campanian break of Garzanti & Hu (2014) in the Dibbling Section, Zanskar Range in Kashmir, located towards the west, not far away from the study area. These Cretaceous-Paleogene hiatuses closely matches with the present data, inferring about the widespread nature and duration of the unconformity not only in the Pakistani and Indian Himalayas but also in southern Tibet, China.

SEA LEVEL FLUCTUATIONS AND SEQUENCE STRATIGRAPHIC CHANGES

The relatively warm climate of the Cretaceous period, owing to abundant CO₂ levels and other multiple reasons (Tabara *et al.* 2017), was associated with elevated eustatic sea levels across the globe. The late Cretaceous has recorded extensive deposits of chalks, marls, planktonic limestones and development of oceanic anoxia (OAE1, OAE2 and probable OAE3) leading to development of black shales around the world (Leckie *et al.* 2002; Jenkyns 2003; Hofmann *et al.* 2003; Sageman *et al.* 2006; Voigt *et al.* 2008). The Dhudial Section represents second

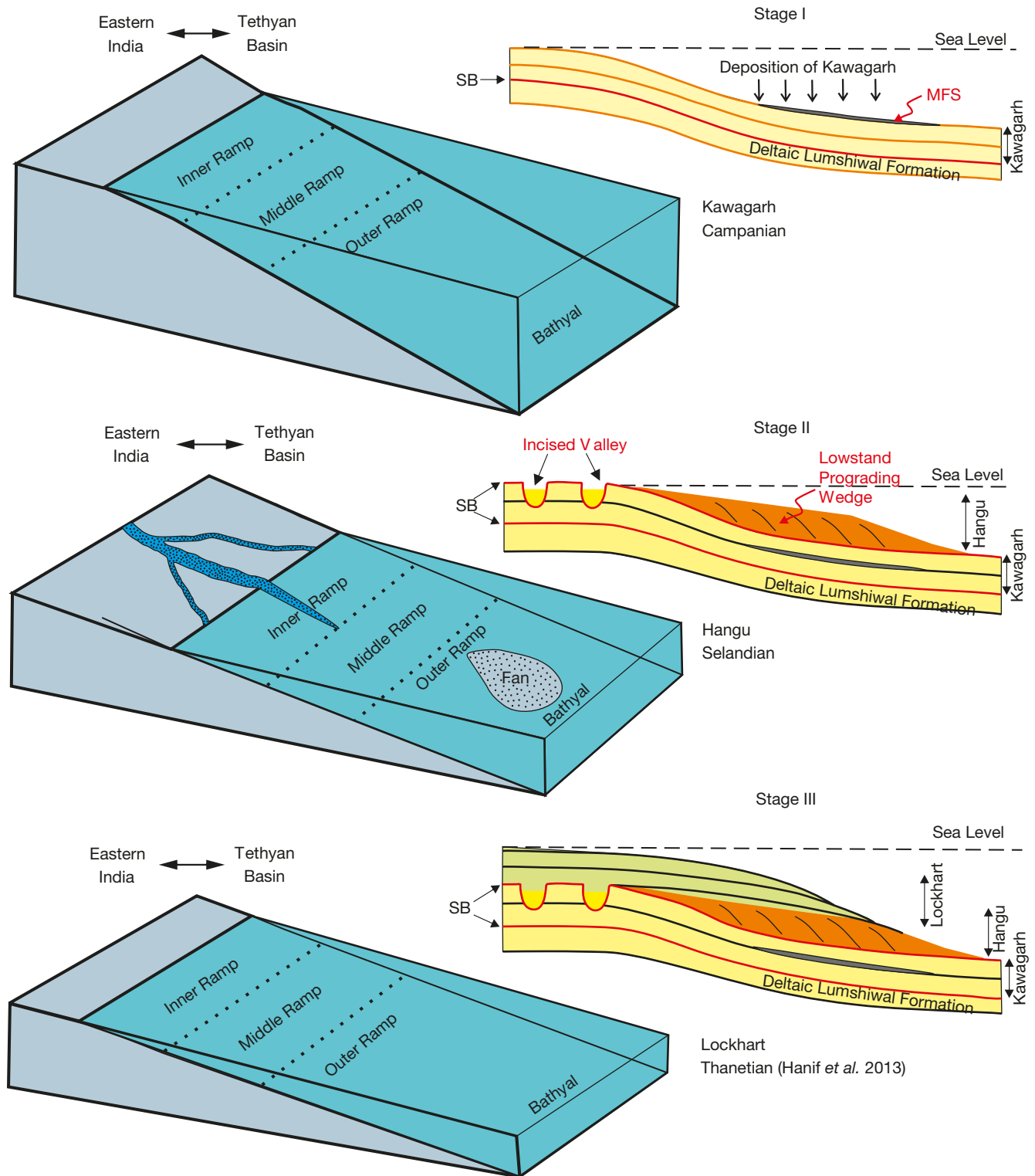


FIG. 9. — Sketch showing the platform evolution through different stages. The platform geometry in the early Campanian time (Stage I) during deposition of Kawagarh Formation was characterized by a steeper ramp. An incised valley and sequence boundary developed in Stage II during Selandian and the platform configuration became much gentler during the Thanetian (Stage III) as the accommodation space filled up.

order cycle in which the Cenomanian-Turonian boundary is associated with the OAE2 in response to increased emissions of greenhouse gases due to extreme volcanism, and reduction in supply of cold oxygenated bottom waters. However, the

OAE2 could not be traced on the outcrop due to absence of organic rich facies and the OAE2 is plausibly marked on the basis of extinction of rotaliporids. The intervening rotalipora bearing planktonic lime mudstone beds (at 12m

and 21m) are concluded as backstepping parasequence sets in progressively deepening upward trends. The eustatic sea levels correspond to 1st and 2nd order of Haq (2014), 1st, 2nd and 3rd orders of Kominz *et al.* (2008) and first order of Müller *et al.* (2008). The Turonian stage is initially marked by drops in sea level corresponding to High Stand System Tract (HST), however, the sea levels didn't fall below the shelf edge and in accordance to 1st, 2nd and 3rd order eustatic falls of Haq (2014), Kominz *et al.* (2008) and Müller *et al.* (2008). Because there is no obvious record of typical Low Stand System Tract (LST) deposits before the inception of overlying TST, a tentative type II sequence boundary could be placed surrounding the radiolarian rich wackestones near the Turonian-Coniacian boundary. However, its equivocal presence makes this sequence boundary ambiguous. The deposits surrounding this boundary experienced brief episode of TST reaching eustatic sea level highs of Kominz *et al.* (2008) and Müller *et al.* (2008), probably as result of Hitchwood event of Jarvis *et al.* (2006) or OAE3. The same time period was also marked by eustatic falls as established by Haq (2014). The onset of HST spanning middle Coniacian-middle Santonian stages continue during eustatic sea level falls of Haq (2014), but contradicts with highs of Kominz *et al.* (2008) and Müller *et al.* (2008). The interbeds of planktonic foraminiferal wackestones occur as prograding parasequence sets reflecting minor sediment production-eustacy permutations. A tentative type-II sequence boundary can be placed at middle Santonian. The rise of keeled globotruncanids in early-middle Santonian manifest inception of TST during the large-scale transgressions of Haq (2014), and continued high of Kominz *et al.* (2008) and Müller *et al.* (2008). The radiolarian rich bed depicts episode of progradational set pattern during large scale transgression. This transgression is followed by a sudden drop in sea level during early Campanian, and leads to deposition of LST induced sub-arkosic, fluvio-deltaic sandstones of the basal part of the Paleocene Hangu Formation in the middle Paleocene (Fig. 7). This drop in sea level (Haq 2014; Kominz *et al.* 2008; Müller *et al.* 2008) accounts for the development of authigenic quartz in freshwater episode leading to introduction of unconformity and incumbent LST. The drop in eustatic sea levels, however, might not be the only reason to account for the absence of Maastrichtian fauna as extinction of some marine species occurred several million years prior to the main event as conjunct with Glasby & Kunzendorf (1996). This presents an ambiguous view whether the absence of Maastrichtian fauna is related with the 'dead ocean model' of Glasby & Kunzendorf (1996), an early drop in sea levels or a combination of both. It can be inferred that in the study area this drop in sea level persisted till Selandian-Thanetian times. In either case, the lateritic sandstones of Hangu Formation represent the type-I sequence boundary (SB). While Umar *et al.* (2015) related this exposure with initial collision of the India with the Kohistan Island Arc at 67 ± 2 Ma, around Maastrichtian-Danian hiatus, but the timings of collisional events along the western margin of the Himalayas are vague. The chronological supposition of Umar *et al.* (2015) is in accordance with Petterson & Treloar (2004) but instead of

collision with India, the later believe that the initial suturing was owing to collision of the Kohistan Island Arc with Asia. The current study reveals no record of the Maastrichtian fauna and highlights a Campanian-Selandian hiatus. Linking this exposure surface and hiatus with the initial Indo-Eurasian collisions, the temporal relationship of present findings is in agreement with Petterson & Windley (1985) who indicated the timing of initial impact as 75 Ma (Campanian). Searle & Treloar (2019) and Green *et al.* (2008) suggested a major phase of shortening and uplift, associated with ophiolite obduction, preceding the Paleocene-Eocene final phase of marine deposition. This, together with the magmatic upwelling of the Indian lithosphere during Campanian (Garzanti & Hu 2014) might be considered for development of this pre-Himalayan hiatus. Though the collision driven causes of this exposure are abstruse, muddled by aging uncertainty, nonetheless it could possibly be explained by significant drop in global sea levels of Haq (2014), Kominz *et al.* (2008) and Müller *et al.* (2008). The overlying bed of sandy limestone containing Paleocene larger benthic foraminifera (LBF) indicates a deepening upward trend and initiation of TST to deposit the overlying LBF limestones of the Thanetian Lockhart Formation of Hanif *et al.* (2013). The resultant infilling of the accommodation space coupled with sea level falls might account for shift in nature and evolution of the platform.

CONCLUSIONS

Sediments deposited in the lesser Himalayas of Pakistan were thoroughly investigated for biostratigraphic, sequence stratigraphic and facies analyses in the Dhudial Section. There is a marked contrast in the facies across the K-Pg boundary zone interval; facies below represent open marine planktonic foraminiferal limestones dated from the Cenomanian to the early Campanian. These late Cretaceous beds are sharply truncated by erosion and followed by deltaic siliciclastics and marine sandy limestones, the latter being of middle Paleocene age. An early Campanian to middle Paleocene hiatus is thus evidenced. The change is also evident in faunal turnover from planktonic foraminifera to larger benthic foraminifera across the boundary zone interval. The study proposes to revise the "Nara Sandstone Member" of the upper part of the late Cretaceous Kawagarh Formation in Hazara area as Paleocene Hangu Formation and making its correlation with the Maastrichtian Pab Sandstone in the Sulaiman Ranges as debatable. On faunal basis, the lower part of the Dhudial Section represents deeper oligotrophic waters which were extinct due to development of oxygen minimum zones possibly related to OAE2, although no organic-rich interval is evidenced here. A brief period of upper photic zone was followed by large scale deeper disphotic zones abrupted by oxic zones during lowered sea levels in early Campanian. Though lower part of the Dhudial Section (Kawagarh Formation) has been deposited during TST, middle portion dominantly represents the HST. The upper portion belonging to the TST is headed by deposits of Selandian Hangu Formation of LST and late Thanetian Lockhart Formation of

TST. The middle and upper parts of Kawagarh Formation bear radiolarian rich limestones, the middle one is associated with an overlying organic rich bed. These radiolarian rich beds are believed to have been deposited in response to high levels of dissolved silica due to the growth of the Kohistan Island Arc within Tethys during the late Cretaceous. However, an origin with open marine environment, associated with the possible presence of OAE3 in the Coniacian-Santonian time periods or latest Turonian Hitchwood event when primary plankton productivity was on a high, is preferred. The abrupt change in facies is due to sea level falls and associated erosion, resulting in absence of Maastrichtian and Danian fauna and the K-Pg boundary encounters on top of the early Campanian beds, just below the sandy limestone beds of Hangu Formation of Paleocene (Selandian) age.

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