Alkalic marine tephra layers at ODP Site 1241-

Plinian eruptions from an oceanic volcano in a pre-shield stage?

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Abstract

We report a series of fourteen marine tephra layers that are the products of large explosive eruptions of Plinian intensities and magnitudes (VEI >4) from Cocos Island, Costa Rica. Cocos Island is a volcanic island in the eastern Central Pacific Ocean ~500 km offshore Costa Rica, and is situated on the northwestern flank of the aseismic Cocos Ridge. Geochemical fingerprinting of Pleistocene (~2.4-1.4 Ma) marine tephra layers from Ocean Drilling Project (ODP) Leg 202 Site 1241 using major and trace element compositions of volcanic glass shards demonstrates unequivocally their origin from Cocos Island rather than the Galápagos Archipelago or the Central American Volcanic Arc (CAVA). Cocos Island and the adjacent seamounts of the Cocos Island Province have alkalic compositions and formed on young (≤ 3 Ma) oceanic crust from an extinct spreading ridge bounded by a transform fault against the older and thicker crust of the aseismic Cocos Ridge. Cocos Island has six times the average volume of the adjacent seamounts although all appear to have formed during the 3-1.4 Ma

time period. Cocos Island lies closest to the transform fault and we explain its excessive growth by melts rising from garnet-bearing mantle being deflected from the thick Cocos Ridge lithosphere toward the thinner lithosphere on the other side of the transform, thus enlarging the melt catchment area for Cocos Island compared to the seamounts farther away from the transform. This special setting favored growth above sealevel and subaerial explosive eruption even though the absence of appropriate compositions suggests that the entirely alkalic Cocos Island (and seamounts) never evolved through the productive tholeiitic shield stage typical of other Pacific Ocean islands, possibly because melt production rates remained too small. Conditions of magma generation and ascent resembled Hawaiian pre-shield volcanoes but persisted for much longer (<1 m.y.) and formed evolved, trachytic magmas. Therefore Cocos Island may be a unique example for a volcanic ocean island that did not pass through the typical growth stages.

Key words

Ocean island evolution, subaerial explosive volcanism, Pleistocene marine tephrostratigraphy, Plinian fallout tephra, tectonically controlled melt ascent

1 Introduction

Growing oceanic intraplate volcanoes eventually reach a mature stage when they breach the sea surface and generate subaerial explosive volcanism (e.g., Canary Islands, Schmincke and Sumita, 1998a; Cape Verde Islands, Holm et al., 2005). Pacific ocean islands typically reach that stage by building most of their volume through a very productive tholeiitic shield phase followed by post-shield alkalic volcanism (Fig. 1; e.g., Hawaiian Islands, Clague and Sherrod, 2006; Galápagos Islands, White et al., 1993). However, at Hawaiian Islands the tholeiitic shield phase was preceded by a pre-shield phase of alkalic volcanism (Fig. 1) considered to

represent incipient melting and melt ascent conditions at the hotspot margin prior to fully established melt production of the shield phase above hotspot center (Calvert and Lanphere, 2006; Lipman et al., 2006). Loihi seamount is an example of an oceanic volcano still in its pre-shield phase. Stratigraphically sampled sections show an increase in tholeiitic lavas at the expense of alkalic lavas towards younger units, and Loihi already has developed structures such as a caldera, rift zone and flank collapse typical of shield volcanoes (Clague and Sherrod, 2006). Representing the first phase of growth on the ocean floor, the alkalic pre-shield volcanoes are submarine structures. However, here we propose the hypothesis that alkalic Cocos Island grew above sealevel and produced highly explosive subaerial eruptions without evolving through the pre-shield, shield, and post-shield stages typical of other Pacific Ocean islands.

The starting point to arrive at that hypothesis is our identification of tephra layers at ODP Leg 202 Site 1241 on the Cocos Ridge offshore Costa Rica (Fig. 2; Mix et al., 2003). We show that these are trachytic products of subaerial, Plinian eruptions on Cocos Island 80 km away from the drill site. We discuss how this marine tephra record modifies published ideas about the subaerial evolution of Cocos Island (Castillo et al., 1988). Our data, and published data on the Cocos Island Seamount Province, show that these volcanoes all have alkalic compositions and formed within the 3-1.4 Ma time window, much younger than the surrounding Cocos Ridge crust. We discuss the special conditions that may have facilitated growth of Cocos Island to a subaerial volcano without passing through the typical shield stage.

2 Geological background

Due to its proximity to the Galápagos (Cocos-Nazca) spreading center (GSC), the Galápagos hotspot has formed two sets of tracks (Fig. 2a): 1) the NE-SW trending aseismic Cocos Ridge

76 and associated seamount provinces (i.e., the Costa Rica Seamount Province, Cocos Island
77 Seamount Province, and Southwest Seamount Province) located on the Cocos Plate extending
78 from the GSC just north of the Galápagos Archipelago to the Central American Trench
79 offshore Costa Rica, and 2) the Carnegie, Malpelo and Coiba Ridges located on the Nazca
80 Plate extending from the Galápagos Archipelago to Central and South American Trenches
81 offshore Ecuador to Panama (Hey, 1977; Lonsdale and Klitgord, 1978; Wilson and Hey,
82 1995) (Fig. 2a). The Cocos Island Seamount Province comprises Cocos Island (~35 km² area
83 above the sea level, roughly rectangular shape, highest peak at 575m asl [Castillo et al., 1988])
84 and an adjacent group of seven large and several smaller seamounts scattered up to 200 km
85 west and ~80 km southwest of the island (Werner et al., 2003).

86 The volcanism on Cocos Island is unusual in that it has a more alkalic composition and is
87 significantly younger (<3 Ma) than the surrounding Cocos Ridge basement (~10 Ma; Castillo
88 et al., 1988; Werner et al., 2003; O'Connor et al., 2007). Cocos Island lies near the eastern end
89 of an approximately east-west trending trough that has been interpreted as an extinct spreading
90 center that was active from ~3.5 to ~1.8 m.y. ago (Meschede et al., 1998). This spreading
91 ridge segment is bounded in the east by a transform fault that juxtaposes the spreading ridge
92 with the Cocos ridge (Figure 4 in O'Connor et al., 2007).

93 The geology of Cocos Island (Fig. 2c) is subdivided into three lithostratigraphic units (Castillo
94 et al., 1988): 1) The Lower Volcanic Series represents the subaerial phase of basaltic lava
95 effusion. 2) This is unconformably overlain by a thick pyroclastic succession mainly of
96 massive breccias, products of phreatomagmatic eruptions and some fallout deposits, and
97 locally capped by in subaerially extruded trachytic domes that may represent the late-stage
98 activity of a previously collapsed caldera (Castillo et al., 1988). 3) The Upper Volcanic Series
99 again comprises subaerial mafic lavas and remnant cinder cones predominantly produced by
100 fissure eruptions. Radiometrically dated lava samples constrain the age of the subaerial part of
101 Cocos Island to 2.3-1.5 Ma (O'Connor et al., 2007).

102

103 3 Methods

104 Marine and terrestrial samples were analyzed for major and trace element glass compositions
105 with electron microprobe (EMP) and laser ablation-inductively coupled plasma-mass
106 spectrometry (LA-ICP-MS). Twenty-four marine tephra samples were selected from core
107 sections archived at the IODP (International Ocean Discovery Program) core repository in
108 College Station, Texas, were wet-sieved and prepared for geochemical analyses following the
109 methods described in Schindlbeck et al. (2015).

110 Two terrestrial Cocos Island tephra samples from RV Sonne cruise SO144-3 (Werner et al.,
111 2000) were prepared for glass analyses of pumice clasts by cleaning and crushing the samples
112 and handpicking fresh pumice particles.

113 Major element glass compositions (26 samples; in total 320 single glass shard analyses) were
114 analyzed with a JEOL JXA 8200 wavelength-dispersive EMP at the GEOMAR Helmholtz
115 Center for Ocean Research Kiel. Analytical procedures follow Kutterolf et al. (2011) and
116 include measuring programs calibrated with international natural and synthetic standards. The
117 accelerating voltage was constantly 15 kV and the beam current was 6 nA for glasses at a
118 defocused beam size of 5 μm . The accuracy has been monitored by standard measurements on
119 Lipari obsidian (rhyolite; e.g. Hunt and Hill, 2001) and Smithsonian basaltic standard VGA
120 and deviations are $\sim 0.5\%$ for major and $< 10\%$ for minor elements. Analyses are normalized
121 to 100% anhydrous to eliminate the effects of variable post-depositional hydration.

122 Trace and selected main element compositions of 49 glass shards were analyzed by LA-ICP-
123 MS at the GEOMAR Helmholtz Center for Ocean Research Kiel (April 2011) and the
124 Academia Sinica in Taiwan (September 2015). At GEOMAR we used a 193 nm Excimer laser
125 ablation system (Coherent, GeoLasPro) equipped with the manufacturer's standard ablation

cell and coupled to a double-focusing, high-resolution magnetic sector mass spectrometer (Nu Instruments, AttoM) in low resolution mode (300 Res, 10% valley definition). Ablation was done under He carrier gas, additionally Argon carrier gas was mixed prior to the plasma torch. Spot analyses were done by 100 s ablation at a laser repetition rate of 3 Hz using a spot diameter of 16 μm and a fluence of 8 J/cm^2 . 50 s gas background were collected prior to each ablation. Gas flows, torch position and ion-optics-focusing were optimized in order to provide a maximum in ion transparency, low oxide production rates ($\text{ThO}/\text{Th} \leq 0.3\%$) and fast sample wash-out. The NISTSRM610 glass (Wise and Watters, 2012) was used for mass calibration. Data evaluation has been performed applying the linear regression slope method (Fietzke et al., 2008). Si was used for internal standardization utilizing data from EMP analyses(see supporting information Table S4 in Kutterolf et al. (2014) acquired during the same analytical runs).

The LA-ICPMS at the Institute of Earth Sciences, Academia Sinica in Taipei, Taiwan, is equipped with a Photon Machines Analyte G2 laser ablation system (Photon Machines, Inc., Redmond, USA) attached to a quadrupole ICP-MS (Agilent 7900). The Photon Machines Analyte G2 laser ablation system utilizes a 193 nm ArF Excimer laser with 5 ns pulse duration. The instrument features a HelEx two-volume ablation cell providing aerosol washout times of <10 s for a signal drop of 3 orders of magnitude. The carrier gas transporting the ablated sample from the ablation cell was mixed with the Ar gas flow prior to plasma torch. Spot analyses were done using a spot diameter of 16 to 30 μm at energy densities of 7 J/cm^2 and 5 Hz repetition rate. Following 45 seconds of blank acquisition, typical ablation times were around 75 seconds. Data reduction was made using Version 4.0 of “real-time on-line” GLITTER© software (van Achterberg et al., 2001) right after each ablation analysis. Calibration was performed using the NISTSRM612 glass as external reference material in conjunction with internal standardization on ^{43}Ca . Internal standardization was used to correct for not only matrix effects and signal drift in the ICP-MS

and differences in the ablation efficiency between the sample and the reference material (Günther et al., 1999). Ca concentrations determined by EMP analyses were used to normalize trace element concentrations. A basaltic glass standard (USGS BCR-2G) was used as reference material. Precision for most trace elements is better than 5 % (1σ) as deduced from comparing the analytical results with literature data for BCR-2G (more details in supplementary material). Replicate measurements at both institutes produced consistent results.

4 Samples

4.1 Marine tephra inventory

A succession of marine ash beds was drilled during Ocean Drilling Project (ODP) Leg 202 at Site 1241 on the Cocos Ridge, ~80 km northeast of Cocos Island (Fig. 2; 5°50.570'N 86°26.681'W). Three holes (A, B and C) were drilled that recovered Late Miocene through Holocene marine sediments containing numerous tephra layers (Fig. 2d; Mix et al., 2003). Many tephra layers originate at the Central American volcanic arc as we will demonstrate on the basis of geochemical compositions elsewhere (Schindlbeck et al. under revision). Herein we focus on fourteen tephra layers that have unusual alkalic compositions (green lines in Fig. 2d). These marine tephra layers in the interval 27.53-52.66 mbsf (meters below sea floor) are 2 to 15 cm thick and range in grain size from medium ash to lapilli (up to 1 cm). The highly vesicular glass shards with round and elongated bubbles are transparent with minor amounts of brown glass shards (Fig. 3). The mineral assemblage comprises feldspar > biotite > pyroxenes > amphiboles, in order of abundance. The structural and textural features of all but one of the ash layers fulfill the criteria for primary fallout deposits from a subaerial eruption (e.g., Fisher and Schmincke, 1984). The ash and pumice lapilli layers are generally well-sorted, massive, and their bottom contacts are sharp with a gradual admixing of sediment at

their tops (Fig.3a), and some layers are normally graded (e.g., Interval 202-1241A 6H-1, 71-79 cm). Additionally, the main bodies of most ash layers show no parallel or cross lamination, nor reworking of volcanic and/or sedimentary material, which would be indicative for flow processes. The one exception is a poorly sorted lapilli-ash layer (Interval 202-1241A-6H-1, 53-55 cm, Fig. 3b) that contains a heterogeneous mixture of pumice lapilli and lithic lava fragments >0.5 cm in diameter, and moderately vesicular blocky glass shards. This bed was likely deposited from a submarine density current rather than from pyroclastic fallout.

Based on petrographic criteria the tephra layers were correlated between the three holes at Site 1241 by the Leg 202 scientific party (Mix et al., 2003). We tested these correlations geochemically by using major and trace element glass compositions (see supplementary dataset).

4.2 Terrestrial samples

During R/V Sonne cruise SO144-3 in 1999 a field party (Tim Worthington, Karen Harpp and Guillermo Alvarado) mapped and sampled shore sections on Cocos Island (Fig. 2c; a log is given in Werner et al., 2000). They described and sampled a several meters thick sequence at West Wafer Bay (Fig. 2c) that consists of volcanoclastic debris flow deposits at top and bottom (>4 m thick) that sandwich four pyroclastic units (0.7-1.5 m thick) in-between from which two samples were collected. Sample Cocos 3 is a light brown, weakly consolidated massive lapilli-tuff with white pumice lapilli taken from a 0.7 m thick massive fine-grained tuff. Sample Cocos 40 was taken from a 1.5 m thick pyroclastic unit that is described as a possible fallout deposit or massive fine-grained flow deposit with white pumice lapilli in an ash matrix, containing accretionary lapilli. This sequence is probably part of the middle pyroclastic series of Castillo et al. (1988) while most other samples taken during this field excursion are thought

to be part of the Lower Volcanic Series. Bulk rock major and trace element analyses, and radiometric ages, of lava samples are presented in Werner et al. (2003), Harpp et al. (2005) and O'Connor et al. (2007). For this study we analyzed the glass compositions of the two pyroclastic samples, since the glass composition is the most useful tool for the correlation with the investigated marine tephra deposits from ODP Site 1241.

5 Results

5.1 Ages

The tephra layers investigated here occur between 27.53-52.66 mbsf in lithologic unit I (0.05-393.67 mbsf). A robust age model for Site 1241 is published in the Initial Report of Leg 202 (Mix et al., 2003). Especially for the upper ~100 m, which includes the investigated tephra layers, there is a good agreement between microfossil datums (calcareous nannofossils planktonic foraminifers and diatoms), the Brunhes-Matuyama boundary (0.78 Ma) as well as the upper Jaramillo transition (1.07 Ma), that leads to a reliable age model for this section. The ages of individual marine tephra layers were calculated with sedimentation rates given in the Initial Report of Leg 202; the methods and errors are described in Kutterolf et al. (2008a, 2013). The calculated ages of the silicic tephra layers range from ~2.4 to 1.4 Ma (supplementary material).

5.2 Geochemistry and Tephra Provenance

Having inferred above that all but one of the marine tephra layers originated from explosive eruptions on land, potential source regions surrounding the position of Site 1241 are the Galápagos Islands, Cocos Island and the CAVA (Fig. 2a). Judging from the ~8.5 cm/yr northeastward motion of the Cocos Plate (DeMets, 2001), the position of Site 1241 at 1.4 to 2.4 Ma was ~720-630 km from the Galápagos islands, and ~630-720 km from the CAVA,

while its position 80 km southeast of Cocos Island remained unchanged. Advection of pyroclastic material by ocean currents was insignificant because the slow (<2 cm/s at 1500 m/bsl; Cravatte et al., 2012) deep ocean currents near Cocos Island would laterally displace particles of the mean grain size (~ 250 μm) by only <2 km during their one day settling through 2000 m water depth (cf. Carey, 1997). The relatively coarse grain sizes of the tephra layers, with an average grain size about 250 μm but lapilli sizes up to 10 mm, favor Cocos Island as the much more proximal source and do not seem compatible with the other source regions at >600 km distances. This conclusion is corroborated by geochemical compositions.

The fourteen ash and lapilli layers have trachy-andesitic to rhyotrachytic compositions (57–72 wt% SiO_2 ; Fig. 4) and form an alkalic trend that is shared by the glass compositions of the pyroclastic samples from Cocos Island. The tephra glasses have typical ocean-island trace-element characteristics (e.g., low $\text{Ba/La} < 15$, $\text{La/Nb} < 1$, high $\text{Nb/Rb} > 0.6$) and do not show typical subduction-zone features such as LILE-enrichment or a Nb-Ta trough (Fig. 5). Although there are some backarc lavas at the CAVA that lack the Nb-Ta trough and are alkalic in composition (e.g., Gazel et al., 2011), there is no evidence that these were associated with widespread, large-volume tephra and thus we exclude the CAVA as potential source region. Volcanic products from the Galápagos Islands and surrounding platform follow a rather tholeiitic trend (e.g., Geist et al., 1995) compared to the distinctly alkalic rocks from the Costa Rica and Cocos Island Seamount Provinces (Werner et al., 2003; Harpp et al., 2005). Moreover, compared to the Galápagos Islands, Cocos Island rocks and the tephra glasses have elevated Ce/Yb ratios at relatively low Zr/Nb (Fig. 6a) as well as elevated La/Sm (Fig. 6b) indicating a general LREE enrichment (Fig. 5). Major and trace element compositions thus support an origin of the investigated tephra layers at Cocos Island.

6 Discussion

6.1 Eruption volumes and frequencies

The grain sizes (up to 10 mm) and thicknesses (up to 15 cm) of the marine tephra layers at 80 km distance from their source on Cocos Island imply their emplacement by Plinian eruptions, because eruption columns of >30 km height are needed to disperse medium ash to fine lapilli fallout to > 30 km crosswind or > 40 km downwind (see Figure 9 in Burden et al., 2011). The generation of a 30 km high, Plinian eruption column, in turn, requires high mass discharge rates of order $\sim 10^7$ to 10^8 kg/s (e.g., Figure 1 in Mastin et al., 2009). Although the occurrence of Plinian eruptions has not yet been recognized on Cocos Island it would be compatible with the caldera formation tentatively inferred by Castillo et al. (1988). Alkalic, Plinian eruptions from calderas are important events at other ocean islands like Gran Canaria (Schmincke and Sumita, 1998b) and Tenerife (Ancochea et al., 1990).

Below we estimate minimum erupted tephra volumes based on the single thickness (T) measurements of individual intervals at the core site. We assume (1) it is the most distal point of that thickness along the dispersal axis, (2) isopachs have teardrop-shapes, and (3) lateral thickness decay is exponential (e.g., Legros, 2000). The teardrop shape allows us to estimate the isopach area for a given distance from source (x) but requires an estimate of the apical angle of the tear-drop (θ) (Fig. S1). Narrower apical angles would correlate with higher windspeeds (e.g., Houghton et al., 2000). We performed volume calculations for a range of θ -values but note that an angle of 60° would be typical approximation for Plinian fallouts from the CAVA (Kutterolf et al., 2008b) which shares the low-latitude wind pattern with Cocos Island although the NOAA database indicates lower average wind speeds over Cocos Island compared to the Central American landmass; therefore a 60° angle would yield a conservative volume estimate. For $\theta=60^\circ$ and $x=80$ km, and for tephra thicknesses of 2-15 cm, we obtain minimum tephra volumes of 0.3 km^3 to 2 km^3 (Fig. 7a), which qualifies the Cocos Island eruptions as at least VEI 4 events (volcanic explosivity index; Newhall and Self, 1982).

Transferring the ash layer thickness and distance from source into the logarithmic thickness versus square root of the isopach area diagram (Fig. 7b) shows that the Cocos Island eruptions would be comparable, for example, to the Hudson 1991 (Chile) or Mount St. Helens 1980 (USA) Plinian eruptions. The erupted tephra volumes of 7.6 km^3 for the Hudson eruption (Scasso et al., 1994) and 1.22 km^3 for the May 18th, 1980, Mount St. Helens eruption (Pyle, 1989) are probably of similar magnitude to those inferred to have erupted from Cocos Island eruptions.

6.2 Geological evolution of Cocos Island

The fourteen highly alkalic tephra layers at Site 1241 originated from Cocos Island between 2.4 and 1.4 Ma; thirteen of the layers were emplaced between 2.4 and 1.8 Ma. The cumulative tephra volume for this 600 kyr period amounts to at least 7 km^3 but is probably much more. One or more voluminous magma reservoirs must have existed during this period in order to feed the Plinian eruptions, which may have been related to the formation of a caldera on Cocos Island that Castillo et al. (1988) tentatively concluded from the occurrence of trachyte lava domes above the pyroclastic sequence on land. The marine tephra layers at Site 1241 may be related to the middle pyroclastic sequence on the island (Castillo et al., 1988). However, that sequence consists of very poorly exposed phreatomagmatic and massflow deposits and the relationship of the Plinian deposits to this succession remains unknown.

The 2.4 to 1.4 Ma age range of the marine tephra layers overlaps completely with the K-Ar ages of 2.4 to 1.9 Ma determined by Bellon et al. (1983) on samples probably taken from rocks in the Lower Volcanic Series of Castillo et al. (1988). O'Connor et al. (2007) Ar-Ar dated samples of lavas from both Lower and Upper Volcanic Series as 2.3 to 1.5 Ma old (samples C-11, 15, 26, 33, 35 collected during SO144 cruise). Overlapping age data for Lower and Upper Series lavas and the widespread tephra strongly support the conclusion that the three-part stratigraphic subdivision of Castillo et al. (1988) needs to be refined with

particular emphasis on the fact that highly explosive pyroclastic events and periods of lava effusion occurred broadly contemporaneous rather than successive.

6.3 Magma generation at Cocos Island

The age data together demonstrate that the subaerial evolution of the island is limited to the narrow age range from 2.4 to 1.4 Ma. This age interval is significantly younger than the surrounding Cocos Ridge crust (~10 Ma), which was formed at the Galápagos hotspot, but overlaps with the age range 3 to 1.8 Ma during which the immediately underlying crust formed at a local E-W trending spreading axis identified by marine magnetic and bathymetric data (Fig. 8a; Meschede et al., 1998; Meschede and Barckhausen, 2001). A number of models have been proposed to explain such young volcanism along the Cocos Ridge.

(1) Castillo et al. (1988) implied a second hotspot 600 km NE of Galápagos because they noted an age progression of volcanic activity from northeast to southwest of Cocos Island. Meschede et al. (1998) and Meschede and Barckhausen (2001) proposed that this second hotspot, presently located 200 km southwest of Cocos Island, also formed the Malpelo Ridge. However, both the age data from Malpelo Island, providing a minimum age for the underlying Malpelo Ridge (15.8-17.3 Ma; Hoernle et al., 2002), and the geochemical composition of the Ridge (Harpp et al., 2005; Werner et al., 2003), indicate an origin of the Malpelo Ridge from the Galápagos hotspot.

(2) Using geochemical comparison with the Mathematican Ridge, Harpp et al. (2005) interpret the Cocos Seamount Provinces as the result of post-abandonment volcanism after the failure of the local spreading ridge between ~3.5-2 Ma.

(3) Werner et al. (2003) propose a viscous layer accreted to the overlying lithosphere that inhibits further ascent of asthenospheric material. However, decompressional melts are

325 enabled to rise,for example,due to extension or spreading of the overlying lithosphere such as
326 near Cocos Island.

327 (4) O'Connor et al. (2007) argued that the Galápagos volcanism generally has not progressed
328 in narrow, time-progressive lines of seamounts and ridges, as predicted by the conventional
329 fixed hotspot scenario. Instead, based on seismic tomographic data showing a low-velocity
330 anomaly beneath the Cocos Ridge extending to depths of 200-300 km, they propose a very
331 broad Galápagos melting anomaly that extends several hundred kilometers underneath the
332 Cocos Ridge and from which melts occasionally rise to produce an overall pattern of none-
333 time progressive volcanism.

334 (5) Herbrich et al. (2015) explain the existence of seamounts offshore Nicaraguasignificantly
335 younger than underlying Cocos Plate ocean crust by northward transport of Galápagos plume
336 material along the base of the Cocos Plate up to 900 km away from the hotspot and 250–500
337 km north of the Galápagos hotspot track. They propose that melting occurs due to
338 decompression as the mantle upwells to shallower depth as it flows northwards, either due to
339 changes in lithospheric thickness or as a result of upwelling at the edge of a viscous plug of
340 accumulated plume material at the base of the lithosphere.

341 Model (1) that considers a second hotspot, has been rejected by Werner et al. (2003) and
342 O'Connor et al. (2007) because it cannot explain young volcanism along Cocos Ridge outside
343 the Cocos Island province. Additionally there is no evidence for lateral age progression.
344 Model (2), assuming post-abandonment volcanism, is no longer plausible for the Cocos Island
345 Seamount Province because the new age data here and in O'Connor et al. (2007) show that the
346 seamounts did not form clearly after the spreading ridge had been abandoned. Models (3-5)
347 provide somewhat different explanations why decompression melting may occur far away
348 from the center of the hotspot. They all agree in that melt formation in garnet-bearing mantle
349 at >70 km depths (e.g., Werner et al., 2003), as required by the high LREE/HREE ratios of the

alkalic magmas (Fig. 5), favors an origin of Cocos Island Seamount Province volcanism from plume-related mantle movements.

6.4 Cocos Island in the framework of Ocean Island evolution

Alkalic eruption phases of plume-related ocean island volcanoes, particularly when associated with explosive subaerial activity, are typically interpreted as post-shield volcanism, following the basaltic shield-forming phase that produced >80% of the edifice volume and has tholeiitic compositions at Pacific ocean islands (e.g., Hawaiian chain, e.g. Clague and Sherrod, 2014; Galápagos, White et al., 1993). However, the Cocos Island seamounts, with edifice volumes of just a few hundred km³ (Fig. 8b), are roughly two orders of magnitude smaller than the Galápagos or Hawaiian islands (e.g., Robinson and Eakins, 2006), and none of the samples dredged in the Cocos Island Seamount Province, or collected on the island itself, has a tholeiitic composition (Harpp et al., 2005; Werner et al., 2003). We therefore propose that the entirely alkalic Cocos Island province volcanoes did not form a tholeiitic shield. This eliminates an interpretation as post-shield volcanism because that develops continuously from shield volcanism (e.g., Clague and Sherrod, 2014). This leaves the pre-shield or rejuvenated options for Cocos Island. Rejuvenated volcanism occurs after a considerable time break following post-shield activity (Fig. 1), at distances up to 550 km from the hot spot at Hawaii, and may persist for ~1 m.y. while producing <1% of the total volcano volume (Clague and Sherrod, 2014). Cocos Island is almost 800 km from the Galápagos hotspot but the ~1 m.y. lifespan would fit. Factors that do not fit the rejuvenated model are (1) that this volcanism formed the entire volume of the volcanic structure (Fig. 9), and (2) that alkali-basaltic to rhyotrachytic compositions differ greatly from the typically basanitic to foiditic, mafic compositions of Hawaiian rejuvenated volcanism. Moreover, Cocos Island formed on oceanic crust not much older than the island itself (Fig. 8a) so that it is questionable if the ~10 Ma old

374 Cocos Ridge crust in the wider surroundings could count as the shield to post-shield precursor
375 stages.

376 The alternative then is to interpret Cocos Island (and the seamounts) as a pre-shield stage
377 volcano analogous to pre-shield growth at Loihi and other Hawaiian volcanoes (Calvert and
378 Lanphere, 2006; Clague and Sherrod, 2014). Pre-shield lavas are hypothesized to originate by
379 low-degree melting at the plume margin producing initially low growth rates that increase to
380 high stable values when the shield stage with tholeiitic lavas is reached with proximity to the
381 hotspot center. The Hawaiian pre-shield volcanoes are submarine edifices (Fig. 1) of
382 moderate volume (e.g., Loihi 1700 km³, Robinson and Eakins, 2006) just like the seamounts
383 west of Cocos Island. Hawaiian pre-shield rocks have basaltic to hawaiitic compositions but
384 trachytic rocks, such as on Cocos Island, have not been described. Loihi has formed its 1700
385 km³ volume during the probably barely 100 k.y. since its birth, compared to the ~1 m.y.
386 required to form the 1900 km³ of Cocos Island. More important, however, Loihi is expected
387 to evolve into the shield stage as did the pre-shields of other Hawaiian islands whereas
388 volcanism at Cocos Island seamount province terminated 1.4 m.y. ago. Hence, while Cocos
389 Island has some features in common with pre-shield volcano growth, it is certainly not a pre-
390 shield volcano in the sense that its growth would herald approach to a hotspot center; note
391 that model (1) has been rejected above and that models (3-5) presently best explain the
392 melting anomaly in this region. In summary, Cocos Island differs from the classical Hawaiian
393 model for ocean island growth in being an island that has not evolved through the different
394 growth stages (pre-shield, shield building, post-shield, rejuvenated stages) (Fig. 1, 9a).
395 Deviations from the Hawaiian model have been observed elsewhere. For example, Guillou et
396 al. (2014) report that the shield to post-shield transition is less clear in the Marquesa Islands as
397 alkalic compositions occur through both stages. The Cocos Island evolution however, differs
398 even more strongly from the Hawaiian pattern, because it formed from a source producing
399 exclusively alkalic magmas through its lifetime (Fig. 9b).

6.5 Growth of Cocos Island

How did Cocos Island become a mature island producing subaerial explosive eruptions of evolved magmas while the adjacent seamounts did not? Most of the oceanic crust underlying the Cocos Island province volcanic edifices (Fig. 8b) is younger than 3 m.y. (Fig. 8a), and there is no evidence for volcanism younger than 1.4 Ma. Hence, all these edifices evolved within ~1.5 m.y., but Cocos Island stands out because of its volume being 2 to 20 times larger than that of the adjacent seamounts (Fig. 8b). Compared to the seamounts, Cocos Island lies in a special situation at the eastern edge of the extinct spreading ridge where it terminates against a transform fault (Fig. 8a). We propose that the transform fault has juxtaposed the thickened lithosphere of the Cocos Ridge against the thinner lithosphere of the spreading axis (Fig. 8a, c); lithosphere thicknesses are not well known but the Cocos Ridge crustal thickness of 21 km (Walther, 2003) is greater than crustal thickness at the spreading ridge and lithosphere thicknesses can be expected to vary accordingly. While all volcanoes of the Cocos Island Province are fed by melts rising from below the spinel-garnet phase boundary, we suggest that Cocos Island has a much larger melt catchment area due to melts rising sideways from beneath the thick Cocos Ridge lithosphere (Fig. 8c). We imply that the larger melt catchment region caused the huge volume of Cocos Island volcano (~1900 km³) as compared to the smaller seamounts (~100-800 km³, Fig. 8b) farther away from the transform and the thick Cocos Ridge lithosphere (Fig. 8c). Thus, while the seamounts remained small and submarine, Cocos Island over the same time became big enough to host large magma reservoirs differentiating to trachytic compositions, and to breach the sea surface to permit subaerial explosive eruptions. Special conditions may thus lead to subaerial explosive volcanism already during a situation comparable to a pre-shield stage. If the cumulative volume of the edifices in Figure 8b (4883 km³) was produced during the about 1.5 m.y. period suggested by available age data, the average growth rate of about 0.003 km³/yr is almost two

orders of magnitude lower than the rate for Hawaiian islands during the last 6 Ma (0.095 km³/yr, Robinson and Eakins, 2006). Melt productivity at the Cocos Island province may just have been too low to ever generate productive shield volcanism, possibly due to processes of models (3-5) discussed above in contrast to proper passage across a hotspot as at Hawaii. Magma generation at the Cocos Island province terminated for unknown reasons about 1.4 Ma ago.

7 Summary

Geochemical fingerprinting of a series of Pleistocene marine tephra deposits from ODP Leg 202 Site 1241 in the eastern Pacific indicates that these layers originated from Cocos Island. Grain-sizes from medium ash to fine lapilli and layer thicknesses of 2-15 cm at ~80 km distance from the source (Cocos Island), as well as the calculated minimum tephra volumes of 0.3 to 2 km³ suggest that these tephras formed by eruptions of Plinian intensities and magnitudes (VEI= 4 to 5). This is the first record demonstrating that such highly explosive eruptions of trachytic magmas occurred regularly on Cocos Island between 2.4 and 1.4 Ma. This age range completely overlaps with ages from lava sequences and implies that dominantly mafic effusive and felsic explosive phases occurred roughly contemporaneously rather than sequentially during the subaerial evolution of Cocos Island.

The volcanic edifices of the Cocos Island province grew on young (mostly ≤ 3 Ma) oceanic crust and have alkalic compositions inherited from low-degree partial melting of garnet-bearing mantle (>70 km depth). No evidence for tholeiitic shield volcanism typical of Pacific Ocean islands has yet been detected at the Cocos Island province. We interpret that Cocos Island and adjacent seamounts did not evolve through the growth stages typical of Pacific volcanic ocean islands but formed entirely under growth conditions similar to Hawaiian pre-

shield volcanism but persisting over much longer time ($>10^6$ vs. $<10^5$ years) and terminating 1.4 Ma ago. We do, however, emphasize that denser sampling of the submarine edifices will be needed to verify the absence of tholeiitic rocks. Cocos Island itself differs from the adjacent seamounts in being more voluminous by about an order of magnitude, which is why it breached the sea surface and produced subaerial explosive eruptions. We relate this large size of Cocos Island to its proximity to a transform fault where thickened lithosphere of the aseismic Cocos Ridge is juxtaposed against thin lithosphere of the now extinct spreading axis. Rising melts channeled up along the edge of the transform significantly enlarged the melt catchment area for Cocos Island compared to seamounts lying farther away from the transform and the thick Cocos Ridge.

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Figure captions

Figure 1 Hawaiian island growth stages: preshield, shield, and postshield stages followed by a hiatus with erosion and subsequent rejuvenated stage. This evolution is commonly considered typical of Pacific Ocean islands. Modified after Clague and Sherrod (2014).

Figure 2a) Overview map of the eastern Pacific showing the location of Cocos Island, seamount provinces and Site 1241 (compare enlarged view of white box in b) on the Cocos Ridge that extends from the Galápagos Islands to Central America, where it subducts beneath the Caribbean Plate (www.geomapapp.org; GMRT-Global Multi-Resolution Topography; Ryan et al., 2009). CAVA= Central American Volcanic Arc. Location of Site 1241=pink dot. Red box marks the area shown in figure 7b. c) Geologic map of Cocos Island showing the distribution of the Lower, Middle and Upper Series of Castillo et al. (1988). d) Simplified core profile (mbsf= meters below seafloor) of the upper 100 m at Site 1241. Green lines mark tephra layers discussed here, black lines are tephra layers of different provenance to be discussed elsewhere. Blue bars mark positions of the Brunhes/Matuyama boundary (B/M) and the Jaramillo transition (JT). Green labels indicate selected tephra layer ages calculated from pelagic sedimentation rates.

Figure 3a) Core scan image with dark gray ash layer (Interval 202-1241B-5H-6, 96-102 cm) and smear slide micro-photographs with transparent highly vesicular glass shards.

b) Core scan image with gray lapilli layer (Interval 202-1241A-6H-1, 53-56 cm) and smear slide micro-photographs with transparent and light brown vesicular and blocky glass shards.

Figure 4 Total alkalis versus silica diagram after Le Maitre (1989). The fourteen investigated marine tephra layers from Site 1241 (green diamonds) and terrestrial samples from Cocos Island (black dots) follow an alkalic trend from trachy-andesite to rhyotrachyte. Each symbol is the mean composition of 4 to 16 single glass shard microprobe analyses per sample. The standard deviations around the means are of the width of their symbols. Reference fields

comprise mainly bulk rock data from Harpp et al. (2005) and Werner et al. (2003) as well as additional references for the Galápagos field that are given in Schindlbeck et al. (2015). AR=Aseismic Ridge.

Figure 5 Chondrite-normalized trace element spider diagram comparing the marine tephra glasses from Site 1241 (green) with bulk rock data from Cocos Island samples that are grouped regarding their silica contents. Steep Ba, Sr and Eu depletions in the more evolved compositions formed by feldspar fractionation. The steep decay from light to heavy rare earth elements indicates low-degree mantle melting with residual garnet.

Figure 6a) Ce/Yb versus Zr/Nb and b) La/Sm versus Zr diagrams serve to distinguish Galápagos and Cocos Island data and to demonstrate the LREE enriched compositions of the Site 1241 alkalic tephra and the Cocos Island samples. Each green star symbol is the mean composition of 1 to 8 glass-shard trace element analyses. Bulk rock data of the aseismic ridges and seamount provinces from Harpp et al. (2005) and Werner et al. (2003). Reference field for the Galápagos islands after references in Schindlbeck et al. (2015). Miocene Galápagos glass-shard data from Site U1381 tephra layers (from Schindlbeck et al., 2015) plot into the Galápagos field and differ significantly in composition from the Site 1241 tephra.

Figure 7a) Plot of estimated minimum tephra volume versus age for the thirteen fallout tephra layers. VEI=Volcanic Explosivity Index after Newhall and Self (1982). Individual tephra layer thicknesses and calculated volumes are given in the supplement. b) Diagram of log thickness versus square root of area showing data from historic subplinian and plinian eruptions compiled by Constantini et al. (2009). Data for the marine tephra layers of Site 1241 compare well with the distal data of, e.g., the Hudson 1991 and May 18th, 1980, Mount St. Helens fallouts.

Figure 8a) Overview map of the Cocos Island Seamount Province displaying the seafloor magnetic anomalies and seafloor ages in vicinity of Cocos Island; modified after O'Connor et al. (2007) following Barckhausen et al. (2001). Black dots mark locations of Cocos Island and the near-by seamounts. b) Close-up bathymetric map (red box area in Fig. 2a) of Cocos Island and neighboring seamounts that are marked in (a). Bathymetry from www.geomapapp.org GMRT-Global Multi-Resolution Topography, Ryan et al. (2009). The volcano volumes noted next to each edifice are calculated using the red contoured base planes and following the methods described in Völker et al. (2011). c) Schematic model to explain excess growth of Cocos Island by enhanced magma catchment where the transform fault has juxtaposed the thickened lithosphere of the Cocos Ridge (where the Moho lies at up to 21 km depth; Walther, 2003) to the thin lithosphere of the Cocos spreading axis.

Figure 9 a) Schematic illustration of a "complete" Hawaiian island that has evolved through the four growth stages shown in Fig. 1. Modified after Clague and Sherrod (2014).. b) Cocos Island as an entirely alkalic edifice with no evidence for several growth stages.

Figure S1 Geometry of the two-dimensional tear-drop shaped isopach area model. Assuming that the distance (x) between the site and the inferred source volcano is known, the area (A) can be calculated for a given apical angle (θ). The area is the sum of the areas of the small triangle (green) and semi-circle (yellow).

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202-1241B-5H-6, 96-102 cm

202-1241A 6H-1, 53-56 cm