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Implications for magmatic processes at Ontong Java Plateau from volatile and major element contents of Cretaceous basalt glasses

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Abstract. [1] Major elements Cl, CO_2, and H_2O were determined in unaltered tholeiitic basalt glasses from Drilling Program Sites 803D and 807C on Ontong Java Plateau (OJP), and Sites 462A in Nauru Basin (NB) and 802A in East Mariana Basin (EMB). Glasses are moderately evolved (Mg# = 5360) with low Na_2O, indicating they formed by large extents of melting. Extent of melting is less for glasses from OJP than from EMB, requiring faster or more focused mantle upwelling or a longer period of activity for OJP. Glasses within each hole define 13 different compositional types. Each compositional type is made up of many glass-bounded flow units whose chemistry is identical within analytical uncertainty, suggesting that each compositional type formed during a single eruptive episode with multiple magma surges. The lack of diversity within each type also suggests that the magmatic system was large and well buffered. Major element chemistry also requires that significant shallow crystallization occurred during plateau formation and limits the amount of deep crystallization that might be responsible for the thick crust of the plateau. Chlorine contents are constant within each compositional type and are 580 ppm in type A and 850 ppm in types CG from Hole 807C, and 360 ppm in glasses from Hole 803D. Glasses from 802A and 462A have ~240 ppm and 1975 ppm Cl, respectively. Cl/K ratios range from 0.2 to 2.33, well above mid-ocean ridge basalt (MORB) mantle values (0.07), suggesting that most liquids assimilated hydrothermally altered material as they ascended and crystallized. High Cl/K ratios also suggest that abundant hydrothermal activity may have accompanied plateau formation. Low dissolved CO_2 (i.e., CO_3^{2-}) contents in one OJP compositional type (54 ppm in 807C type A) are similar to those on shallow sections of the mid-ocean ridge and are consistent with a maximum depth of eruption of 1320 ± 250 m, in contrast to the greater eruption depths inferred from fossils in the overlying sediments. Our proposed eruption depth falls in the MORB field on a plot of depth versus $Na_{8.0}$ supporting a ridge or near-ridge origin. The shallow depth estimate allows for more postformation subsidence and diminishes the importance but does not eliminate the need for additional processes such as prolonged underplating to account for OJP's shallow depth. Higher CO_2 in other glasses, including types CG from Hole 807C, suggests that some liquids were oversaturated with CO_2 when emplaced and therefore did not travel far from their point of eruption. H_2O contents are similar to MORB and range from 0.13 wt% for NB, to 0.25 wt% (807C, types CG) to 0.49 wt% (807C type A). Ratios of H_2O to rare earth elements (REE) are slightly higher than most MORB and are similar to those from northern Mid-Atlantic Ridge (MAR), which has distinctively high H_2O. H_2O/Ce (constant in depleted through enriched MORB) ranges from 200 to 400, compared to 220380 for northern MAR glasses and 150250 for MORB elsewhere and >500 for arc-related glasses. The slightly higher H_2O contents in the OJP mantle source could have contributed very little to the extensive melting that built the plateau, so a substantial temperature anomaly is still required.

Keywords: Ontong Java Plateau; basalt volatiles; magmatic processes.

Index terms: Minor and trace element composition; major element composition; composition of the crust; mid-ocean ridge processes.

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1. Introduction

[2] Large igneous provinces (LIPs) are huge outpourings of basalt that are among the most prominent geological features in the world. Geochronological studies [*Mahoney et al.*, 1993a; *Tarduno et al.*, 1991] indicate that LIPs formed very rapidly and thus may have had profound impacts on global geology, climate, and biology [*Larson*, 1991]. There is growing consensus from geochemical and geochronological studies [*Mahoney*, 1987], as well as laboratory and numerical modeling [*White and Mackenzie*, 1995; *Campbell and Griffiths*, 1990], that LIPs form when the large inflated head of mantle plume, initiated as a boundary layer

instability, intersects the lithosphere.

[3] Ontong Java Plateau (OJP) is the largest LIP. Because it is submarine and covered by sediments, very little is known about the processes attending its formation. However, analysis of basement samples of OJP from drill cores [*Mahoney et al.*, 1993a] and from upthrusted subaerial parts of the plateau [*Tejada et al.*, 1996; *Neal et al.*, 1997] has improved the understanding of OJP. Unlike continental flood basalts these samples have not been influenced by transport through reactive continental crust and can provide unambiguous information about the mantle source and melting characteristics. Geochemical studies of OJP lavas show that they formed by large extents of melting of plume-like mantle [*Mahoney et al.*, 1993a]. The lavas underwent extensive crystal fractionation, suggesting that much of OJP's >30-km-thick foundation is plutonic in character [*Neal et al.*, 1997].

[4] In this work, major elements and volatile contents in fresh glasses from OJP and nearby basins were determined to further explore how this LIP was generated. Knowledge of the liquid composition permits better comparisons of source chemistry, extent of melting, and depths and extents of crystallization between sites and with modern basalts. Crystallization depths provide some limits on the amount of magmatic underplating that might have occurred during plateau formation. Cl contents and Cl/K ratios show whether magmas interacted with the crust during their ascent and also provide insight as to whether shallow magma chambers and hydrothermal systems were associated with volcanism. Concentrations of CO_2 that are dissolved in glasses were determined to estimate the maximum eruption depths. This information is used to constrain the subsidence history of the plateau which is important in determining if volcanism occurred in a single short episode or whether there was prolonged underplating [*Ito and Clift*, 1998]. Knowledge of the depths of eruption also helps us to understand whether OJP was formed as a near-ridge plume like Iceland or erupted far from a ridge [*Coffin and Gahagan*, 1995]. H_2O contents in the glasses were measured to estimate H_2O contents of the mantle sources and to speculate about the importance of H_2O in promoting the extensive melting that formed OJP. Along with major and trace elements this helps to constrain melting temperatures and the magnitude of the temperature anomaly: How different is OJP from modern plumes? Are high temperatures and a plume head required? H_2O/rare earth elements (REE) ratios are also used to compare OJP's mantle source with sources of other submarine basalts.

2. General Features and Tectonic Setting of Ontong Java Plateau

[5] OJP is made up of a basement of Cretaceous tholeiitic basalts [*Kroenke et al.*, 1991] that is overlain by Cretaceous to modern pelagic sediments (> 1000 m thick in places). Its shallowest part lies at 1700-m depth, but most of the plateau is at 2000-3000 m depth and stands roughly 2000-3000 m above the surrounding seafloor [*Kroenke*, 1972]. The southern and southwestern edges have collided with the Solomon Islands arc and are exposed above sea level in Malaita, Santa Isabel, and other islands [*Tejada et al.*, 1996; *Neal et al.*, 1997]. Estimates of crustal thickness vary widely, but recent studies [*Gladczenko et al.*, 1997] suggest that the depth to the MOHO is 3032 km, that the plateau-generated crust is up to 26 km thick, and that OJP's volume is between 44 and 56 x 10^6 km^3 [*Coffin and Eldholm*, 1994].

[6] Paleontologic and paleomagnetic data [*Tarduno et al.*, 1991] as well as $^{40}Ar/^{39}Ar$ ages of basalts from widely separated parts of the plateau (Ocean Drilling Project, ODP, Hole 807; Deep Sea Drilling Program, DSDP, Hole 289; Santa Isabel Island and Malaita Island; see Figure 1) [*Mahoney et al.*, 1993a; Tejada et al., 1996] indicate that plateau-forming eruptions ended simultaneously over areas that are up to 1600 km away and suggest that volcanism occurred over a short time interval ~122 ± 3 Ma. A second, less voluminous, interval of volcanism occurred around 90 ± 4 Ma and is represented by basalts from Hole 803D and Santa Isabel Island (Figure 1) [*Mahoney et al.*, 1993a; *Tejada et al.*, 1996]. Volcanic events recorded in basins proximal to OJP and at distant reference sites suggest a short duration and therefore a very high rate of volcanic output [*Tarduno et al.*, 1991] at 122 Ma. The initiation of what is now the Louisville hotspot may have triggered OJP's construction [*Tarduno et al.*, 1991; Mahoney and Spencer, 1991; Richards et al., 1991], although there are inconsistencies with this theory [*Neal et al.*, 1997]. Despite the strength of the evidence supporting a plume head origin for OJP, special models such as nearly zero absolute plate motion from 125 to 90 Ma or a double plume head [*Bercovici and Mahoney*, 1994] are required to account for the later phase of volcanism (i.e., 90 Ma) of essentially identical chemical character and the apparent lack of a hotspot track.

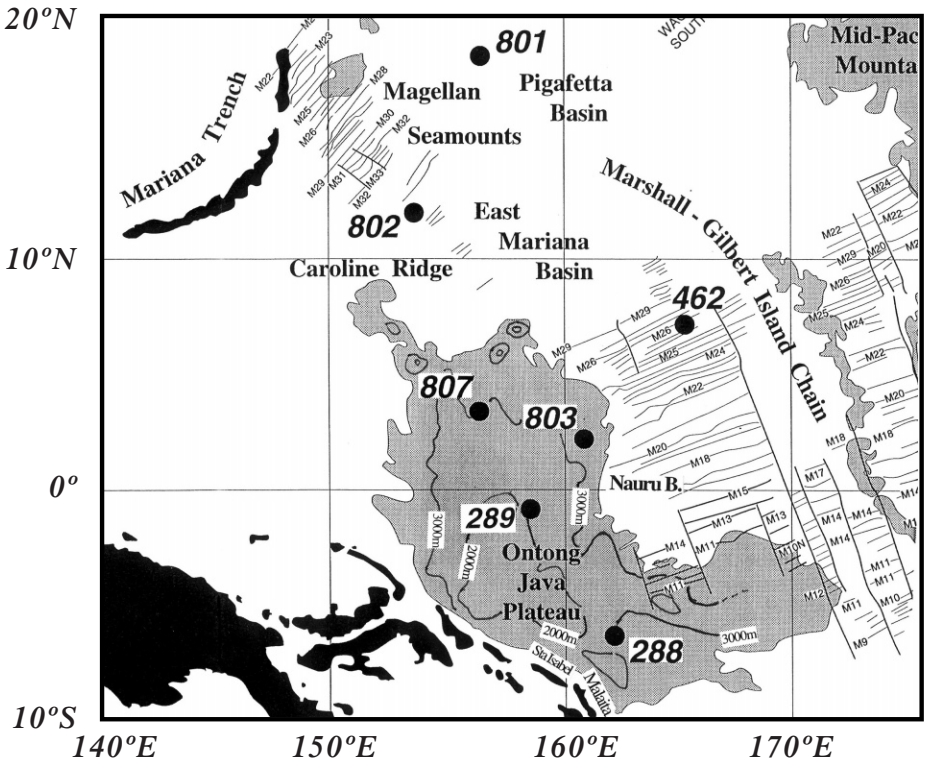

Figure 1. *Ontong Java Plateau and surrounding basins and islands, modified from* Castillo et al. [1994]. *Large numbers are Deep Sea Drilling Project (DSDP) and Ocean Drilling Project drill sites. Additional samples are from Malaita Island. Approximate contours of 2000- and 3000-m depth on Ontong Java Plateau are adapted from* Mahoney et al. [1993a]. *The gray pattern roughly outlines the 4000-m isobath, except on the southwest side, where the plateau is uplifted.*

[7] It is not known how close OJP formed to a mid-ocean ridge (MOR) [*Coffin and Gahagan*, 1995]. Basalt chemistry suggests that extents of melting were similar to those beneath Iceland [*Mahoney et al.*, 1993a], the site of a ridge-centered mantle plume. However, the 20003000 m water depth inferred for emplacement of OJP lavas [*Sliter and Leckie*, 1993; *Saunders et al.*, 1994] is inconsistent with the modern relationship between vigorous ridge-centered plumes and depth on the MOR. The new estimates of eruption depth herein therefore have some bearing on this question. Diverse evidence "weighs lightly against" the generation of OJP at a mid-ocean ridge similar to Iceland [*Coffin and Gahagan*, 1995]. Extrapolation of magnetic anomalies from the surrounding seafloor suggests that OJP was emplaced on crust that ranged from 9 to 28 Myr old [*Gladczenko et al.*, 1997]. A plume head responsible for OJP may have eventually become associated with a nearby ridge axis through its influence on ridge propagation [*Mahoney et al.*, 1993a; *Mahoney and Spencer*, 1991], and this might account for the similarity of basalts' major elements with those of Iceland or Kolbeinsey Ridge (Figure 2). This might also account for the lack of a hotspot track. All of the lavas collected thus far from OJP, EMB, and NB underwent extensive fractional crystallization, possibly from picritic or komatiitic primary magmas [*Neal et al.*, 1997; *Storey et al.*, 1991].

[8] Basalts that are very similar in age and chemical and isotopic composition to those from OJP were recovered from EMB to the north of OJP and NB to the east of OJP, at ODP Hole 802A and DSDP Hole 462A, respectively (Figure 1) [*Castillo et al.*, 1986]. EMB basalts are dated at 114.6±3.2Ma [*Castillo et al.*, 1994], while those of NB yielded ages of 110.8±1.0 Ma [*Castillo et al.*, 1991]. These ages are within error of each other and are consistent with ages of overlying sediments but younger than the Jurassic ages predicted

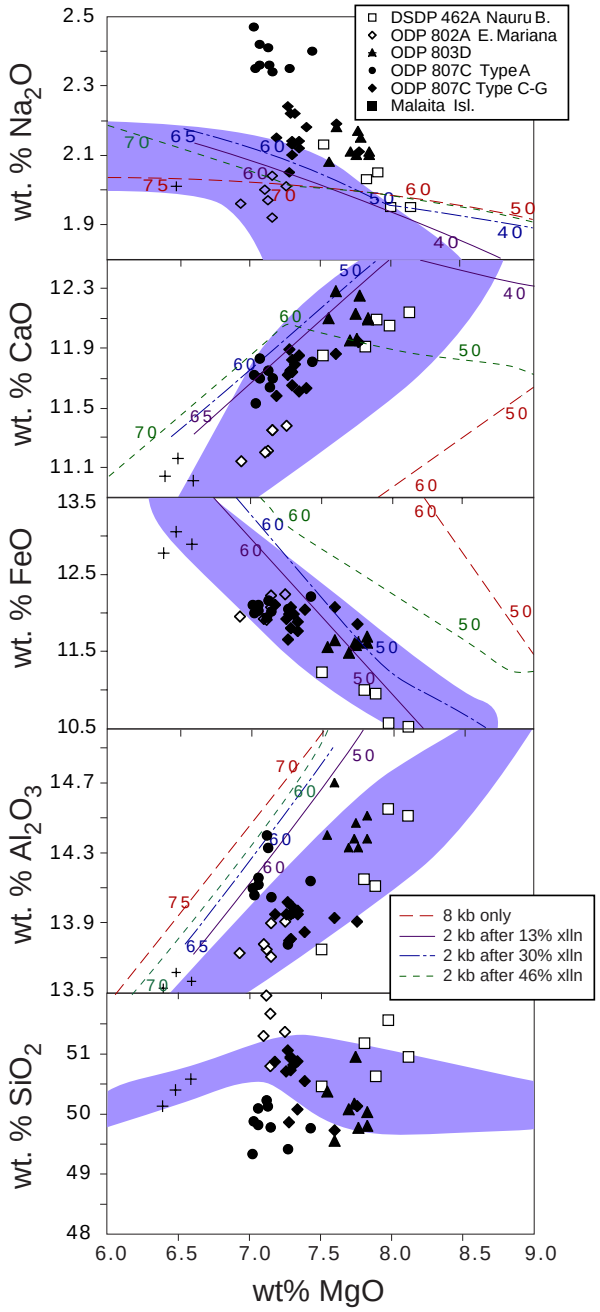

Figure 2. *Major element chemistry of basalt glasses. The very thin (magenta) lines outline the fields of mid-ocean ridge basalt (MORB) from Kolbeinsey Ridge* [Devey et al., 1994] *for comparison. The scatter for each symbol mostly reflects analytical uncertainty. Only the symbols for DSDP 462A show compositional diversity that is significantly greater than analytical uncertainty and appear to define trends. The overall pattern displayed by the different sample types reflects varying extents of fractional crystallization of parental magmas that were broadly similar in composition. Lines show expected trends of liquids derived by fractional crystallization using the model of* Weaver and Langmuir [1980]. *Numbered ticks on the lines correspond to the percent crystallized from an initial picritic liquid with 16.25% MgO, 47.8% SiO_2, 0.6% TiO_2, 14% Al_2O_3, 8.5% FeO, 11% CaO, and 1.4% Na_2O. Besides being run entirely at 8 kbar, the model was run with 13%, 30%, 46% crystallization occurring at 8 kbar, followed by additional crystallization at 2 kbar. (The trend for crystallization at 2 kb-only is the same as the trend for 13% crystallization at 8 kbar, because both involve OL-only crystallization.) The primary liquid and high-MgO parts of the trends are not shown because they would render the rest of the diagram too small. Note that the high SiO_2 and CaO and low FeO of the glasses require that they could not have more than ~3040% crystallization at 8 kbar, before ascending to shallower depths and crystallizing further. Na_2O was not plotted for two Malaita glasses because they were so low, probably because of alteration.*

by extrapolation of marine magnetic anomalies into the basins. The basalts were probably formed during Cretaceous rifting of earlier Jurassic basins that was related to the formation of OJP [*Castillo et al.*, 1994], and may have occurred at a MOR.

3. Samples

[9] Glass intervals were sampled from cores to obtain the freshest, most crystal-free and most representative material possible from each core. Hole 289 contained no glass suitable for analysis. Previous studies [*Castillo et al.*, 1991, 1994; *Mahoney et al.*, 1993b; *Batiza*, 1981; *Saunders*, 1986] that identified different chemical types were used as guides. Sample material that was left over from these studies was also analyzed.

[10] All of the basalts are fine-grained tholeiites and are crystal-poor to aphyric. Vesicles are rare or absent. Glasses are slightly to mostly altered to clays or palagonite. Transparent wafers show 1001000 μm zones of glass that are completely free of alteration. Most analyzed glasses were free of crystals and microlites, but a few had scattered microlites because no better material was available from certain parts of the cores. Three glasses from Malaita were also analyzed [*Tejada et al.*, 1996; *Neal et al.*, 1997]. They are similar to the type A of Hole 807C. Their color and H_2O contents suggest that they have suffered alteration, even though their major element chemistry was consistent with mag-

matic values.

4. Analytical Methods

[11] The freshest and clearest glass shards available were mounted in Buehler Transoptic® and then ground and polished on both sides to a thickness suitable for analysis by Fourier transform infrared (FTIR) spectroscopy: 50200 μm. While still mounted, the transparent glass wafers were analyzed by electron microprobe for major elements Cl and K. The transparency of the wafers allowed us to avoid areas of alteration. Major elements Cl and K were determined using the Cameca MBX electron microprobe at the University of Tulsa following the methods described earlier *Michael et al.*, 1989; *Michael and Cornell*, 1998]. Cl and K were counted for 30 min per sample to obtain a reproducibility of about ±20 ppm Cl or K at a concentration of 400 ppm.

[12] After microprobe analysis the wafers were dismounted and cleaned with acetone and then mapped for thickness (±1 μm precision) using a digital micrometer. Concentrations of volatiles dissolved in glasses (not in vesicles), including CO_3^{2-}, molecular CO_2, molecular H_2O (H_2O_M), and OH^-, were analyzed by FTIR at the University of Tulsa, using published methods and calibrations [*Dixon et al.*, 1988; 1995] with slight modifications [*Michael*, 1995]. CO_3^{2-} appears as an absorbance doublet at 1435 and 1515 cm^{-1} [*Fine and Stolper*, 1986]. Estimation of the complex background in this region is often the largest source of analytical error, so the spectrum of a volatile-free reference glass from Juan de Fuca Ridge [*Dixon et al.*, 1988] was subtracted from each sample spectrum and the background of the resulting spectrum was determined following published methods closely [*Dixon et al.*, 1988, 1995; *J. Dixon*, personal communication, 1998]. Since the solubility versus pressure determinations were done using this method [*Dixon et al.*, 1995], this ensures that our depth constraints will be as accurate as possible. Concentrations are reproducible to better than ±20% of the amount present. The presence of microlites and devitrification may influence CO_2 abundances [*Dixon et al.*, 1988]. No significant differences in CO_2 were noted between glasses with microlites and those with none. Replicate analyses of H_2O in different wafers from the same specimen were reproducible to ±5%. A few samples with H_2O <0.2% were reproducible to ±15% of the amount present. Replicate analyses of the same 80 80 μm spot were closer: ±3%.

[13] The new analyses of glasses (Table 1) are similar to the published analyses of whole-rock powders [*Mahoney et al.*, 1993a; *Tejada et al.*, 1996; *Neal et al.*, 1997, *Castillo et al.*, 1991, 1994] that showed that the lavas are all tholeiitic and are fairly evolved and have low Na_2O and very high FeO (Figure 2) indicative of large extents of partial melting and a large average depth of melting [*Mahoney et al.*, 1993a; *Langmuir et al.*, 1992]. The glasses are similar to MORB from Kolbeinsey Ridge (Figure 2). The glass analyses confirm earlier proposals [*Mahoney et al.*, 1993a, b] that there are two different chemical types in Hole 807C: type A (upper) with higher Na_2O, K_2O, TiO_2, and Al_2O_3 and types CG (lower) with higher SiO_2 and MgO. Each chemical type comprises many different cooling units. A similar arrangement with three chemical types was described for Hole 462A [*Saunders*, 1986], but our glass analyses suggest that their middle type is made up of three chemical types that differ slightly in their degree of evolution. These are so close in composition that they are not discernible in the figures. Glasses 462A-47R-2 and 462A-47R-3 are the first type. Glasses 462A-51R-4 and 462A-56R-1 are a second type, and the single glass from 462A-69R-1 is a third type. No glasses were found or analyzed in the upper and lower chemical types of Hole 462A defined by *Saunders* [1986]. The glass data confirm earlier studies that showed Holes 802A and 803D each contain one chemical type. The glass data are unaffected by crystal accumulation and alteration and show that all of the flows within a chemical type, which may be >50 m thick and comprise 50 different flows, have nearly identical chemistry within analytical error. This is also true for the precise analyses of Cl and K. This remarkable interflow homogeneity is discussed further in section 6.1.1. It is also remarkable how small the range of evolution is within each site and between all of the sites: Mg# ranges from 53 to 60, and K/Ti ranges from 0.13 to 0.18 (Figure 3) for the four different holes (seven chemical types) studied.

5. Results

5.1. Major Elements

[14] The variations in glass chemistry displayed by the eight different chemical types plus Malaita (Figure 2: increasing FeO, Na_2O, K_2O, and TiO_2 and decreasing CaO and Al_2O_3 with decreasing MgO) are consistent with variable extents of fractionation of parental picritic or tholeiitic liquids that were broadly similar

Table 1. Chemistry and Crystallization Pressures of Glasses From Nauru Basin, Ontong Java Plateau, and East Mariana

	SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	Cl	H_2O	CO_2	Total	Ce	Crystallization P PD	PY	PWN	$Na_{8.0}$	P Eq. CO_2
Nauru Basin																				
462A-47R-2, 54-55	50.95	0.91	14.51	10.52	0.20	8.12	12.14	1.95	0.084	0.07	0.197	0.134	0.0096	99.89	6.4	1.6	3.3	2.5	1.98	209
462A-47R-3, 75-76	51.56	0.92	14.55	10.57	0.21	7.98	12.05	1.95	0.084	0.06	0.197	0.143	0.0095	100.39	6.4	1.0	3.0	2.5	1.94	207
462A-51R-4, 28-30	51.80	1.01	14.15	11.00	0.20	7.81	11.91	2.03	0.950	0.07	0.197	0.152	0.0105	99.92	7.1	1.0	2.6	2.3	1.98	229
462A-56R-1, 9-10	50.63	1.00	14.11	10.95	0.19	7.89	12.09	2.05	0.095	0.07	0.199	0.160	0.0104	99.55	6.1	1.1	2.7	1.9	2.02	227
462A-69R-1, 12-13	50.46	1.08	13.75	11.23	0.18	7.51	11.85	2.13	0.100	0.07	0.198	0.186	0.0110	98.87	6.3	0.7	1.9	1.4	2.00	241
Average 462A	**50.96**	**0.98**	**14.21**	**10.85**	**0.20**	**7.86**	**12.01**	**2.02**	**0.092**	**0.07**	**0.198**	**0.155**	**0.0102**	**99.72**	**6.4**	**1.1**	**2.7**	**2.1**	**1.98**	**223**
East Mariana Basin																				
802A-58R-2, 117-122	51.97	1.17	13.75	11.91	0.18	7.12	11.21	1.97	0.116	0.09	0.240	0.282	0.0102	99.89	7.1	0.2	2.0	2.8	1.73	227
802A-58R-2, 142-149	52.02	1.15	13.73	11.95	0.21	6.93	11.14	1.96	0.116	0.08	0.024	0.297	0.0109	99.72	8.2	0.2	1.7	2.5	1.67	243
802A-58R-4, 65-70	51.30	1.17	13.78	11.92	0.21	7.10	11.20	2.00	0.117	0.08	0.031	0.289	0.0107	99.31	8.3	0.4	2.3	2.9	1.76	238
802A-58R-4, 67-68	51.37	1.14	13.91	12.24	0.25	7.25	11.38	2.01	0.111	0.07	0.020	0.289	0.0107	100.14	8.3	0.7	2.8	2.9	1.81	238
802A-59R-1, 131-132	51.67	1.18	13.90	12.22	0.25	7.15	11.35	1.92	0.114	0.09	0.023			99.96		0.2	2.6	2.7	1.69	
802A-59R-2, 56-57	50.80	1.16	13.71	12.07	0.23	7.15	11.35	2.04	1.109	0.07	0.021			98.80		0.5	2.3	2.6	1.81	
Average 802A	**51.52**	**1.16**	**13.80**	**12.05**	**0.22**	**7.12**	**11.27**	**1.98**	**0.110**	**0.08**	**0.024**	**0.289**	**0.0106**	**99.64**	**8.0**	**0.3**	**2.3**	**2.7**	**1.74**	**236**
Ontong Java Plateau																				
803D-69R-2, 1-2	50.03	1.25	14.51	11.61	0.21	7.83	12.09	2.10	0.125	0.09	0.037	0.250	0.0182	100.28	10.1	1.9	4.0	2.6	2.05	397
803D-69R-3, 54-55	49.80	1.30	14.38	11.69	0.24	7.83	12.10	2.11	0.127	0.09	0.036	0.251	0.0184	100.11	12.3	1.9	3.6	2.5	2.06	401
803D-69R-3, 110-111	50.08	1.28	14.33	11.48	0.22	7.70	11.95	2.11	0.133	0.10	0.037	0.274		99.82	12.3	1.8	3.3	2.5	2.03	
803D-70R-1, 112-113	50.95	1.24	14.47	11.58	0.22	7.75	11.96	2.17	0.130	0.07	0.038			100.72	10.6	1.7	3.0	2.8	2.10	
803D-70R-2, 119-120	49.77	1.26	14.33	11.62	0.21	7.77	12.25	2.15	0.127	0.11	0.036	0.280	0.0197	100.06	10.6	1.7	3.6	1.8	2.09	430
803D-70R-2, 6-7	50.37	1.21	14.40	11.55	0.21	7.55	12.10	2.08	0.129	0.07	0.036			99.84	10.6	1.0	2.8	1.7	1.96	
803D-71R-2, 10-11	50.17	1.23	14.38	11.60	0.21	7.74	12.13	2.10	0.133	0.11	0.036	0.284	0.0218	100.28	10.2	1.6	4.0	2.1	2.03	475
803D-71R-3, 9-10	49.55	1.28	14.70	11.64	0.22	7.60	12.28	2.18	0.130	0.10	0.038	0.292	0.0223	100.17	10.2	1.5	3.8	1.9	2.07	486
Average 803D	**50.09**	**1.26**	**14.44**	**11.60**	**0.22**	**7.72**	**12.11**	**2.13**	**0.129**	**0.09**	**0.037**	**0.272**	**0.0201**	**100.16**	**10.9**	**1.6**	**3.5**	**2.2**	**2.05**	**438**
807C-74R-4, 98-100	49.77	1.60	14.14	12.21	0.20	7.43	11.81	2.40	0.202	0.11	0.058	0.444	0.0052	100.52	15.0	2.4	3.0	2.5	2.25	130
807C-74R-4, 98-100	49.88	1.60	14.06	12.00	0.21	7.03	11.53	2.35	0.202	0.11	0.058	0.444	0.0048	99.69	15.0	1.5	2.3	2.3	2.09	121
807C-75R-2, 73-75	50.10	1.62	14.12	12.10	0.24	7.06	11.70	2.36	0.202	0.14	0.059	0.492	0.0042	100.34	14.7	1.1	2.4	1.9	2.11	112
807C-76R-1, 57-58	50.23	1.61	14.40	11.94	0.31	7.12	11.75	2.41	0.193	0.11	0.058	0.400	0.0061	100.54	14.7	1.5	2.6	2.2	2.17	146
807C-77R-2, 108-110	49.78	1.61	14.05	12.02	0.21	7.15	11.70	2.34	0.207	0.10	0.058	0.385	0.0049	99.75	15.2	1.4	2.5	2.2	2.11	119
807C-77R-36-37	50.13	1.64	14.33	12.16	0.24	7.13	11.64	2.36	0.196	0.13	0.055	0.401	0.0055	100.42	14.4	1.6	2.8	2.6	2.13	133
807C-77R-2, 107-108	49.82	1.60	14.16	12.03	0.25	7.06	11.83	2.42	0.193	0.12	0.059	0.399	0.0057	99.95	14.4	1.3	2.2	1.5	2.17	137
807C-78R-1, 33-34	49.34	1.63	14.10	12.10	0.30	7.02	11.72	2.47	0.196	0.12	0.059	0.388	0.0049	99.45	15.2	1.6	2.5	1.8	2.21	119
807C-79R-4, 21-23	49.42	1.61	13.78	12.05	0.17	7.27	11.74	2.35	0.206	0.11	0.056	0.372	0.0076	99.28	15.4	1.7	2.4	1.8	2.15	177
Average 807C A	**49.83**	**1.69**	**14.13**	**12.07**	**0.24**	**7.14**	**11.71**	**2.38**	**0.200**	**0.12**	**0.058**	**0.414**	**0.0054**	**99.99**	**14.9**	**1.6**	**2.5**	**2.1**	**2.15**	**132**

Table 1 (continued). Chemistry and Crystallization Pressures of Glasses From Nauru Basin, Ontong Java Plateau, and East Mariana

	SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	Cl	H_2O	CO_2	Total	Ce	Crystallization P PD	PY	PWN	$Na_{8.0}$	P Eq. CO_2
807C-80R-4, 81-81	50.87	1.12	13.95	12.10	0.20	7.18	11.58	2.15	0.108	0.10	0.087	0.253	0.0149	99.72	9.4	0.6	2.3	2.2	1.93	327
807C-81R-1, 135-137	49.73	1.16	13.93	12.07	0.21	7.60	11.86	2.19	0.110	0.15	0.085	0.233	0.0150	99.46	9.4	1.8	3.5	2.4	2.08	328
807C-82R-1, 31-32	50.95	1.14	13.95	12.00	0.20	7.29	11.65	2.13	0.111	0.07		0.223	0.0129	99.73	9.6	0.7	2.3	2.2	1.94	283
807C-82R-1, 1-2	50.82	1.13	13.96	11.98	0.25	7.31	11.79	2.22	0.110	0.07		0.262	0.0153	99.92	9.4	0.7	2.2	1.8	2.03	335
807C-82R-4, 36-37	49.86	1.16	13.80	11.98	0.27	7.28	11.75	2.22	0.111	0.06	0.089	0.239	0.0152	98.83	9.4	1.0	2.4	1.7	2.03	332
807C-82R-5, 11-12	50.71	1.17	13.95	11.92	0.24	7.26	11.72	2.24	0.106	0.06	0.087	0.258	0.0164	99.74	9.4	0.7	2.1	1.9	2.04	359
807C-82R-6, 120-121	50.55	1.16	13.85	12.04	0.20	7.39	11.63	2.18	0.110	0.07	0.084	0.250	0.0166	99.53	9.4	1.1	2.5	2.4	2.02	363
807C-85R-7, 131-132	50.93	1.14	14.00	12.07	0.29	7.29	11.82	2.14	0.118	0.06	0.088			99.95		0.4	2.2	1.7	1.95	
807C-86R-2, 2-3	50.08	1.13	13.97	11.88	0.22	7.34	11.85	2.14	0.104	0.08	0.084	0.241	0.0165	99.13		0.8	2.5	1.6	1.96	360
807C-86R-2, 136-137	50.73	1.14	13.81	11.80	0.24	7.29	11.74	2.10	0.102	0.08	0.084	0.242	0.0163	99.37		0.3	2.0	1.6	1.91	356
807C-90R-3, 60-63	50.13	1.09	13.91	11.85	0.23	7.76	11.94	2.11	0.110	0.09	0.081	0.223	0.0161	99.66	8.9	1.5	3.3	2.3	2.05	351
807C-90R-2, 118-119	50.88	1.14	13.95	11.76	0.24	7.34	11.61	2.12	0.110	0.07	0.080	0.252	0.0171	99.57	9.1	0.8	2.4	2.4	1.94	373
807C-92R-2, 113-117	51.06	1.09	14.02	11.65	0.23	7.27	11.89	2.05	0.108	0.15	0.082	0.225	0.0174	99.96	8.5	0.0	1.9	1.2	1.85	379
Average 807C C-G	**50.56**	**1.14**	**13.93**	**11.93**	**0.23**	**7.35**	**11.76**	**2.15**	**0.109**	**0.09**	**0.085**	**0.242**	**0.0158**	**99.58**	**9.5**	**0.8**	**2.4**	**2.0**	**1.98**	**345**
Malaita																				
Malaita 1	50.58	1.63	13.57	12.89	0.23	6.59	11.01	1.66	0.190	0.11	0.056	1.250		99.91	15.7	0.1		2.4	1.28	
Malaita 2	50.13	1.60	13.53	12.77	0.22	6.39	11.04	1.48	0.170	0.11	0.056	0.710		98.35	16.0	-0.5			1.05	
Malaita 3	50.40	1.66	13.62	13.05	0.21	6.48	11.16	2.01	0.170	0.11	0.055	0.770		99.85	16.0	-0.4		1.9	1.60	
Average Malaita	**50.37**	**1.63**	**13.57**	**12.90**	**0.22**	**6.49**	**11.07**	**1.72**	**0.180**	**0.11**	**0.060**	**0.910**		**99.37**		**-0.3**		**2.1**	**n.d.**	
Standard JDF-D2	50.59	1.89	13.78	12.14	0.26	6.75	10.66	2.70	0.291	0.23	0.030	0.344	0.0268	99.60						
1 σ (std. dev.)	0.30	0.03	0.11	0.10	0.02	0.15	0.09	0.04	0.009		0.001	0.020	0.0020							

All oxides and volatiles are reported in weight percent except Ce, in ppm. $Na_{8.0}$ is calculated using equation by *Klein and Langmuir* [1987] and slope drived from Kolbeinsey lavas: Na/Mg equals 0.27. Averages are provided because most samples within a group are within analytical uncertainty, except 462A. Ce (in ppm) is taken from nearest analysis in core as reported by *Mahoney et al.* [1993a]. Crystallization pressures are in kilobars and were obtained using three different models: PY, *Yang et al.* [1996]; PD, *Danyushevsky et al.* [1996]; PW, *Weaver and Langmuir* [1990]. P equil CO_2 is the pressure (in bars) at which liquid would be saturated with a CO_2 plus H_2O volatile phase. It should correspond to the maximum eruption pressure [*Dixon et al.*, 1995].

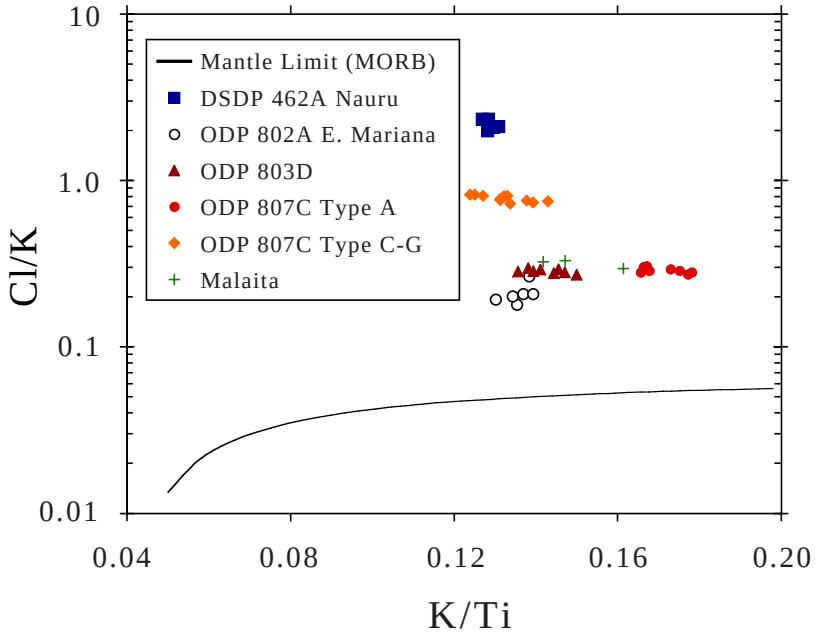

Figure 3. *Cl/K versus K/Ti. Note the logarithmic scale for Cl/K. K/Ti is an indicator of mantle source enrichment: K/Ti < 0.13 in N-MORB and K/Ti > 0.22 in E-MORB. Note the limited range of K/Ti. The mantle trend is derived from N-MORB through E-MORB (K/Ti = 0.020.7) that have erupted on ridges with low magma flux. Mid-ocean ridges with high magma flux often have Cl/K above this trend, up to 0.7. The higher values are presumed to be related to assimilation of seawater-derived components* [Michael and Cornell, 1998].

in composition. However, the long distances between sites and differences in trace element and isotopic ratios between sites preclude the glasses from actually being related to each other. (Crystallization models are discussed in section 6.1.2). There are small but significant differences in K/Ti (Figure 3), an indicator of source composition that should be nearly unaffected by fractionation. Hole 462A and Hole 807C types CG have the lowest K/Ti, while type A in Hole 807C has the highest ratio and Holes 802A and 803D are intermediate. There are also deviations from simple trends that show up as scatter in diagrams of MgO versus oxides. For example, Hole 807 type A has higher K_2O, TiO_2 (not shown), Na_2O, and K/Ti than the trend for other glasses. Glass from Hole 802A has lower Na_2O and CaO and higher SiO_2 at a given MgO than the other glasses. The relation of these differences to variations in source composition and extent of melting is discussed in section 6.3.

5.2. Chlorine

[15] Cl contents range widely from 200 to 1990 ppm (Table 1). These values were obtained on clear, fresh glass fragments and are constant within each glass layer (and even between adjacent glass layers), so they represent magmatic concentrations that were not influenced by posteruptive alteration. The Cl concentrations are high compared to most MORB (20400 ppm) and are remarkably constant within a core, even for the different chemical types of 462A. The altered glasses from Malaita have similar Cl and Cl/K as the 807 type A glasses, suggesting that alteration did not affect Cl or K significantly.

[16] Cl/K is unaffected by partial melting and simple fractional crystallization, and has been defined for mantle sources (Figure 3) [*Michael and Cornell*, 1998]. Cl/K above these values (>0.07) has been attributed assimilation of hydrothermally altered material, possibly brine [*Michael and Schilling*, 1989]. Cl/K is very high for all glasses analyzed, compared to MORB and ocean island basalts (OIB) that have not been affected by assimilation (Figure 3). Cl/K is not correlated with MgO for the OJP/EMB/NB glasses as a group, unlike MORB from fast-spreading ridges. Cl/K ratios are actually higher at a given MgO than those for most MORB. Cl and Cl/K are the highest ever measured in

basalts for all three chemical types from Hole 462A. It is surprising that all three chemical types share this trait, even though their major element chemistry varies slightly.

5.3. CO_2

[17] Dissolved CO_2 is present completely as CO_3^{2-}. No molecular CO_2 was detected. CO_2 concentrations in Hole 807C are 54 ± 10 ppm in type A and 158 ± 20 ppm in types CG basalts. Concentrations are 103 ± 7 ppm in glasses from Hole 802A and 462A. Glasses from Hole 803D had 201 ± 19 ppm CO_2. Analytical errors are significant (20%) because concentrations are low. Standard deviations for multiple analyses within the same flow and within the same chemical type are smaller than the analytical error, suggesting that the measured abundances are reliable magmatic values and have not been strongly influenced by alteration or by the formation of microlites.

5.4. H_2O

[18] Dissolved H_2O is present almost entirely as OH^-, the hydroxyl ion, and varies from 0.13 in NB glasses to 0.5 in OJP glasses (Table 1). Areas of alteration were avoided during analysis. Concentrations are fairly homogeneous within flows and within chemical types, except for 807C, type A. Moreover, the very low (or zero) values of $H_2O_M/(H_2O_M+OH^-)$ are consistent with H_2O speciation expected for magmas [*Dixon et al.*, 1988; *Stolper*, 1982], and not altered glass. Thus H_2O contents represent magmatic values, unaffected by surficial alteration or uptake during eruption. The measureable CO_2 contents, the proposed eruption depths, and the lack of vesicles in the lavas all suggest that H_2O did not degas significantly during eruption. The assimilation of altered material (see section 6.1.3) probably did not significantly affect H_2O contents based on similar studies of MORB [*Michael and Cornell*, 1998; *Michael and Schilling*, 1989] and on the fact that the glasses with the highest Cl have the lowest H_2O. Thus the H_2O contents represent primary magmatic values that can be used to describe the glasses' mantle source. H_2O contents are broadly similar to MORB and are lower than arc-related magmas.

[19] For most MORB, H_2O/Ce ratios are constant from N-MORB to E-MORB, and fall mostly in the range of 150250, while MORB around Iceland and the north-

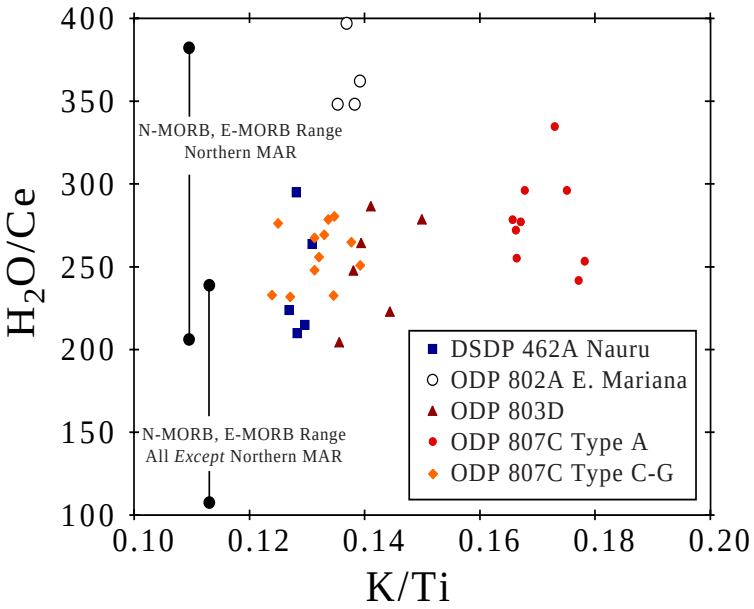

Figure 4. *H_2O/Ce versus K/Ti. H_2O/Ce is roughly constant from N-MORB to E-MORB but varies regionally. In particular, the Mid-Atlantic Ridge north of 30°N has higher H_2O/Ce than the rest of the ridge system* [Michael, 1995]. *H_2O/Ce >500 in glasses associated with island arcs* [Stolper and Newman, 1994]. *Note that the Cretaceous glasses have H_2O/Ce ratios that exceed those of most MORB and are similar to the northern MAR. Glasses from Malaita are not shown because they are affected by alteration. Note the limited range of K/Ti compared to MORB.*

ernmost MAR are higher: from 220 to 380 [*Michael*, 1995] (Figure 4). H_2O/REE for arc related magmas are much higher [*Stolper and Newman*, 1994], typically >500. H_2O/Ce ratios in OJP, NB, and EMB glasses are high compared to most MORB but not as high as arc-related glasses and are similar to the anomalously high northern MAR glasses [*Michael*, 1995] (Figure 4).

6. Discussion

6.1. Magma Chamber Characteristic

6.1.1. Magma homogeneity

[20] Within each chemical type (up to 100 m thick, with >40 cooling units) the glasses show considerably less geochemical variability than similar drilled sections of MORB glasses [*Fisk et al.*, 1996] or even continental flood basalts [*Hooper*, 1997]. The fact that many different cooling units have nearly identical chemistry suggests that each chemical type represents a single eruptive episode that was made up of many lava flows or surges, or lobes, in the terminology of *Self et al.* [1997]. In Hole 807C, ~7 cm of sediment separates basalt layers with identical chemistry: units 4C and 4E [*Kroenke et al.*, 1991], suggesting some time elapsed during this eruptive episode. If this does represent a time interval, such homogeneity would require a very large, well-mixed, and/or strongly buffered magma chamber. It also suggests that the magmas did not travel very far from this reservoir. It is impossible to estimate the areal extent or volumes of any of the basalt chemical types to compare with terrestrial flood basalts.

6.1.2. Crystallization depth.

[21] By analogy with chemically similar MORB and based on petrography of relict or altered phases [*Kroenke et al.*, 1991] these fractionated lavas must have evolved by crystallization of a three phase assemblage consisting of olivine (OL), plagioclase (PL), and clinopyroxene (CPX). By assuming that the glasses are in equilibrium with these three phases, depths of crystallization can be calculated for each glass (Table 1) using three different major element models that describe MORB crystallization *Danyushevsky et al.*, 1996; *Weaver and Langmuir*, 1990; *Yang et al.*, 1996]. The methods rely on the fact that the dT/dP of CPX saturation is significantly greater than that of OL or PL, so as a magma ascends, CPX becomes relatively more undersaturated. A calculated depth represents a minimum for the initiation of crystallization because a magma may have started crystallizing deeper, and then continually crystallized and equilibrated to some extent as it ascended. The depth represents a maximum for the completion of crystallization, because a magma may have been crystallizing OL+PL+CPX at depth, and then ascended to shallower levels without reattaining saturation (equilibrium) with OL+PL+CPX. The three different methods and associated errors are discussed fully by *Michael and Cornell* [1998]. Calculated crystallization pressures vary depending on the model used, but in each case they are in the low end of the range of pressures calculated for MORB [*Michael and Cornell* , 1998]. The pressures are similar to those from fast-spreading ridges like EPR or robust slow-spreading ridges like Kolbeinsey Ridge north of Iceland. Surprisingly, there are no significant differences in crystallization pressures between glasses from OJP and NB and EMB. Calculated crystallization pressures (Table 1) require that a significant amount of crystallization took place at shallow levels. Fast seismic velocities in the lower crustal levels of OJP have been attributed to olivine-rich cumulates that formed by crystallization at high pressure [*Farnetani et al.*, 1996]. Deep and shallow crystallization are not mutually exclusive. Indeed, *Farnetani et al.* [1996] preferred a two-stage model in which picritic magmas first crystallized at MOHO levels and then at shallower levels.

[22] The modeling and discussion below help constrain (1) the amounts of total crystallization, (2) the amount of shallow crystallization that is required to account for the glasses' current compositions, and (3) the amount of OJP's thickness that may be attributed to cumulates formed by crystallization of magmas at deeper crustal depths.

[23] Similar to earlier models [*Farnetani et al.*, 1996], major element compositional evolution was modeled as a two-stage process, with the first stage at 8 kbar and the second stage at 2 kbar. The primary magma composition is unknown, so as a plausible starting point, a picrite magma with ~16% MgO (similar to the composition used previously [*Farnetani et al.*, 1996]) is used. Uncertainty in the primary magma composition leads to uncertainty in the amounts of crystallization and probably leads to some misfit between the models and data (Figure 2). By using a similar composition as the earlier studies [*Neal et al.*, 1997; *Farnetani et al.*, 1996] the results can be compared. The model used here [*Weaver and Langmuir*, 1990] was designed for MORB, so it should be applicable to these tholeiitic compositions. It uses experimentally constrained

distribution coefficients and phase proportions to calculate a LLD, so it should be more accurate than models using arbitrary or observed phase proportions [*Neal et al.*, 1997]. If all crystallization occurred at 8 kbar, the resulting derivative liquids would have CaO and SiO_2 that are much lower and FeO that is much higher than the observed compositions (Figure 2). If all crystallization occurs at 2 kbar, the derivative liquids match the observed compositions much more closely (Figure 2). These results are consistent with the calculated crystallization pressures in Table 1. Total crystallization to attain the fractionated compositions (e.g., 7.2% MgO) from the picrite ranges from ~65% at 8 kbar to ~55% at 2 kbar. Model simulations with varying amounts of early crystallization at 8 kbar show that the observed compositions and trends can be reproduced with up to 3035% crystallization at 8 kbar, followed by the remainder of crystallization (~2530%) at 2 kbar. Greater amounts of crystallization at 8 kbar, such as 46% crystallization at 8 kbar in Figure 2, are unable to reproduce the observed compositions or trends. Thus a significant amount of shallow crystallization is required. A maximum of ~1/3 of the initial melt could have crystallized at 25-km depth. This raises the possibility that some of the plateau was built up by the vertical stacking of successive shallow eruptive systems. In this case the high seismic velocities at the base of OJP might be partly due to the metamorphic grade of the rocks [*Gladczenko et al.*, 1997].

6.1.3. Cl and assimilation.

[24] Very high Cl/K in every flow type suggests that ascending magmas encountered and assimilated rocks that had been affected by hydrothermal alteration [*Michael and Cornell*, 1998]. This, in turn, suggests that hydrothermal activity was widespread on OJP, EMB, and NB. High Cl in MORB is associated with low crystallization pressures and often with a shallow axial magma chamber [*Michael and Cornell*, 1998]. The necessary conditions are probably modulated by magma flux so are favored by a fast-spreading rate but are also present on slow-spreading ridges formed by large extents of melting [*Michael and Cornell*, 1998]. Very high Cl in OJP lavas shows that extremely vigorous shallow magmatic and hydrothermal systems may have been formed by the voluminous magmas that constructed the plateau. The lack of correlation of Cl/K with Mg# for OJP, NB, and EMB is unlike the correlation observed along the southern EPR and suggests that crustal alteration might have been heterogeneous on a timescale and/or distance scale that was larger than that of the magma chambers.

[25] The constant high Cl (1975 ± 8 ppm) in the three compositional types analyzed from Hole 462A is a curious feature that seems to be inconsistent with assimilation, which should be a more random process. For the three sample types analyzed from Hole 462A, TiO_2, FeO, Na_2O, K_2O, and H_2O increase while MgO and Al_2O_3 decrease. These trends are consistent with 510% fractional crystallization and suggest that the three types may be variably differentiated liquids from the same magmatic system. The only problem with this explanation is that Cl should also increase, from 1975 to ~2100 ppm. It is not likely that Cl saturation accounts for the constant Cl, since Cl^- saturation would be >2% in a H_2O-poor basalt melt like this [*Webster et al.*, 1999]. An alternative explanation for high Cl/K in Hole 462A (and the other glasses as well) is that the mantle source of the basalts is enriched in Cl, possibly from recycled subducted altered material [*Jambon et al.*, 1995]. However, Cl/K in OJP and related glasses is not correlated with any other features that are indicative of recycling. Pb and Sr isotopic ratios and trace element ratios (e.g., Ba/Ta) in glasses from Hole 462A are within errors of the Hole 807A types CG ([*Mahoney et al.*, 1993a], and references therein). In summary, an assimilation origin is favored for the high Cl/K in all of these glasses, but the extreme and constant values in Hole 462A are difficult to explain.

6.2. Depths of Eruption

[26] Maximum depths of eruption are calculated for these Cretaceous glasses in Figure 5. by using experimentally determined saturation curves [*Dixon et al.*, 1995] and by assuming that glasses are saturated or oversaturated with a combined CO_2 + H_2O volatile phase at their depth of eruption. This method relies on the strong pressure dependence of CO_2 solubility in magmas, and the fact that modern submarine basalt glasses are rarely undersaturated with respect to CO_2 [*Dixon et al.*, 1988] unless they are associated with subaerial or shallow submarine volcanoes [*Dixon et al.*, 1991]. Deeply erupted tholeiitic glasses typically lose significant amounts of CO_2 but very little H_2O during ascent from the mantle because of the way these gases partition between melt and vapor [*Dixon et al.*, 1995] and because initial H_2O concentrations are low. The calculated depths are maxima because some glasses are oversaturated as they did not have time to degas be-

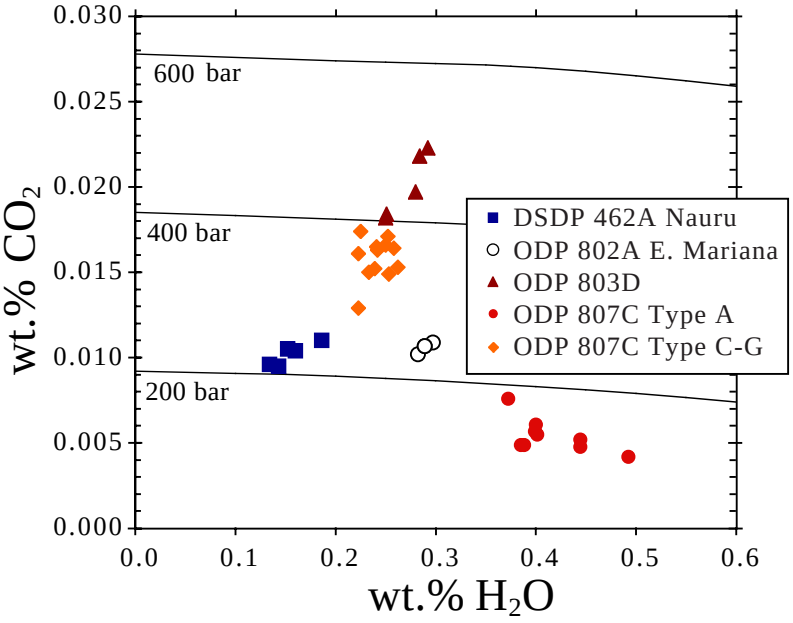

Figure 5. *H_2O versus CO_2, showing contours of pressure at which various combinations of H_2O and CO_2 will be saturated. The curves are excerpted from* Dixon et al. [1995]. *For seawater, depth in meters is 10 pressure in bars. Eruption pressures presented in Tables 1 and 2 are estimated from this diagram and the equations on which it is based. If glasses are oversaturated, their eruption depth will be too high.*

fore cooling, and therefore record a pressure that is too high.

[27] The maximum depth of eruption (Tables 1 and 2) calculated for Hole 807C type A glasses, even accounting for the analytical uncertainty in CO_2 is 1320 ± 250 m, which is shallower than the lower bathyal depth range (20002800 m; [*Sliter and Leckie*, 1993]) that was estimated based on planktonic foraminifers in overlying sediments. The greater eruption depths are based on estimations of the Cretaceous carbonate compensation depth (CCD) and perturbations of the CCD near a submarine plateau. Eruption depths below the CCD were proposed for the Malaita sites by *Neal et al.* [1997] and *Saunders et al.* [1994] on the basis of the lack of calcareous sediments. It is remotely possible that CO_2 was undersaturated during eruption, but this is rare for MORB [*Dixon et al.*, 1988; *P.J. Michael*, unpublished data, 1999] and is highly unlikely here. It is more likely that depth estimates for Hole 807C that are based on sediments are too great. The depth calculated using CO_2 in the glass is nearly within error of the shallowest depth estimated for this site (1800 m) by *Ito and Clift* [1998].

[28] One way to account for the discrepancy between paleoeruption depths calculated from CO_2 in 807C type A glasses and the estimates based on sediments would be to suppose that the lavas were erupted from a shallower edifice and then flowed downslope for hundreds of km to their current location. The most likely source is ~1000 m shallower than Site 807C and 450 km distant. Such long flow distances may be possible for the 28-m-thick flow of unit F but are much less likely (not impossible [*S. Self*, personal communication, 1999]) for the many flows that are <2 m thick in units A and CG [*Mahoney et al.*, 1993a; *P. Hooper*, personal communication, 1998]. The fact that types CG lavas give much greater eruption depths (3420 m) than type A is consistent with the type CG glasses having been oversaturated with CO_2 upon emplacement and cooling. If they were in fact oversaturated, they could not have traveled far from their eruptive source; otherwise, they would have degassed.

[29] The 1320-m eruption depth of 807C type A glasses falls within the broad limits of the present-day $Na_{8.0}$ versus depth curve for hotspot-influenced MORs (Figure 6) [*Langmuir et al.*, 1992], which is permissive evidence for eruption at or near a ridge. The greater depths estimated from sediments (2000-2800 m) also fall inside this array but is considerably deeper than the best analog for a ridge above a large plume: Iceland [*Cof-*

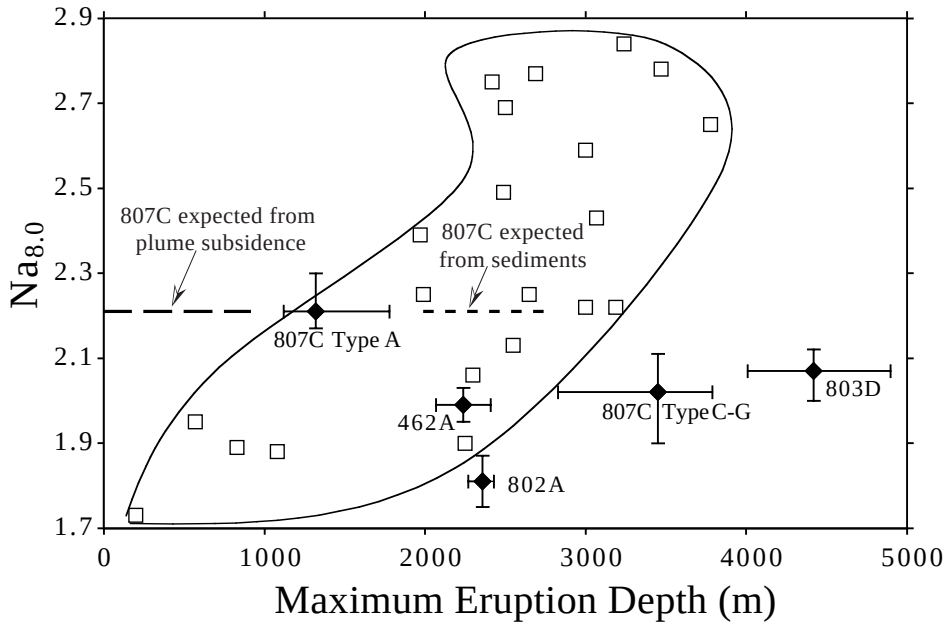

Figure 6. *$Na_{8.0}$ versus calculated maximum eruption chemical type corresponds to average $Na_{8.0}$ as reported in Table 1 and average eruption depth calculated from CO_2 and H_2O content as derived in Figure 5 and reported in Table 2. For each point, the horizontal bars show the range of CO_2 equilibration pressures and the vertical bars show the range of $Na_{8.0}$ for all of the samples of a particular group. The open squares and the field enclosing them denote the averages of hotspot-influenced ridges globally* [Langmuir et al., 1992]. *The short-dash line is the depth range estimated from sediments for Site 807C* [Sliter and Leckie, 1993]. *The long-dashed line is the depth range (9001000 m above sea level) that would be expected for Site 807C at 122 Ma, if it is assumed that the site subsided to its present depth after being over a plume at 122 Ma* [Ito and Clift, 1998]. *Notice the depth estimate from (H_2O + CO_2) is much closer to this estimate.*

fin and Gahagan, 1995]. The shallower depth strengthens arguments in favor of the formation of OJP at or near a MOR.

[30] Lavas of Hole 803D give maximum eruption depths of 4420 m, which is deeper than the present reconstructed basement depth. It is likely that the glasses are oversaturated with CO_2, and therefore do not constrain the paleoeruption depth. It is unlikely that this site has become shallower in the past 90 Myr, as it is distant from the tectonically uplifted southern part of the plateau. The oversaturation suggests that the flows did not erupt far from their present position.

[31] Maximum paleoeruption depths calculated for basin Sites 462A and 802A are 2200 and 2400 m, respectively. These depths are considerably shallower than expected for eruption of basalt onto seafloor that was perhaps already 3570 Myr old. (Extrapolation of magnetic anomalies [*Nakanishi et al.*, 1992] in EMB and NB suggests an age of 145170 Ma. Basalt ages from Hole 802A and 462A are 110115 Ma [*Castillo et al., 1994*]. The shallower eruption depths suggest that both basins were affected by a hotspot swell from OJP. In a plot of $Na_{8.0}$ versus depth (Figure 5) the basin lavas fall at the deep end of the range of hotspot-influenced mid-ocean ridges. However, OJP stands 23 km above the surrounding seafloor, so it would be difficult to have both OJP and EMB and NB conform to the $Na_{8.0}$ versus depth relationship, unless they have followed very different subsidence histories [*Ito and Clift*, 1998].

6.2.1. Implications for plateau subsidence.

[32] Because our calculated eruption depths are maxima, our calculated subsidence amounts are minima. OJP's subsidence was calculated previously [*Ito and Clift*, 1998] by subtracting estimates of eruption depth from calculated basement depths in the absence of sediments (reconstructed basement depths). The amounts of subsidence were not sufficient to account for the thermal subsidence expected after passage of the lithosphere over a mild (T = 150°C) or high-

Table 2. Sample Site and Core Characteristics

Hole	DSDP 462A	ODP 802A	ODP 803D	ODP 807C	ODP 807C (C-G)
Present seafloor depth	5187	5968	3412	2806	2806
Basement depth, mbsf	994	509	631	1380	1380
Basalt drilled, m	215	51	26	44	105
Basalt recovered, m	155	15.3	9.6	22	65
Percent basalt recovered	54	30	38	50	62
Number of cooling units	56	17	25	29	38
Number of chemical types	3+2	1	1	1	1
Thickest flow, m	30	3.2	1.3	2	28
Average flow thickness, m	3	3	1	1.5	2.1
$Na_{8.0}$	1.99	1.81	2.07	2.21	2.02
Max. eruption depth (from CO_2)	2240	2360	4420	1320	3450
Approx. reconstructed basement	5515	6136	3620	3261	3261
Minimum calculated subsidence (n)	3275	3776	-800	1941	-189
Subsidence expected from cooling (1)	2400	2400	1500	2400	2400
Subsidence expected from cooling (2)	4400	4400	2500	4200	4200

Separate columns are provided for 807C chemical types A and C-G. Percent basalt recovered is meters of basement drilled/meters of basalt recovered; number of cooling units: estimated subjectively from DSDP volumes 89 and ODP volumes 129 130; number of chemical types is number of different glass types within analytical uncertainty of major element chemistry plus other types from parts of core with no glass (462A); thickest flow is the widest interval of basalt without a cooling unit break; average flow thickness is meters of basalt drilled/#cooling units (does not include 28-m-thick flow for 807C C-G); approximate reconstructed basement depths are from *Ito and Clift* [1998] directly (803D and 807C), or by using their method (462A and 802A); minimum calculated subsidence is the approximate reconstructed basement depth-maximum eruption depth; subsidence expected from cooling is based on the cooling models of *Ito and Clift* [1998] , and a (1) 150°C or (2)350°C plume that affected the site at 110-120 Ma (90 Ma for 803D).

temperature (T = 350°) plume at 122 Ma. It was proposed that OJP had not subsided as much as expected because it had formed over a long time interval, perhaps by repeated episodes of magmatism and underplating after the 122 Ma event [*Ito and Clift*, 1998]. Our shallower eruption estimate for 807C allows greater subsidence of OJP than *Ito and Clift* [1998]. Using their reconstructed basement depth (3300 m) and our 1320-m eruption depth estimate, subsidence is 1980 m for Hole 807C, which is nearly enough to account for continuous cooling of lithosphere that was influenced by a mild plume at 122 Ma (~2500 m) [*Ito and Clift*, 1998]. Greater subsidence (~ 4300 m) would be expected for a higher-temperature plume [*Ito and Clift*, 1998].

[33] Subsidence for Holes 462A and 802A is calculated similarly in Table 2, using eruption depths from CO_2 and reconstructed basement depths that are roughly estimated using the methods of *Ito and Clift* [1998]. The minimum subsidence amounts calculated for Nauru Basin and EMB (3275 and 3775 m) fall within the broad limits (2200 - 4400 m) that are predicted for lithospheric cooling after passage of a plume at 110 Ma [*Ito and Clift*, 1998, Figure 4], and are ~1300 and 1800 m greater than the subsidence amounts calculated above for OJP. (Sediment-based eruption estimates for OJP of 3000 m [*Saunders et al.*, 1994] provide 2000 - 2700 m greater subsidence for EMB or Nauru compared to OJP). In any case, if OJP was centered on a hotspot swell, it should have subsided more, not less, than the surrounding basins, unless it underwent late stage uplift [*Ito and Clift*, 1998]. In summary, our estimates of shallower eruption depth and greater subsidence reduce the amount but do not eliminate the need for later magmatic activity [*Ito and Clift*, 1998] or other mechanisms [*Neal et al.*, 1997] to account for the anomalously low subsidence of OJP.

6.3. Melting and Source Characteristics

[34] For liquids formed by large extents of partial melting, K/Ti is a good indicator of the source's degree of incompatible element enrichment. All of the glasses fall within a narrow range of K/Ti (Figure 4). The source for Hole 807C type A is slightly more enriched, while that for Hole 462A is more depleted and is similar to that from Hole 807C types CG. Glasses from Holes 802A and 803D have intermediate K/Ti in their sources. This order is consistent with previous suggestions for Holes 803D and 807C based on incompatible trace elements [*Mahoney et al.*, 1993a].

[35] Once the variations in source composition are con-

sidered, it is possible to examine differences that may be consistent with variations in the extent and depth of mantle melting. Hole 802A has lower Na_2O at a given MgO than other chemical types (Figure 2); yet it has higher K/Ti (Figure 3) and similar K_2O and TiO_2 concentrations (Table 2) compared to 807C types CG. These differences cannot be due to differences in crystallization depth. They most likely reflect higher concentrations of K and Ti (and to a lesser extent, Na) in the source of 802A but a greater extent of melting that leads to similar K_2O and TiO_2 and lower Na_2O at the same MgO in the glasses. Considering that $Na_{8.0}$ varies from 3.5 to 1.9% in MORB as the extent of melting increases from 8 to 22% [*Klein and Langmuir*, 1987], the difference in $Na_{8.0}$ between 802A and 807CG (1.81 versus 2.02%) would correspond to roughly 2% (absolute) additional melting. If the total extent of melting is 20%, then this is a 10% (relative) increase in the amount of melting. The difference is probably even larger since the source of 802A probably had a higher Na_2O content. Higher SiO_2 and lower CaO in Hole 802A glasses [*Langmuir et al.*, 1992; *Jaques and Green*, 1980], which seems likely for these lavas. If the voluminous lavas of OJP formed by lower extents of melting than the adjacent and less voluminous EMB lavas, then either there must have been much faster or more focussed mantle upwelling beneath OJP, or OJP must have been built over a longer time interval.

6.3.1. Source concentrations of H_2O and their influence on melting.

[36] Could the relatively high H_2O in the mantle have contributed to the large extent of melting that formed these lavas? If the influence of H_2O on melting was large, then a smaller temperature anomaly would be necessary to account for the large extents of melting. Data from island arc basalts have been used to show that melting increases by ~6% (absolute) for every additional 0.1% H_2O in the source [*Stolper and Newman*, 1994]. The H_2O content of the sources of the OJP can be calculated after correction for 5065% fractional crystallization (estimated from Figure 2) and 1924% partial melting estimated from $Na_{8.0}$ and the range of melting of MORB [*Klein and Langmuir*, 1987]. Calculated H_2O contents of the mantle source are roughly 250 ppm for EMB and 200300 ppm for OJP lavas (or 300450 ppm if the 3045% crystallization estimates of *Neal et al.* [1997] are used). These H_2O contents are higher than N-MORB sources and are lower than E-MORB sources [*Michael*, 1995] and more enriched Hawaiian sources (525 ppm [*Dixon et al.*, 1997]). The additional H_2O content can account for only ~2% (absolute) of the melting compared to N-MORB. Therefore a substantial temperature anomaly, similar to or greater than that of Iceland is still required. In models of plateau formation and subsidence, larger temperature anomalies result in larger initial excess bathymetry and greater subsequent subsidence. This strengthens arguments that some other process is necessary to account for the small amount of subsidence that OJP has undergone subsequently [*Ito and Clift*, 1998]. H_2O differences in the mantle source are also unable to account for the larger extents of melting of Hole 802A glasses compared to glasses from Hole 807C, since H_2O contents are slightly lower for the source of 802A glasses than for 807C type A glasses.

7. Conclusions

[37] For basalts from OJP, NB, and EMB the strong homogeneity of many flow units within each compositionally defined type suggests that each chemical type represents a single eruptive episode with many flow lobes. The homogeneity suggests that magmatic systems (chambers?) were large and well buffered. Magmas at all locations fractionated extensively at shallow pressures. Deep crystallization cannot be ruled out, but it must have been limited, and it cannot account for all of the crystallization. Very high Cl concentrations and Cl/K ratios in all of the glasses suggest that there was abundant hydrothermal activity associated with magmatism and that magmas assimilated hydrothermally influenced material, probably at shallow crustal levels.

[38] Eruption depth at Site 807C derived from CO_2 and H_2O dissolved in glass is 1320 ± 250 m, which is shallower than previous estimates. It is remotely possible that the magmas erupted far away and flowed long distances to this site. Greater amounts of subsidence are calculated using the shallower eruption depth. This reduces the amount but does not eliminate the need for subsequent magmatism to explain the anomalously shallow nature of the plateau. The shallower eruption estimates also provide support (although weak) for eruption of OJP at or near a MOR. Some of the glasses are oversaturated with CO_2, suggesting they did not flow far from their eruptive vent.

[39] H_2O in the mantle source of OJP is similar to the northern MAR source, which is distinctively higher than other MORB sources. The additional H_2O is not

sufficient to have caused extensive melting that generated OJP. A substantial temperature anomaly is still required. EMB liquids formed by the greatest extents of melting. The greater volume of OJP, yet lesser extents of melting compared to EMB means that mantle upwelling beneath OJP was faster or more focused or of longer duration.

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