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1 Neogene to Quaternary clay mineral fluxes in the Central Indian basin

2
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8 9 **Abstract**

10 The late Cenozoic clay sedimentation in the Central Indian Basin was investigated by
11 mineralogical analyses of five very long cores (33-49 m long), which were recovered between
12 1 ° and 10°S along a transect at about 80°E and compared with data from adjacent ODP and
13 piston cores. Progressive changes occur from North to South, and are especially marked by a
14 decrease of illite and chlorite amounts balanced by increased contents of smectite and
15 kaolinite. These changes are attributed to a southwards reduction of Himalaya-derived detrital
16 fluxes, and to a correlative augmentation both of smectite reworked from coastal Indian,
17 Indonesian and volcanic submarine sources, and of kaolinite supplied from Australia and
18 blocked by the equatorial divergence. The stratigraphic changes are more diversified than the
19 geographic changes and develop discontinuously. High fluxes of illitic and chloritic materials
20 during the middle-late Miocene reflect major activity of the Himalayan Main Central Thrust.
21 A tectonic relaxation developed during the latest Miocene-early Pliocene, allowing alternating
22 supply of illite-rich and smectite-rich sediments, which suggests periodical changes from
23 distant to more local supply partly controlled by climate and by Indian intraplate
24 deformations. The more abundant and constant amounts of smectite in late Pliocene-early
25 Pleistocene sediments are chiefly attributed to erosion of Indian coastal zones favored by a
26 worldwide sea-level drop. The middle to late Quaternary increase of the illite group resulted
27 from the resumption of the tectonic activity in the Himalaya-Tibet range associated with a
28 general climatic cooling.

29 30 **Introduction**

31 The question of the nature and origin of fine-grained minerals in the clay-rich
32 sediments which constitute the major deposits of the Indian Ocean has been debated since the
33 late sixties. The first syntheses were devoted to the general distribution of the main clay

1 mineral groups, and provided preliminary arguments on their sources and genetic processes
2 (eg., Gorbunova, 1966; Griffin et al., 1968; Goldberg and Griffin, 1970; Aoki and Sudo,
3 1973; Kolla and Biscaye, 1973). Illite, chlorite and kaolinite were mostly attributed to detrital
4 supply, whereas smectite was supposed to mainly derive from submarine volcanic activity.
5 Palygorskite, which was recorded especially in the Western Indian Ocean, was considered to
6 result essentially from aeolian supply (Kolla et al., 1976).

7 Subsequent investigations were essentially performed during the eighties and
8 combined mineralogical, geochemical and electron microscopic data in order to better
9 characterize the origin of clay minerals, especially in the Northern Indian Ocean (eg.,
10 Fröhlich, 1982; Tlig and Steinberg, 1982; Rao and Nath, 1988; Bouquillon et al., 1989, 1990;
11 in Chamley, 1989; Nath et al., 1989; Brass and Raman, 1990). Special attention was paid to
12 the nature and chemical characters of smectitic minerals, which appeared to be diversified and
13 to result from several sources and processes. For instance Rao and Nath (1988) identified in
14 sediments from manganese nodule fields two smectite minerals, i.e. a montmorillonite
15 probably reworked from volcanic areas and a Fe-smectite attributed to an in situ formation in
16 the metal-rich environment. Bouquillon et al. (1989) showed, from a multidisciplinary study
17 of 400 samples from the northeastern Indian Ocean compared with data from continental
18 India, that smectites are of various types (Al-Fe, Al, Fe, Si-Fe, Na-types) and mostly
19 inherited from exposed land masses; only Si-Fe smectites and oxides seem to have been
20 formed in their host sediments. Nath et al. (1989, 1992) found evidence of terrigenous
21 influence in deep-sea sediments up to 8°S in the Central Indian Basin, and recognized obvious
22 terrigenous signals among several rare-earth element (REE) patterns. Similarly Pattan and
23 Banakar (1993) recognized from REE investigations that alumino-silicates in buried Mn
24 nodules were of detrital origin as far as 14°S and 74°E.

25 As clay minerals in the northern Indian Ocean are proved to derive often from a
26 detrital origin, their variations in space and time may be used to identify the mineral sources
27 and the paleoenvironmental controls on the distribution of fine-grained sediments. In addition
28 the N-S replacement from continental to pelagic and chemical influences allows to investigate
29 the characteristics of a major transitional sedimentary environment. As the northern Indian
30 Ocean was submitted since the early Cenozoic to major changes in the tectonic context, the
31 marine circulation, the climate and the sediment supply (in Kennett, 1982), it constitutes a
32 working area of an outstanding interest. The mineralogical distribution of clay assemblages is
33 able to provide much information, as already shown by many investigations performed on
34 surface and piston core sediments. Our purpose is to investigate a much larger time interval

1 than in previous studies, by considering the clay mineral data from sedimentary columns
2 encompassing most of late Neogene and Pleistocene periods.

4 **Materials and methods**

5 During the MD 90/SHIVA cruise of the Marion- Dufresne conducted by the Geology
6 Department of the "Museum National d'Histoire Naturelle" (Paris), five piston cores (33-49 m
7 long) were recovered along a N-S transect (from 1 to 10°S) at about 80 ° East in the Central
8 Indian Ocean (Fig. 1, Table 1), at water-depths ranging from 4.8 to 5.4 km, i.e. below the
9 modern carbonate compensation depth. The five very long cores bottomed in late Miocene to
10 late Pliocene deposits (Caulet, 1992). In order to extend the N-S transect from the Equator to
11 12.5°S, two additional cores of MD cruise 81 (Bout, 1991) and Hole 717 C of ODP Leg 116
12 (Cochran, Stow et al., 1989c) have also been considered (Table 1).

13 The five very long MD90 cores have been lithostratigraphically described visually and
14 by microscopic techniques. Biostratigraphic information was provided essentially by means
15 of radiolarian assemblages (Caulet, 1992; Sanfilippo et al., 1985; Johnson et al., 1989). The
16 nature and origin of the organic matter have been investigated on selected samples with Rock-
17 Eval and Leco equipments (Bout, 1991).

18 The mineralogical investigations have been performed on about six hundred sediments
19 sampled every 30-40cm. The clay-mineral data were obtained using X-ray diffraction of the
20 <2 µm size carbonate-free fraction (XRD). Samples were decalcified with 0.2 N HCl, and the
21 <2 µm fraction was separated by settling. Oriented mounts were prepared on glass slides and
22 were scanned three times (Cu, K, radiation): from 2.5° to 28.5° 2theta on untreated samples,
23 from 2.5° to 14.5° 2theta on glycolated samples, and from 2.5° to 14.5° 2theta on samples
24 heated for 2 hr at 490°C. Analytical techniques, determination, and quantitative estimation of
25 clay minerals are detailed in Holtzapffel (1985). Analytical uncertainties in the clay-mineral
26 abundances are ±5% (absolute) for abundances above 20%, + 3% (absolute) for abundances
27 between 5% and 20%, and ±2% (absolute) for abundances below 5%.

29 **Results**

30 *Lithology*

31 The dominant lithologies in MD 90/SHIVA cores consist of light brown, oxidized,
32 siliceous clayey mud superimposed to green-blackish, reduced, silty to sandy mud (Caulet,
33 1992). The dark color is due to organic matter (eg., 1.2-1.7% total organic carbon in core
34 943), of essentially terrigenous origin (Bout, 1991; see also Cochran, Stow et al., 1989b). The

1 sediment is statistically enriched in pelagic, biosiliceous and fine clayey components from
2 North to South and from the base to the top of the cores, (Figs. 2-4). On average the transition
3 from underlying reduced facies to super imposed oxidized facies (i.e., redox boundary) is
4 Pleistocene in age close to the equator (core 947, Fig. 3), and early Pliocene South to 6°S
5 (Fig. 2). Precise correlations do not exist between the age of the redox boundary and either the
6 latitudinal location, the thickness of the deposits or the appearance and development of
7 biosiliceous components (Fig. 2). The transition from reduced to oxidized bottom conditions
8 therefore resulted from other causes than distance from sources, sedimentation rate or marine
9 productivity.

10 The three southern cores (10-6°S) display very similar lithologic facies (Fig. 2). Their
11 lower parts, late Miocene to early Pliocene in age (944:31-40.2 m, 4.9-6.3 Ma; 943:29.3-44
12 m, <5->6.3 Ma; 942:37.03-47.90 m, > 5 Ma) consist of dark green to blackish clayey mud
13 rich in dominantly terrigenous organic matter. From 5.1-4.9 Ma upwards occur alternating
14 centimeter- to meter-thick levels of biosiliceous and silty-clayey muds (944:20.7-31 m,
15 3.8/3.9-4.9 Ma; 943:21.8-29.3 m, 4.3/4.4-5 Ma; 942:28.50-37.03 m, 3.5/3.7-5 Ma). These
16 levels are green to black below the redox boundary, and brown above (i.e., at about 4.7 Ma in
17 cores 942 and 944, and 3.8/3.7 Ma in core 943). The blackish lithofacies appear to frequently
18 result from redeposition processes. The brownish levels are enriched in clayey horizons to the
19 North, and constitute transitional facies which extend up to the end of the early Pliocene. Late
20 Pliocene and Pleistocene sediments (944: 0-20.7 m, 0.9-3.8/3.9 Ma; 943:0-21.8 m, 0.9-4.3/4.4
21 Ma; 942:0-28.5 m, 0.3-3.5/3.7 Ma) essentially comprise light brown biosiliceous, clayey
22 hemipelagites. Notice that the lithology of Pliocene-Pleistocene deposits (0-17.5 m, <0.9->4.1
23 Ma) in cores MD 81 374/375 located at 12.5°S is very similar to that of contemporaneous
24 sediments at SHIVA southern sites (Fig. 4). The sedimentation rates appear to strongly differ
25 in the three SHIVA cores during the early Pliocene, Table 2). Pleistocene deposits are 4-5 m
26 thick at Sites 944 and 943, but more than 18 m thick in core 942 (Fig. 2).

27 The lithology of core 946 (3°S) is more homogeneous than in other cores (Fig. 3).
28 Light brown oxidized facies occur throughout the sedimentary column, anoxic facies being
29 restricted to the lowermost part of the core (25.6-34.3 m, > 6.3-6.5 Ma, i.e., undated greenish
30 mud at the bottom, and one single blackish organic level at about 32 m depth). Clay-rich
31 facies occur principally in late Miocene deposits (25.6-34.3 m >6.3/6.5 Ma), most other
32 sediments being enriched in biosiliceous debris (0-25.6 m, < 1-6.3/6.5 Ma). The
33 sedimentation rate at site 946 is similar to that of core 942 (6.5 m/m.y.) and much lower than

1 in cores 943 and 944 during the early Pliocene, and similar to that of the three SHIVA
2 southern cores (4-5.5 m/m.y.) during the late Pliocene (Table 2).

3 Core 947, which was recovered to the northwest of site 946 at 1.5°S, surprisingly
4 displays very heterogeneous and variable lithofacies (Fig. 3). Most of the 33.9 m thick
5 sedimentary column corresponds to alternating greenish and blackish reducing clayey to silty
6 muds. A few calcareous reworked facies are interbedded in the reducing sediments, and
7 probably result from slumping on adjacent submarine highs such as the Chagos- Laccadive
8 Rise (Cochran, 1990). Brownish oxidized, biosiliceous facies occur only in the 4 m deposited
9 during the middle-late Pleistocene since 0.17 Ma. The late Pliocene and Pleistocene
10 sedimentation rates at site 947 are much higher than at southeastern site 946 (Fig. 3).

11 ODP Hole 717C, drilled during Leg 116 at 0.5°S and 81°E to the north of the SHIVA
12 core transect in the distal part of the Ganges deep-sea fan (Fig. 1), is dominated by reworked
13 sediments largely derived from the North (Cochran, Stow et al., 1989a). Latest Miocene
14 (Messinian) to Pleistocene deposits are more than 500 m thick and mostly comprise sandy to
15 clayey turbidites interbedded with hemipelagites (Fig. 4). The turbidites are either of light or
16 dark gray color, and include some terrigenous organic matter. The coarsest lithofacies
17 deposited during the middle- late Miocene and the middle-late Pleistocene, which corresponds
18 to the highest sedimentation rates (up to 350 m/Ma in the late Pleistocene).

19 *Clay mineralogy*

20 Southern sites

21 The cores SHIVA 942, 943 and 944 (Fig. 1) contain very similar clay assemblages in
22 the < 2 μ m sedimentary fraction (Fig. 2). Smectite minerals constitute the dominant species
23 (35-90%) followed by illite (5-45%), kaolinite (5-25%), chlorite (0-10%) and random mixed-
24 layers (0-10%, mainly of illite-smectite types). Associated non clay minerals include
25 ubiquitous quartz and frequent feldspars, as well as occasional other species (e.g., pyrite, iron
26 oxides).

27 *Geographically* the number and intensity of illite pulses are decreasing towards the
28 south (Fig. 2). By contrast the average percentages of smectite and of kaolinite increase
29 progressively southwards. The same trend extends in a larger geographic area, smectite and
30 kaolinite amounts continuing to increase south of 10°S in MD 81 cores 374/375 (Fig. 4a).
31 There is no general relationship between lithology and clay mineralogy, and the redox
32 transition which occurs at different periods in the three cores is not associated with any
33 durable change in clay mineral proportions.
34

1 *Stratigraphically*, two successive clay mineral zones are recognized in late Cenozoic
2 sediments of SHIVA cores 942 to 944 (Fig. 2). From the late Miocene (ca. 6.3 M.a.) to early
3 Pliocene (3.7-3.5 Ma) strong alternations of smectite-rich and illite-rich layers occur,
4 characterizing the *mineralogical zone 1*. The percentages of kaolinite also fluctuate in this
5 zone, but independently of those of the two other clay groups. This suggests distinct origins
6 for the three main clay minerals. Notice that the illite and smectite alternations recorded in
7 SHIVA cores from 1.5 to 10°S do not exist any longer more southwards, as shown by their
8 absence in early Pliocene sediments deposited at 12.5°S in MD 81 cores 374 and 375 (Fig. 4a,
9 Bout, 1991).

10 The mineralogical zone 1 can be divided in two subzones. *Subzone la* corresponds to
11 late Miocene-earliest Pliocene series, in which the black, organic-rich, often redeposited
12 levels correlate with higher percentages of smectite. As the organic matter is essentially of a
13 continental nature (Bout, 1991), this suggests the dominantly terrigenous origin of the clay
14 mineral. *Subzone lb* corresponds to sediments progressively enriched in biosiliceous
15 components, interbedded with silty clayey detrital pulses usually enriched in illite, chlorite
16 and random mixed-layers. The illite group in subzone lb corresponds dominantly to
17 resedimented deposits, whereas smectite-rich levels rather characterize hemipelagites. The
18 relative amounts of the sum of illite + chlorite + random mixed-layers are higher in the
19 mineralogical zone lb (late early Pliocene) compared to surrounding zones la and 2. Notice
20 that the transition between mineralogical subzones I a and 1 b in cores 942 and 944 occurs
21 during the early Pliocene ca. 4.9-4.7 Ma ago, at a period marked by a high sea-level stage
22 (Haq et al., 1987). The uppermost levels of subzone 1 a are devoid of chlorite, a mineral
23 commonly associated with illite in Himalaya-derived products (Bouquillon et al., 1989,
24 1990).

25 *Mineralogical zone 2* encompasses the late Pliocene (from 3.6 m.y. ago) and
26 Pleistocene stages. Illite-smectite alternations have disappeared and been replaced by a fairly
27 homogeneous and constant clay mineral suite largely dominated by smectite (70-80% of the
28 clay fraction; Fig. 4). Smectite amounts reach maximum values in late Pliocene sediments. In
29 the course of the Pleistocene the illite group amount increases slightly again, and is associated
30 in core 942 (10°S) with a moderate augmentation of kaolinite.

31

32 Equatorial sites

33 Average percentages of illite and chlorite are higher at northern locations than at
34 southern sites, whereas those of smectite and kaolinite are lower. Respectively located at 3 °

1 and 1.5°S, the cores SHIVA 946 and 947 display different clay mineral zonation (Fig. 3).
2 Core 946 (78°E), which consists almost exclusively of oxidized lithofacies, comprises the
3 mineralogical zones 1 (alternating illite-rich and smectite-rich levels) and 2 (more constant
4 smectite-rich deposits) which have already been recognized at southern sites 942 to 944 (Fig.
5 2). The transition between zones 1 and 2 occurs in the middle part of a lithologic unit
6 composed of biosiliceous clayey muds, stressing the absence of any close correspondence
7 between lithology and clay mineralogy. The boundary is located close to the early Pliocene-
8 late Pliocene passage. Zone 1 can be divided in two subzones 1a and 1b as in cores to the
9 South, the boundary being located at the transition from greenish reducing clayey muds to
10 brownish oxidized biosiliceous muds. In addition it is possible to identify a mineralogical
11 zone 3 in the upper 4 m (Pleistocene) of core 946, which is marked by significantly increased
12 amounts of illite and chlorite.

13 Core 947 (76°E), which essentially comprises reducing lithofacies, is marked by an
14 unusual extension of mineralogical zone 1 (Fig. 3). Smectite-rich levels and sediments
15 enriched in illite, chlorite and often kaolinite alternate throughout this 34.4 m long core. Only
16 the uppermost 3 m of Pleistocene sediments display oxidized facies marked by more
17 homogeneous clay assemblages, which are referred to as mineralogical zone 2. A total of 75%
18 of the dark levels containing organic matter of a dominant continental origin are enriched in
19 smectite, supporting the mainly terrigenous origin of the clay mineral. A few calcareous
20 turbidites rich in illite suggest reworking from submarine highs. For example since the
21 Middle Miocene the bioclastic turbidites from the Chagos-Laccadive aseismic ridge are
22 enriched in primary minerals from the Indus discharge (Weser, 1974; DSDP Leg 23, Site
23 219).

24 Hole 717C of ODP Leg 116 is located northeast of cores 946 and 947 and very close
25 to the equator (Fig. 1) and shows an amplification of the mineralogical changes recorded at
26 the SHIVA cores (Bouquillon et al., 1990; Brass and Raman, 1990). The average contents of
27 Himalaya-derived clay minerals (illite and chlorite) display both very high values (up to 80%
28 of the clay minerals) and strong temporal variations (Fig. 4b). Despite these differences with
29 SHIVA core data, a similar zonation can be established. After middle to late Miocene strong
30 fluxes of illite and chlorite attributed to intense tectonic activity in the Himalaya-Tibet range
31 (Bouquillon et al., 1990; Aoki et al., 1991), the latest Miocene-early Pliocene shows
32 alternating smectite/organic matter-rich and illite-rich levels which resemble mineralogical
33 zone 1. Late Pliocene to early Pleistocene deposits contain more constant and often smectite-
34 enriched clay assemblages evoking the mineralogical zone 2. A systematic increase of the

1 illite group is recorded from mineralogical zones 1 to 2 when moving from southern site MD
2 946 to the northern site ODP 717. The middle to late Pleistocene sediments display a further
3 strong increase of the illite group which resembles with much more intensity the
4 mineralogical zone 3 identified in core 946 (Fig. 3). Notice that the kaolinite percentage
5 increases at Site 717 relative to southern sites, suggesting that this mineral here results from
6 sources located around the northern part of the Indian Ocean. The quantitative variations of
7 kaolinite often parallel those of smectite and are in opposition with illite and chlorite
8 percentages (Fig. 4b).

9

10 **Discussion**

11 *Major clay mineral sources and distribution*

12 In the Central Indian Basin the amounts of *illite*, *chlorite* and *random mixed-layers* as
13 well as the number of illite pulses increase to the north throughout the late Cenozoic and
14 reach maximum values in the Bengal deep-sea fan where Ganges/Brahmaputra derived
15 channels feed directly the marine sedimentation. Numerous references document the high
16 abundance of these minerals in the Himalaya-Tibet mountain range (metamorphic and
17 magmatic rocks) and their river drainage systems (e.g. in Bouquillon et al., 1989, 1990;
18 Chamley, 1989; Brass and Raman, 1990; Raman et al., 1993). As a consequence the main
19 modern source of illite, chlorite and associated random mixed-layers (three minerals referred
20 to as the "illite group") is located to the north of the Indian Ocean in Himalayan chains. The
21 southwards decrease of the abundance of these minerals reflects their dispersion among
22 plankton-derived components and fine mineral particles derived from other sources. The
23 Himalayan influences, which are clearly recorded as far as 10°S (core 942, Fig. 2), especially
24 by repeated fluxes of the illite group in late Miocene-early Pliocene deposits (see also Nath et
25 al., 1989), are lacking beyond this latitude in cores 374/375 (Fig. 4, a) suggesting (1) that the
26 equatorial divergence acted during this period as a barrier against northern terrigenous inputs
27 or (2) an ultimate stage of dilution.

28 Contrary to the illite group, the *kaolinite* amount tends to increase significantly
29 towards the south (Fig. 4), suggesting distant terrigenous sources located southeast of the
30 Central Indian Basin. The most probable source of kaolinite lies in the soils and rocks of
31 Western Australia which are eroded by westwards winds, as documented in Pye (1987) (see
32 also Bouquillon et al., 1989; Chamley, 1989). The particularly high amounts of kaolinite (up
33 to 25%) in SHIVA cores 942 (10°S) and 374/375 (12.5°S), relative to the more northern cores
34 (SHIVA 943-947, Fig. 3), could reflect the role of an oceanic barrier played by the equatorial

1 divergence relative to the transportation of clay minerals towards the north (see Fig. 6).
2 Notice that the fairly high amounts of kaolinite in ODP Hole 717c sediments drilled in the
3 Ganges deep-sea fan result from other sources (Indogangetic plain and eastern coast of India,
4 Sri Lanka (Bouquillon et al., 1989)) which are close to those of smectite, since kaolinite
5 fluctuations mostly parallel smectite variations (Figs. 4, 6, 7).

6 As documented by previous investigations, *smectite* minerals in Northern Indian
7 Ocean sediments largely result from detrital supply and transportation (see Introduction
8 section). Sources of smectite are diversified and vary according to the chemical nature of the
9 mineral (e.g., Bouquillon et al., 1989). In the Central Indian Basin a terrigenous origin for a
10 large part of smectite is supported by frequent increases of the mineral and associated
11 essentially land-derived organic matter. The occurrence of smectite-rich sediments in
12 redeposited facies (slumps, etc) corroborates such an assertion. As redeposited facies rich in
13 smectite especially appear to occur in western cores (e.g. SHIVA 947; see also Bouquillon et
14 al., 1989), a dominant source of smectite is supposed to be located in Southern India and Sri
15 Lanka. The contribution of the Deccan Traps is always present (Amano et al., 1992; Aoki et
16 al., 1991). The mineral is especially produced by weathering of Indian basalts and other rocks
17 submitted to tropical humid climate. Smectite is known to be abundantly supplied on the
18 eastern Indian margin from weathered rocks outcropping in India (Raman et al., 1992). Other
19 sources consist of Indo-Gangetic plains, Indonesia and reworked products of altered
20 submarine volcanics (Bouquillon et al., 1989). The southwards increase of smectite therefore
21 appears to result from both a diminution of Himalaya-derived products (illite group) and an
22 augmentation of materials from alluvial plains and volcanic areas. In addition differential
23 settling processes tend to favor the deposition of smectite in open sea environments. The
24 proportion of authigenic smectite in the clay assemblages of the northern part of the Central
25 Indian Basin remains indeterminate, and this question needs further geochemical
26 investigations.

27

28 *Continental control of clay mineral successions*

29 The very large abundance of the illite group, which is recorded at ODP Hole 717C in
30 the *middle-late Miocene* (Fig. 6b) and at other ODP Leg 116 Sites since the early Miocene
31 (Cochran, 1990), reflects a major erosion stage in the Himalayan mountain range (Fig. 6a).
32 An uplift phase linked to major activity in the Himalayan Main Central Thrust induced strong
33 continental erosion and rapid growth of the Ganges deep-sea fan (Gansser, 1966; Curray and
34 Moore, 1971), development of turbidite facies (Stow et al., 1990), and high fluxes of sand-

1 sized feldspars and micas (Copeland et al., 1990). The clay mineral expression of this major
2 step of Himalayan collision consists in abundant illite and chlorite supply in the Central
3 Indian Basin (Bouquillon et al., 1990; Brass and Raman, 1990; Aoki et al., 1991).

4 During the upper part of *late Miocene* and the *early Pliocene* the fluxes of Himalaya-
5 derived illite group still remained important but diminished in abundance and alternated with
6 smectite inputs (mineralogical zone 1; Figs. 2, 3, 4). The decrease of illite and chlorite supply
7 reflects a relaxation stage of the Himalayan tectonic activity (Gansser, 1966). As smectite is
8 often associated with redeposited lithofacies and land-derived organic matter, it is considered
9 to frequently result from the supply to the ocean of soil-derived materials. Terrigenous
10 smectite is a mineral primarily formed in soils of the downstream zones of river basin systems
11 under warm climate (in Chamley, 1989). Kaolinite, which is often associated with smectite
12 detrital discharges, especially at ODP Site 717 (Fig. 4b), supports the hypothesis of soil-
13 derived mineral fluxes. The decrease of tectonic activity during the latest Miocene-early
14 Pliocene therefore allowed clay assemblages to express the tropical climatic conditions which
15 characterized the Indian subcontinent. The abundance of detrital smectite around southern
16 India and Sri Lanka (Bouquillon et al., 1989; Raman et al., 1992), and the northwestern origin
17 of smectite-bearing slumps and turbidites at ODP Sites drilled in the Ganges distal deep-sea
18 fan (Stow et al., 1990; Sager and Stuart, 1990), suggest that smectite inputs largely derive
19 from eastern and southern Indian land masses (Fig. 6b). According to Aoki et al. (1991), the
20 dioctahedral Fe-beidellite in the ODP Leg 116 originated largely from the augite-basalt on the
21 Indian Deccan Traps. As a consequence the clay sedimentation in the northern part of the
22 Central Indian Basin at the Miocene-Pliocene transition is mainly attributed to alternating,
23 aperiodic, detrital fluxes issuing from the North (Himalayan range) and from the West (South
24 India). Additional illitic inputs could have been determined by the lateral migration of feeding
25 channels of the deep-sea fan (Yokohama et al., 1990). In the southernmost part of the study
26 area beyond 10°S the illite/smectite alternations are no longer identified, which probably
27 results from both the dilution of north-derived detrital fluxes in the pelagic biosiliceous
28 sedimentation, and the role of barrier to northern supply played by the equatorial divergence
29 (Fieux, 1987). Notice that current high-resolution investigations suggest that decimetric,
30 second-order illite/smectite fluctuations are controlled by earth's orbital cycles (N. Fagel,
31 pers.comm., 1992).

32 *The late Pliocene-early Pleistocene* period (Fig. 6c), which is marked in most cores by
33 more abundant and more constant amounts of smectite in the clay sedimentation
34 (mineralogical zone 2), corresponds to strong sea-level fluctuations resulting in a global drop

1 (Prell, 1982; Zimmerman et al., 1984; Haq et al., 1987). These sea level fluctuations favored
2 the active erosion of coastal soils and sedimentary materials accumulated on the shelf, and
3 primarily the smectite-rich soils and sediments of Southern India and Sri Lanka regions which
4 are closest to the study area (Bouquillon et al., 1989; Raman et al., 1992). Such reworking of
5 coastal pedogenic and shelf sedimentary materials is consistent with the association in deep-
6 sea deposits of smectite (and partly of kaolinite) with land-derived organic matter and with
7 nannofossil shells reworked from the western upper margin (Stow et al., 1990). Of course
8 smectites were also eroded from other coastal regions surrounding the Gulf of Bengal (Indo-
9 Gangetic plain, Indonesia, submarine volcanic highs; Fig. 5), which is expressed by
10 differences in the particular chemical composition of this mineral group in the Northeastern
11 Indian Ocean (in Bouquillon et al., 1989, 1990).

12 *The middle-late Pleistocene* corresponds, from about 0.8 Ma ago, to increasing supply
13 of illite and chlorite minerals, particularly in cores located to the north of the study area (MD
14 946 and 947, ODP 717C. Figs. 3, 4). This mineralogical change correlates with a resumption
15 of activity along the Main Central Thrust in the Himalayan range. The illitic flux reflects this
16 important tectonic activity (Aoki et al., 1991; Fig. 6d), as does the intense turbidite deposition
17 in the Ganges deep-sea fan (Gansser, 1966; Stow et al., 1990). Sea-level drops related to
18 global climatic cooling (Pisias and Moore, 1981) during Pleistocene times favored the direct
19 erosion of rock-derived products relative to soil-derived products (in Chamley, 1989), which
20 still increased the supply to the ocean of illite and chlorite. The absence of significant clay
21 mineral changes at this time in the southernmost cores (e.g. SHIVA 942, Fig. 2) is somewhat
22 surprising, and suggests that a submarine oceanic circulation control of the clay sedimentation
23 partly obliterated the continental control.

24

25 *Submarine control of clay mineral successions*

26 In addition to the sedimentary formation of *autochthonous smectite* (especially Fe-
27 smectite associated with iron-oxides), which is supposed by several authors (e.g., Fröhlich,
28 1982; Bouquillon et al., 1989; Rao and Nath, 1988; Nath et al., 1989) but is still unquantified,
29 the distribution of clay mineral assemblages in the Central Indian Basin appears to partly
30 depend on submarine circulation processes. As already stated above, the equatorial
31 divergence seems to have played an essential role in the dilution of clay particles among the
32 siliceous planktonic remains, in impeding the southwards transfer of Himalaya-derived
33 minerals (e.g., illite, chlorite) and the westwards and northwards transfer of Australia-derived

1 minerals (e.g., kaolinite). It allows also lateral mineral transportations through its migration in
2 the course of the time (see Discussion section: mineral sources and distribution).

3 Some other lithological and clay mineralogical features in the Central Indian Basin
4 appear to depend on morphological and tectonical characteristics. The progressive
5 replacement of anoxic facies by oxidized facies, which takes place at different periods
6 according to the core location and is neither related to the development of biosiliceous facies
7 nor to the clay mineral composition, sets up an intriguing problem. The major lithological
8 changes seem to strongly depend on the intraplate deformations which affected the Northern
9 Indian Ocean because of the India-Asia collision processes (Sclater and Fisher, 1974; Weissel
10 et al., 1980; Geller et al., 1983; Stein et al., 1989; Cochran, Stow et al., 1989c; Karner and
11 Weissel, 1990; Leger and Loudon, 1990). For instance the early development (late Miocene-
12 early Pliocene) of oxidized lithofacies in core SHIVA 946, opposed to their very late
13 appearance (middle Pleistocene) in core 947 probably resulted from the location of the two
14 cores on either side of a submarine relief. Such a submarine high could have favored
15 oxidation and low sedimentation at the eastern site, and active, organic- and turbidite-rich
16 sedimentation at the western site. The abundance and strong variations of the Himalaya-
17 derived illite group at site SHIVA 947 probably resulted from the same cause, feeding
18 channels of the Ganges deep-sea fan running preferentially through Site 947 area. A similar
19 explanation has been put forward to explain the very different sedimentation rates recorded
20 during the Pliocene at ODP Sites 717 and 718 (Table 2), which are located only 10 km apart
21 in an area of active faults and tilted blocks (Cochran, 1990).

22 The deformation due to the Indian intraplate compression also favored structural
23 instability and earthquakes on the continental margins, responsible for turbiditic currents and
24 slumpings into the adjacent oceanic basins. Such sediment reworking developed especially on
25 the eastern India margin and on the flanks of submarine highs (Stow et al., 1990). The fluxes
26 of smectite-rich sediments, which are particularly developed in the easternmost SHIVA core
27 947 located at the foot of the Chagos-Laccadive rise, could result from such gravity flows
28 controlled by plate compression (Fig. 7). Finally one should notice that the clay mineral
29 associations are very similar from late Miocene to Pleistocene times in SHIVA cores 942, 943
30 and 944, which are located on a unique structural block bordered by Indrani and Indira
31 fracture zones (Fig. 7). This similarity also suggests a regionally developing structural control
32 of the clay sedimentation.

33 34 **Conclusion**

1 (1) Mineralogical, lithological and biostratigraphical investigations on five very long
2 cores recovered along a 80°E transect in the northern part of the Central Indian Basin,
3 compared with data from previous studies, allow to clarify the different factors controlling the
4 clay sedimentation since the late Miocene. The geographic and stratigraphic distribution of
5 clay mineral associations depend on both *continental and oceanic environmental factors*.

6 (2) The *clay mineral sources* are located in both *continental and submarine domains*.
7 Most minerals are reworked from land, continental margins or
8 aseismic ridges. The amount of authigenic minerals and especially of smectitic minerals is
9 still unknown, but does not seem to be dominant as stated by the close correlation recorded
10 between smectite contents and essentially terrigenous organic matter pulses. Illite, chlorite
11 and random mixed-layers come essentially from the Himalaya-Tibet range, through the river
12 systems and the Ganges deep-sea fan tributaries. Kaolinite is dominantly supplied from
13 Australia by aeolian and marine current, and accessorially from southern India through soil
14 erosion. Smectite is abundantly reworked from weathering profiles developed on southern
15 India basalts and other rocks, as well as from the upper continental margin of eastern India
16 and Sri Lanka. Additional smectite is supplied from the Indo-Gangetic plain, Indonesia and
17 submarine volcanic areas.

18 (3) *The Himalayan tectonic activity* is responsible for two major inputs of the illite
19 group in the marine clay sedimentation, first during the early to late Miocene, and secondly
20 since the middle Pleistocene. These inputs, which induced the obliteration of other
21 paleoenvironmental messages born by clay assemblages, were determined by renewed erosion
22 of the mountain range due to activation of the Main Central Thrust. In addition to continental
23 tectonics in Asia, *Indian Ocean intraplate movements* resulted during the late Cenozoic in
24 important local changes in the redox budget, the sedimentation rates and illitic fluxes, as well
25 as in discharges of smectite- and organic matter-bearing sediments reworked during the late
26 Miocene-early Pliocene from eastern India margins and submarine highs.

27 (4) The expression by clay assemblages of *continental climate* is expressed only
28 during tectonic relaxation stages, especially in late Pliocene-early Pleistocene times where
29 abundant smectite associated with kaolinite reflects the tropical-humid conditions responsible
30 for active chemical weathering in South India.

31 (5) The clay mineral sedimentation reflects the *sea-water movements* in two ways. The
32 sea-level drops developing since the late Pliocene globally favored the reworking of smectite
33 accumulated in coastal soils and on the upper part of continental margins. Furthermore the
34 equatorial divergence acted regionally at different latitudes according to the period as a hydro

1 dynamical barrier against the transfer to the south or to the north of suspended minerals, and
2 induced the mixture in the sediments of land-derived minerals and plankton-derived siliceous
3 biogenics.

4 (6) High-resolution mineralogical and geochemical investigations currently per
5 formed should help in answering some *unsolved questions* such as the quantification of
6 authigenically formed smectite, precise petrographic sources of clay minerals, and earth's
7 orbital control on late Cenozoic clay sedimentation.

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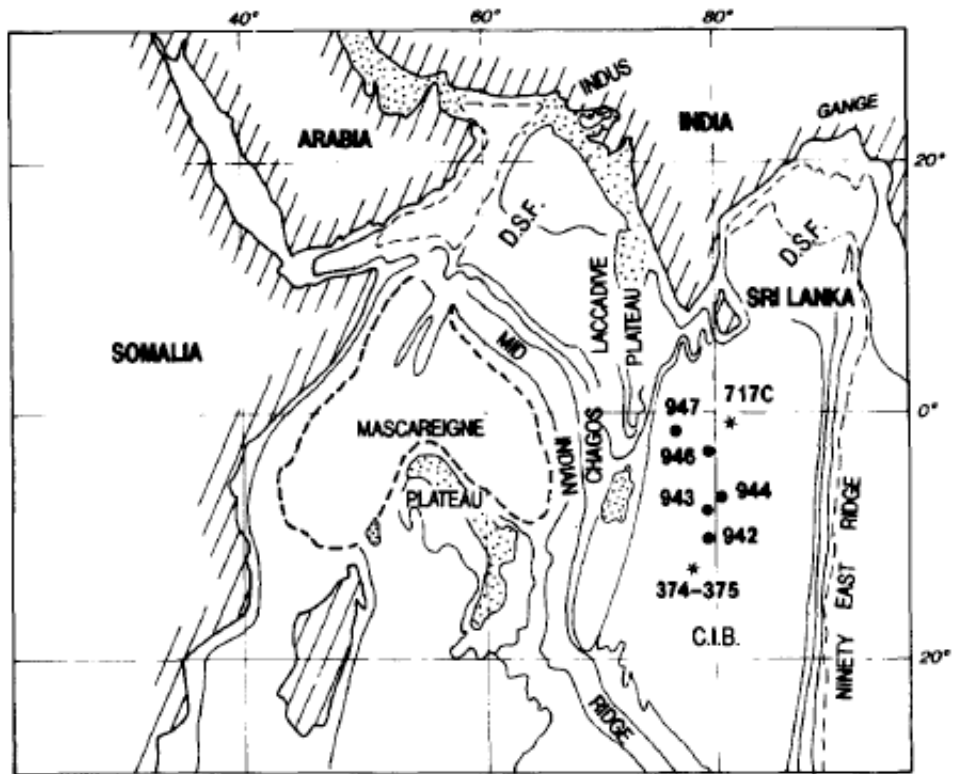
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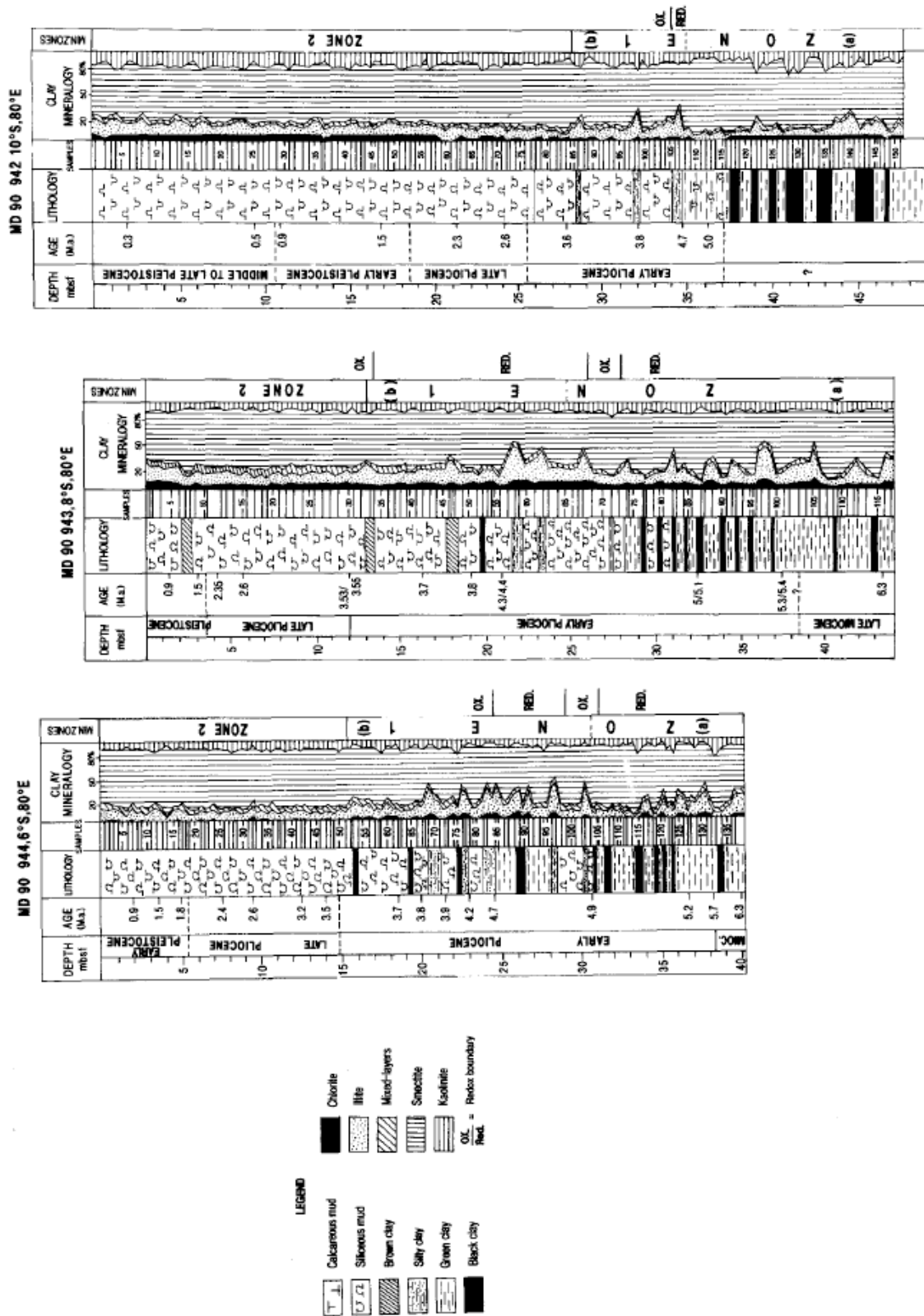
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1 Figure Captions

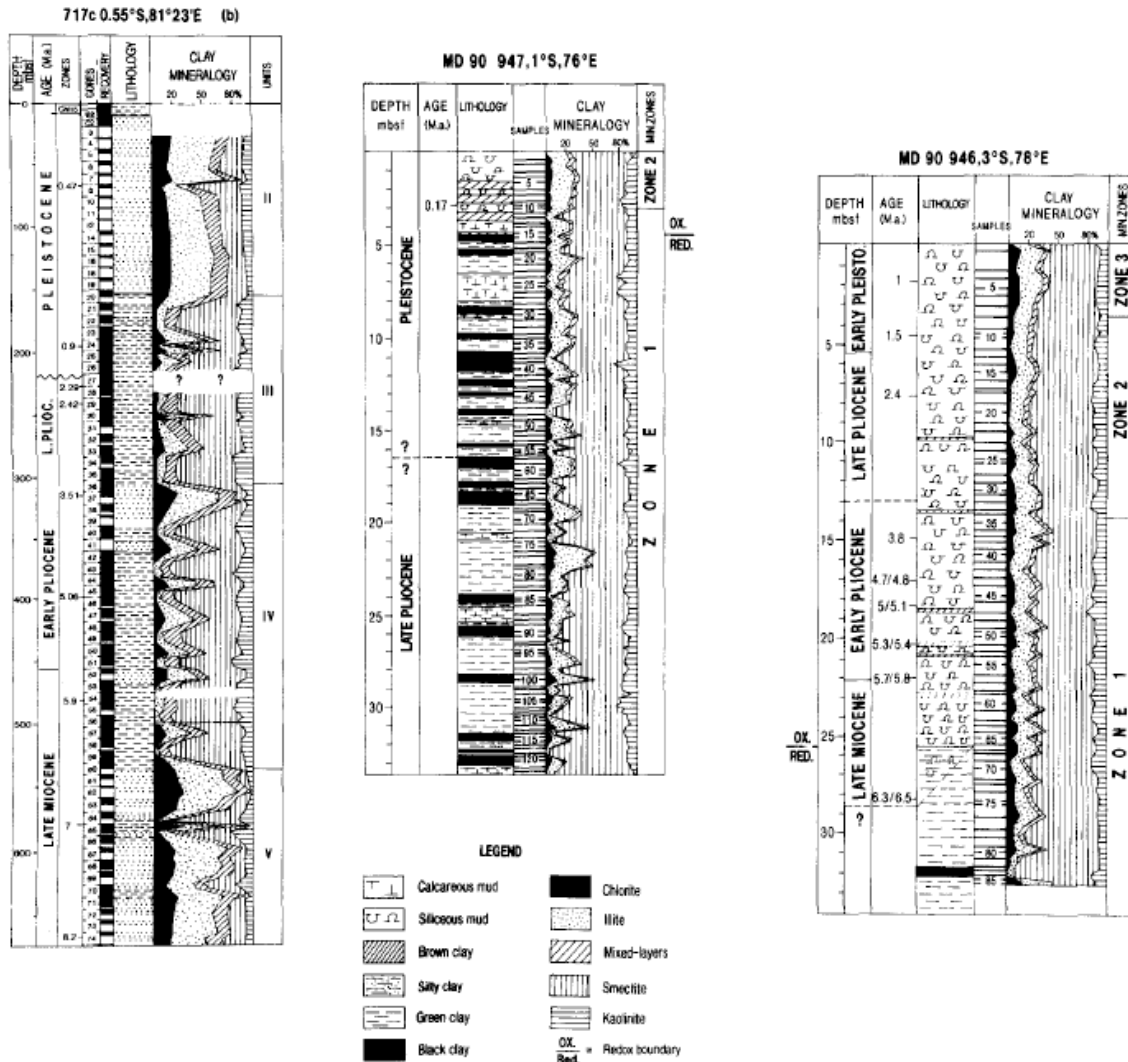


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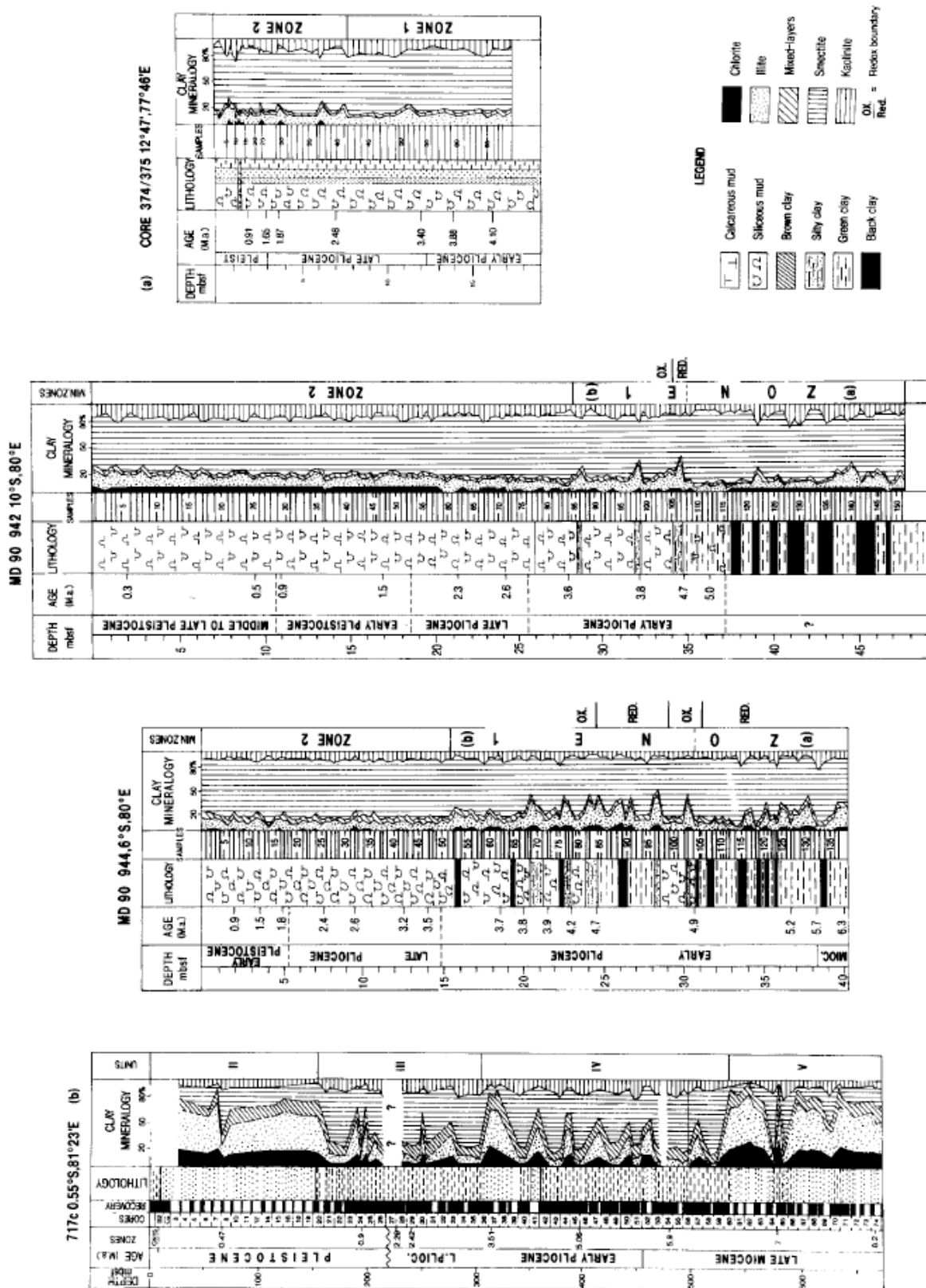
3 Fig. 1. Location of the five very long SHIVA cores (MD 90-942 to -947) recovered in the
4 Central Indian Basin (C.I.B.), and of reference cores. D.S.E = deep-sea fans.



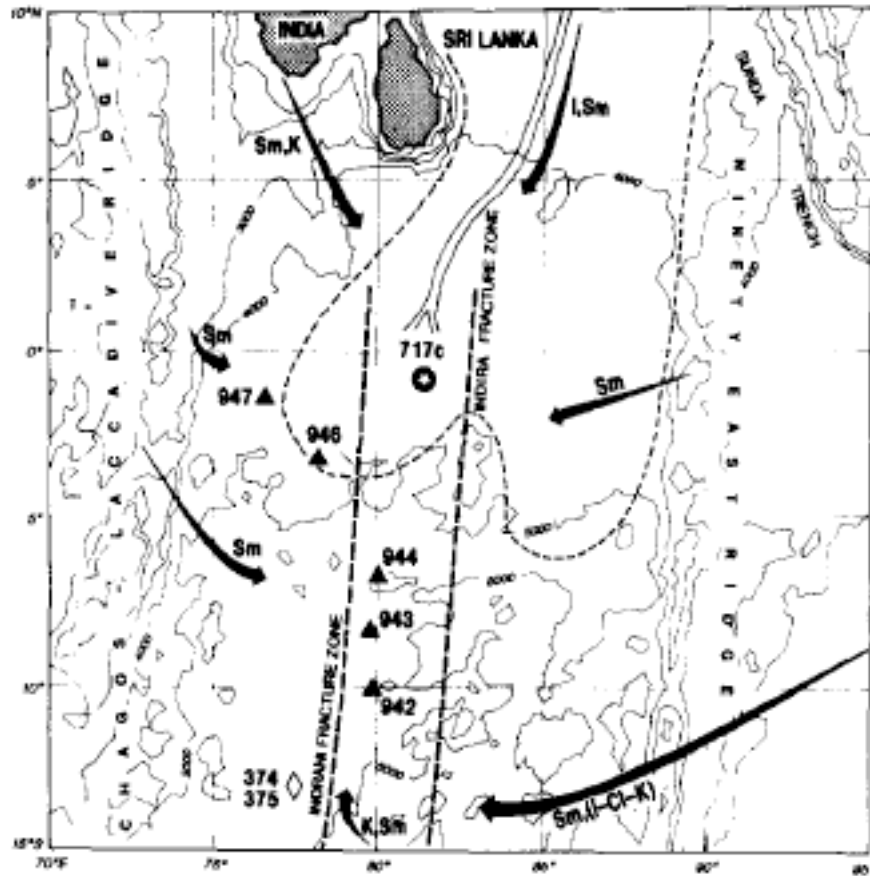
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 2 Fig. 2. Detailed lithology and clay mineralogy of southern SHIVA cores (MD 90-942 to -944)
 3 in the Central Indian Basin between 6° and 10°S at 80°E. Age (in Ma) deduced from
 4 radiolarian associations. Zonation from clay mineral successions. *OX./RED.* = redox
 5 boundary.



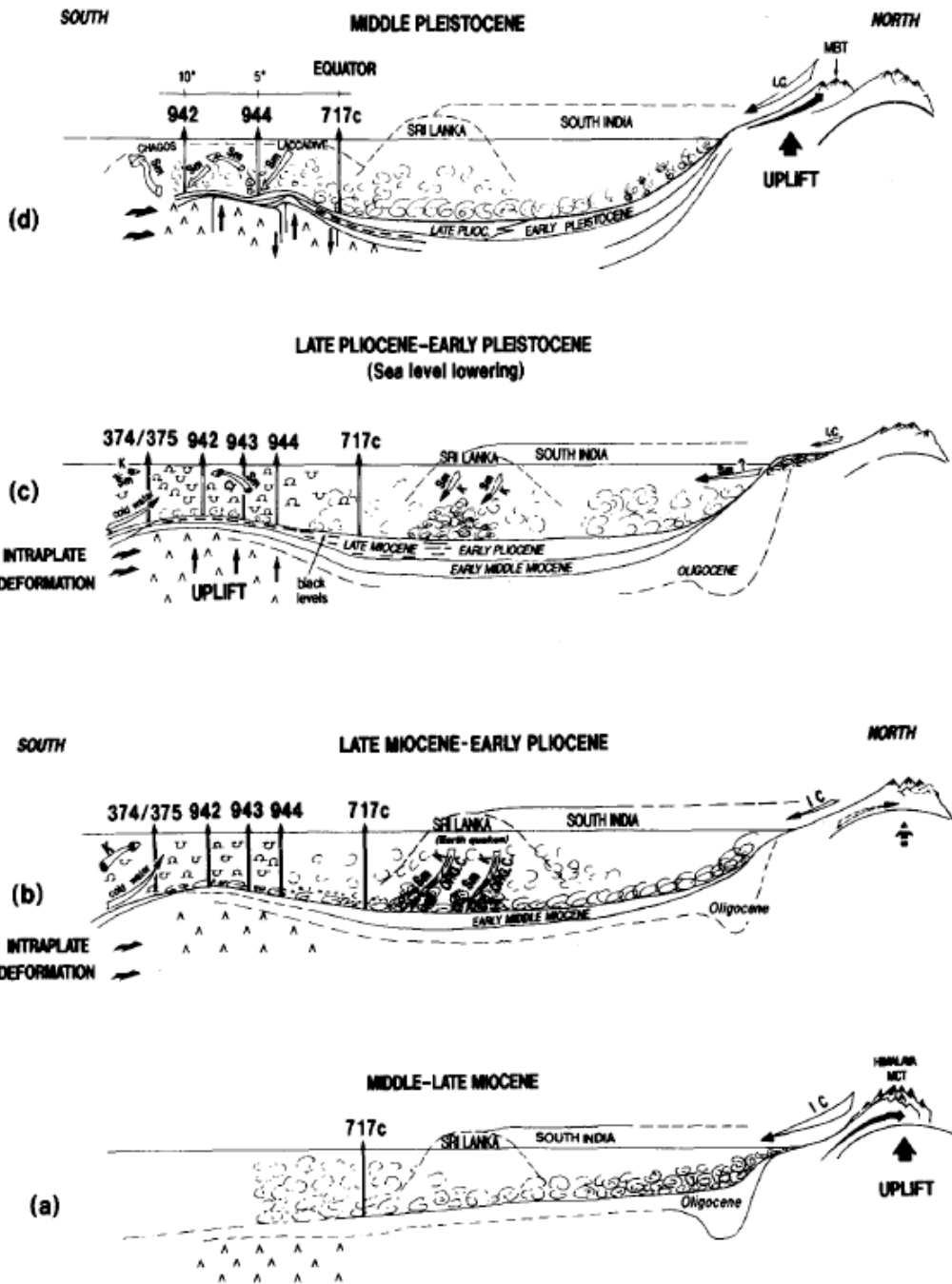
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 2 Fig. 3. Detailed lithology and clay mineralogy of northern SHIVA cores (MD 90-946 and -
 3 947) in the Central Indian Basin between 3° and 1°S at 78° and 76°E. Age (in Ma) deduced
 4 from radiolarian successions. Zonation from clay mineral successions. OX./RED. = redox
 5 boundary; comparison with ODP Hole 717C (0.5°S; Bouquillon et al., 1990).



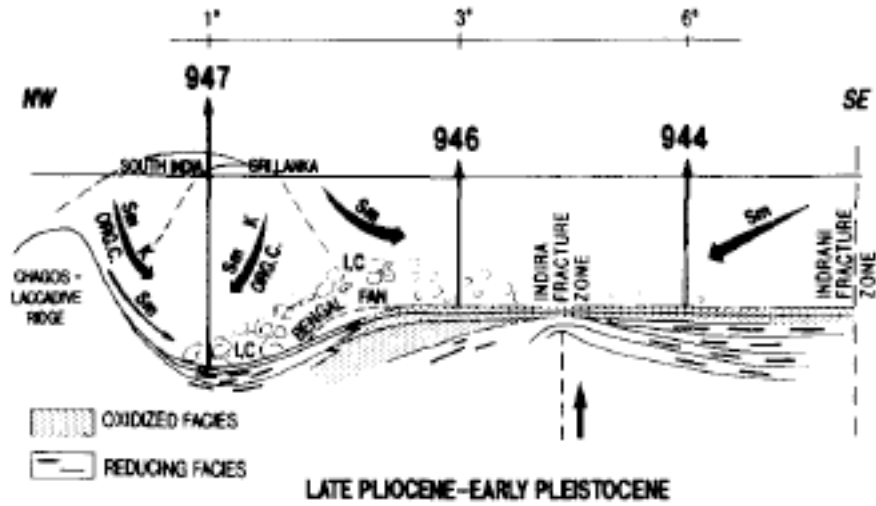
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 2 Fig. 4. Detailed lithology and clay mineralogy in SHIVA cores 942 (10°S), 944 (6°S) and in
 3 cores MD 81 374/375 (12°S ; Bout, 1991) ; comparison with ODP Hole 717C (0.5°S ;
 4 Bouquillon et al., 1990).



1
 2 Fig. 5. Tentative interpretation of the main clay mineral sources in Central Indian Basin
 3 sediments. Map after Cochran (1990); dashed line: Ganges deep-sea fan boundary.



1
 2 Fig. 6. Interpretative sketch of clay mineral fluxes in the northern part of the Central Indian
 3 Basin since the upper middle Miocene.
 4



1
 2 Fig. 7. Possible tectonic control of the lithology and clay mineralogy in the equatorial zone of
 3 the Central Indian Basin during the late Pliocene. 944-947= SHIVA cores. Vertical arrow =
 4 positive intraplate deformations.

1 Table 1

2 Characteristics of the five very long cores recovered during the MD 90 SHIVA cruise, and of
3 MD cores and ODP hole used for comparison

Cores	Latitude	Longitude	Water depth (m)	Penetration (m)	Age of oldest sediments
MD 90-947	01°24'S	76°37'E	4781	34.37	Late Pliocene
MD 90-946	03°16'S	78°08'E	4796	33.15	Late Miocene
MD-90-944	06°39'S	80°07'E	4902	40.17	Late Miocene
MD-90-943	08°19'S	79°49'E	5249	43.75	Late Miocene
MD 90-942	10°03'S	79°48'E	5356	48.83	Late Miocene
MD 81-374	12°48'S	77°43'E	5332	13.50	Late Pliocene
MD 81-375	12°47'S	77°46'E	5279	17.50	Early Pliocene
ODP 717C	0°34'S	81°14'E	4735	828.2	Late Miocene

4

5

6 Table 2

7 Average sedimentation rates (m/m.y.) of Pliocene sediments in the Central Indian Basin. ODP
8 data from Cochran, Stow et al. (1989c)

	Sites	Early Pliocene	Late Pliocene
North (Equator)	ODP 717c	58	58
	MD 947	39	39
	MD 946	6.5	4
	MD 944	9.5	5.5
	MD 943	14	4.5
South (10°S)	MD 942	6.5	4

9