Synthesis of morphotectonics and volcanics of the Central Indian Ocean Basin

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The Central Indian Ocean Basin (CIOB) is an enigmatic ocean basin in the young and tectonically complex Indian Ocean. Major tectonic and volcanic forms identified are fracture zones, abyssal hills, seamounts and ridges and a unique zone of intraplate deformation. Petrology indicates the presence of basalts, ferrobasalts and spilites. The basalts have erupted from moderately evolved magmas, the ferrobasalts from a long residing and shallow seated magma body while the spilites have formed due to the hydrothermal effect on the pre-existing subalkaline basalts. Besides these rocks, there is a wide occurrence of pumice of probably *in situ* origin. A distinct relation occurs between the morpho-tectonic forms and the volcanics. For example, in and around the topographic highs, pillow and massive basalts are common while close to fracture zones, flow lavas occur. We perceive two significant volcanic activities: one during the formation of the near-axis generated seamounts and the other in an intraplate environment.

INVESTIGATION of ocean floor volcanism reveals the type, nature and origin of volcanics. Additionally, the geologic settings in which the volcanics occur could indicate a relation between the volcanics and the morpho-tectonics, on a local to regional scale. Such studies help to understand related phenomena like plate tectonics, seafloor spreading and hydrothermal mineralizations, among others.

Volcanological studies of Pacific and Atlantic Oceans commenced a century ago[1]. Expeditions on-board the research vessels *Challenger* (1873–1876) and *Valdivia* (1888–1889), recognized the geomorphology and petrology of the Indian Ocean. Detailed explorations were carried out by *RV Dana* (1928–1930), *RV Snellius* (1929–1930), *RV Mabahiss* (1933–1934), *RV Albatross* (1950–1952) and *RV Ob* (1956–1958). The outcome of the International Indian Ocean Expedition (1959–1965), was a remarkable physiographic map of the Indian Ocean[2]. Under the Deep Sea Drilling Project (DSDP), the *DV Glomar Challenger* (1972) drilled 52 sites in 6 legs (22 to 27) to understand the breakup of the Gondwanaland. In 1975, the Russians published a well-documented atlas of the Indian Ocean[3]. Subsequently, the Ocean Drilling Project (under the auspices of the Joint Oceanographic Institutions for Deep Earth Sampling) drilled 62 sites in 9 legs (115 to 123) between 1987 and 1989 (ref. 4). Figure 1 depicts the physiography of the Indian Ocean and the various drilled sites.

In 1981, an on-going programme, 'Surveys for Polymetallic Nodules' (SPN) in the Central Indian Ocean Basin (CIOB) was conceived at the National Institute of Oceanography, Goa. Since then, in-depth research revealed many interesting and unknown facets of the CIOB. In this article, we collate and synthesize the various findings of the morpho-tectonics and associated volcanics and suggest directions for future investigations.

Physiographic settings

The Indian Ocean is characterized by three active spreading mid-oceanic ridges (MOR) that integrate with the global ridge system. The ridges, Central Indian (CIR), South East Indian (SEIR) and South West Indian (SWIR), meet at the Rodriguez Triple Junction, forming an inverted 'Y'. Besides continental islands, numerous minor ridges, plateaus and basins also occur (Figure 1). Here we are concerned with the general features of the CIOB *per se* than the models pertaining to the lithospheric plates' reorganization of the Indian Ocean[5]. A most notable peculiarity in the CIOB is the mid-plate[6] or intraplate[7] deformation zone, so termed due to its location within the Indo-Australian lithospheric plate, which is suggested to be a part of a 'diffuse' plate boundary between the Australian and Indian plates[8]. This zone (5°N–10°S and 75°E–90°E) has been deformed under approximately N–S stresses[9] and is related to the ongoing collision of India and Eurasia. The intraplate deformed area is characterized by contortions of the oceanic crust and most of the overlying sediments into long wavelength (100–300 km), peak to trough amplitudes of 1–3 km, E–W trending undulations and associated high seismicity. Superposed on the undulations are faulted and rotated blocks, 5–20 km wide, that are separated by high-angle reverse faults[7,9]. An Indo-Soviet project (Integrated Long Term Projects, ILTP) reviewed earlier informations and provided newer evidences for this enigmatic deformed area. It was concluded that the intraplate deformation

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could be modeled using the 2-level plate tectonic theory[10], based on the following findings:

1. Revelation of blocky mosaic structure and associated small elongated hills, 15–20 m high and about 45 km wide with gentle slopes on the flanks of the deformed blocks.
2. Identification of two new seamounts that strike NE–SW. The seamounts chain and the deformed blocks are the result of repeated tectonism and the related intraplate volcanism.
3. Mapping magnetic chrons 32, 32B and 34, which suggest a spreading rate of 4.4 cm/yr.

Two long-lived hotspots (Kerguelen and the Reunion) are located in the Indian Ocean that have been active since at least 80 Ma and 66 Ma, respectively[11]. The evolution of this ocean since the late Jurassic occurred during three main periods of seafloor spreading separated by two major plate boundary organizations and the last phase resulted in the opening of the Gulf of Aden (1.0 Ma) and the onset of intraplate deformation (~ 7 Ma)[8].

A detailed bathymetric map of the CIOB (Figure 2) using over 420,000 line km of echo-sounding data[12], shows an average water depth of ~ 5100 m. The 3 broad areas from west to east, rugged, plain and high to medium relief. Fracture zones (FZs), seamounts and abyssal hills have been located[13,14] (Figure 2) which were earlier either demarcated or postulated. The N–S oriented FZs are 75°E and 83°E (ref. 13), 86°E (ref. 15) and 75°45′E, parallel to 86°E (ref. 16) (Figure 2). The trend of the 75°45′E FZ was subsequently suggested to be 76°30′E (ref. 17). Two FZs that were previously noted and termed as La Boussole (at 73°E) and L'Astrolabe (at 79°E (ref. 17), also known as the Indrani fracture zone[13]) were studied using the multi beam swath (MBS) system[18,19]. A re-interpretation of the earlier data[16] led to the identification of the Triple Junction Trace on the Indian plate (TJT–In), which corresponds to an N 38°E offset of the magnetic lineation (A23 and A22), oblique to the SEIR and the CIR[20].

The deeper parts of the CIOB are rugged due to a mountainous relief and about 25 seamounts are explored here which were generally conical and of volcanic origin[14]. Tall seamounts have formed due to repeated and focused volcanism, whereas abyssal hills formed during single and short-lived volcanism[21]. The seamounts could also be of tectonic origin, especially those near the FZs[14].

Seamounts and hills were identified using single beam echosounders and MBS system (Figure 2). Nineteen major ($h > 1000$ m) and 109 minor ($h = 300$–1000 m) seamounts were located[22]. Of the 19 seamounts, 6 trends NNE–SSW and 13 are in an arc-like manner close to the CIR. The summits of some of these seamounts have collapsed (by ~ 100 to 200 m) either due to subsidence

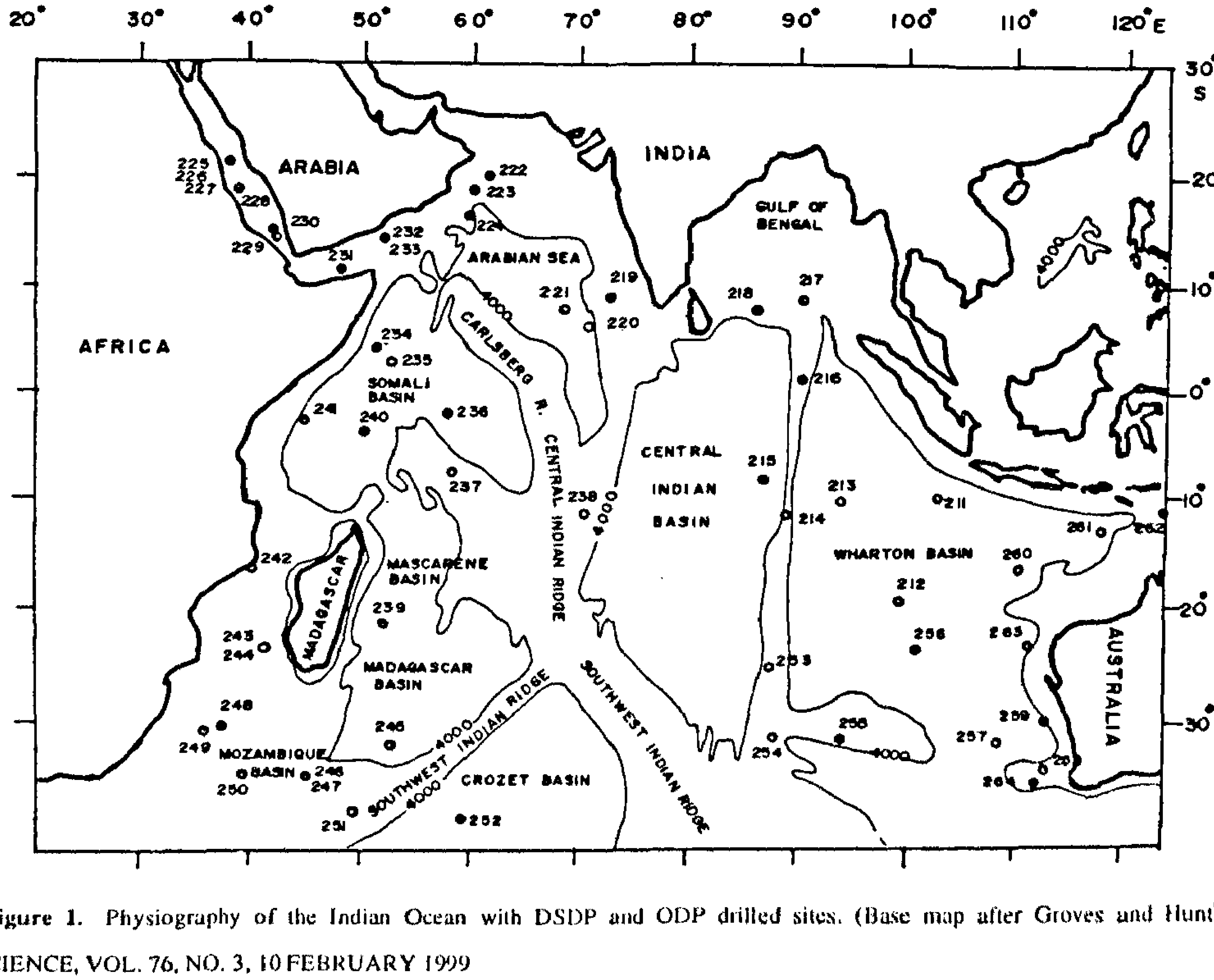

Figure 1. Physiography of the Indian Ocean with DSDP and ODP drilled sites. (Base map after Groves and Hunt[94].)

because of lithosphere cooling, as in the Pacific[23] or by crater formation. A large seamount, at 12°35′S and 76°16′E, trends NNE–SSW[24]. It has an elevation of 1275 m, an aerial coverage of 1289 km^2, a minimum basal extension of ~ 24 km, a high slope angle (up to 35°) and a cratered summit. A genetic relation between this seamount and the 75°45′E FZ was suggested[24], similar to such an association of FZs and seamounts in the Pacific Ocean[25].

MBS study of the 79°E FZ revealed 14 seamounts within its trough and a tectonic reactivation along this FZ was suggested as a cause for their formation[26], a mechanism earlier proposed for seamounts formation in the Indian[14] and Pacific Oceans[25]. On the basis of the seamounts morphology and rate of seafloor spreading, it was opined that the linear seamount chains in the CIOB could have originated from the SEIR at 50–60 Ma during a phase of fast spreading rate (11–15 cm/yr; full rate)[27].

Types of volcanisms

Volcanic activities are integral to and complement the morpho-tectonic forms in the world oceans. Volcanic forms of relief are greatly diversified and have a general mode of development. Certain rock types and/or volcanism are typical of a structural or geomorphic feature, for e.g. the frequent occurrence of ultramafic rocks in transform faults and lava flows along FZs and ridges. The association of the volcanic forms and activities show two common types: central and ridge (including FZ) eruptions. A definite regularity is noted in the distribution of relief forms in the Indian Ocean and the eruptive sources; the central eruptions are confined to abyssal basins (hills and seamounts) and the ridge types at the MORs. In the case of central type eruptions, the volumes and ages of small isolated seamounts from both FZs and normal oceanic crust can be related to the lava chemistry[28]. The smaller volcanoes at or near the ridge crest are composed of large-ion lithophile depleted tholeiites and are later capped successively by alkalic basaltic and alkali differentiated rocks as they drift away from the ridges. Further, the seamounts may host fairly dense lava flows, capped by low-density materials like hyaloclastite[29,30] or pumice[25]. In contrast to central eruptions, the widely developed linear type of volcanism related to fissure eruptions, form narrow ranges, crests and ridges. Petrologically, a great bulk of the rocks is low-K tholeiites produced by the MORs. An example of linear type of eruptions in the CIOB is the 14 flows of pillow lava at DSDP site 215, which were rapidly emplaced[31].

Volcanics

The MOR basalts (MORBs) which cover ~ 60% of the earth's surface and are the most voluminous igneous

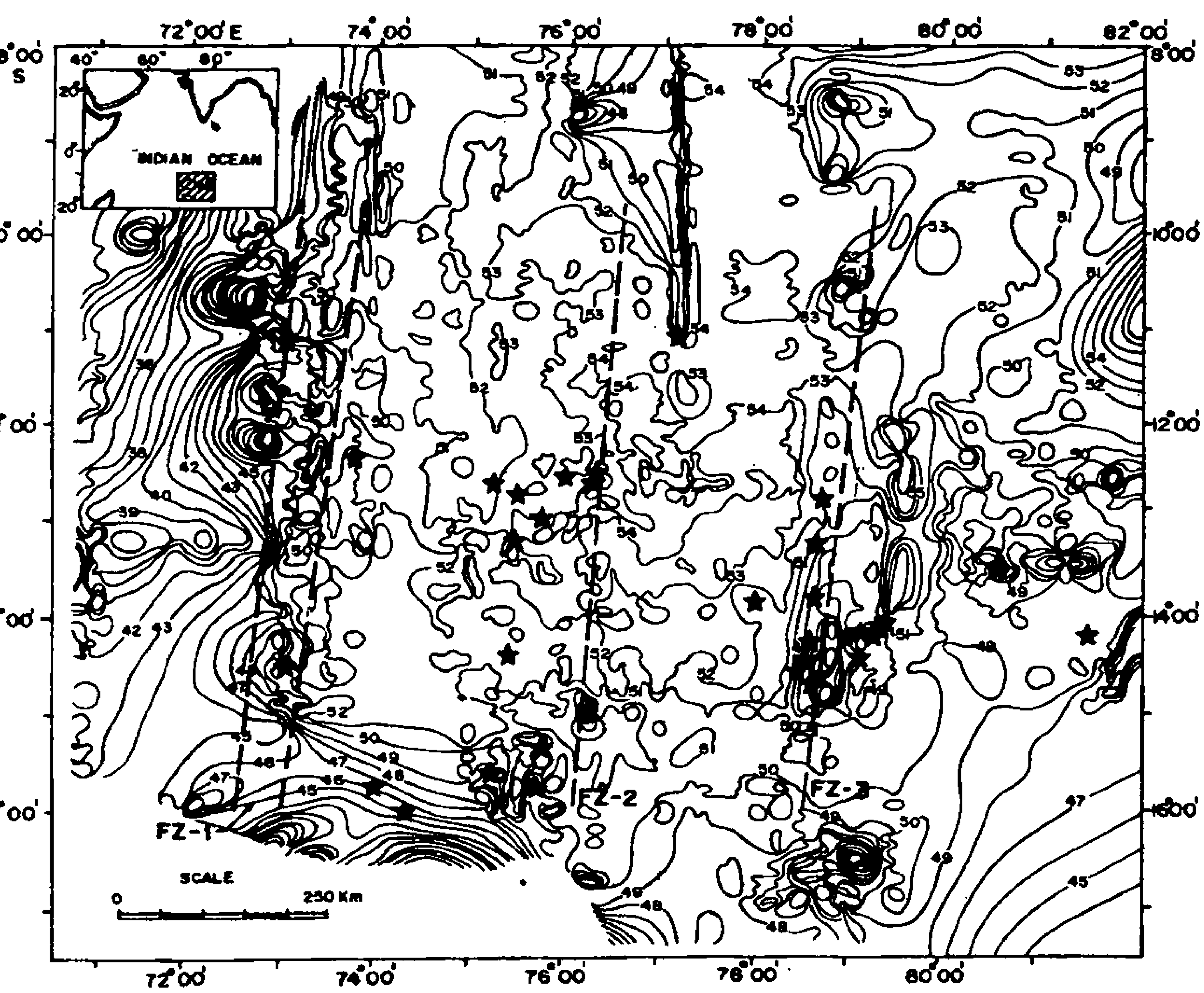

Figure 2. Bathymetry and morpho-tectonic features in the CIOB (see text for source). FZ = fracture zones, Stars = seamounts, contour interval = 100 m.

rock, are typically low-K tholeiites (avg. $K_2O = 0.2\%$) as compared to continental basalts[32,33] and they also have very low Ba, P, Pb, Sr, Th, U and Zr contents and show < 0.2 Fe_2O_3/FeO and > 10 Na/K ratios.

Early petrologic studies in the Indian Ocean, confined to the CIR, showed tholeiitic basalts to be most common[32] while spilites, gabbros, anorthosites, serpentines and dunites and aplitic veins and dikelets occur in FZs transecting the CIR[34–36]. The Ninety East Ridge basalts have high contents of pyroxenes, Fe_2O_3/FeO ($> 11\%$) and TiO_2 (2–3%) and variable K_2O (0.2–1.5%). It is concluded that the type of volcanism here is similar to the islands near to or associated with this ridge system[37,38].

DSDP site 215 (Figure 1), although located at the foot of the Ninety East Ridge, is bathymetrically and geochemically distinct[38]. A conspicuous chemistry of the successions of 14 pillow lavas drilled, is the K_2O (0.8–1.0%; #1, Table 1) which depends on the extent of weathering of these basalts[31]. It is however, believed that the K abundance is a primary feature of site 215 basalt, since the neighbouring site 214 basalts, though intensively altered (4–7% loss on ignition) has much lower K. Hence, the presence of a potassic–phosphorous (K–P) magma was suggested[39].

Volcanism along the Indonesian Volcanic Arc (IVA) arises from known sources during the last 5 Ma (ref. 40), but their presence prior to this has not been proven[41]. The sediments since the last 150 m.y. are essentially volcanogenic and derived from volcanisms at the MOR, hotspot, subduction related along the IVA and also from terrestrial sources[42]. More than the MOR, the Reunion and Kerguelen hotspots largely contributed of tephra to the Indian Ocean throughout their evolution while the IVA is the dominant source at least since the early Pliocene, particularly to the northern Indian Ocean[42]. A simple technique by utilizing the ratios of clay minerals to evaluate the volcanogenic components in the Indian Ocean has been proposed. Here, the change in the type of volcanic materials with time has been used in as much as basaltic volcanism was dominant *before* the late Cenozoic while silicic volcanism intensified *during* the late Cenozoic; and such changes were correlated with the tectonics of the Indian Ocean floor[40,43]. It has been suggested that clay supplies in the CIOB are controlled by periodic late Miocene climatic changes while during the late Pliocene, an irregular, probably tectonic control, appeared[44].

Evidently volcanics, ranging from typical low-K MORBs to sub-alkaline to ultramafic and even dikelets of granitic affinity, exist in the Indian Ocean and various views are advanced concerning the source materials and magmatic processes responsible for their formation. The volcanics of the CIOB, with respect to their nature, type, origin and relationship with the morpho-structural features, are now discussed.

Basalts

CIOB basalts occur as pillow lavas and massive outcrops near topographic highs and as fragments in the plains. The freshness of specimens depends on the degree of alteration and is reflected by their colouration: fresh rocks have shades of gray and the altered ones are brown and in shades of brown. Glass selvages with plagioclase phenocrysts are conspicuous in some samples. The basalts also form nuclei and substrate for FeMn oxides.

The basalts reveal different textural and mineralogical characteristics[45]. Plagioclase (most dominant) occurs as long, acicular needles, phenocrysts, microphenocrysts, as microlites in the groundmass and occasionally as twinned and/or zoned phenocrysts. Olivine (next dominant) occurs as anhedral to euhedral crystals, at times are fractured, but phenocrysts, zoned or quenched forms are rare. Clinopyroxenes (augite) are scanty. Opaques are as thin, fine needles forming dendrites and trichites and as stubby magnetite crystals, reddish–brown hematite and rarely spinel/chrome spinel. Vesicles are few and mostly empty but at times are lined with chalcedony and/or epidote.

Alteration of oceanic basalts is ubiquitous and varies depending on the presence/absence of glass, mineralogy, occurrence of fractures on the rocks and form of the rocks (pillows, flows, fragments). CIOB basalts display minimal to highly alteration and show the glassy groundmass to be replaced with cholorophaeite, palagonite and smectite, while plagioclase crystals are sericitized and olivine exhibits ramifying veins of iddingsite.

The basalts are depleted in incompatible elements and have moderate Mg# (61.7 to 63.5)[46] (#1, Table 1). The compositions are typical of N–MORB and comparable to seamount lavas from the Mid-Atlantic Ridge (MAR), East Pacific Rise (EPR) (#2, 3; Table 1) and from other Indian Ocean sites (#4, 5; Table 1). These ancient seamount basalts are nearly indistinguishable from the young, near-axis originated seamounts of the MAR and the EPR. It is suggested that the melts that were responsible for the formation of the CIOB basalts were derived from a heterogeneous source, similar to the EPR seamounts[47,48] and that the moderately evolved magmas experienced a low rate of effusion.

Under the ILTP, several seamounts in the intraplate deformed zone (crustal age > 80 Ma BP), were sampled which revealed sub-alkaline transitional basalts, tuff-breccias, hawaiites and mugearites, while later volcanic episodes resulted in the eruption of trachy-basalts[10]. This suggests episodic outpourings of progressively fractionated basalts in three distinct stages: N-MORB at the base, moderately alkaline basalts at the flanks and slopes, and trachyandesites and tuffs at the summit[10,49]. Such a volcanic sequence has been observed in the Pacific Ocean seamounts[25] and needs confirmation in the CIOB.

Table 1. Representative chemical composition of the Central Indian Ocean Basin (CIOB) volcanics. For comparison are included similar volcanics reported from other sites in the Indian Ocean, Mid-Atlantic Ridge (MAR) and East Pacific Rise (EPR)

	1	2	3	4	5	6	7	8	9	10	11
SiO_2	50.53	50.10	49.89	49.40	48.37	48.02	48.10	46.85	50.67	60.28	70.26
TiO_2	1.27	1.25	1.36	1.65	1.63	2.28	2.35	1.96	2.18	0.64	0.55
Al_2O	15.44	16.73	15.52	16.43	16.18	16.21	14.90	16.44	13.69	13.09	13.07
FeO*	9.81	9.20	9.42	8.06	8.44	14.88	14.60	15.43	13.05	7.88	3.74
MnO	0.19	na	na	0.18	0.16	0.43	na	0.26	0.19	0.49	0.29
MgO	7.73	8.13	7.66	6.59	6.20	3.75	6.45	3.69	4.81	2.48	1.22
CaO	12.30	11.45	12.40	10.38	10.08	9.07	9.04	6.27	5.38	5.13	2.18
Na_2O	2.80	2.79	3.58	2.94	3.04	3.18	2.75	4.36	4.61	5.92	4.43
K_2O	0.08	0.06	0.10	0.85	0.85	0.89	0.37	0.84	0.80	3.93	4.13
P_2O_5	0.11	0.12	0.13	0.29	0.29	0.25	0.19	0.33	0.30	0.15	0.13

FeO* = Total iron; na = not available.
1. Microprobe analyses of CIOB basaltic glass from ancient fast generated seamounts[46]. 2. Microprobe analyses of MAR basaltic glass from young slow generated seamounts[91] 3. Microprobe analyses of EPR basaltic glass from young fast generated seamounts[92]. 4. Basalts from DSDP site 215 (ref. 93). 5. Basalt from the lower flank of a seamount, eastern Indian Ocean[33]. 6. Ferrobasalts from the CIOB[50]. 7. Ferrobasalts from the DSDP site 214 (ref. 39). 8. Spilitized basalts from the CIOB[58]. 9. Spilitized basalts from the Sao Paulo Ridge, MAR[60]. 10. Trachy-andesitic pumice from the CIOB[72]. 11. Rhyodacitic pumice from the CIOB[72].

Ferrobasalts

Ferrobasalts (FeTi-enriched basalts) are reported from DSDP sites 214, 216, 254 and 256 in the Indian Ocean, in segments of SEIR and SWIR and in the CIOB[50]. FeTi basalts are useful for telechemical predictions, since the amplitude of magnetic intensities is proportional to the FeTi content of the source rocks[51]. The CIOB FeTi basalts occur in two areas, with high amplitude magnetic (HAM) anomaly values (– 150 to + 500 nT), near chrons A24 and A25 which depict spreading rates of 190 and 110 mm/yr, respectively.

These basalts show phyric plagioclase laths and sheaf-like aggregates intervening with the glass while the opaques are mainly hematite and magnetite. The samples have 43.37 to 51.67% SiO_2 and TiO_2 and FeO* are between 1.93–2.99% and 11.92–19.13%, respectively (#6, 7; Table 1). CaO and MgO contents suggest fractionation of plagioclase and olivine. Although the high K_2O contents might indicate alteration, similar values are reported from DSDP site 216 (ref. 39) and Spiess Ridge, SWIR[52], that are considered to be unaltered. The trace element values are akin to FeTi basalts from other oceanic locations[50].

Various views have been advanced for FeTi basalts, such as variable degree of shallow crystallization[53,54], differences in the depth of magma generation[55] and magma mixing and fractionation[56]. It is generally opined that long storage time of a well-fractionated, shallow-seated magma could give rise to FeTi basalts. This resulted in the concept of the 'horizon of neutral buoyancy' (HNB)[57]. The HNB, defined as that 'depth interval within which the magma density and the aggregate country rock density are equal' is situated at sublithospheric depths, being confined by a narrow, vertical and a wide lateral extent. Below the HNB, the magma that ascends under the influence of positive buoyancy, is stabilized between 2 and 4 km depth, while above this region the magma descends by negative buoyancy. The conducive conditions for the CIOB FeTi basalts might have been facilitated due to: a change in the rate of spreading, effect of topographic highs and presence of a long-lived and shallow-seated magma. Subsequent slow and continuous fractionation of this trapped magma in a zone of HNB, resulted in FeTi enrichment and ferrobasaltic eruptions during chrons A24 and A25.

Spilites

Spilites are less abundant compared to basalts[37]. In the CIOB, they occur near the 79°E FZ[58]. The rocks are fresh to highly altered, sometimes exceed 25 cm, have sparse glass selvages and are fine to medium grained with albitic plagioclase (few as phenocrysts), clinopyroxene, olivine and little of chlorite, epidote and hematite. The typical texture is microlitic intersertal formed by the chlorite minerals[59], whilst porphyritic, hyalopilitic, intersertal and glomeroporphyritic are also seen.

Ten samples, (3 highly altered, having low CaO and high K_2O) and others with lower CaO than N–MORBs (#8, Table 1), were analysed which concur with spilitic basalts from the Sao Paulo region, Atlantic Ocean[60] (#9, Table 1). The binary plots show the characteristics of the samples (Figure 3). A variety of origins for the spilites have been proposed: a primary, auto hydrothermalism or auto metamorphism, secondary, diagenetic and metamorphic[61] and unmixing of a parent magma[62]. In addition, there could be an influence of tectonic structures by way of percolation of hydrothermal solutions through cracks, fissures and FZs. Hydrothermal alteration and metamorphism have been noted at MORs where hydrothermal regimes exist at temperatures of 200–300°C between 1 and 2 km depth[37,60]. Similar conditions can occur at FZs that are away from the ridge crests, promoting spilitization of the rocks[63]. Thus, hydrothermal circulations, which are not uncommon within FZs, may lead to alteration and spilitization. The CIOB samples were dredged near 79°E FZ that has a throw of ~100 m towards the east[26] and was probably reactivated[64]. These tectonism here might have facilitated the

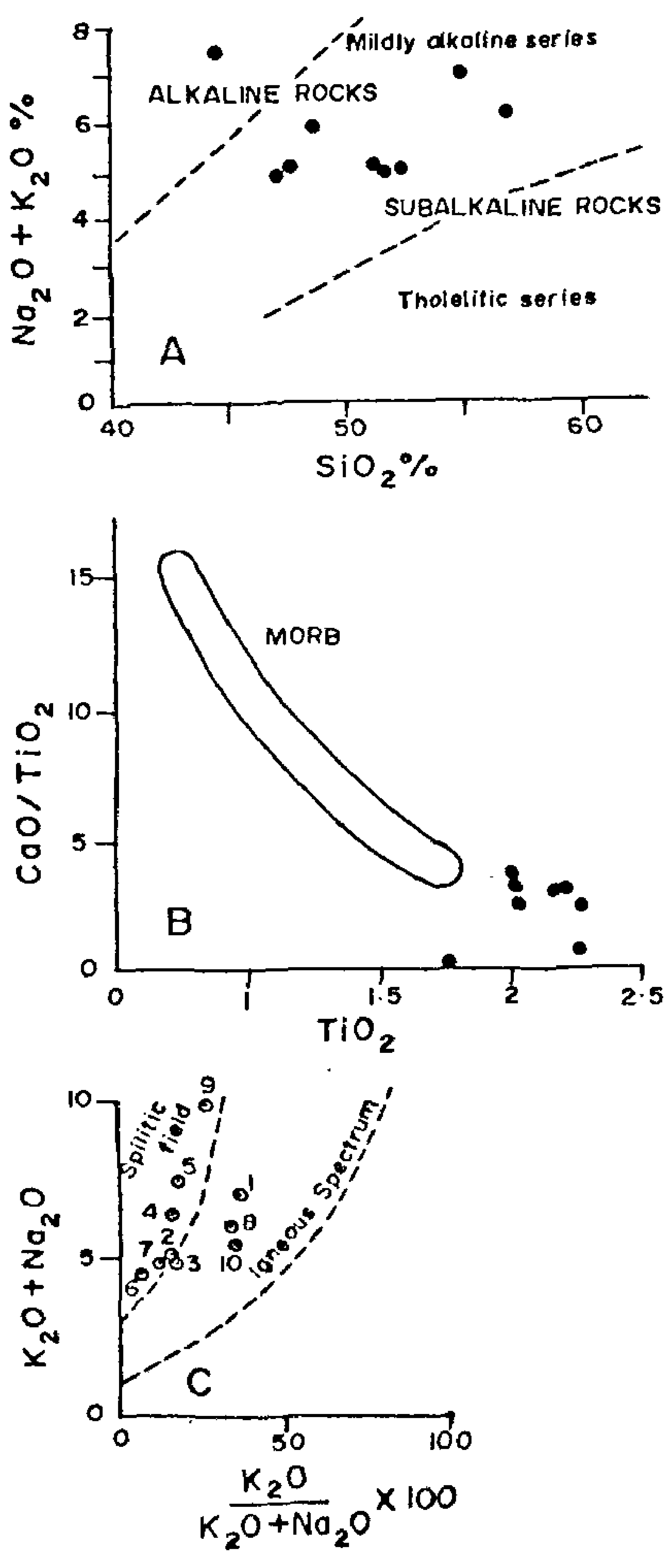

Figure 3. Variation diagram for the spilitic rocks of the CIOB[58].

seepage of hydrothermal solutions to bring about alteration and spilitization noticeable in some, though not all, of the pre-existing basalts. Signatures of hydrothermal effect are noted in zeolitic slabs[65] and FeMn crusts dredged from seamounts near the 75°30′E FZ[66].

Pumice

Pumice is widely distributed in the world oceans but large pumice fields are relatively rare[67]. Presence of drift pumice has been recorded in and around the Indian Ocean near Sri Lanka, Madagascar, Reunion, etc.[68]. The distribution, morphology[69] and chemical composition of the CIOB pumice and their probable *in situ* origin are now described. A field of pumice extends from 9° to 20°S and 72° to 84°E, (based on 1925 station samples), covering an area of ~600,000 km^2 (ref. 70). Of the 3083 samples studied, 94% are in the 0–4 cm size range and about 1% are >8 cm (ref. 71), with the largest clast being ~36 cm long. The pumice exhibit varied colours (buff, grey, black and brown) and shapes (oval, lineated, subrounded, rounded and irregular) and are either coated fully or partially by FeMn oxides. Thin sections show silicic glassy webs with few phenocrysts of plagioclase and/or pyroxenes and vesicles. Quartz crystals (confirmed by XRD), tests of radiolarians, phytoliths and diatoms and FeMn micronodules occur within the glassy interstices. These materials make the pumice heavy and sink to the seafloor[69].

Analyses of pumice reveal two types[72]: trachyandesite (group 1) and rhyodacite (group 2), with average SiO_2 of ~60.0% and ~70.0%, respectively (#10, 11; Table 1). On a plot of solidification index vs the major oxides (Figure 4), it is noted that SiO_2 increases with decreasing SI while TiO_2 and FeO* of group 1 pumice increase with increasing SI but decrease in group 2 pumice. CaO and MgO of the former increase with increasing SI but has nearly constant CaO for variable SI for the latter type. K_2O is relatively lower and Na_2O nearly constant with increasing SI for group 1 pumice compared to group 2.

Pumice in the Indian Ocean[68] and in the CIOB[73] were largely derived from the 1883 eruption of Krakatoa, but there are no reports of Krakatoan products in any significant amount in the recent sediments to the east of 80°E[74]. Considering the topographic expressions and high heat flow values at the central part of the Ninety East Ridge, the source of pumice from a seamount near this ridge was suggested[75]. Previous discoveries have provided strong evidences that pumice forming eruptions can occur at great depths. For example, over Atlantic seamounts at depths varying from 1500 m and more[76], near Tonga at depths > 1500 m (ref. 77), Okinawa Trough, South of Japan[78], and trachytic pumice flows from intraplate volcanoes in the Society and Austral hotspot

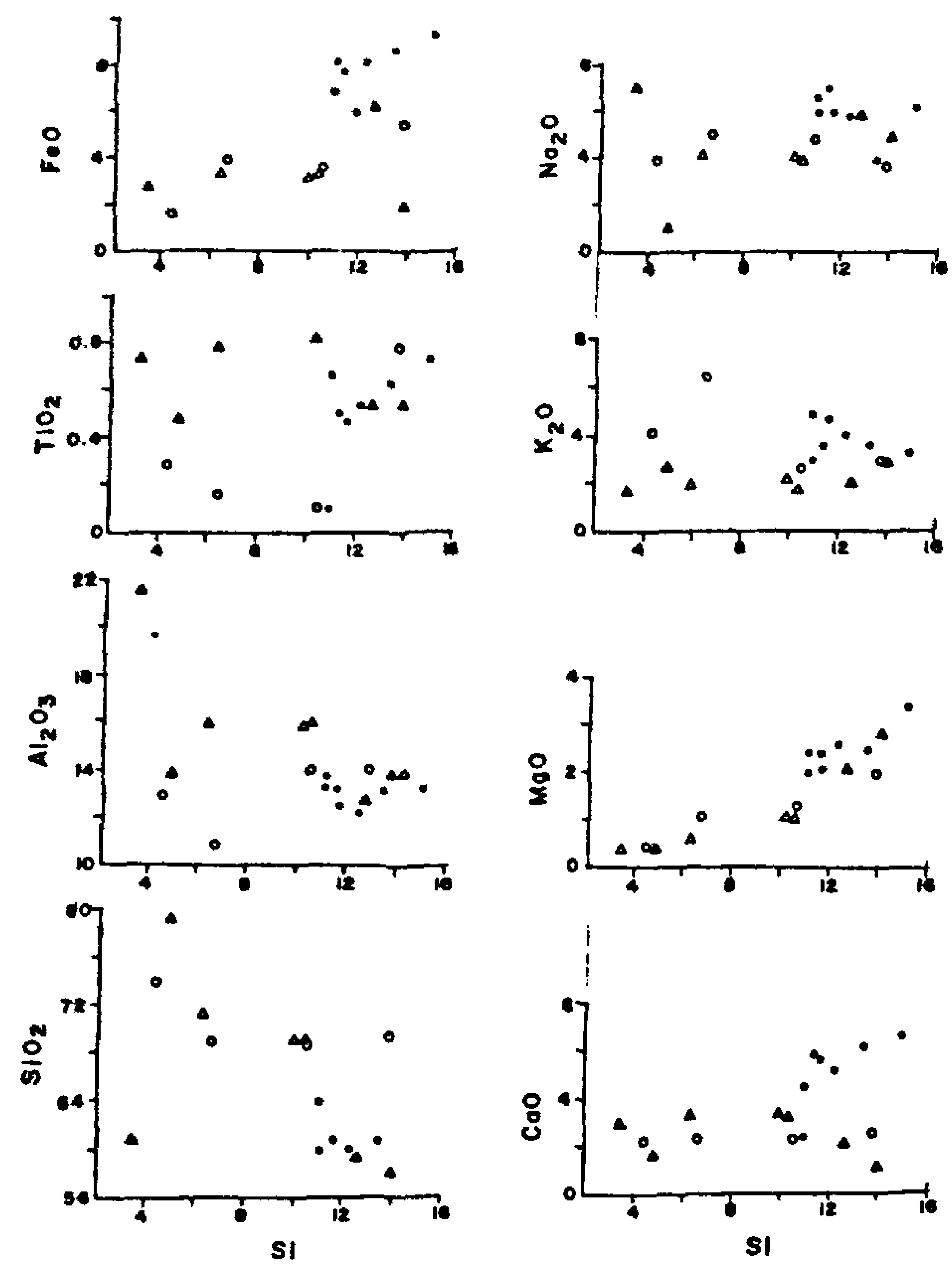

Figure 4. Variation diagram for the pumice of the CIOB[72]. Symbols: filled circles = trachyandestic pumice; open circles = rhyodacitic pumice; open triangles = Frick and Kent[68]; solid triangles = Martin–Barajas and Lallier–Verges[73].

regions (depth = 3500 m)[79]. Collectively these studies suggest that widely distributed pumiceous fragments have recently erupted from submarine volcanoes.

The above would suggest an *in situ* origin for pumice at abyssal depths to account for its vast occurrence especially, in the vicinity of seamounts, hills and FZs in the CIOB[70]. MBS depict that many CIOB seamounts have summit craters[24]. Their spatial locations and the associated pumice indicate that cratering could have been caused by explosive silicic eruption. Since hot pumice sinks faster and near to the source compared to colder ones[80], the larger pumice clasts in the CIOB may be rapidly sunk hot clasts near their eruptive sources. Moreover, the frequent occurrence of pumice to the west of the Ninety East Ridge and the absence of favourable surface current circulation patterns to the CIOB from the Indonesian margin[81] also seem to favour an *in situ* origin.

The CIOB pumice occurs in a region dominated by tholeiitic basalts[46,72] that could have fractionated to more silicic melts. Extreme fractional crystallization of abyssal tholeiites probably formed layered or stratified magma chambers with alkali- and volatile-enriched silicic melts, a mechanism proposed for the Tonga Ridge pumice[82]. Basaltic and silicic glass shards in ash layers from the Kerguelen Plateau, point to simple fractionation of the basaltic magmas to silicic types[83]. Magma mixing[84] and zoning of magma chambers by viscosity differences[85] could be the other possible mechanisms.

Silicic volcanic events occurred in the Indian Ocean during Cretaceous–Eocene as indicated by the abundance of zeolites (phillipsite and clinoptilolite) in the CIOB[86]. A strong input of silicic glass did not begin until late Miocene, which continued up to Pleistocene in the Indian Ocean. Also, basaltic volcanism was predominant before late Cenozoic and silicic volcanism increased during late Cenozoic[40]. All these can be related to compressional tectonism along the IVA.

The above discussion leads us to remark that *in situ* silicic volcanism(s) could have resulted in pumice formation. However, we do not negate some contributions of pumice from earlier eruptions in the IVA or from the cataclysmic Krakatoan eruption[69]. It is pertinent to note that pumice at many times have FeMn oxides and also occur as nucleus for the manganese nodules[69–71]. It is well known that accretion of 1 mm of FeMn oxides requires 1 m.y. and a conservative thickness of 1 cm of oxide coating would require 10 m.y. This indirectly testifies the coated pumice to be older than the 1883 Krakatoan event. Alternatively, one could visualize a rapid accretion of FeMn oxides as a result of hydrothermal episodes, as discernible from recent evidences[58,65,66,87,88]

Summary

The morpho-tectonic forms in the CIOB are isolated seamounts to chains of seamounts, abyssal hills and large FZs with substantial displacements and the zone of intraplate deformations. These forms, which occur in conjunction with the volcanics (N-MORBs to pumice), influence the distribution and variety of the volcanics. The seamounts and possibly the FZs served as conduits for magma ascension and eruptions. The seamounts are dominantly oriented in an N–S manner but some have a distinct E–W bulge[87] that indicates addition of magmatic mass during later eruptions. The linear seamount chains and the occurrence of the volcanics in and around them, suggests that the seamounts could have shared a common magmatic source, resulting in a trade-off in the production between adjacent seamount chains[89]. We believe that there are definite indications of different volcanic episodes in the CIOB. For instance, while the near-axis originated seamounts produced N-MORBs, the spilites, ferrobasalts and pumice are the manifestations of intraplate volcanism and associated activities.

The present study reconfirms the existence of a large volcanic province[90] in the CIOB. Although our investigations are not exhaustive, the findings help to understand the origin of the volcanics, role of the structural features in their distribution and a strong probability of different volcanic events.

Future research problems

There is a need to conduct geophysical investigations, particularly in the uncovered areas, complemented by extensive sampling. The geological problems that require to be addressed are: demarcating areas of specific volcanic types and their mutual relation, their precise radiometric dating, their alteration and geochemical contribution to the sediments and seawater, association of volcanics and manganese nodules and crusts, origin of pumice which form nucleus and substrate for FeMn oxides and identifying areas of hydrothermal deposits. We expect that further efforts would help us understand better the geodynamics and volcanic history of the CIOB and their inter-relationships.

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ACKNOWLEDGEMENTS. We thank the reviewer for suggesting pertinent inclusions. S.D.I. thanks the Director for the facilities provided. The SPN project is supported by the Council of Scientific and Industrial Research (Govt of India) and Department of Ocean Development (Govt of India). The paper forms a part of S.D.I's Ph D thesis. This is NIO's contribution #2590.

Received 9 February 1998; revised accepted 21 November 1998