

Genesis of Central Indian Ocean Basin Seamounts: Morphological, Petrological, and Geochemical Evidence

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Abstract

We present the petrological investigation carried out of the seamounts located between water depths of 4300 and 5385 m in the Central Indian Ocean Basin (CIOB). The seamounts have variable shapes (conical and elongated) and heights (625 -1200 m). The basalts have a glassy veneer that forms the outer rind while the holocrystalline interior shows variable textures. The basalts are plagioclase phyric and compositionally have low FeO* (8.0-10.5 wt%) and TiO₂ (1.3-2.0 wt%), and variable K₂O (0.1-1.0 wt%) contents and are slightly enriched in the light rare earth elements. These characteristics are similar to the basalts from the CIOB seafloor and the Central Indian and Southeast Indian Ridges. These facts attest to the simultaneous formation of the CIOB seafloor and associated seamounts that shared a common source between 56 and 51 Ma when the spreading (half) rate was 95 mm/yr. Similar to the East Pacific Rise (EPR), the source melt was perhaps ferrobasalts which over a period of time fractionated to N-MORB during the emplacement of the seamounts. The production of the seamounts may have involved a periodic tapping of a regularly replenished and shallow seated source melt. These basalts from the older seamounts of the CIOB are analogous to their present-day counterparts that form at the fast-spreading EPR and other locales in the world oceans.

Keywords: seamounts, morphology, geochemistry, origin, CIOB, EPR

1.0 Introduction

Seamounts are distributed in space and time, and occur in groups, chains or in isolation. Further, seamounts are considered as windows to the mantle and reflect processes that occur at the crust-mantle boundary. A majority of seamounts also occur along convergent plates, in areas of vertical tectonic movement, at ridge-transform fault intersections, overlapping spreading centers and hotspots. It has been estimated that ~50 million tons/year of basalts are utilized in the production of seamounts (Lisitzin, 1996). Several factors influence seamount production and disposition including mantle upwelling associated with superfast spreading, off-axis mantle heterogeneities, mini-plumes, local upwelling and the vulnerability of the lithosphere to magma penetration

(Batiza, 2001; Iyer et al., 2012 and references therein). Seamounts that, over a period of time, are transported into the subduction zone (e.g. West Junggar, Central Asian orogenic belt) not only add trace elements but also affect the mantle composition, chemistry of the arc and back-arc lavas and influence hydrothermal activity (Yang et al., 2015).

Seamounts are generally composed of basalts and hyaloclastites (e.g., Helo, 2014) and sometimes gabbros (Bideau and Hekinian, 2004) and serpentinites (Ueda et al., 2011). Non-hotspot seamounts occur on a young, thin and hot lithosphere and host tholeiites while plume-generated seamounts on the other hand, lie on a thick, older lithosphere and are composed of either tholeiitic or alkaline basalts or more evolved basalts (Batiza, 2001). Seamounts are characterized by unique biodiversity, influence the water circulation patterns and currents, cause seismicity and are sites for hydrothermal deposits (Fe, Mn, Co), rare metals and phosphorites (Iyer et al., 2012 and references therein).

The morphology, distribution, and petrology of seamounts help to understand and constrain models of seafloor spreading and provide evidence for their temporal growth, the role of local tectonics on magma production and their emplacement mechanism. Seamounts not only characterize the flanks of modern fast spreading mid-ocean ridges (MOR) but also in areas that formed during a phase of fast spreading in the geological past. This is exemplified by the seamounts present in the Central Indian Ocean Basin (CIOB).

2.0 Previous Studies

The CIOB extends from latitudes 9° S to 19° S, longitudes 72° E to 80° E and has an average water depth of 5,000 m. The various aspects of the basin that have been investigated are volcanics, sediments, ferromanganese deposits (FeMn nodules and crusts), micropaleontology, bathymetry, magnetic and gravity. The morpho-tectonic features are represented by lineations, fracture zones

(FZ), faults, abyssal hills and seamounts (Kodagali, 1998; Mukhopadhyay and Batiza, 1994; Das et al., 2005). The major fractures in the CIOB that originate from the Southeast Indian Ridge (SEIR extends up to 30° S) (Fig. 1a) traverse through the basin. The reorganization of the tectonic plates occurred between 60 and 50 Ma and during this time period the seamounts were formed (Fig. 1b) (Sclater and Fisher, 1974; Patriat and Ségoufin, 1988; Mukhopadhyay and Batiza, 1994; Mukhopadhyay et al., 1995; Müller et al., 2016). Several seamounts in the CIOB were emplaced in an approximate N-S linear fashion (Fig. 2) along eight fractures that developed in the basin due to the northward movement of the Indo-Australian Plate coupled with Indo-Eurasian collision events (Das et al., 2007). As evident between anomalies 24 and 21 propagating rifts played a crucial role in the formation of the Triple Junction in the Indian Plate (TJT-In) and a ridge-fault-fault configuration led to the westward receding of the CIR while the SEIR advanced westward (Dyment, 1993).

The prerogative of India's polymetallic nodules project was to identify a First Generation Mine site in the CIOB and this inhibited exclusive sampling for rocks. Studies have revealed the occurrence of basalts, spilites, ferrobasalts, pumice, glass shards and volcanogenic-hydrothermal material (Iyer and Karisiddaiah, 1990; Karisiddaiah and Iyer, 1992; Mislankar and Iyer, 2001; Kalangutkar et al., 2011; Mascarenhas-Pereira et al., 2016; Mukhopadhyay et al., 2018). A holistic view of the investigations suggests the prevalence of a volcanic province in the CIOB, the possibility of hydrothermal activity, intraplate volcanism and a nexus between the volcanics and FeMn deposits (Iyer, 2005; Jauhari and Iyer, 2008).

Due to the above-mentioned constraints in sampling, basalts collected from the CIOB seafloor were better investigated in contrast to only limited samples recovered from a few seamounts. Based on the disposition of the seamounts and analysis of basaltic glass Mukhopadhyay et al. (1995) opined that the seamounts were of a near-axis origin, similar to that in the East Pacific Rise (EPR) and Mid-Atlantic Ridge (MAR). This study was followed by the report concerning the ferrobasalts (high Fe-Ti basalts) recovered from some of the CIOB seamounts (Iyer et al., 1999). Using Artificial Neural Network method, Das and Iyer (2009) found that most of the CIOB basalts are N-MORB (Normal-MOR basalts) while a few have compositional signatures of E (Enriched)-MORB and OIB (Ocean Island Basalts). Das et al. (2012) discussed in detail the petrography, chemistry and origin of the CIOB basalts, while Hemond et al. (2012) reported the preliminary results of the isotopes of a few of these basalts.

We present the investigations that pertain to the petrology and genesis of basalts dredged from four seamounts of the CIOB. We compare our basalt data with those from the CIOB (Das et al., 2012), the SEIR (Michard et al., 1986; Dosso et al., 1988), Central Indian Ridge (CIR, Ray et al.,

2007) and from the DSDP Sites 214, 215, 216 and 254 (all near the Ninetyeast Ridge; Frey et al., 1977; Thompson et al., 1978) and the Afanasy-Nikitin Seamount located to the north of the CIOB (Kashintsev et al., 1987; Mahoney et al., 1996). This comparison is made because the CIOB evolved as a result of the collective effect of plate movements and formation of the CIR, SEIR and Triple Junction in the Indian Plate (TJT-In; Fig. 1a) (Dyment, 1993; Das et al., 2007).

3.0 Materials and Methods

Twelve dredging operations were carried out and basalts were recovered near the summit, flanks and base of four seamounts (Fig. 3a,b). The locations and geologic setting of the earlier and presently seamounts are tabulated (Table 1) and plotted (Fig. 1b). The basaltic glass was chipped and the fresh exposed part of the samples was used for thin section preparation and chemical analyses. Thin sections were observed under Carl Zeiss and Nikon microscopes and photomicrographs were obtained with an SLR camera at magnifications between 20× and 100×. The samples were sliced using a diamond embedded saw and the slices were abraded with sandpaper to remove the traces of sawdust and visible alteration and were cleaned with acetone and distilled water. The slices were coarsely crushed in a hydraulic piston crusher prior to powdering in a tungsten-carbide ring mill.

Major element oxides were measured using an X-ray Fluorescence (XRF) machine (Philips PW2400 at the Wadia Institute of Himalayan Geology, WIHG, India) following the methods described by Rhodes (1996). The contents of minor, trace and rare earth elements (REE) were determined using an inductively coupled plasma-mass spectrometry (ICP-MS, Perkin Elmer's Elan DRCe and Perkin Elmer's Elan DRC-II) at the National Institute of Geophysical Research, Hyderabad India and the WIHG, India). About 0.05 g of powder of twelve basalts was digested with a mixture of HF, HNO₃ and HClO₄ to remove silica. Diluted (1:1) HNO₃ was added to dissolve the digested material and the aliquot was made to 100-mL with distilled water, for ICP-MS analysis. A blank solution was run for each set of five samples and the reading was used to correct the results for contamination (if any) during solution preparation. Repeated measurements of rock standards BHVO-1, BRC-1 and JB-2 were made for calibration and calculation of the accuracy of the analysis. The relative standard deviation for the major oxides was < ±1% while for the trace elements and REE analysis it was ±0.6 to 3%.

4.0 Results

4.1 Seamount morphology

The seamounts have been numbered as #1, 2 and 3 (Fig. 3a) and #4 (Fig. 3b) and we would use these numbers for further discussion. The underlying oceanic crust in the study area formed due

to tensional stress at a superfast (half) rate of 95 mm/yr during chron 25 and 23 (~56-51 Ma) (Royer et al., 1989; Mukhopadhyay et al., 2002). The four seamounts lie on magnetic anomaly A24 and have variable morphology and dimensions. Seamounts #1, 2, and 3 are along the 75° 30' E FZ and seamount #4 is to the south of the 79° E FZ (Das et al., 2007). The seamounts range in height between 625 and 1200 m, basal width 7 and 17 km and basal area 40 and 236 sq km (Table 2). The seamounts #1, 3 and 4 are conical and have an N-S elongated base while seamount #2 has an E-W elongated base. We ascribe the addition of an excess of 700 km of oceanic crust in the CIOB between anomaly 29 and 24 (duration of 10 m.y.), when the CIR and SEIR were created (Dyment, 1993), to the initiation and production of the seamounts.

Das et al. (2007) identified four types of relations between flatness (i.e., summit width / basal width) and slope angle for the CIOB seamounts. Type 1: low HW (height-width) ratio <0.08 with low slope angle (<10°) and low flatness (<0.12), type 2: low to intermediate HW ratio of 0.08 to 0.16 with variable slope angle (6° to 15°) and flatness (0.08 to 0.3), type 3: an intermediate HW ratio 0.16 to 0.23 with high slope angle (>10°) and high flatness (>0.2) and type 4: high HW ratio >0.23 with an high slope angle (>10°) and low to moderate flatness (<0.2).

4.2 Petrography:

The basalts (Fig. 4) are fresh with a veneer of glass but sometimes the glassy rind was oxidized and overlain by FeMn oxides. The studied basalts are not extensively altered or palagontised as compared to those obtained from the seafloor (Iyer, 1999). A cross-section of some of the basalts displays three zones, with different textures and minerals, starting from the outer glassy (zone A) to an intermediate (zone B) to the interior (zone C) (Table 3; Fig. 5). In general, plagioclase is the most abundant and occurs as laths (<0.25 mm), microphenocrysts or tabular grains (0.25 to 0.50 mm), phenocrysts (0.5 to 1.0 mm) and megacrysts (>1.0 mm) (Mislankar and Iyer, 2001). The quenched zone A (glassy) have sheaf-like radial texture with reddish-brown globular structure in which small plagioclase crystal form nuclei (Fig. 5A). These features resemble the red-feathery alteration noticed in MAR basalts of DSDP Sites 332 and 335 (Baragar et al., 1977). Phenocrysts and microphenocrysts of olivine and sector zoned plagioclase occur in the interstices of glass as are common in MORB glass (e.g. Iyer and Banerjee, 1993). The transitory zone B is hemicrystalline, with skeletal crystals of plagioclases in glass, and this zone grades into the holocrystalline zone C. The zone C has intergranular, rarely intersertal, glomeroporphyritic and incipiently formed flow textures with profuse plagioclase in the groundmass. In contrast, olivine grains occur in the groundmass only if these are present as phenocrysts.

Occasionally, there are ‘ghost’ or skeletal plagioclase grains that show only the outline while the interior is hollow. In contrast to zoned plagioclase, twinned ones are more frequent and may contain inclusions of glass that extends as embayment into the plagioclase and result in fine reaction /alteration margins. Olivine phenocrysts are few and unzoned and the groups of fresh olivine in zone A (Fig. 5D) may indicate a cumulate origin. In the dyke samples (HRX 13) there are coarse grains of pyroxene and hornblende and opaque minerals occur in irregular forms and streaks (Fig. 4F). In all the samples, vesicles are scanty and are mostly empty.

4.3 Compositional variations:

The major oxides of the seamount basalts (Table 4) show a narrow range of SiO₂ (49.0-51.5 wt%), Al₂O₃ (13.6-18.6 wt%), Fe₂O₃ (2.5-3.5 wt%), FeO (8.0-11.5 wt%), MgO (3.5-9.0 wt%), significant CaO (9.5-12.0 wt%), moderate Na₂O (2.5-3.5 wt%), moderate to slightly high TiO₂ (1.3-2.1 wt%) and low P₂O₅ (0.1-0.2 wt%). The closely similar values of oxides of Si, Al, Ca and Na and low to moderate Mg (3.5-6.5 wt%) attest to the abundance of plagioclase and rarity of olivine and augite. Sample PD 26 has the lowest K₂O (0.1 wt%) and highest MgO (~9.0 wt%) (Table 4) while the other samples have moderate to high K₂O (0.4-1.0 wt%). The enhanced K₂O content may indicate seawater alteration of the basalts but this may not be a primary reason since the samples were made free of visible alteration prior to chemical analyses. Additionally, the low Fe⁺³ and MnO (<0.25 wt%) contents indicate less alteration and freshness of the studied samples. On these bases, the high K could be an inherent property of the CIOB basalts as reported for the seafloor basalts (Das et al., 2012) and those from the DSDP sites in the Indian Ocean (Table 5) (Frey et al., 1977; Thompson et al., 1978).

5.0 Interpretations and Discussion

5.1. Seamount Morphology:

As stated previously, based on their morphological characteristic there are four types of seamounts. Type 1 seamounts indicate point source with flow of lava along the flanks and type 2 are complex seamounts with volcanic mass added in phases. Type 3 represents seamounts with collapse summits that formed through fissure eruptions whilst type 4 indicates seamounts that had a point (?dyke) source of eruption. Based on the above parameters, the seamounts 1 and 2 are complex while seamounts 3 and 4 are a combination of fissure and dyke eruptions.

5.2. Petrography:

The petrographical features and mineral assemblages of the presently studied seamount basalts are similar to the earlier reports (cf. Mukhopadhyay et al., 1995). The CIOB basalts are of

four types: aphyric, plagioclase dominant, plagioclase + olivine and plagioclase + olivine + augite (Mislankar and Iyer, 2001). On this basis the presently investigated seamount basalts are plagioclase dominant variety and exhibit characteristics of the HPPB and MPPB varieties (i.e., Highly- and Moderately- Plagioclase Phyric Basalts) of the MAR basalts (Bougault and Hekinian, 1974).

The plagioclase morphotypes in the CIOB basalts represent different stages in their cooling. The sheaf-like radial clusters and spherulitic textures (Fig. 5A, B) form due to rapid growth during quenching of the lava (Gill, 2010) and/or resorption because of disequilibrium with the melt (Gill, 2010). In oceanic basalts, several types of twinned and zoned plagioclase have been recognized that are diagnostic recorders of their growth and of the magmatic conditions (Gill, 2010). In the CIOB basalts (seafloor and seamounts) a profusion of zoned plagioclases (Mislankar and Iyer, 2001) is absent in contrast to the MORB (cf. Iyer and Banerjee, 1993; Ray et al., 2007).

The factors that favor the growth of unzoned crystals are: very slow cooling of the magma; rapid stirring of the magma; large supply of the magma that does not change the bulk composition; growth from a bulk composition of minimum melting temperature; interruption of zoning by settling or floating of uniform crystals that form a compact rock; and rapid eruption of a volcanic rock with inhibition of reaction of early formed phenocrysts with the host magma (Gill, 2010). For e.g., the varieties of plagioclases in the FAMOUS region (French-American Mid-Ocean Undersea Study), Iceland and Carlsberg Ridge suggest magma mixing, which is common at slow- to intermediate-spreading ridges (e.g., Iyer and Banerjee, 1993).

The amount of zoning could be approximately proportional to the ascent rate of magma and the magma chamber size could affect the abundance of the unzoned crystals (Fisk, 1984). On these tenets, it appears that the magma for the CIOB basalts (seafloor and seamounts) upwelled during fast rate of crustal spreading and ascended rapidly from small chambers with little time to form zoned crystals. Alternatively, unzoned minerals may form due to removal of the residual host liquid prior to its complete crystallization and this result in glassy inclusions in the plagioclase phenocrysts (Mislankar and Iyer, 2001). The phenocrysts probably formed when the magmatic mush underwent re-melting and re-crystallization and show the effects of magmatic corrosion.

In general, the predominance of plagioclase suggests the CIOB seamount basalts to be moderately evolved HPPB/MPPB and implies a process of fractional crystallization of a plagioclase-rich magma, an observation that concurs with the other reported oceanic basalts (e.g., Iyer and Banerjee, 1993).

5.3 Major oxides:

The CIOB seamount samples are compared with the CIOB seafloor that have $\text{TiO}_2 < 2.0 \text{ wt}\%$ and $\text{TiO}_2 > 2.0 \text{ wt}\%$ (Das et al., 2012), the CIR (Ray et al., 2007), the Afanasy-Nikitin Seamount (Kashintsev et al., 1987; Mahoney et al., 1996) and ferrobasalts from the CIOB (Iyer et al., 1999), SEIR (Michard et al., 1986; Dosso et al., 1988) and Ninetyeast Ridge (Frey et al., 1977; Thompson et al., 1978) (Table 5). On the TAS diagram (Fig. 6), the CIOB seamount samples and from other locations all fall in the field of basalt, except the Afanasy-Nikitin Seamount which lie in the field of basanite.

In the binary plots of MgO vs. oxides two distinct clusters are noticed (Fig. 7). Cluster C1 has high MgO (6.0-10.0 wt%) and low $\text{TiO}_2 (< 2.0 \text{ wt}\%)$ and low FeO ($\sim 10.0 \text{ wt}\%$) and consist of from the presently studied seamounts, CIOB seafloor, CIR, SEIR and DSDP Sites 215 and 254. Cluster C2, with lower MgO (2.0-6.0 wt%) and higher $\text{TiO}_2 (> 2.0 \text{ wt}\%)$ and FeO ($> 10.0 \text{ wt}\%$), contains the CIOB seafloor basalts and ferrobasalts from the Afanasy-Nikitin Seamount, SEIR, CIOB and DSDP Sites 214 and 216 near the Ninetyeast Ridge. (It needs to be noted that the CIOB seafloor basalts have TiO_2 with either less than or more than 2.0 wt% and FeO either $> 10.0 \text{ wt}\%$ or $< 10.0 \text{ wt}\%$, cf. Das et al., 2012. Hence, these basalts fall either in cluster C1 or C2, respectively). The plots, show a scatter for SiO_2 and Al_2O_3 for C1 and C2 (Figs. 7A,B), with the higher Al_2O_3 indicating an increase in plagioclase grains.

The behaviour of FeO and TiO_2 is similar (Figs. 7C,D). TiO_2 in C1 is lower ($< 2.0 \text{ wt}\%$) and decreases with increasing MgO while in C2 it is higher ($> 2.0 \text{ wt}\%$) and increases with decreasing MgO. These characteristics point to the ferrobasaltic nature of the C2 basalts. The TiO_2 and FeO contents of the seamount basalts suggest the onset of Ti-magnetite crystallization but thin sections did not reveal significant Fe-Ti-minerals, unlike the CIOB ferrobasalts that have distinct opaque minerals (Iyer et al., 1999). Perhaps the C1 basalts crystallized under low oxygen fugacity ($f\text{O}_2$) condition that inhibited the formation of large sized Fe-Ti minerals (Toplis and Carroll, 1995). Incidentally, the opaque minerals are in the glassy groundmass as minute single domain ferromagnetic grains that are 1 to $0.06 \mu\text{m}$ in size (Das et al., 2012).

The positive trend of CaO with MgO (Fig. 7E) of both the clusters indicates removal of Ca-pyroxene that resulted in iron enrichment in the basalts. The K_2O contents show a strong inverse relation with MgO for C1 basalts ($\text{K}_2\text{O} < 0.5 \text{ wt}\%$) while C2 basalts ($\text{K}_2\text{O} > 0.6 \text{ wt}\%$) show a weak relation (Fig. 7F). In terms of K_2O , Zr/Nb and $(\text{La}/\text{Yb})_N$ contents, the C1 basalts resemble those of the SWIR segment, where the K-rich basalts formed by fractionation and/or assimilation processes

(le Roex et al., 1982, 1992; Michard et al., 1986; Dosso et al., 1988). The antithetic relation between Na₂O and MgO (Fig. 7G) can be accounted by the accumulation of tiny crystals of Na-plagioclase in the glassy groundmass. The distribution of P₂O₅ (Fig. 7H) is akin to that of K₂O. The characteristics of the C2 basalts with high TiO₂ and a moderately high P₂O₅ and K₂O (Table 4) may be caused by differences in bulk distribution coefficient of these elements during partial melting of the sub-oceanic mantle i.e., $D^K < D^P < D^{Ti}$ (cf. Sun and McDonough, 1989).

5.4 Minor, trace and rare earth elements:

The average contents of minor, trace and RE elements of the CIOB seamount basalts are compared with those from the CIOB seafloor, CIR, SEIR and Afanasy-Nikitin Seamount and the ferrobasalts of the CIOB, SEIR and DSDP Sites near the Ninetyeast Ridge (Table 5). The Co and Cu values are on par with MORB and indicate the rarity of olivine and augite that host these elements. In the plots of V vs. FeO (Fig. 8A) and V vs. TiO₂ (Fig. 8B) the C1 basalts have lower FeO (<10 wt%) and V (200-350 ppm) while those of C2 have higher FeO (>10 wt%) and V (300-400 ppm). The positive trends indicate accumulation of FeO and TiO₂ with progressive crystallization and incorporation of V in the opaque minerals (Fig. 4F) and glassy phase.

The basalts of both the clusters have a negative trend on the plot of Ba/Nb vs. MgO (Fig. 8C). The enhanced Ba/Nb ratio (13-59), in contrast to N-MORB (~5.7) and Primitive Mantle (PM 9.56), and its decrease with increasing MgO suggest partial melting of the source region that resulted in the CIOB seamount basalts. The reverse trend of increasing Ba/Nb ratio with decreasing MgO indicates that fractional crystallization of the magma controlled the distribution of the incompatible elements. The CaO/Al₂O₃ ratio (~0.5 to 0.9) points to the presence of Ca-rich plagioclase (Fig. 8D). This is in contrast to the CIOB ferrobasalts that have very low CaO/Al₂O₃ ratio (0.1 to 0.2) (Iyer et al., 1999).

On plots of Zr vs. FeO (Fig. 9A) Zr vs. TiO₂ (Fig. 9B) the clusters C1 and C2 depict a positive trend. The Zr values are slightly low in basalts of C1 (50-150 ppm) as compared to C2 (75-200 ppm). The C1 basalts show a positive trend between Y and Zr (Fig. 10A) and have lower Y (20-40 ppm) relative to the higher Y (40-80 ppm) in C2 basalts. In case of Rb and Sc contents, C1 basalts have Rb <20 ppm and Sc <50 ppm for Zr between <50 and 150 ppm while C2 basalts have Rb >20 ppm and Sc >50 ppm for Zr >100 ppm and both these elements have a flat distribution (Fig. 10B,C). The plot (La/Sm)_N (normalized with chondrite values of McDonough and Sun, 1995) vs. Zr/Nb (Fig. 10D) indicates that the C1 basalts have lower (La/Sm)_N (<1 to 2) and C2 has (>1.5 to 4.5) for a similar spread of Zr/Nb (30 to 150). The above observations together with the small range of Ce/Y

(0.15-0.34) and Nb/Zr (0.01-0.02) ratios indicate that partial melting of the source region produced the seamount basalts.

The relative enrichments of Pb, Th and U (Table 4) in the CIOB seamount basalts relative to the N-MORB are enigmatic. The U/Pb ratio (0.03-0.58) of the CIOB seamount basalts is similar to N-MORB whereas Ce/Pb and Ce/U ratios (0.64 to 5.75 and 8 to 31, respectively) are remarkably low vis-à-vis the N-MORB (~25, >150; Hofmann et al., 1986) and PM (28, 97; McDonough and Sun, 1995; Hannigan et al., 2001). It appears that U is more incompatible than Pb during mantle melting in the CIOB basalts as compared to the other oceanic basalts (Halliday et al., 1990). This may be ascribed to recycling of the oceanic crust and sediment in the Indian Ocean (Rehkamper and Hofmann, 1997; Hemond et al., 2012; Mukhopadhyay et al., 2016).

The plot of $(La/Sm)_N$ against TiO_2 (Fig. 11A) shows a lateral spread for C1 basalts for a range of $(La/Sm)_N$ (<1.0 to 3.0) and suggests a larger extent of partial melting of a near homogeneous source magma that had a low TiO_2 content (<2.0 wt%). This is in contrast to the C2 basalts that have a narrow range of $(La/Sm)_N$ (1.0-2.0) and higher TiO_2 content (>2.0 wt%) and perhaps formed due to a smaller extent of partial melting. The plot of Ba/Nb vs. V (Fig. 11B) portrays a flat distribution for C1 than the C2 basalts. The plot of Ce/Y vs. Zr/Nb (Fig. 11C) shows C1 to have a higher range of Zr/Nb (>80) similar to N-MORB (>30; Wilson, 1989). This is analogous to the SEIR basalts (Michard et al., 1986; Dosso et al., 1988) and points to a genetic relation between the basalts of the CIOB seamounts and the SEIR. The La/Yb and Ce/Y ratios (~0.8-2.0 and ~0.15-0.34, respectively) of the CIOB seamount basalts are close to the chondrite values ($La/Yb \approx 1.39$ and $Ce/Y \approx 0.39$; Sun and McDonough, 1989) and were probably affected because of crystallization of olivine and pyroxene.

The concentrations of incompatible and RE elements of the CIOB seamount basalts were plotted (Figs. 12-14) after normalizing to N-MORB and PM compositions (after McDonough and Sun, 1995), and also shown are the values of N-MORB, E-MORB and OIB (Ocean Island Basalts, Sun, 1980). In general, the Ba, U and Pb contents are higher and Th and Nb are relatively lower than N-MORB whereas the less incompatible elements (e.g., Nd, Zr, Sm, and Yb) show a flat N-MORB normalized distribution. The basalts from the CIOB seafloor, CIR and SEIR (Fig. 12), also reflect this pattern. The CIOB seamount basalts are slightly depleted in Light REE (LREE) (Fig. 13) with the exception of sample HRX 13 that shows a prominent positive Ce anomaly while the High REE (HREE) shows a near flat pattern. The average ratio of LREE/HREE is quite alike being 1.63, 1.75 and 1.63 for the seamount basalts, CIOB seafloor basalts ($TiO_2 < 2.0$ wt %) and CIOB seafloor basalts ($TiO_2 > 2.0$ wt %), respectively.

The above observations are on par with the N-MORB, DSDP Site 215 basalts (Thompson et al., 1974; 1978; Reddy et al., 1978; Frey et al., 1977), SEIR basalts and “normal” MAR and EPR basalts (Dosso et al., 1988). The behavior of the trace and RE elements and similar LREE/HREE ratio suggest that the variations and trends for the CIOB seamount and seafloor basalts are comparable irrespective of the location of the samples, and suggest a similar source and mode of evolution. Considering the seamount morphology, and petrography and composition of the basalts, we now discuss their origin, mode of formation and relation with the seafloor basalts.

As mentioned earlier, the CIOB seamount basalts are geochemically analogous to those from the seafloor (with <2.0 wt% TiO_2), CIR (Ray et al., 2007) and SEIR (Michard et al., 1986) but for the high K_2O content (0.63 wt%) and low P_2O_5 (0.11 wt%) that resemble the basalts from the SEIR (0.66 wt%; Dosso et al., 1988) (Table 5). K-enrichment is also noted in the ferrobasalts at DSDP Site 216 (0.90 wt%), Spiess Ridge (0.5-1.11 wt%), Chain Ridge (0.77-0.83 wt %), Iceland (0.74 wt%), Broken Ridge (0.37-2.06 wt %), Kerguelen Plateau (0.72-1.95 wt%) and the CIOB ferrobasalts (le Roex et al., 1982; Dosso et al., 1988; Mahoney et al., 1995; Iyer et al., 1999). It is noteworthy that all the above sites with enhanced K_2O have a ferrobasaltic affinity, except DSDP Site 215 which lies to the east of the CIOB. Similar to the high K_2O values as the CIOB seamount basalts are the K-P-rich basalts from the DSDP Site 215 that have enhanced K_2O (0.90 wt%) and (P_2O_5 0.31 wt %) and low TiO_2 (1.71 wt%) and FeO^* (8.43 wt%) (Table 5). This fact indicates a common source for the basalts of the DSDP Site 215 and CIOB seamounts and seafloor (<2.0 wt% TiO_2). The high-K content could be an inherent property (Frey et al., 1977; Thompson et al., 1978) and indicate the mildly alkalic nature of the basalts from the CIOB seamounts and seafloor. Such basalts occur in the rift zones e.g. Iceland (cf. Thy, 1989) and off the west coast of India (Karisiddaiah and Iyer, 1991).

The range of Zr/Nb ratio of the studied basalts is high (40-143) (Table 4) and the average Zr/Nb ratio of the CIOB seamounts are 54.53; CIOB seafloor (<2.0 wt% TiO_2) 48.6; CIOB seafloor (>2.0 wt% TiO_2) 56.8, CIR 23.4 and SEIR 28.5. High Zr/Nb ratio (30-110) has been reported for the basalts present along the different segments of the MAR and for the abyssal basalts from the Indian and Pacific oceans (Erlank and Kable, 1976). Considering the similar Zr/Nb ratio of the CIOB basalts (seamounts and seafloor) it appears that the source areas were uniformly depleted.

In the world oceans, ferrobasalts have been dredged from propagating spreading centre (Christie and Sinton, 1981) and from the fast and intermediate spreading ridges, such as Juan De Fuca Ridge (half rate 30 mm/yr; Vogt and Byerly, 1976); the Galapagos Spreading Center (GSC; 30

mm/yr; Byerly, 1980), Reykjanes Ridge (23 mm/yr; Rona and Scott, 1993), and the Nazca Plate (DSDP Leg 34; 85 mm/yr; Mazzullo and Bence, 1976). Therefore, it is apparent that ferrobasalts are not only associated with propagating rifts but are also found in high amplitude magnetic (HAM) zones, fast spreading ridges, or a combination of such sites. The CIOB seafloor (>2.0 TiO₂ wt %) and some seamounts have ferrobasalts that formed near the HAM zone which represents the TJT-In (Iyer et al., 1999). This observation points to the occurrence of two types of mafic volcanism in the CIOB: ferrobasaltic and N-MORB.

The CIOB seamount basalts with moderate FeO and MgO contents indicate significant fractionation of the magma. Furthermore, the narrow CaO/Al₂O₃ ratio, moderate to high abundances of K₂O and P₂O₅ and comparable ratios of highly incompatible elements, point to a single or uniform magmatic source. The fairly enriched LREE of the CIOB seamount basalts (Fig. 14) suggests an absence of extensive partial melting of the source region as this would have resulted in lower LREE or a flat REE pattern (Wilson, 1989; Feigenson et al., 1996; Das et al., 2012). Hence, it is evident that the CIOB basalts formed by fractional crystallization of a ferrobasaltic melt that had undergone low partial melting. Our observation is similar to that made for the EPR basalts for which Sinton and Detrick (1992) opined the source melt was ferrobasaltic in composition.

The formation of ferrobasalts has been suggested to be due to fractional crystallization of melts occurring at shallow-level (Vogt and Byerly, 1976; Christie and Sinton, 1981; Anderson et al., 1980). This mechanism has been considered to account for the ferrobasalts of the DSDP sites 214 and 216 (Thompson et al., 1978), GSC (Byerly, 1980), the Conrad fracture zone (le Roex and Dick, 1981) and Spiess Ridge (le Roex et al., 1982). Ferrobasalts could also form from a Fe-Ti rich parent (Klein et al., 1991), magma mixing (Wilson et al., 1988) or from a large molten magma chamber (e.g. 13° and 23° S in the EPR with a spreading rate of 75 mm/year) (Sinton and Detrick, 1992).

The along-axis petrologic variations of fast-spreading ridges, such as the EPR the GSC and the moderate to fast-spreading ancient SEIR, are suggested to be due to variable degrees of shallow crystallization or by differences in the depths of melt generation and residence (Byerly, 1980; Langmuir et al., 1986; Klein et al., 1991). The mechanism of ‘neutral buoyancy zonation’ structure of magma reservoirs coupled with fractional crystallization could explain the retention of magma at shallow depths for long duration of time (Ryan, 1994). The horizon of neutral buoyancy (HNB), defined as that ‘depth interval within which the magma density and the aggregate country rock density are equal’ occurs sub-lithospherically and has a narrow vertical and a wide lateral extent. Beneath the HNB region, magma ascends due to positive buoyancy and is stabilized at a shallow depth (2–4 km), while above this region the magma descends by negative buoyancy. The HNB of

tholeiitic melts thus provides congenial conditions for the long-term stability of magma reservoirs and the nearly constant magmatic compositions over millions of years (Ryan, 1994; Thompson and Petford, 2008).

The CIOB seafloor formed during a phase of superfast spreading rate (95 mm /yr) between ~56 and 51 Ma (Mukhopadhyay et al., 2002). The similarity in compositions of the basalts from the CIOB seafloor and the seamounts suggest sharing of a common magmatic source. This is may be because of episodic eruptions due to changes in relative plate motions that may have produced the seamounts, as at Iceland (cf. Gaina et al., 2016). Such a process could be envisaged if we consider that the trapped off-axis melts in older oceanic crusts (cf. Ryan, 1994) i.e. the crusts of the CIOB and the SEIR were underlain by small magma chambers at relatively shallow depths (Mukhopadhyay et al., 1995; Das et al., 2005; Das et al., 2007). The fast spreading rate of the plates perhaps restricted the residence of the melt for a long time. Under this circumstance the melt may not have differentiated extensively and largely resulted in uniform basaltic magma compositions both, at the seafloor and seamounts. Based on clinopyroxene–melt barometry it was suggested that an underplated zone at 7–14 km depth could exist for Cumbre Vieja volcano, La Palma (Canary Islands; Klügel et al. 2005). In this underplated zone, similar to the HNB, the ascending magmas are either ponded prior to eruption or are temporarily retained during eruption and thus resulting in long- and short-term storage of magma.

Besides the spreading rate and residence time, the other factors, such as rate of magma supply, low degree of partial melting or presence of small relatively stable magma chamber could also produce basalts of similar compositions in the CIOB. Additionally, a lateral transport of the melts and their subsequent ascension through a thin lithosphere could result in focusing of melts at the seamounts, similar to the observed along-axis variation at the Knipovich Ridge (Norwegian-Greenland Sea; Schlindwein et al., 2013).

The basalts from the older seamounts of the CIOB resemble those from the present day younger, near-axis located seamounts of the MAR and EPR (Mukhopadhyay et al., 1995). Shen et al. (1995) reported that in the EPR (15° and 19° S) the production of seamounts in closely spaced (20-25 km) seamount chains could share a common source and this results in an increased magmatic flux to one chain and decrease in the adjacent chain. Based on the Na₈ (extent of melting) and Fe₈ (depth of melting) values, of more than 2000 basalts from the EPR and seamounts, Zhang (2011) concluded that the fast spreading EPR and the nearby seamounts have a common source that passed through the lithosphere with different thermal structures and underwent crystallization. In the CIOB the seamount chains are spaced between 50 and 55 km (Das et al., 2007) and therefore, it is highly

conceivable that the seamounts and the seafloor may have tapped a common source region from the then spreading ridges.

We propose that during the creation of the CIOB seafloor at a fast rate of spreading (95 mm/yr), the formation of large magma chambers was hindered. On the other hand, pockets of small magma bodies were retained that were uniformly depleted, regularly tapped and replenished. This led to concurrent or penecontemporaneous production of the seafloor and the associated seamounts. The seamounts were emplaced along the N-S trending fractures that were formed as a consequence of the Indo-Eurasian collision. At a later stage the seamounts witnessed episodes of low intensity hydrothermal activity between 625 ka and 100 yr age (Iyer, 2005, Nath et al., 2008). The addition of such materials resulted in either an N-S or E-W trending bulge of the base of some of the seamounts. This fact indicates that the CIOB is still unstable – tectonically and volcanically – and prone to episodic hydrothermal activities.

6.0 CONCLUSIONS

The salient findings from the investigations of four of the CIOB seamounts are:

1. The seamounts of variable shapes (conical and elongated) and heights (625 to 1200 m) were produced when the CIOB experienced a half-spreading rate of 95 mm /yr between the period 56 and 51 Ma.
2. The basalts have similar petrographical features and are highly- to moderately- plagioclase phyrlic.
3. The basalts are quite homogeneous with low FeO* and TiO₂, and have elevated K₂O contents and are slightly enriched in the LREE.
4. The seamount basalts are geochemically comparable with those from the CIOB seafloor (TiO₂ <2.0 wt%), CIR and SEIR but are unlike those from the CIOB seafloor (TiO₂>2.0 wt%), and typical ferrobasalts from the CIOB, SEIR and DSDP Sites near the Ninetyeast Ridge.
5. The CIOB seafloor and the associated seamounts were generated from the unified crusts that formed the CIR and the SEIR and these shared a common ferrobasaltic melt source that in course of time differentiated to N-MORB.

In the future, geophysical, petrological, geochemical and age determination of additional seamounts could help to address and further clarify the processes that were responsible for the formation and evolution of the seamounts in the CIOB.

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Table 1. Locations of samples collected from four seamounts in the CIOB (#1 to 12). Seamount 1 samples #1 to 5, seamount 2 samples #6 to 10, seamount 3 sample #11 and seamount 4 sample #12. Details of previously studied seamounts are also shown (samples #13-15 basalts, Mukhopadhyay et al., 1995; samples #16-22 ferrobasalts, Iyer et al., 1999; sample #23 VHM volcanogenic-hydrothermal material, Iyer et al., 1993). na = not available. TJT-In = Trace of Triple Junction on the Indian Plate.

Table 2. Characteristics of the four studied seamounts of the CIOB wherefrom the samples were dredged. HW = height-width ratio, Sigma = 1-f, where f is the flatness of the seamount. For locations of seamounts 1 to 4, see Fig. 2a,b.

Table 3. Petrographic details of the seamount samples. Samples #PD 4, PD 9, HRX 15 from seamount 1; samples PD 12, PD 33, PD 37; Sample HRX#1 from seamount 3; sample HRX#13 from seamount 4. See figure 2 for sample locations. RFA = red feathery alteration.

Table 4. Whole-rock composition of basalts from four CIOB seamounts. Major oxides are in wt% and elements are in ppm.

Table 5. CIOB seamount basalts compared with those from the seafloor, CIR and SEIR. Oxides are in wt% and elements in ppm. 1 = Present study; 2 = avg of 16 CIOB basalt (<2 TiO₂); 3 = avg of 25 CIOB basalts (>2 TiO₂) (both from Das et al., 2012); 4 = avg of 15 CIR basalts (Ray et al., 2007), 5 = average of 10 SEIR basalts (Dosso et al., 1988); 6 = avg of 2 basalts from Afanasy-Nikitin (Mahoney et al., 1996); 7 = avg of 3 basalts from Afanasy-Nikitin seamount (Kashintsev et al., 1987); 8 = avg of 10 CIOB ferrobasalts (Iyer et al., 1999); 9 = ferrobasalts of SEIR (Dosso et al., 1986); 10 = Ferrobasalt from site 214; 11 = DSDP Site 215 basalt; 12 = DSDP Site 216 ferrobasalt; 13 = DSDP Site 254 ferrobasalts (Thompson et al., 1978); 14 = avg of 7 ferrobasalts from SEIR (Michard et al., 1986), na = not available.

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Figure 1. Generalized map of the Central Indian Ocean Basin and the adjacent regions showing (a) isobaths and the tectonic elements and (b) age of the ocean floor. The thick red line represents the present day spreading system. Continuous grey lines and black dotted lines represent isobaths (IOC-IHO-BODC, 2003) and fracture zones (Wessel et al., 2015), respectively. The triple junction trace is shown as dashed purple line. Colour-coded image of the ocean floor is plotted with the latest available age grid of the ocean floor (Müller et al., 2016). The locations of the samples used in the

present study are marked with red solid stars and those used by previous studies in this region are marked with blue triangles (Details are given in Table 1). SR: Sheba Ridge; CR: Carlsberg Ridge; CIR: Central Indian Ridge; ABB: Arabian Basin; ESB: Eastern Somali Basin; CIOB: Central Indian Ocean Basin; NER: Ninetyeast Ridge; DT: Deccan Trap.

Figure 2. General multibeam map of the CIOB studded with several seamounts, several of which have an approximate N-S alignment.

Figure 3. Bathymetry of the four studied seamounts and locations of the dredged samples. The numbers in red are the seamounts (1 to 4) discussed in the text.

Figure 4. Hand specimen photographs of the basalts recovered from the seamounts. A = Pillow basalts (PD 4) with a thick rind of oxidized glass. B = Pillow basalts (PD 12) with a fresh rind of glass and unaltered interior. C (sample HRX 1) and D (sample HRX 13) = Fragments of pillow basalts with variable stages of freshness of the glassy rind and interior.

Figure 5. Photomicrographs of the seamount basalts. A = Radial sheaves of plagioclase in a glassy groundmass ($\times 100$). B = Submicroscopic sheaves of plagioclase with tiny elliptical blebs of glass surrounded by plagioclase microlites to form a reddish alteration. Spherulitic texture formed in altered glass. C = Transition zone between fresh glass (upper left) and fresh interior. D = Fresh euhedral olivine phenocrysts in a glassy matrix. E = Intersertal and ill-developed flow textures formed by plagioclase laths in a glassy matrix. F = Coarse grains of hornblende and pyroxene in a basaltic dyke (sample HRX 13). A string and mass of opaque are present near the pyroxene. Magnification for B to F $\times 40$. Scale = 0.1 mm. A-E under plane light, F under crossed polars.

Figure 6. TAS diagram show the rock samples from the CIOB seamounts to be basalts as are those from DSDP and SEIR sites. The samples from the Afanasy-Nikitin Seamount are basanite. Abbreviations: Ph = Phonolite; TPh = Tephriphonolite; PhT = Phonotephrite, T = Tephrite; Ba = Basanite; PBas = Picro-basalt; TrBas = Trachybasalt; BaTrAnd = Basaltic trachyandesite; TrAnd = Trachyandesite; BasAnd = Basaltic andesite.

Figure 7. Interrelationship between MgO and the major oxides. The groupings of basalts in cluster C1 are shown as full oval and those of cluster C2 as dashed oval.

Figure 8. The relation between V vs. FeO (Fig. 8A), V vs. TiO₂ (Fig. 8B) Ba/Nb vs. MgO (Fig. 8C) and CaO/Al₂O₃ vs. MgO (Fig. 8D). Legends as in Fig. 7.

Figure 9. Plots of Zr v/s FeO and TiO₂. Both the clusters depict a positive trend that point to partial melting of the source magma. Legends as in Fig. 7.

Figure 10. Y and Zr (Fig. 10A). Cluster C1 basalts have Rb <20 ppm and Sc <50 ppm for Zr between <50 and 150 ppm while cluster C2 basalts have Rb >20 ppm and Sc >50 ppm for Zr >100 ppm and both these elements depict a flat distribution (Fig. 10B,C). Plot of (La/Sm)_N (normalized with chondrite values of McDonough and Sun, 1995) vs. Zr/Nb (Fig. 10D). Legends as in Fig. 7.

Figure 11. Plot of (La/Sm)_N against TiO₂ shows the clusters C1 and C2 (Fig. 11A). Ba/Nb vs. V (Fig. 11B), Ce/Y vs. Zr/Nb (Fig. 11C).

Figure 12. Spidergrams of incompatible trace elements of the CIOB seamounts. A: sample /N-MORB. B: sample /primitive mantle. Also plotted are the values for N-MORB and E-MORB (McDonough and Sun, 1995) and OIB (Ocean Island Basalts, Sun, 1980).

Figure 13. Spidergrams of incompatible trace elements of basalts from the CIOB seamounts and other areas (See Table 5). A: average sample /N-MORB. B: average sample /primitive mantle.

Figure 14. Plots of rare earth elements of the CIOB seamounts. A: sample /N-MORB. B: sample /primitive mantle.

Table 1. Locations of samples collected from four seamounts in the CIOB (#1 to 12). Samples #1 to 5 are from seamount 1, samples #6 to 10 are from seamount 2, sample #11 is from seamount 3 and sample #12 is from seamount 4. Details of previously studied seamounts are also shown (samples #13-15 basalts, Mukhopadhyay et al., 1995; samples #16-22 ferrobasalts, Iyer et al., 1999; sample #23 VHM volcanogenic-hydrothermal material, Iyer et al., 1993). na = not available. TJT-In = Trace of Triple Junction on the Indian Plate.

S.No.	Sample #	Latitude (°S)	Longitude (°E)	Water depth (m)	Sample position at seamount
1	PD4	12° 30'	76° 23'	5320	at foot eastern side
2	PD9	12° 30'	76° 20'	na	at summit
3	PD15	12° 30'	76° 24'	5385	at western base
4	PD17	12° 28'	76° 20'	4800	along northern upper flank
5	HRX15	12° 28'	76° 18'	5100	at northwest flank
6	PD12	12° 35'	76° 15'	4400	along western flank
7	PD26	12° 34'	76° 22'	5375	at foot eastern side
8	PD29	12° 37'	76° 23'	5250	at southeast flank
9	PD33	12° 33'	76° 18'	4300	at northern flank
10	PD37	12° 32'	76° 14'	5250	at northwest base
11	HRX1	12° 31'	76° 12'	5150	at southwestern upper flank
12	HRX13	14° 05'	79° 20'	4820	at northeast upper flank
13	GML 36	12° 35'	76° 38'	5100	along western flank
14	GML 4	12° 49'	76° 33'	5440	along northern flank
15	GML 26	13° 04'	76° 06'	5400	along western flank
16	GML 16	11° 00'	77° 55'	5250	at foothill/slope
17	GML 28	11° 00'	78° 00'	5300	at foothill
18	GML 30	11° 12'	77° 49'	5400	at base
19	GML 02	12° 33'	77° 12'	5210	at base
20	GML 17	13° 00'	76° 30'	5400	at slope
21	GML 26	12° 33'	77° 12'	5200	at foothill
22	GML 37	12° 59'	76° 30'	4975	at base, near E-W lineaments and TJT-IN
23	VHM 93	14°	76°	4440	elongated base

Table 2. Characteristics of the four studied seamounts wherefrom the samples were dredged from the CIOB. HW = height-width ratio, Sigma = 1-f, where f is the flatness of the seamount. For locations of seamounts 1 to 4, see Fig. 2a,b.

Seamounts	height (m)	basal area (sq. km)	summit area (sq. km)	Volume (cu. km)	flatness	slope angle (degree)	Summit width (km)	HW ratio	Sigma	basal width (km)	1-f
1	750	65.00	0.80	22.50	0.11	12.90	1.01	0.16	0.19	9.10	0.89
2	1200	236.00	1.00	101.00	0.07	8.40	1.13	0.14	0.15	17.3 3	0.93
3	625	40.00	0.30	9.10	0.09	10.90	0.62	0.18	0.19	7.14	0.91
4	1150	88.61	0.66	33.97	0.08	12.02	0.92	0.21	0.23	10.8 0	0.92

Table 3. Petrographic details of the seamount samples. Samples #PD 4, PD 9, HRX 15 from seamount 1; samples PD 12, PD 33, PD 37; Sample HRX#1 from seamount 3; sample HRX#13 from seamount 4. See figure 2 for sample locations. RFA = red feathery alteration.

Sample #	Petrographical features	Seamount morphology
PD 4	Medium to fine grain, textures (crude flow, glomeroporphyritic, intersertal), fresh to altered glass with RFA, plagioclase laths plenty, few euhedral plagioclase microphenocrysts and phenocrysts, twinned, stressed, sector zoned. Few olivine.	Medium size, conical, steep flanks, single summit, N-S elongated
PD 9	Medium grain, plagioclase very abundant, laths, microphenocrysts, phenocrysts, sector zoned. Thick streak of opaque.	

HRX 15	Medium to coarse grain, fresh to less altered glass, abundant plagioclase laths, crude flow texture, euhedral plagioclase phenocrysts and microphenocrysts, stressed, fractured, twinned, cross-cutting. Few vesicles, patches of opaque.	
PD 12	Altered glass, reddish-brown, spherulitic, 3 fresh euhedral twinned plagioclase phenocrysts.	Large, E-W elongated, gentle slope.
PD 33	Glass with RFA, abundant plagioclase laths few stressed, sector zoned, microphenocrysts.	
PD 37	Quite fresh glass, medium to coarse grain, sub-ophitic, spherulitic, intersertal, abundant plagioclase laths, 2 megacrysts with glass inclusions, round and irregular vesicles.	
HRX 1	Glass is partly altered to palagonite, RFA, profuse plagioclase phenocrysts and laths show cross-cutting and sector zones. Megacrysts have glass inclusions. Below the glass is holocrystalline, plenty plagioclase laths, intersertal, porphyritic, ophitic and glomero-porphyritic textures. Twinned and zoned plagioclase. Fine grain olivine and augite in matrix. A few olivine microphenocrysts. Irregular empty vesicles.	small, conical, steep slope.
HRX 13	Two samples. (1) Fresh glass with abundant fresh euhedral olivine phenocrysts. Glass has spherulitic texture. Below glass is crystalline, intersertal, crude flow and intergranular textures. Profuse plagioclase laths, megacrysts, swallow-tailed, twinned, and stressed. Few empty vesicles. (2) Coarse grain, plagioclase. Hornblende replaces pyroxene, expulsion of large opaque grains. Dyke rock.	Part of a twin seamounts, conical, single summit, steep slope, flank cones.

Table 4. Composition of twelve basalts from four seamounts of the CIOB. Major oxides are in wt% and elements are in ppm. nd = not determined.

←-----1-----→←-----2-----→
 --→ ←-- 3--→ ←-- 4--→

wt %	PD4	PD9	PD1 5	PD1 7	HRX 15	PD1 2	PD2 6	PD2 9	PD3 3	HRX 37	HRX 1	HRX 13
SiO ₂	51.41	50.2 5	49.62	51.4 5	51.05	49.14	49.5 5	49.49	50.27	48.78	51.1	50.22
TiO ₂	1.36	1.56	1.41	1.61	1.38	1.96	1.47	1.51	1.52	2.12	1.33	1.46
Al ₂ O ₃	13.64	16.3 2	15.26	16.8 3	18.61	14.67	14.3 5	15.88	16.11	17.23	15.4	17.65
Fe ₂ O ₃	2.89	3.25	3.03	3.09	2.57	3.53	2.64	3.19	3.21	2.99	2.5	2.61
FeO	9.4	10.5 7	9.85	10.0 5	8.37	11.47	8.58	10.37	10.45	9.71	8.12	8.48
Mn O	0.16	0.15	0.15	0.18	0.25	0.17	0.16	0.12	0.1	0.12	0.20	0.10
Mg O	6.2	3.52	5.28	3.44	3.64	5.19	8.87	4.31	3.68	5.64	6.94	5.47
CaO	11.75	10.0 2	11.8	9.75	9.84	10.05	11.7 1	11.2	10.72	9.57	10.2	10.18
Na ₂ O	2.68	2.86	2.72	2.83	2.6	2.8	2.49	2.77	2.73	3.46	3.74	3.48
K ₂ O	0.37	0.99	0.68	0.67	0.91	0.88	0.1	0.93	0.9	0.37	0.43	0.37
P ₂ O ₅	0.16	0.14	0.21	0.11	0.19	0.15	0.1	0.22	0.18	nd	nd	nd
Tota l	100.0 2	99.6 3	100.0 1	99.4 1	100.0 1	100.0 2	99.9 9	100.0 1	100.0 2	99.99	99.96	100.0 2

ppm	PD4	PD9	PD15	PD17	HRX15	PD12	PD26	PD29	PD33	HRX37	HRX1	HRX13
La	3.11	3.40	5.92	5.06	6.73	3.50	3.50	2.84	7.37	9.09	2.73	7.15
Ce	7.20	7.43	6.79	6.91	10.57	10.33	10.45	6.64	6.70	18.15	7.99	44.85
Pr	1.15	1.32	1.95	1.73	1.68	1.99	1.87	1.53	1.76	2.83	1.29	1.90
Nd	8.53	9.50	10.04	9.31	11.18	9.55	8.63	8.55	11.58	17.93	9.23	10.92
Sm	2.89	3.19	3.39	3.33	3.46	3.86	3.37	3.19	3.46	5.16	3.03	2.77
Eu	1.12	1.26	1.28	1.30	1.28	1.36	1.17	1.24	1.34	1.73	1.21	1.02
Gd	4.63	5.11	4.56	4.52	5.32	5.20	4.46	4.30	5.59	7.71	4.83	4.41
Tb	0.88	0.98	0.85	0.86	1.01	0.97	0.82	0.81	1.05	1.42	0.94	0.72
Dy	5.04	5.56	5.75	5.68	5.80	6.49	5.37	5.70	6.09	7.99	5.34	3.99
Ho	1.09	1.23	1.32	1.27	1.23	1.44	1.20	1.30	1.34	1.73	1.17	0.83
Er	3.56	3.81	3.56	3.31	3.82	3.92	3.19	3.46	4.27	5.54	3.76	2.57
Tm	0.62	0.63	0.55	0.51	0.64	0.61	0.50	0.53	0.72	0.93	0.63	0.42
Yb	3.32	3.39	3.75	3.37	3.32	4.05	3.23	3.60	3.88	4.85	3.54	2.31
Lu	0.51	0.52	0.56	0.49	0.50	0.65	0.52	0.55	0.61	0.74	0.54	0.36
Y	36.47	37.72	39.90	35.37	37.54	40.19	33.89	38.28	44.53	52.65	37.57	22.62
Sc	44.24	47.53	45.68	46.04	47.46	45.96	34.67	46.87	46.56	48.97	46.69	42.69
Nb	0.95	1.02	0.63	0.67	1.33	1.53	2.56	0.73	1.01	2.81	1.76	3.87
Th	0.12	0.15	0.04	0.08	0.32	0.13	0.16	0.03	0.12	0.76	0.26	0.23
Sr	130.10	152.28	130.50	149.59	159.03	119.65	nd	138.69	140.79	nd	137.54	132.33
Rb	6.00	12.38	9.18	14.98	14.58	16.66	2.39	9.66	10.33	22.08	9.04	3.15
Ba	52.92	42.58	13.91	51.58	62.19	38.19	5.73	14.96	59.68	64.12	31.45	51.72
Co	45.48	38.49	39.52	39.42	58.34	34.74	39.37	24.76	22.84	73.63	41.31	107.72
Cu	123.68	128.31	142.53	133.13	159.15	60.22	54.82	105.73	189.52	nd	117.61	133.06
V	288.88	279.43	300.54	276.52	273.18	434.21	297.9	315.61	271.90	310.5	295.30	247.52
Ga	14.13	14.75	14.93	16.78	15.09	24.04	13.94	16.01	14.42	17.39	13.98	11.74
Zn	125.28	127.03	73.40	66.27	127.00	78.64	44.16	67.04	125.72	164.8	120.53	73.38
Zr	77.63	85.87	78.82	95.78	85.34	98.21	103.4	63.77	84.12	nd	83.71	36.46
U	0.40	0.86	0.24	0.84	0.72	0.59	0.33	0.42	0.74	1.00	0.62	0.28
Pb	6.35	2.47	1.84	1.45	4.00	2.82	1.81	3.23	10.48	3.60	5.76	10.84

Table 5. CIOB seamount basalts compared with those from the seafloor, CIR and SEIR. Oxides are in wt% and elements in ppm. 1 = Present study; 2=avg of 16 CIOB basalt (<2 TiO₂); 3=avg of 25 CIOB basalts (>2 TiO₂) (both from Das et al., 2012); 4=avg of 15 CIR basalts (Ray et al., 2007), 5=average of 10 SEIR basalts (Dosso et al., 1988); 6=avg of 2 basalts from Afanasy-Nikitin (Mahoney et al.,1996); 7=avg of 3 basalts from Afanasy-Nikitin seamount (Kashintsev et al.,1987); 8=avg of 10 CIOB ferrobasalts (Iyer et al., 1999); 9=ferrobasalts of SEIR (Dosso et al., 1986); 10=Ferrobasalt from site 214; 11=DSDP Site 215 basalt; 12=DSDP Site 216 ferrobasalt; 13= DSDP Site 254 ferrobasalts (Thompson et al., 1978); 14=avg of 7 ferrobasalts from SEIR (Michard et al.,1986), na = not available.

wt %	1	2	3	4	5	6	7	8	9	10	11	12	13
SiO ₂	50.2	49.70	49.41	49.13	49.62	48.65	44.62	47.75	50.17	48.1	50.4	49.5	47.66
TiO ₂	1.56	1.56	2.28	1.37	1.19	2.01	3.26	2.58	3.05	2.35	1.71	2.75	2.12
Al ₂ O ₃	16	15.54	15.41	16.03	16	17.47	17.04	15.91	13.21	14.9	16.8	13.5	15.37
Fe ₂ O ₃ ^T					10.05				15.26				
Fe ₂ O ₃	2.96	1.84	2.09	9.98	nd	7.45	nd	nd	nd	nd	nd	nd	nd
FeO	9.62	9.38	10.66	nd	nd	nd	12.04	14.88	nd	14.6	8.43	13.8	12.9
MnO	0.16	0.17	0.22	0.20	0.17	0.14	0.15	3.75	0.21	nd	nd	nd	nd
MgO	5.18	5.76	5.08	7.28	8.14	1.38	1.71	9.07	4.22	6.45	6.48	6.57	8.92
CaO	10.57	11.15	9.80	10.91	11.91	8.78	6.37	3.18	9.03	9.04	10.95	8.79	8.53
Na ₂ O	2.93	3.07	2.97	3.03	2.45	4.13	3.12	0.89	3.15	2.75	3.17	2.57	2.69
K ₂ O	0.63	0.53	0.70	0.34	0.17	3.61	2.12	0.43	0.66	0.37	0.9	0.9	0.32
P ₂ O ₅	0.16	0.15	0.18	0.17	0.11	1.37	1.15	0.31	0.37	0.19	0.31	0.22	0.24
LOI	nd	1.09	1.22	0.74	nd	4.50	nd	nd	nd	nd	nd	nd	nd
H ₂ O							6.07						
Total	99.97	100.95	101.05	99.19	99.81	99.45	97.65	98.76	99.33	98.75	99.15	98.6	98.75

ppm	1	2	3	4	5	6	7	8	9	10	11	12	13	14
La	5.03	7.46	8.34	na	3.5	47.5	37	na	19	8.4	16.2	13.2	9.9	3.43
Ce	12	12.29	15.41	na	na	89.85	72	na	na	na	36	na	na	9.67
Pr	1.75	2.18	2.62	na	na	9.9	na	na	na	na	na	na	na	na
Nd	10.41	12.68	15.79	na	7.99	41.05	na	na	23	na	20.1	na	na	7.45
Sm	3.43	3.68	4.8	na	2.66	8.59	6.7	na	5.9	5.1	4.5	5	4.8	2.62
Eu	1.28	1.31	1.64	na	1.07	2.54	na	na	2.35	1.7	1.5	1.45	1.63	0.99
Gd	5.05	5.39	7.02	na	na	8.21	na	na	na	na	na	na	na	3.64
Tb	0.94	1	1.31	na	0.64	1.105	na	na	1.14	na	0.8	na	na	na
Dy	5.73	5.78	7.7	na	na	6.15	na	na	na	na	na	na	na	4.02
Ho	1.26	1.23	1.67	na	na	1.2	na	na	na	na	na	na	na	na
Er	3.73	3.71	5.1	na	na	3.18	na	na	na	na	na	na	na	2.55
Tm	0.61	0.61	0.84	na	na	0.445	na	na	na	na	na	na	na	na
Yb	3.55	3.45	4.74	na	na	2.585	na	na	na	na	na	3.6	3.3	2.53
Lu	0.55	0.53	0.72	na	na	0.375	na	na	na	na	0.51	na	na	na
Y	38.06	37.54	50.21	35.48	29.29	36.5	na	60.13	47.3	26	36	31	50	na
Sc	45.28	43.29	47.46	37.27	na	13.5	na	na	na	46	47	42	42	na
Nb	1.57	2.09	2.56	4.59	2.97	35	na	na	34	na	na	na	na	na
Th	0.2	0.45	0.51	1.13	0.26	4.43	na	na	2.22	na	na	na	na	na
Sr	135.73	139.56	139.76	141.11	120.8	644.5	959	na	245.2	265	390	235	115	140
Rb	10.87	15.3	16.39	6.09	2.2	62.5	na	na	8.27	na	na	na	na	2.67
Ba	40.75	45.12	68.1	52.31	na	1145.5	2530	59.55	na	45	450	140	56	na
Co	47.14	46.71	53.09	92.33	na	na	na	72.93	na	65	50	53	39	na

Cu	124.28	160.72	149.91	73.89	na	na	na	114.32	na	na	65	na	na	na
V	299.29	280.44	323.2	264.98	249	159.5	na	370.73	365	525	245	445	342	na
Ga	15.6	14.93	16.83	16.56	na	na	na	na	na	na	na	na	na	na
Zn	99.44	122.53	142.69	81.58	na	na	na	188.85	na	na	na	na	na	na
Zr	85.62	101.64	133.55	107.17	84.6	279.5	331	191.35	197	120	160	159	156	na
U	0.59	0.57	0.68	0.52	na	1.145	na	na	na	na	na	na	na	na
Pb	4.56	4.44	5.27	3.57	na	9.7	na	na	na	na	na	na	na	na
Ni	na	na	na	na	129	47.5	27	116.4	24	60	100	44	194	na
Cr	na	na	na	na	328	na	na	118.4	6	38	250	45	469	na
Li	na	na	na	na	na	na	na	na	na	na	13	na	na	na
B	na	na	na	na	na	na	na	na	na	na	4	na	na	na