Anomalous uplift and subsidence of the Ontong Java Plateau inferred from CO_2 contents of submarine basaltic glasses

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ABSTRACT

The Ontong Java Plateau in the western Pacific is anomalous compared to other oceanic large igneous provinces in that it appears to have never formed a large subaerial plateau. Paleoeruption depths (at 122 Ma) estimated from dissolved H_2O and CO_2 in submarine basaltic glass pillow rims vary from ~1100 m below sea level (mbsl) on the central part of the plateau to 2200–3000 mbsl on the northeastern edge. Our results suggest maximum initial uplift for the plateau of 2500–3600 m above the surrounding seafloor and 1500 ± 400 m of postemplacement subsidence since 122 Ma. Our estimates of uplift and subsidence for the plateau are significantly less than predictions from thermal models of oceanic lithosphere, and thus our results are inconsistent with formation of the plateau by a high-temperature mantle plume. Two controversial possibilities to explain the anomalous uplift and subsidence are that the plateau (1) formed as a result of a giant bolide impact, or (2) formed from a mantle plume but has a lower crust of dense garnet granulite and/or eclogite; neither of these possibilities is fully consistent with all available geological, geophysical, and geochemical data. The origin of the largest magmatic event on Earth in the past 200 m.y. thus remains an enigma.

Keywords: large igneous province, basalt, volatiles, mantle plume, meteorite impact.

INTRODUCTION

The Ontong Java Plateau is the most voluminous large igneous province on Earth (Coffin and Eldholm, 1994). Unlike other oceanic plateaus believed to be formed by mantle plumes, Ontong Java appears never to have formed a large subaerial plateau. In fact, all basaltic lavas recovered on the plateau by drilling and on land in the Solomon Islands were erupted well below sea level (Mahoney et al., 2001). The only exceptions are subaerial volcaniclastic rocks located far to the southeast of the main plateau at Ocean Drilling Program (ODP) Site 1184 (Thordarson, 2004). The lack of subaerial volcanism on the main plateau is unusual because introduction of a hot mantle plume beneath oceanic lithosphere and thickening of the basaltic crust should produce significant uplift (Olson and Nam, 1986). There is clear evidence that many oceanic large igneous provinces formed subaerial landmasses that later subsided below sea level as they moved away from the plume heat source (Detrick et al., 1977; Coffin, 1992).

In this paper we use eruption depth estimates for basalts on the Ontong Java Plateau to test whether the initial uplift of the plateau and subsequent subsidence due to cooling of the lithosphere are consistent with effects of a high-temperature plume. We estimate eruption depths by using the H_2O and CO_2 contents of basaltic glass pillow rims. Our results demonstrate that initial uplift and postemplacement subsidence of the Ontong Java Plateau are significantly less than are observed for other oceanic large igneous provinces.

GEOLOGICAL SETTING

The Ontong Java Plateau is located in the western equatorial Pacific and, together with obducted portions, covers an area of 2×10^6 km^2 (Tejada et al., 2004). Crustal thickness is 30–35 km in the central part of the plateau, thinning to near-normal oceanic crustal thickness (~10 km) along its flanks (Gladczenko et al., 1997). The ^{40}Ar-^{39}Ar geochronology suggests that most of the plateau formed ca. 122 Ma (see Fitton and Godard, 2004; Tejada et al., 2004, and references therein).

Major and trace element compositions of Ontong Java Plateau basaltic magmas require that they formed by relatively large degrees of melting (~30%; Fitton and Godard, 2004). These large degrees of melting could not have been caused by anomalously high mantle H_2O content because the basalts have low H_2O, similar to depleted mid-oceanic ridge basalt (MORB) (Michael, 1999; Roberge et al., 2004). The hypothesis that the plateau was formed by an upwelling plume therefore requires a mantle potential temperature >1500 °C (>220 °C hotter than the ambient mantle potential temperature of 1280 °C; Fitton and Godard, 2004).

SAMPLE DESCRIPTION AND ANALYTICAL METHODS

We present H_2O and CO_2 data for unaltered glass from pillow basalt rims (ODP Sites 1183, 1185, 1186, and 1187) and from non-vesicular glass shards in volcaniclastic rocks (Site 1184). All glasses were analyzed for H_2O and CO_2 using Fourier transform infrared (FTIR) spectroscopy (Table 1). Our results complement previously published data for glasses recovered from ODP Leg 130 Sites 803 and 807 (Michael, 1999). For consistency, we reanalyzed the glasses from Sites 803 and 807 because we used a different data reduction procedure

TABLE 1. AVERAGE H_2O AND CO_2 CONTENTS SUBMARINE BASALTIC GLASSES FROM ONTONG JAVA PLATEAU

ODP site	H_2O (wt%)	CO_2 (ppm)	Vapor saturation pressure (MPa)	Eruption depth (m)	Subsidence (m)
1183	0.23 (0.04)	46 (7.6)	10.7 (11)	1071 (110)	1500 (133)
1184	0.16 (0.04)	30 (7.9)	54.0 (11)	543 (110)	1400 (323)
1184*				Sea level	1900 (304)
1185	0.19 (0.09)	100 (13.4)	22.1 (18)	2208 (180)	1900 (254)
1186	0.22 (0.4)	99 (12.8)	21.5 (23)	2154 (230)	1200 (229)
1187	0.20 (0.3)	111 (5.2)	24.4 (18)	2448 (180)	1600 (224)
803	0.27 (0.07)	130 (35.0)	28.9 (50)	2899 (500)	900 (501)
807A	0.41 (0.07)	43 (5.9)	11.1 (10)	1111 (100)	2600 (162)
807C-G	0.24 (0.02)	137 (22.0)	30.4 (24)	3041 (240)	600 (249)

Note: ODP—Ocean Driling Program. H_2O and CO_2 were analyzed by Fourier transform infrared spectroscopy using band assignments and absorption coefficients as described in Roberge et al. (2004). Peak height measurements for CO_2 were calculated using a peak fitting program (S. Newman, unpublished). This method yields values for experimental glasses (Dixon et al., 1995) that are comparable to the reference-glass subtraction and hand-drawn background method upon which the CO_2 solubility relations have been established (J. Dixon, 2003, personal commun.) The H_2O and CO_2 values reported for each site are averages ($\pm 2\sigma$ in parentheses) of multiple glass chips (see Roberge et al., 2004, for complete data and analytical uncertainties). Vapor saturation pressures were calculated using VolatileCalc 1.1 (Newman and Lowenstern, 2002). Uncertainties (in parentheses) for saturation pressures, eruption depths, and subsidence values are based on propagation of 2σ uncertainties in the average H_2O and CO_2 values. The subsidence uncertainties also include uncertainties in depths of the samples in the drill hole.

*Based on interpretation that Site 1184 volcaniclastic rocks were erupted in a shallow marine environment but deposited subaerially (Thordarson, 2004).

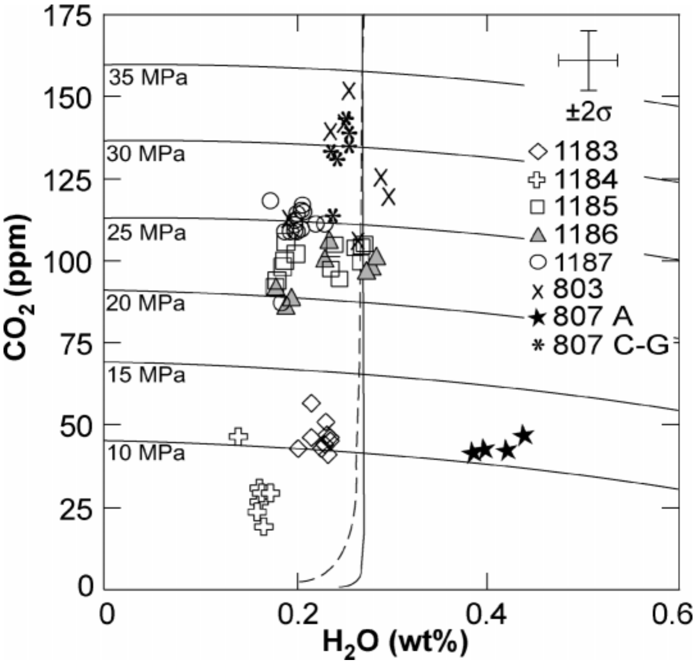

Figure 1: H_2O vs. CO_2 for Ontong Java Plateau basaltic glasses. Symbols correspond to Ocean Drilling Program (ODP) site numbers. Vertical lines represent degassing paths for basaltic melts with initial CO_2 contents of 200 ppm (solid line) and 2000 ppm (dashed line). Also shown are vapor saturation curves for basaltic melts at pressures from 10–35 MPa. All calculations were made using VolatileCalc 1.1 (Newman and Lowenstern 2002).

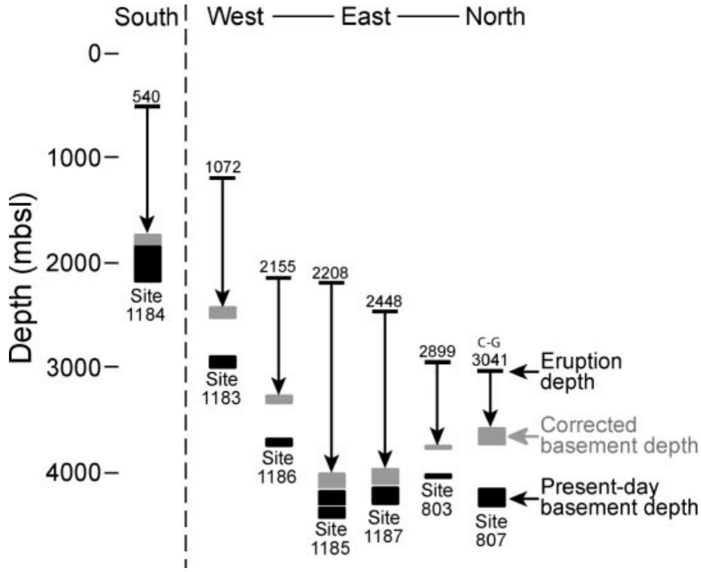

Figure 2: Eruption depth estimates (in meters below sea level, mbsl) for Ocean Drilling Program sites on Ontong Java Plateau. For all sites, present-day depth of top of igneous basement has been corrected for sediment loading. Corrected basement depth (D_c) is obtained from equation of Crough (1983): $D_c = d_w + t_s(\rho_s - \rho_m)/(\rho_w - \rho_m)$, where d_w is water depth (in meters), t_s is sediment thickness (in meters), ρ_s is average sediment density (1.9 g/cm³), ρ_m is upper-mantle density (3.3 g/cm³), and ρ_w is seawater density (1.03 g/cm³).

for our CO_2 analyses (see Table 1). Our new analyses and data reduction procedure result in CO_2 values that are ~25% lower than those of Michael (1999).

ESTIMATING PALEOERUPTION DEPTHS

Vapor saturation pressures were calculated for all sites and converted into eruption depths (0.1 MPa = 10 m water depth) assuming equilibrium solubility of H_2O and CO_2 at the depth of quenching (Fig. 1; see Table 1 caption for details). Glass shards from the volcaniclastic deposits at Site 1184 have low vapor saturation pressures, indicating an average quenching depth of 540 ± 210 m (Fig. 2). Site 1183 glasses, which come from the shallowest water site on the central high plateau, also yield shallow eruption depths (1070 ± 90 m). The deeper-water Sites 1185, 1186, and 1187 yield eruption depths of 2150–2450 (±200) m. Samples from Site 803 yield an eruption depth of 2900 ± 90 m. At Site 807, very different CO_2 contents were found in glasses from the upper (unit A) and lower (units C–G) parts of the hole (Table 1; Michael, 1999), making an eruption depth estimate highly uncertain (see following discussion).

Estimated eruption depths for all sites should be viewed with caution. Submarine pillow rims, particularly MORB samples, are commonly supersaturated with CO_2 (Dixon and Stolper, 1995), so measured CO_2 contents could potentially overestimate true eruption depths. However, submarine Ontong Java Plateau lava flows are likely to have much larger volumes and longer flow distances than MORB flows; geochemical data suggest that some flows on the plateau may have traveled hundreds of kilometers (P.J. Michael, 2004, personal commun.). This would allow time for dissolved CO_2 to reach equilibrium at the seafloor depth before quenching. In fact, such long downslope flow distances could have caused lavas to be vapor saturated near their eruption (vent) depth, which would be shallower than the final depth of emplacement (Michael, 1999). Thus we argue that our eruption depths calculated from CO_2 data are minimum values because true emplacement depths could have been deeper. This provides a plausible explanation for the differences in CO_2 contents and inferred eruption depths of glasses from Site 807 unit A and units C–G (Table 1). The low CO_2 contents of unit A glasses suggest that this may have been part of a very long lava flow that had an original vent in much shallower water. In contrast, units C–G represent multiple flows, all of which have much higher CO_2, and their CO_2 contents probably more closely represent their original emplacement depth (3041 ± 240 m).

For glass shards in the Site 1184 hydrovolcanic deposits, equilibrium degassing probably did not occur because of rapid quenching during ascent and eruption. The characteristics of these deposits suggest subaerial deposition (Thordarson, 2004), so in our subsidence calculations we assume that Site 1184 deposits were emplaced at sea level.

UPLIFT AND SUBSIDENCE OF THE ONTONG JAVA PLATEAU

The estimated eruption depths allow us to constrain the maximum initial uplift of the Ontong Java Plateau. Mesozoic marine magnetic anomalies in the adjacent Nauru Basin suggest that the plateau formed on 10–35 m.y.-old oceanic crust (Ingle and Coffin, 2004). This oceanic crust would have been at 3600–4700 m below sea level(mbsl) (Stein and Stein, 1992; Parsons and Sclater, 1977). Evidence to date indicates that the central high plateau never underwent subaerial volcanism. Using an eruption depth of 1070 ± 90 m for Site 1183 on the high part of the plateau (Table 1), we estimate a maximum elevation for the plateau of 2500–3600 m above the surrounding seafloor.

The arrival of a hot buoyant plume at the base of the lithosphere, combined with crustal thickening due to eruption and intrusion of a large volume of magma, should produce substantial surface uplift (Olson and Nam, 1986). Dynamic models for the Ontong Java Plateau predict an uplift of ~1000–3000 m above the surrounding seafloor (Neal et al., 1997, and references therein). Models that explain hotspot uplift by isostatic compensation of thermally expanded mantle rather than dynamic effects of a rising plume yield similar results (Ito and Clift, 1998). The isostatic effect of crustal thickening has also been calculated; using an average crustal density of 3.08 g/cm³ yields an

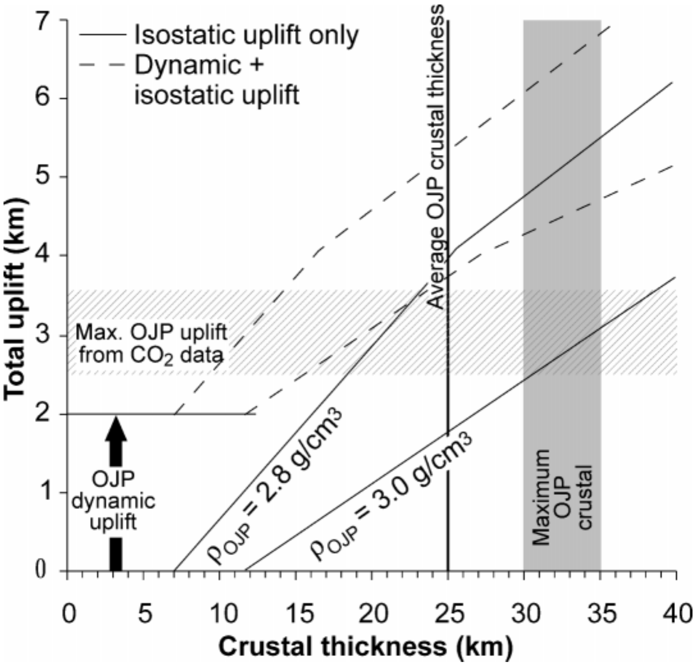

Figure 3: Comparison of predicted vs. observed uplift of Ontong Java Plateau (OJP). Solid lines show estimated isostatic uplift due to crustal thickening. Dashed lines represent isostatic uplift plus average of 2000 m of dynamic uplift based on plateau-specific models (Neal et al., 1997; Ito and Clift, 1998). For calculations below sea level, water-corrected isostasy was calculated using $\Delta h = [h_{OJP}(\rho_m - \rho_{OJP}) + h_{oc}(\rho_{oc} - \rho_m)]/(\rho_m - \rho_w)$ where Δh is amount of uplift above seafloor, h_{OJP} is OJP crustal thickness, h_{oc} is thickness of normal oceanic crust (7 km), ρ_w is water density (1.03 g/cm^3), ρ_m is mantle density (3.3 g/cm^3), and ρ_{OJP} is OJP crustal density (2.8–3.0 g/cm^3). For calculations above sea level, water- and air-corrected isostasy were calculated by using $\Delta h = [h_{OJP}(\rho_m - \rho_{OJP}) + h_{oc}(\rho_{oc} - \rho_m) + h_w \rho_w]/\rho_m$, where h_w is water depth for normal 10–35 Ma oceanic crust (4.1 km). Ruled area shows maximum plateau uplift inferred from paleoeruption depths based on CO_2 data.

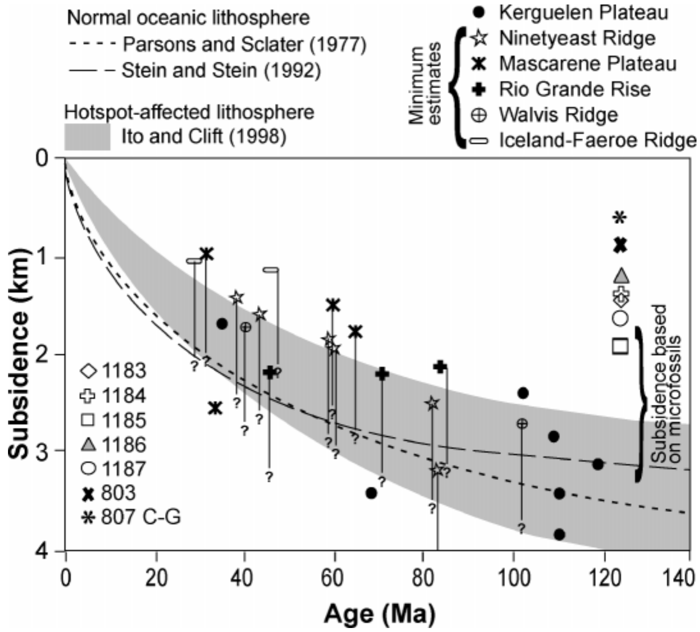

Figure 4: Subsidence estimates vs. age for Ocean Drilling Program (ODP) sites on Ontong Java Plateau. Subsidence estimates based on microfossils are from Ingle and Coffin (2004). Subsidence estimates for other large igneous provinces (Detrick et al., 1977; Coffin, 1992) except for Kerguelen Plateau (Wallace, 2002) are minimum values and assume that these features originally formed at sea level; true subsidence for these could be 1000–2000 m greater than values plotted. Subsidence of hotspot-affected lithosphere (shaded field; Ito and Clift, 1998) is calculated for plume excess temperatures (ΔT) ranging from 200 °C (minimum subsidence) to 350 °C (maximum subsidence). Symbols correspond to ODP site numbers.

additional isostatic uplift of 1800–3000 m (Neal et al., 1997). However, Gladczenko et al. (1997) calculated the average Ontong Java Plateau (OJP) crustal density to be lower (2.86 g/cm^3) on the basis of combined seismic velocity analyses and gravity modeling. Given uncertainties in velocities and the nonunique nature of gravity modeling, we calculate isostatic uplift for a range of densities (2.8–3.0 g/cm^3). Using the high plateau (Site 1183) crustal thickness of ~30 km (Gladczenko et al., 1997), we estimate isostatic uplift ranging from 2400 m (ρ_{OJP} = 3.0 g/cm^3) to 4700 m (ρ_{OJP} = 2.8 g/cm^3) above the surrounding seafloor due to effects of crustal thickening (Fig. 3). Adding the initial dynamic uplift (2000 ± 1000 m), and correcting for changes in seafloor water depth, maximum total uplift would be 4300–6100 (±1000) m above the surrounding seafloor (Fig. 3). These estimates are significantly larger than the estimated maximum uplift (2500–3600 m) based on H_2O and CO_2 data for the basaltic glasses from Site 1183.

After their emplacement, oceanic plateaus subside as a result of cooling and contraction of the lithosphere (Detrick and Crough, 1978; Coffin, 1992). Thermal subsidence curves for normal oceanic lithosphere as it moves away from the ridge crest and for 10–35 m.y. old lithosphere that is thermally rejuvenated as it passes over a hot mantle plume suggest that the 122 Ma Ontong Java Plateau should have subsided ~2700–4100 m since its formation (Fig. 4). After correcting for sediment loading (Fig. 2), we calculate total subsidence of the plateau by subtracting the present-day basement depth from the original eruption depth estimated from H_2O and CO_2 data. The subsidence estimates vary from 900 m (Site 803) to 1900 m (Sites 1184 and 1185) with an average of 1500 ± 400 m over much of the plateau (Fig. 4; Table 1). We have excluded Site 807 from our subsidence average because of the large differences in CO_2 content between units A and C–G glasses, but our preferred eruption depth based on the C–G glasses as described here suggests 600 ± 250 m of subsidence at this site. Our estimated average subsidence for the Ontong Java Plateau is lower than previous estimates based on microfossils (Fig. 4; Ito and Clift, 1998) and CO_2 in glasses from Site 807 unit A (Michael, 1999).

DISCUSSION

Our data clearly show that both the initial uplift (Fig. 3) and post-eruption subsidence (Fig. 4) of the Ontong Java Plateau are significantly less than predictions from thermal models of oceanic lithosphere, and less than what is observed for other oceanic large igneous provinces (Fig. 4). One possibility is that uplift was tempered by the presence of dense garnet granulite and possibly eclogite in the plateau's lower crust that formed from cumulates and intruded and underplated gabbros (Neal et al., 1997). Direct evidence for garnet granulite in the lower crust comes from xenoliths in 34 Ma alnöites on the island of Malaita (Neal et al., 1997). However, seismic velocities, gravity data, phase equilibria, and crustal thickness estimates based on geophysical data do not support the widespread presence of eclogite (Gladczenko et al., 1997; Richardson et al., 2000). These data do not eliminate the possibility of high-density hidden cumulates in the lower crust, but they indicate that the contribution of such rocks to the average crustal density of the plateau is significantly less than Neal et al. (1997) estimated. The upper limit for average crustal density (3.0 g/cm^3) in our modeling allows for the presence of significant dense garnet granulite in the lower crust, but still predicts more initial uplift than is observed (Fig. 3).

If crustal characteristics of the plateau are not responsible for the anomalous uplift and subsidence, then the cause may be in the underlying mantle. The production of a pleateau-scale volume of basaltic magma would produce an enormous melt-depleted, relatively buoyant residuum in the upper mantle (Neal et al., 1997; Fitton and Godard, 2004). Seismic tomography shows the presence of a seismically slow upper-mantle root extending to ~300 km beneath the plateau, and the seismic characteristics of this root suggest that it is chemical or mineralogical rather than thermal in origin (Richardson et al., 2000; Gomer and Okal, 2003). However, the volume of the root is much larger than can be explained by melt extraction needed to form the pleateau (Neal et al., 1997). Given the enigmatic nature of this low-velocity root, its role in causing the anomalous uplift and subsidence behavior of the plateau is unclear. Another possibility is that large-scale magmatic underplating for ~30 m.y. after formation of the plateau provided a continued heat source and thus reduced subsidence (Ito and Clift, 1998). While there is evidence of some younger volcanic events, the lack of voluminous volcanism after 122 Ma seems inconsistent with this hypothesis.

As an alternative to the mantle plume hypothesis, the plateau may have been formed by a large bolide impact (Rogers, 1982; Ingle and Coffin, 2004). Both Ingle and Coffin (2004) and Tejada et al. (2004) proposed that this could explain the anomalous uplift and subsidence because the impact hypothesis does not require a positive mantle temperature anomaly to generate large degrees of melting, and hence it would neither buoy the lithosphere nor lead to subsequent lithospheric cooling and contraction. However, the thermal effects of a bolide impact are difficult to model and could be comparable to those of a mantle plume in creating surface uplift and later subsidence (J. Korenaga, 2004, personal commun.). From a geochemical perspective, the impact hypothesis is also controversial because it requires an Early Cretaceous upper mantle in the region of the plateau that was richer in ocean-island–like isotopic components than average Pacific upper mantle (Ingle and Coffin, 2004; Tejada et al., 2004). Although the Early Cretaceous paleoenvironmental record has not yet yielded evidence for a bolide impact (Tejada et al., 2004), the record has not been scrutinized for such evidence, and differences between subaerial, shallow-water, and deep-water impacts in producing such evidence are not well understood (Ingle and Coffin, 2004). More work is clearly needed to determine whether the world's largest large igneous province was formed by a mantle plume, bolide impact, or some other process.

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REFERENCES CITED

Coffin, M.F., 1992, Emplacement and subsidence of Indian oceanic plateaus and submarine ridges, *in* Duncan, R.A. et al., eds., Synthesis of results from scientific drilling in the Indian Ocean: American Geophysical Union Geophysical Monograph 70, p. 115–125.

Coffin, M.F., and Eldholm, O., 1994, Large igneous provinces; crustal structure, dimensions, and external consequences: Reviews of Geophysics, v. 32, p. 1–36.

Crough, S.T., 1983, The correction for sediment loading on the seafloor: Journal of Geophysical Research, v. 88, p. 6449–6454.

Detrick, R.S., and Crough, S.T., 1978, Island subsidence, hot spots, and lithospheric thinning: Journal of Geophysical Research, v. 83, p. 1236–1244.

Detrick, R.S., Sclater, J.G., and Thiede, J., 1977, The subsidence of aseismic ridges: Earth and Planetary Science Letters, v. 34, p. 195–196.

Dixon, J.E., and Stolper, E.M., 1995, An experimental study of water and carbon dioxode solubilities in mid-ocean ridge basaltic liquids. Part II: Applications to degassing: Journal of Petrology, v. 36, p. 1633–1646.

Dixon, J.E., Stolper, E.M., and Holloway, J.R., 1995, An experimental study of water and carbon dioxode solubilities in mid-ocean ridge basaltic liquids. Part I: calibration and solubility models: Journal of Petrology, v. 36, p. 1607–1631.

Fitton, J.G., and Godard, M., 2004, Origin and evolution of magmas on the Ontong Java Plateau, *in* Fitton, J.G., et al., eds., Origin and evolution of the Ontong Java Plateau: Geological Society [London] Special Publication 229, p. 151–178.

Gladczenko, T.P., Coffin, M.F., and Eldholm, O., 1997, Crustal structure of the Ontong Java Plateau: Modeling of new gravity and existing seismic data: Journal of Geophysical Research, v. 102, p. 22,711–22,729.

Gomer, B.M., and Okal, E.A., 2003, Multiple-ScS probing of the Ontong-Java Plateau: Physics of the Earth and Planetary Interiors, v. 138, p. 317–331.

Ingle, S., and Coffin, M.F., 2004, Impact origin for the greater Ontong Java Plateau?: Earth and Planetary Science Letters. v. 218, p. 123–134.

Ito, G., and Clift, P.D., 1998, Subsidence and growth of Pacific Cretaceous plateaus: Earth and Planetary Science Letters, v. 161, p. 85–100.

Mahoney, J.J., Fitton, J.G., Wallace, P.J., et al., 2001, Proceedings of the Ocean Drilling Program, Initial reports, Volume 192: College Station, Texas, Ocean Drilling Program, 75 p.

Michael, P.J., 1999, Implications for magmatic processes at Ontong Java Plateau from volatile and major element contents of Cretaceous basalt glasses: Geochemistry, Geophysics, Geosystems, v. 1, doi: 10. 1029/1999GC000025.

Neal, C.R., Mahoney, J.J., Kroenke, L.W., Duncan, R.A., and Petterson, M.G., 1997, The Ontong Java Plateau, *in* Mahoney, J.J., and Coffin, M.F., eds., Large igneous provinces: Continental, oceanic, and planetary flood volcanism: American Geophysical Union Geophysical Monograph 100, p. 183–216.

Newman, S., and Lowenstern, J.B., 2002, VOLATILECALC: A silicate melt-H_2O-CO_2 solution model written in Visual Basic for Excel: Computers & Geosciences, v. 28, p. 597–604.

Olson, P., and Nam, I.S., 1986, Formation of seafloor swells by mantle plumes: Journal of Geophysical Research, v. 91, p. 7181–7191.

Parsons, B., and Sclater, J.G., 1977, An analysis of the variation of ocean floor bathymetry and heat flow with age: Journal of Geophysical Research, v. 82, p. 803–827.

Richardson, W.P., Okal, E.A., and Van der Lee, S., 2000, Rayleigh-wave tomography of the Ontong-Java Plateau: Physics of the Earth and Planetary Interiors, v. 118, p. 29–51.

Roberge, J., White, R.V., and Wallace, P.J., 2004, Volatiles in submarine basaltic glasses from the Ontong Java Plateau (ODP Leg 192): implications for magmatic processes and source region compositions, *in* Fitton, J.G., et al., eds., Origin and evolution of the Ontong Java Plateau: Geological Society [London] Special Publication 229, p. 239–257.

Rogers, G.C., 1982, Oceanic plateaus as meteorite impact signatures: Nature, v. 299, p. 341–342.

Stein, C.A., and Stein, S., 1992, A model for the global variation in oceanic depth and heat flow with lithospheric age: Nature, v. 359, p. 123–129.

Tejada, M.L.G., Mahoney, J.J., Castillo, P.R., Ingle, S.P., Sheth, H.C., and Weis, D., 2004, Pin-pricking the elephant: Evidence on the origin of the Ontong Java Plateau from Pb-Sr-Hf-Nd isotopic characteristics of ODP Leg 192 basalts, *in* Fitton, J.G., et al., eds., Origin and evolution of the Ontong Java Plateau: Geological Society [London] Special Publication 229, p. 133–150.

Thordarson, T., 2004, Accretionary-lapilli–bearing pyroclastic rocks at ODP Leg 192 Site 1184: A record of subaerial phreatomagmatic eruptions on the Ontong Java Plateau, *in* Fitton, J.G., et al., eds., Origin and evolution of the Ontong Java Plateau: Geological Society [London] Special Publication 229, p. 275–306.

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