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Crustal structure of the Ontong Java Plateau: Modeling of new gravity and existing seismic data

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Abstract. Seismic refraction and gravity-based crustal thickness estimates of the Ontong Java oceanic plateau, the Earth's largest igneous province, differ by as much as 18 km. In an attempt to reconcile this difference we have evaluated available seismic velocity data and developed a layered crustal model which includes (1) a linear increase in velocity with depth in the Cenozoic sediments and the uppermost extrusive basement and (2) a reinterpretation of deep crustal and Moho arrivals in some deep refraction profiles. Previously, Moho had commonly been interpreted from later arrivals and in some cases constrained by precritical arrivals. However, if first arrivals at distal offsets are interpreted as Moho refractions, the maximum depth to Moho is reduced by about 10 km. Two-dimensional gravity modeling along two transects from well-determined oceanic crust in the Nauru Basin across the central Ontong Java Plateau to the Lyra Basin, based on the reinterpreted crustal model, is regionally consistent with satellite altimetry derived and shipboard gravity fields yielding a 8.0 km/s Moho velocity at a depth of ~32 km under the central plateau. The crust features a thick oceanic, three-layer igneous crust comprising an extrusive upper crust, a 6.1 km/s middle crust and a ~15 km thick 7.1 km/s lower crust. The total Ontong Java Plateau crustal volume is calculated at 44.4 x 10^6 km^3 and 56.7 x 10^6 km^3 for off- and on-ridge emplacement settings, respectively. On the basis of velocities and densities we interpret the lower crust on the plateau to consist of ponded and fractionated primary picritic melts, which due to deformation and/or fluid invasion may have recrystallized to granulite facies mineral assemblages. The melts were emplaced during lithospheric breakthrough of a mantle plume in an oceanic, near-ridge plate tectonic setting.

1. Introduction

The Ontong Java Plateau (OJP) in the western central Pacific (Figure 1 and Table 1) is the world's most voluminous large igneous province (LIP). LIPs are emplaced during episodes of massive mafic magmatism and comprise continental flood basalts, volcanic passive margins, oceanic plateaus, and ocean basin flood basalts [*Coffin and Eldholm*, 1994]. In particular, oceanic plateaus, by virtue of their tectonic setting, offer excellent potential for studying the velocity structure and dimensions of LIPs. These parameters are particularly sensitive to reliable models of crustal structure and emplacement setting, questions that are not yet properly resolved for the OJP LIP. For example, seismic refraction and gravity data yield maximum Moho depths of 43 and ~25 km, respectively [*Furumoto et al.*, 1976; *Sandwell and Renkin*, 1988]. In this study, we discuss available crustal data in an attempt to resolve this discrepancy and present suggestions about emplacement setting and crustal composition.

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The work is partly based on detailed data analysis and modeling by *Gladczenko* [1994].

The OJP, defined by the 4000 m depth contour, is adjacent to the >4500 m deep Lyra, East Mariana, and Nauru ocean basins and bounded to the southwest by the >3000 m deep North Solomon Trench (Figure 1). The transition between the plateau and East Mariana Basin is marked by the Caroline Seamounts, of Miocene and younger age [*Keating et al.*, 1984]. The plateau proper comprises a 1700-2500 m deep central part where, except for some atolls and seamounts toward the trench, seafloor undulates gently. The central plateau contains 0.8-1.2 s thick sediments thinning to ~0.6 s on the flanks and 0.5-0.2 s in the ocean basins [*Mosher et al.*, 1993]. The western flank is block-faulted (sheared?) toward the Lyra Basin [*Ewing et al.*, 1968; *Kroenke*, 1972], which is characterized by sediment thicknesses of 0.7-0.9 s. The northern and eastern flanks are incised by submarine canyons [*Kroenke*, 1972]. The eastern flank linking the plateau with the southern Nauru Basin is characterized by ridges and seamounts rising up to ~1.5 km above the surrounding seafloor. Although seismic data show abundant erosional features in the area, the ridges and seamounts appear to reflect basement topography. The gravity field shows that the ridges, troughs, and

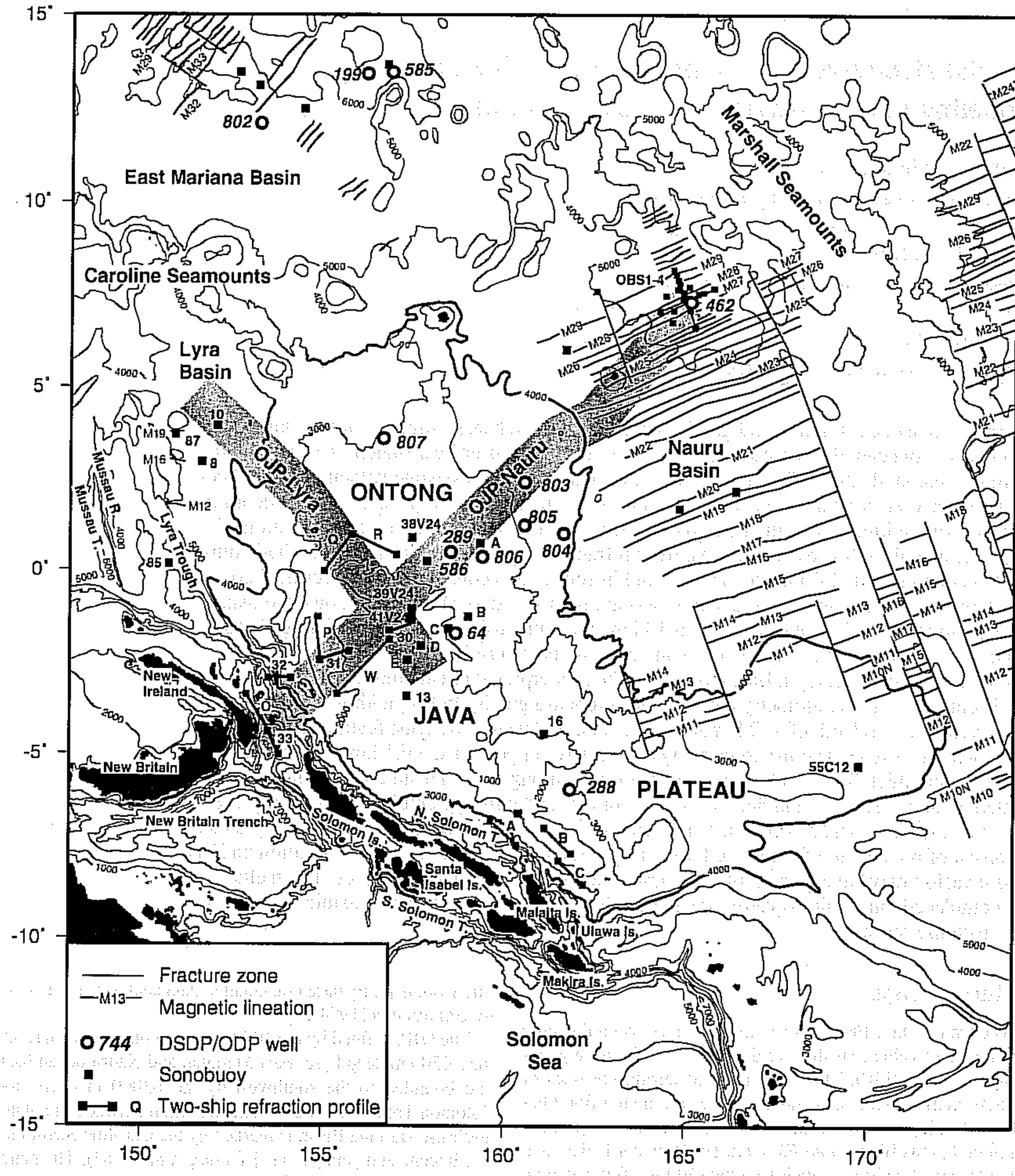

Figure 1. Simplified bathymetry from the *Iwabushi* [1994] and *Mohanan et al.* [1994]. Magnetic lineations and fracture zones are from *Cande et al.* [1989] and *Nakanishi et al.* [1992]. The 4000 and 3000 m contours (bold) outline the Ontong Java Plateau (OJP). Key for sonobuoy and refraction profile locations is given in Table 1. OJP-Nauru and OJP-Lyra, are crustal transects discussed in the text.

seamounts have a northeast trend (Plate 1), and thus are oblique to the interpreted seafloor spreading magnetic anomalies [*Nakanishi et al.*, 1992]. Offset of basement across these features suggests that they are fracture zones; however, they may also represent subparallel seamount chains.

Three of seven Deep Sea Drilling Project/Ocean Drilling Project (DSDP/ODP) plateau drill sites terminated in basaltic basement, with 149 m as the maximum penetration at Site 807 (Figure 1 and Plate 1). The 9 and 26 m of basalts drilled at Sites 289 and 803, respectively, consist of successive pillow lavas and thin flows, whereas pillow lavas and massive flows were recovered at Site 807 [*Shipboard Scientific Party*, 1975, 1991a, b]. All samples are tholeiitic basalts, broadly similar in chemical and isotopic composition to interpreted Cretaceous lavas on Santa Isabel, Malaita, Ulawa, and Makira islands, to Lower Cretaceous flows and sills in the Nauru Basin, and to dredged

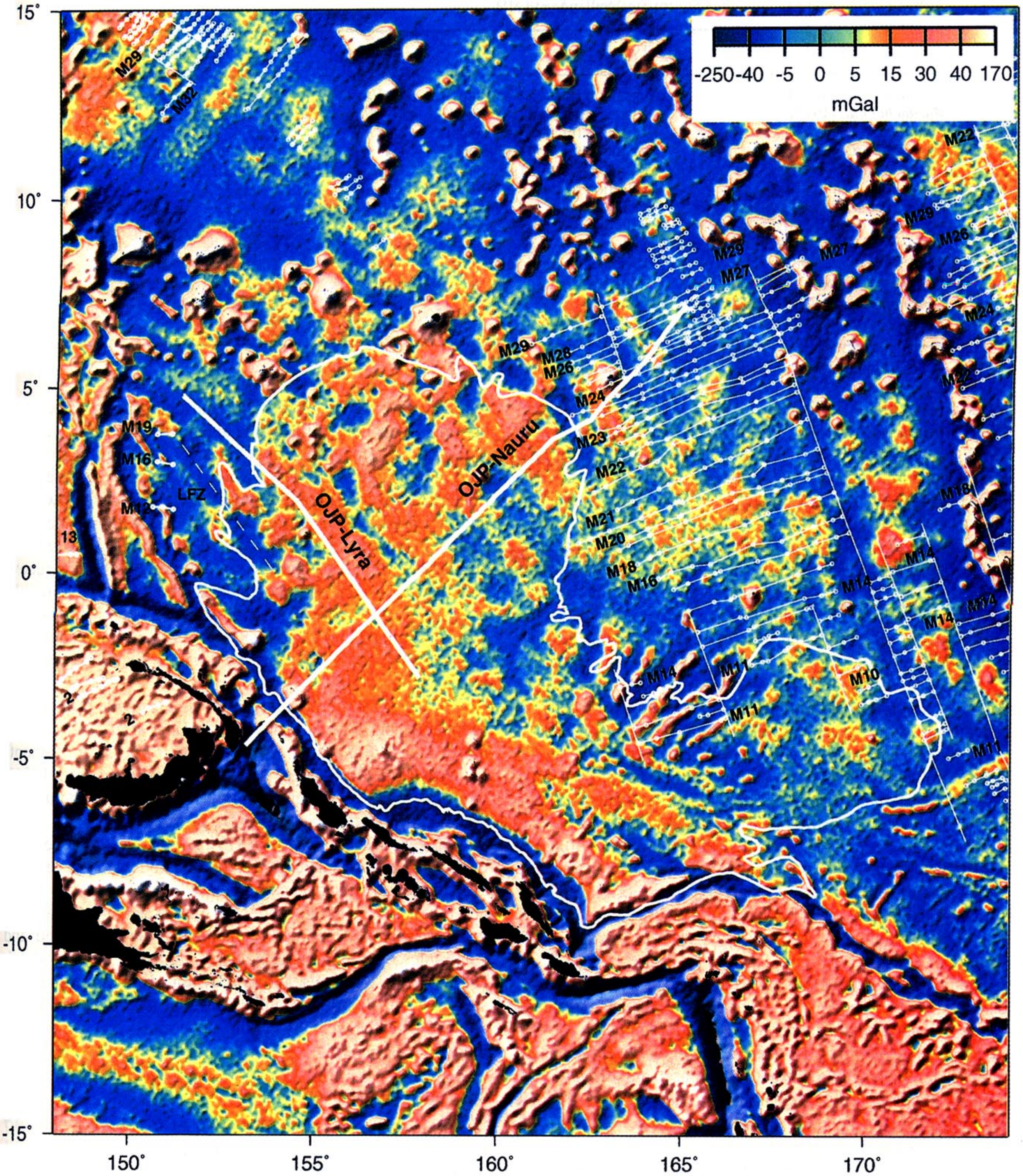

Plate 1. OJP, delineated by white line on the basis of bathymetry, as it appears on the satellite-derived gravity field (D. T. Sandwell and W. H. F. Smith, Marine gravity from satellite altimetry, digital file, version 7.2, available at anonymous ftp to baltica.ucsd.edu, 1995). Magnetic anomaly picks (white circles), magnetic isochrons (white lines connecting picks), fracture zones (unlabeled white lines) are from *Cande et al.* [1989] and *Nakanishi et al.* [1992]. LFZ, possible fracture zone trend in the Lyra Basin.

Table 1. Seismic Surveys and Scientific Drilling on OJP

Reference	Type of Survey	Map Abbreviation
Furumoto et al. [1970]	seismic refraction (two ships)	S-A, S-B, S-C
Murauchi et al. [1973]	seismic refraction (two ships)	30, 31, 32, 33
Furumoto et al. [1976]	seismic refraction (two ships)	P, Q, R, W
Hussong, et al. [1979]	seismic refraction (sonobuoy)	A,B,C,D,E,8,10,13,16,85,87
LDEO database	seismic refraction (sonobuoy)	xV24, xV28
Ewing et al. [1968]	reflection seismic	xV24, xC12
Winterer et al. [1971]: DSDP Leg 7	well, reflection and refraction (sonobuoy) seismic	64
Andrews et al. [1975]: DSDP Leg 30	wells and reflection seismic	288, 289
Larson et al. [1981]: DSDP Leg 61	well	462
Wipperman et al. [1981]	seismic refraction, OBS	OBS1-4
Moberly et al. [1986]: DSDP Leg 89	wells	462, 586
Lancelot et al. [1989]: ODP Leg 129	wells	800, 801, 802
Kroenke et al. [1991]: ODP Leg 130	wells	803, 804, 805, 806, 807

Location is given in Figure 1. OBS, ocean bottom seismometer.

rocks on the Manihiki Plateau [e.g., *Mahoney and Spencer*, 1991; *Mahoney et al.*, 1993; *Parkinson et al.*, 1995; *Petterson*, 1995; *Tejada et al.*, 1996]. At Sites 803 and 807, major and trace element abundances lie within the range for mid-ocean ridge basalts (MORB) values. The basalts have $^{40}Ar/^{39}Ar$ plateau ages of ~90 Ma (Site 803) and 122.4±0.8 Ma (Sites 289, 807) [*Mahoney et al.*, 1993], whereas biostratigraphic ages of basal sediments are 113 Ma at Site 803 and 117.5 Ma at Sites 289 and 807 [*Sliter and Leckie*, 1993]. Paleomagnetic and micropalaeontologic data have been interpreted in terms of rapid plateau formation, in <3 m.y. at the onset of the Cretaceous Normal Polarity superchron [*Tarduno et al.*, 1991].

Some drill samples, including a ~28 m thick single flow at Site 807 [*Shipboard Scientific Party*, 1991b], suggest a hotspot-like mantle source, very high degrees of melting, and lack of discernible age progression among sites. This has led to a model of rapid plateau construction above a plume head, the Cretaceous superplume [*Larson*, 1991], possibly associated with the Louisville hotspot [*Richards et al.*, 1989; *Mahoney et al.*, 1993; *Tarduno and Gee*, 1995]. Recently, the younger basalt age at Site 803 and from Santa Isabel has been interpreted as a second flood basalt event caused by separation of the original plume head and its trailing conduit; a second plume head then formed above the conduit and ascended to the surface [*Bercovici and Mahoney*, 1994].

Defining acoustic basement in the absence of multichannel seismic (MCS) data, over large parts of the plateau and flanks is problematic due to its smoothness [e.g., *Ewing et al.*, 1968], and the presence of intrabasement reflectors [e.g., *Hagen et al.*, 1993] which possibly represent "real" basement. *Hagen et al.* [1993] suggested that intrabasement reflectors may indicate multiple phases of volcanism, thus creating several "basement" surfaces. This may explain the discrepancy in ages between basalts drilled at Sites 803 and 807, where intrabasement reflectors are common in the vicinity of the former and absent around the latter.

2. Crustal Velocities

2.1. Sediment

We have compiled sonic log (Sites 586, 803, 805-807) and laboratory core measurements (Sites 288, 289, 804) to develop a velocity-depth function for the OJP sediments (Figure 2). All logs were normalized using lithostratigraphic units to Site 807 depth. No correction for compaction was made because the sediment layer is <1.5 km thick, and compaction effects are minimal, especially in limestone and chert. The upper seismic sequence was represented by a linear relationship, $v=1.6+1.0z$ km/s. A constant velocity, 3.5 km/s, is chosen to represent the Eocene to lower Campanian limestone layers of the lower sequence [*Shipboard Scientific Party*, 1991b], with a velocity inversion in the thin basal limestone. The function is based mainly on Site 807 which has the best log coverage and which recovered the thickest and most complete sediment section [*Kroenke et al.*, 1993]. In view of the similarity between Site 807 and Malaita stratigraphies [*Petterson*, 1995; also personal communication, 1996], ~1500 km apart, we consider Site 807 representative for much of the OJP.

Wide-angle profiles, mainly sonobuoys, provide data on both sediments and upper igneous crust [*Houtz et al.*, 1970; *Shipboard Scientific Party*, 1971; *Hussong et al.*, 1979; Lamont-Doherty Earth Observatory (LDEO), unpublished data, 1991], whereas images of the middle and lower crust are exclusively from two- or three-ship refraction profiles [*Furumoto et al.*, 1970, 1976; *Muruachi et al.*, 1973]. All profiles have been reduced by the slope-intercept method. Sonobuoy data reveal that a major sediment sequence boundary, characterized by chert, and the top of the basalt produce refracted arrivals. If these interfaces were not recorded, their depths were estimated and slope-intercept solution recalculated. A linear velocity-depth curve, $v=1.7+1.0z$ km/s, above a 3.4 km/s layer appears representative for all profiles, indicating that sonobuoy data may overestimate the velocity in the uppermost 100 m of sediments.

2.2. Igneous Basement

Basalt core velocities at Sites 289 and 807 average 5.43±0.7 km/s and 5.2±0.8 km/s, respectively, but refer commonly to homogenous hand samples of solid basalts and thus are not representative of the entire extrusive unit. The Site 807 sonic log is incomplete [*Shipboard Scientific Party*, 1991b]; however, resistivity, density, and gamma ray logs reveal a cyclicity presumed to represent porosity variations associated with flow heterogenities and boundaries, comparable with logs in flow basalt units drilled at ODP Sites 642 and 917 on the Vøring and East Greenland volcanic margins, respectively [*Planke*, 1994; *Planke and Cambray*, 1997]. The cyclicity suggests comparable physical properties of the upper extrusive crust in these LIPs. At Site 642 the central 7 m of each homogeneous massive 16 m thick basalt flow has a constant velocity of ~5.25 km/s, comparable

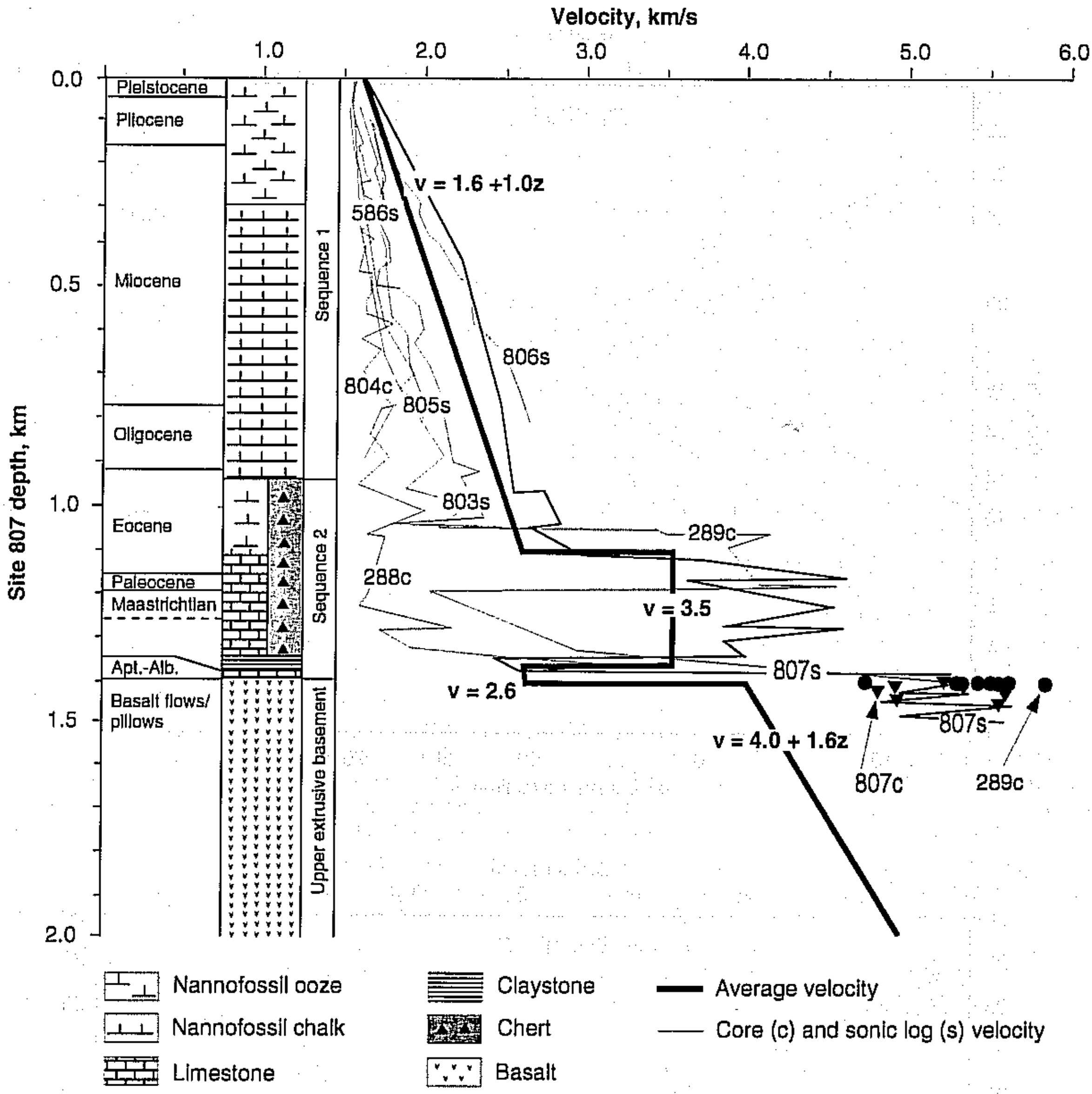

Figure 2. Velocity-depth plots from DSDP/ODP sites normalized to lithology from ODP Site 807. The core values and digitized sonic logs were smoothed before plotting. The average velocity-depth function (shown in bold) is based on core, sonic log, and sonobuoy data. Note that the thin Aptian-Albian low-velocity layer is only resolvable in well data.

to core velocities at Sites 289 and 807, contrasting with the 4.1±0.8 km/s average log velocity and the 4.5 km/s average refraction velocity at Site 642 [*Planke and Eldholm*, 1994]. In fact, the OJP's uppermost basement also exhibits low refraction velocities, ~4.0 km/s, gradually increasing to ~5.0 km/s at ~0.6 km depth. Hence a 4.5 km/s average velocity was chosen for this region of rapidly increasing velocity in uppermost basement (Figure 2). This "layer" velocity accounts for both interbedded sediments and porosity variations within single lavas units.

Refraction data from the middle and deep crust have been analyzed and interpreted in four steps. (1) Published solutions were duplicated based on published water depths, apparent velocities and intercept times. (2) The upper crustal velocity structure was analyzed, and in most cases, new average velocities were introduced, based on the more detailed model illustrated in Figure 2. (3) Published travel time graphs were evaluated to determine the quality of identified "straight-line" refractor segments according to number of first and later arrivals. Where applicable, alternative refractors were interpreted using some high-velocity arrivals. (4) The new solutions were ray traced to satisfy published arrivals and intercept times.

The modified water depths and refined upper crustal model (Figure 2) introduce only minor changes in lower crust and Moho refractor depths. We note, however, that in the *Furumoto et al.* [1970, 1976] profiles the Moho refractors commonly have been interpreted from second arrivals, and that some refractors appear to be constrained by precritical arrivals. We have therefore examined whether some distal offset first arrivals may actually originate from Moho, thus yielding a Moho depth more consistent with gravity calculations.

For profiles A, B, C [*Furumoto et al.*, 1970], and *R* [*Furumoto et al.*, 1976], the above procedure resulted in some modification of the lower crustal layer without major changes in crustal thickness. The key profiles, however, are P, Q (Plate 2) and W of *Furumoto et al.* [1976]. No 8+ km/s first arrivals were interpreted in profile P. An 8.1 km/s apparent velocity fits the

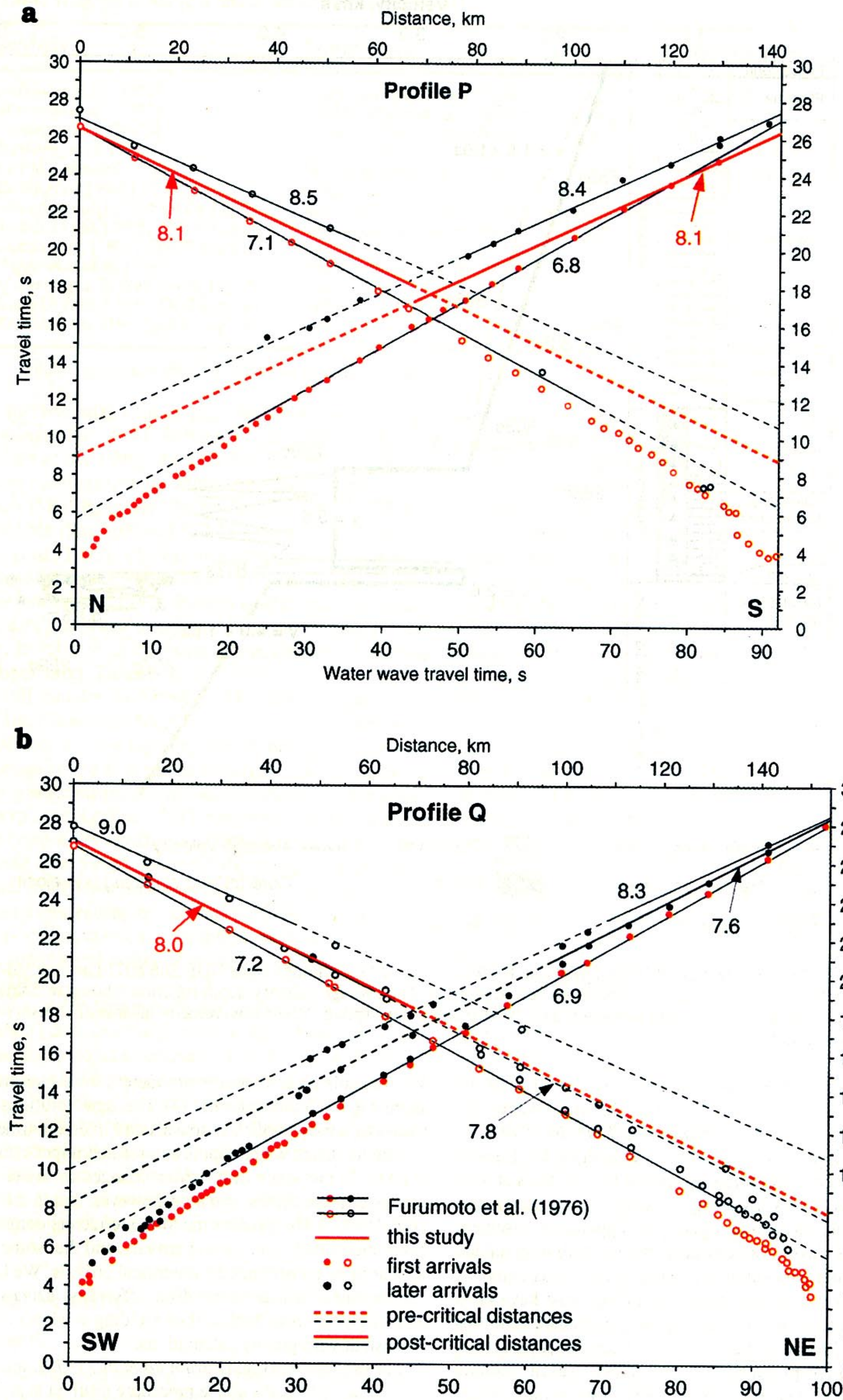

Plate 2. Travel time plots of seismic refraction profiles P and Q from *Furumoto et al.* [1976]. Reinterpreted Moho refractors are shown in red.

three far arrivals recorded at the north end. A reversed refractor of the same velocity through the reciprocal point fits the distal offset arrival recorded at the south end reasonably well, yielding a Moho depth of 28.6 km. In profile Q all refractors with apparent velocities >7.6 km/s are based on later arrivals; in fact, the 8.3 and 9.0 km/s apparent Moho velocities refer only to one and three postcritical second arrivals, respectively. We have therefore excluded these refractors and assign the 7.6 and 7.8 km/s apparent velocities to Moho, noting that the latter becomes slightly greater, 8.0 km/s, without the precritical arrivals (Plate 2). A new reversed velocity of 7.8 km/s and one new intercept time also eliminate the 5 km westward dip interpreted for the 7.7 km/s refractor by *Furumoto et al.* [1976]. Ray tracing and reciprocal points of the lower crustal 6.9 km/s refractor in profile W reveal unresolvable ambiguities. Thus we have used the original lower crustal model of *Furumoto et al.* [1976], with corrections for the 0.7 km increased water depth at far southwest end of the profile.

The 4.5 km thick upper crust contains a zone of rapidly increasing velocity with depth (Figures 2 and 3). The average velocity is 5.4 km/s. The middle crust exhibits 5.8-6.3 km/s velocities and an average thickness of 6.5 km. Profiles A, B, D, E, and 13 [*Hussong et al.*, 1979] and 31 [*Murauchi et al.*, 1973], on the central and western plateau, respectively (Figure 1), yielded 6.5-6.7 km/s velocities at depths equivalent to our middle crust (Figure 3 and Table 2). The 6.5-6.7 km/s refractors do not appear to be regional features because profiles 30 and W, in the same area, show a 5.4-5.6 km/s layer above 6.1 and 6.9 km/s refractors. *Hussong et al.* [1979] noted that the ~6.6 km/s refractors, commonly the deepest detected, are poorly determined, and therefore used results of *Furumoto et al.* [1976] in their OJP model (Figure 4). This velocity might originate from precritically reflected arrivals; thus the middle crust velocity is mainly based on work by *Furumoto et al.* [1970, 1976].

The lower crust velocities range from 6.9 to 7.5 km/s with an average of 7.1 km/s. The average Moho velocity is 8.0 km/s. Moho, however, is poorly constrained in all profiles, and a 7.6 km/s layer was recorded only in profiles R and Q [*Furumoto et al.*, 1970, 1976; *Murauchi et al.*, 1973]. The maximum Moho depth of 43 km refers to profile Q [*Furumoto et al.*, 1976].

Thick OJP crust is flanked by interpreted Jurassic and Early Cretaceous oceanic crust in the Nauru, East Mariana, Pigafetta, and Lyra Basins (Figure 1). However, contrary to Jurassic ages predicted by magnetic anomalies, drill sites recovered Lower Cretaceous rocks from the first three basins, in addition to Jurassic rocks from one site in the southeastern Pigafetta Basin. No drill sites have been drilled in the Lyra Basin. Furthermore, two out of three tentative seafloor magnetic anomaly interpretations in the Lyra Basin [*Taylor*, 1978] correlate spatially with positive topographic features and gravity anomalies, possibly seamounts (Figure 1 and Plate 1). For each adjacent basin, we have compiled published velocity profiles and made simplified crustal models (Figure 5). Similarities in acoustic basement character, age (130-110 Ma), and composition of drilled Lower Cretaceous basalts from OJP and the basins suggest a common origin [*Mahoney et al.*, 1993; *Castillo et al.*, 1994].

Smooth basement in the Nauru Basin corresponds to Lower Cretaceous basalt flows, drilled at Site 462, with a velocity >3.6 km/s [*Shipley et al.*, 1993]. Ocean bottom seismometer (OBS) profiles reveal a low-velocity zone below the basalts [*Wipperman et al.*, 1981], which probably consists of Lower Cretaceous sediments, and *Shipley et al.* [1993] infer up to 1.7 km of volcanics on oceanic crust in the northern basin, thinning to ~100 m south of 2.5°N. The ~9 km thick crust in Figure 5 includes a 3.7 km thick upper crust comprising Lower Cretaceous volcanics, older sediments, and oceanic layer 2. We arrive at this thickness by assuming that the layer 2-3 transition coincides with a velocity of 6.5-6.6 km/s [e.g., *Maynard*, 1970; *White et al.*, 1992]. A thickness estimate of the overlying Cretaceous rock unit (LK in Figure 5) may be obtained by assuming a ~5.2 km/s velocity for the top of Jurassic layer 2 [*Houtz and Ewing*, 1976].

A similar procedure is used for the East Mariana and northwest Pigafetta Basins where seismic profiles record intrabasement features [*Whitman* 1986; *Abrams et al.*, 1992, 1993; *Shipley et al.*, 1993]. Acoustic basement, horizon B of *Ewing et al.* [1968], correlates with the Lower Cretaceous flows/sills that overlie Jurassic/Lower Cretaceous sediments and oceanic crust in the East Mariana Basin and part of the Pigafetta Basin [*Abrams et al.*, 1992, 1993]. If we only consider the Jurassic crust, layers 2 and 3 correspond to the old Pacific crust *of White et al.* [1992] in the Nauru, East Mariana, and Pigafetta Basins (Figure 5).

Horizon B extends into the Lyra Basin, where seismic sediment structure is similar to OJP [*Ewing et al.*, 1968; *Kroenke*, 1972]. Upper and middle crustal velocities are only reported from the four sonobuoy profiles in Figure 5, but velocities and layer thickness vary [*Hussong et al.*, 1979]. The ~2 km thick 4.6-5.4 km/s layer appears to be a normal layer 2, except in profile 10 where the >4 km thickness resembles Nauru Basin crust (Figure 5). Hence it is not possible to make a representative velocity-depth curve. After correction for sediment loading [e.g., *Renkin and Sclater*, 1988], depth to top of oceanic crust in the Lyra Basin is ~5 km. Thus the Lyra Basin may be of Early Tertiary age, or more probably, the ~1 km depth anomaly and the presence of a reflector resembling horizon B indicate an extrusive cover similar to the Nauru Basin. A genetic relationship to the Nauru Basin and the OJP is suggested by the gravity lineaments which in the Lyra Basin (e.g., LFZ, Plate 1) trend nearly parallel to fracture zones east of the plateau.

3. Ontong Java Plateau Crust

The central and eastern plateau have anomalies from 0 to +25 mGal increasing to +25 to +75 mGal in the west and southwest, while -200 to +150 mGal anomalies dominate the Solomons collisional zone. Adjacent basins are typically characterized by 0 to +25 mGal anomalies (Plate 1 and Figure 6). A 19-23 km thick plateau crust was calculated by *Rose et al.* [1968], using the two-dimensional (2-D) gravity algorithm of *Talwani et al.* [1959] on a two-layer model with densities of 2870 and 3330 kg/m^3. Applying the same method and densities, *Furumoto et al.* [1976] noted that their thickest refraction crust would produce a -200 mGal anomaly. *Sandwell and Renkin* [1988] used the satellite-determined geoid-height-to-topography ratio to calculate a maximum Airy compensation (Moho) depth of 25 km, and *Schubert and Sandwell* [1989] calculated an average Airy thickness of 12.24 km for the plateau, defined by the 4.65 km depth contour, from satellite altimetry data.

We have constructed two intersecting crustal transects from the OJP to the Nauru and Lyra Basins by projecting velocity profiles within a ~100 km wide zone (Figures 1 and Plate 1). The transects were located (1) to obtain optimal deep and shallow crustal coverage, (2) to avoid seamounts and other features unrelated to the emplacement of the OJP, (3) to avoid areas introducing major 3-D gravity effects, and (4) to include the well-

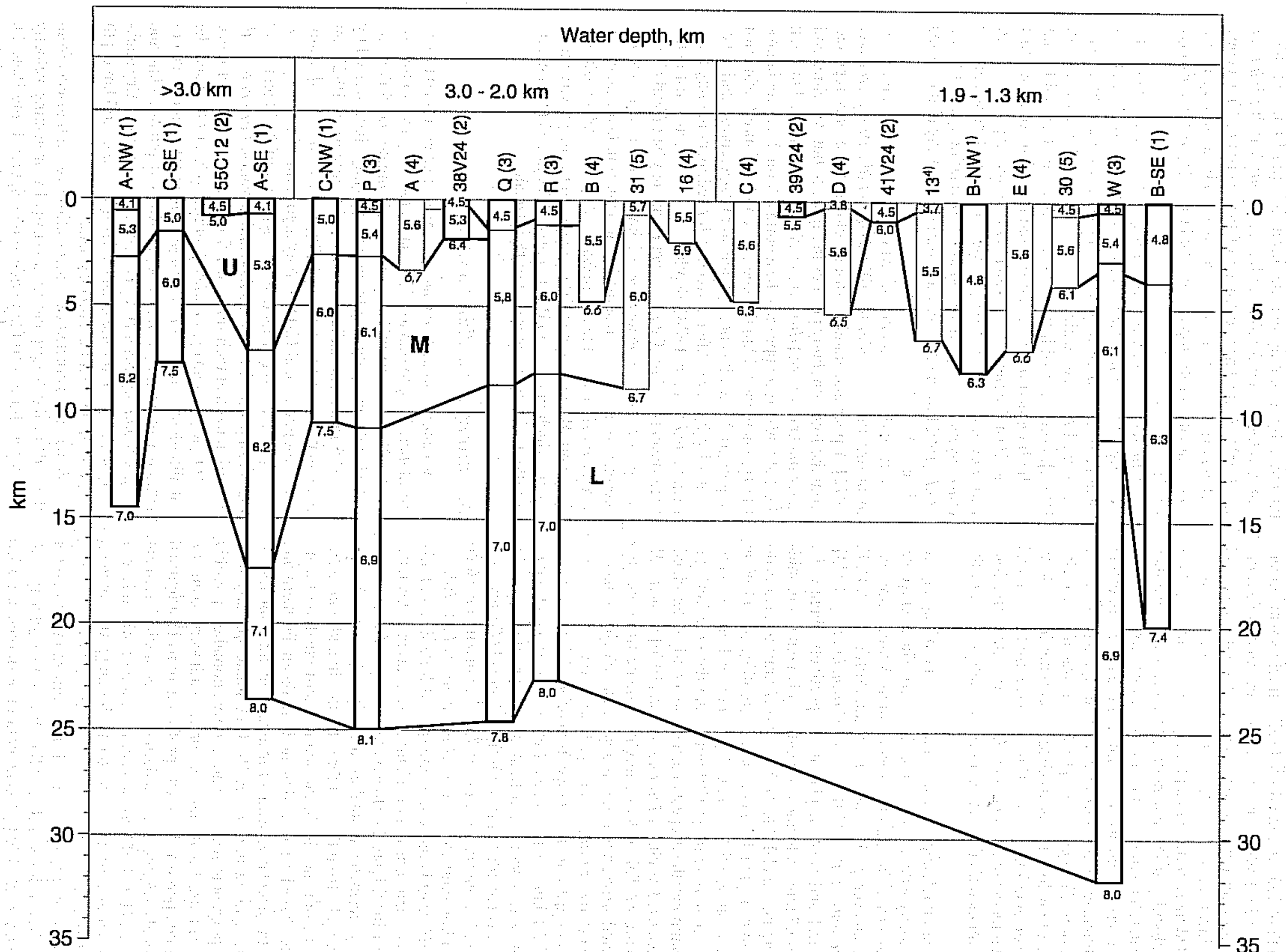

Figure 3. Layer velocities and thickness for all reinterpreted (thick columns) and originally published (thin columns) refraction profiles on the OJP, grouped according to water depth. U, upper crust; M, middle crust; L, lower crust. Anomalous 6.5-6.7 km/s middle crust velocities in italics. Sources are (1) *Furumoto et al.* [1970], (2) *Houtz et al.* [1970], (3) *Furumoto et al.* [1976], (4) *Hussong et al.* [1979], (5) *Murauchi et al.* [1973].

Table 2. OJP Average Velocities and Layer Thicknesses

Crustal Unit	Number of Profiles	Average Velocity, km/s	Velocity Range, km/s	Average Thickness, km	Thickness Range, km
Unit I: Upper crust	20	5.4	4.8-5.7	3.7	1.5-8.0
Transition zone	-	4.5	-	0.6	-
Above 6.1 km/s refactor	14	5.3	4.8-5.6	3.5	1.5-8.0
Above 6.6 km/s refractor	6	5.6	5.5-5.7	4.8	2.4-7.0
Unit II: Middle crust	14	6.1	5.8-6.3	6.5	3.1-16.2
6.6 km/s refractor	6	6.6	6.5-6.7	-	-
Unit III: Lower crust	9	7.1	6.9-7.5	14.8	6.1-19.0
Moho	5	8.0	7.8-8.1	-	-

See Figures 2 and 3. Six refraction stations recorded 6.5-6.7 km/s velocities [*Murauchi et al.*, 1973; *Hussong et al.*, 1979], which are intermediate between the middle and lower crustal velocities for the majority of profiles and therefore are treated separately.

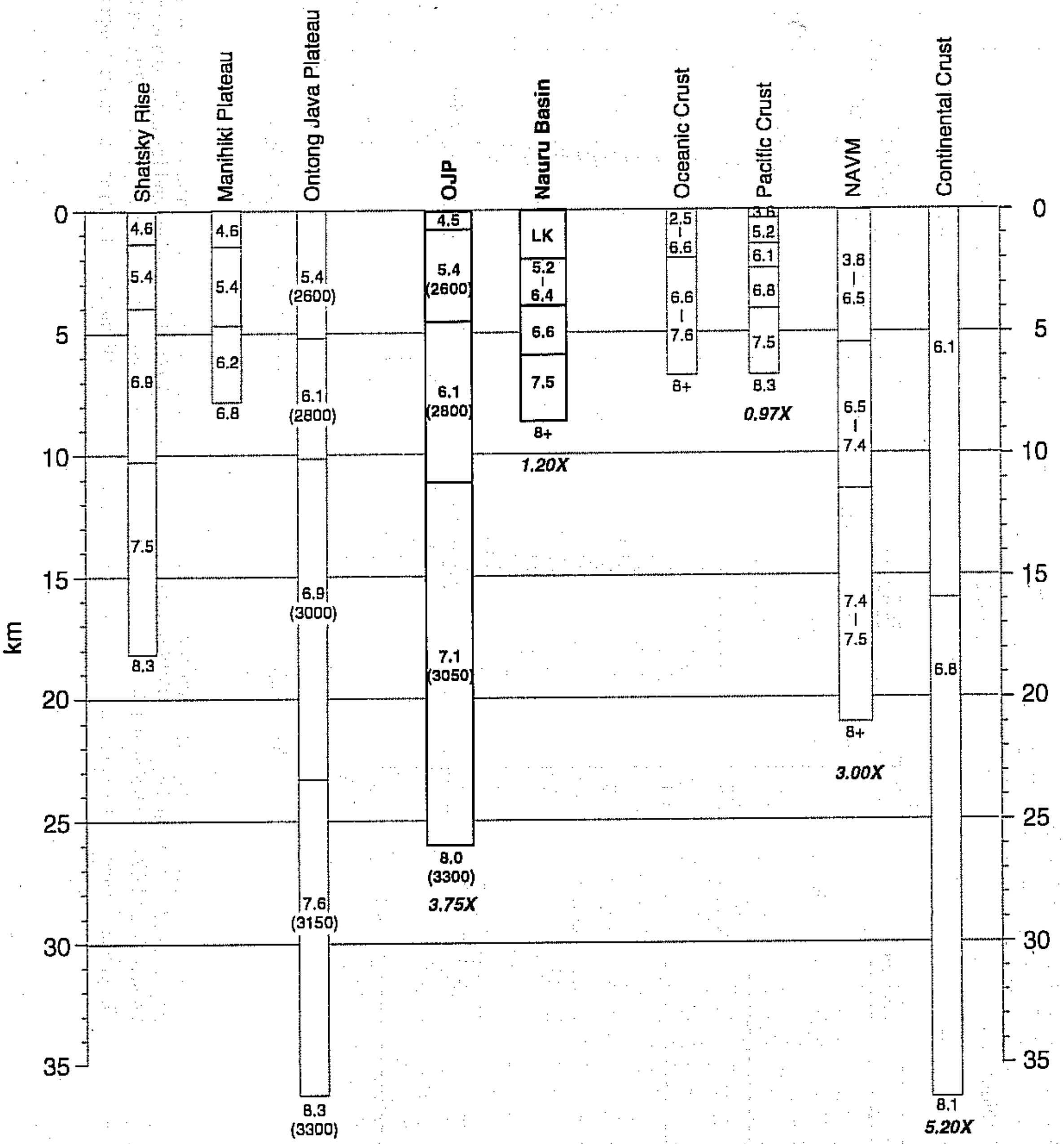

Figure 4. Average crustal columns for OJP, water depth < 3 km, and Nauru Basin compared to earlier models for OJP, Manihiki Plateau, Shatsky Rise, and Pacific crust [*Hussong et al.*, 1979], global average oceanic crust [*White et al.*, 1992], North Atlantic volcanic margin crust [*Eldholm and Grue*, 1994], and continental crust [*Nur and Ben-Avraham*, 1982]. Average densities used in gravity modeling are shown in parentheses. The factors shown at the base of the columns refer to multiples of normal oceanic crustal thickness. LK, Lower Cretaceous volcanics.

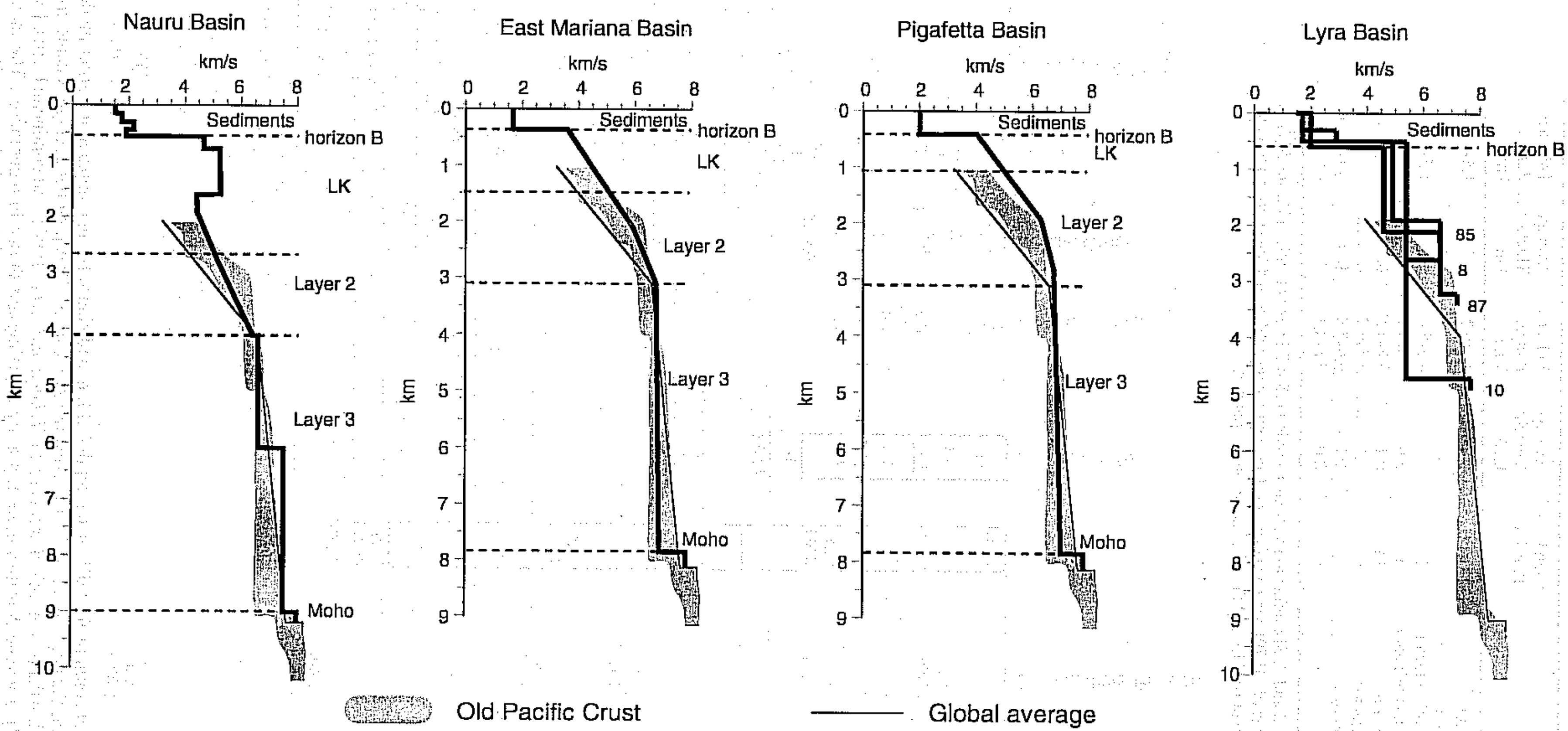

Figure 5. Typical velocity-depth functions (bold) for basins adjacent to OJP. Nauru Basin crust data are from *Houtz* [1976], unpublished LDEO sonobuoys, and OBS profiles [*Wipperman et al.*, 1981]. East Mariana and Pigafetta basins crust data are from unpublished LDEO sonobuoys and *Abrams et al.* [1993]. Only four sonobuoy stations [*Hussong et al.*, 1979] exist in the Lyra Basin. The global oceanic crust average and old (29-140 Ma) Pacific crust curves [*White et al.*, 1992] are shown for comparison. LK, Lower Cretaceous volcanics.

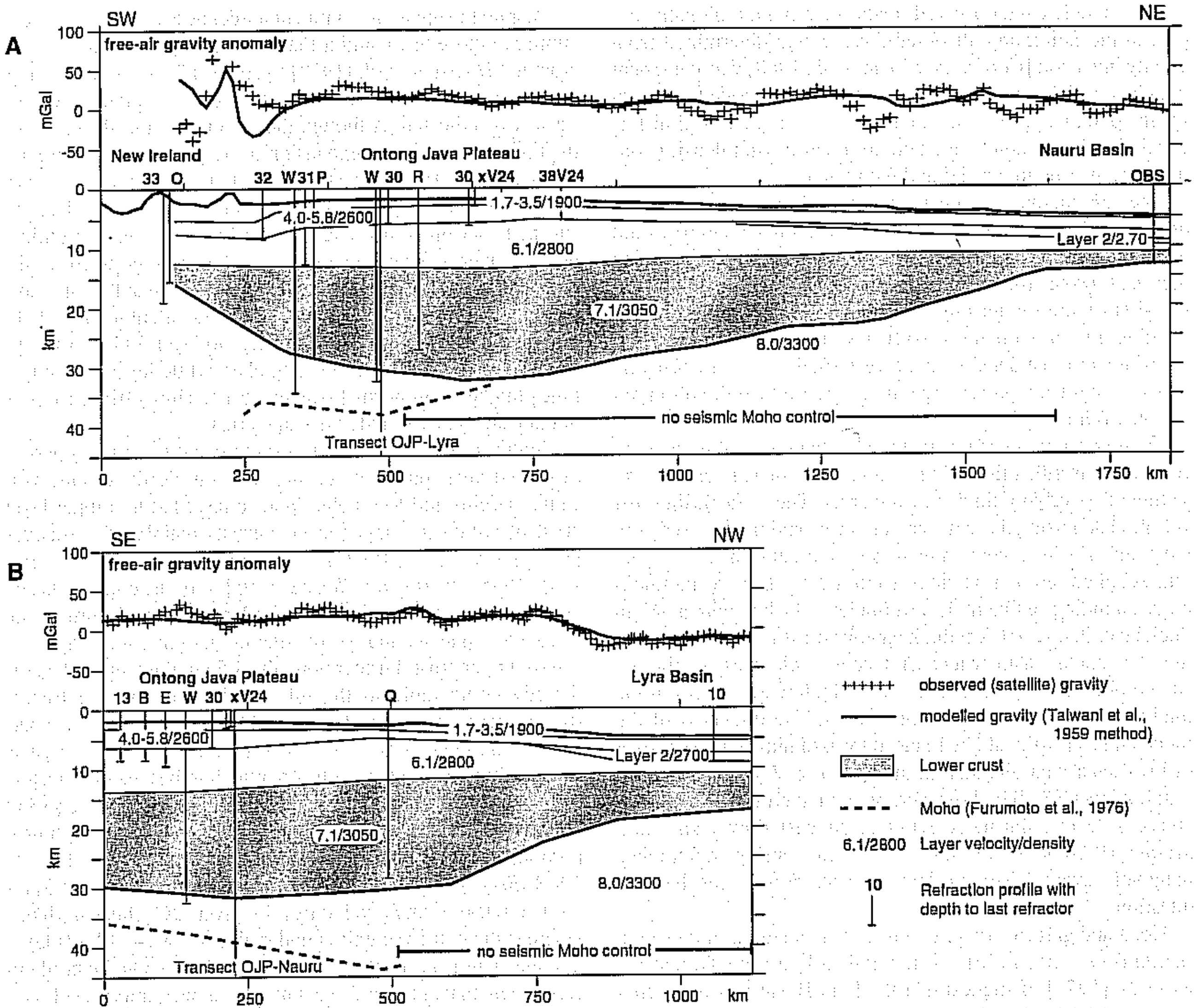

Figure 6. Multilayered crustal models along transects OJP-Nauru and OJP-Lyra (Figure 1), based on seismic refraction data and gravity modeling. Maximum depth yielded by each refraction profile projected onto the transects is indicated by a vertical bar and profile identification. See text for explanation.

determined oceanic crust in the Nauru Basin (Figure 5) [*Wipperman et al.*, 1981]. Then, 2-D gravity was modeled along each transect [*Talwani et al.*, 1959] by assigning crustal layer densities to the seismic model (Figure 6), keeping the Nauru Basin crust as a reference. A 45 km Airy compensation depth was used to account for the thickest crust of *Furumoto et al.* [1976]. The transect bathymetry and free-air anomalies are from ship track data and 2 arc min x 2 arc min gridded values from satellite alitmetry [*National Geophysical Data Center* (NGDC), 1988; *Wessel and Smith*, 1991; *NGDC*, 1995; D.T. Sandwell and W.H.F. Smith, digital file, available at anonymous ftp to baltica.ucsd.edu, 1995].

The 1900 kg/m^3 average sediment density is primarily based on Site 807 log and core data [*Shipboard Scientific Party*, 1991a, b], whereas upper and middle crust values of 2600 and 2800 kg/m^3 are based on general velocity-density relations for oceanic crust [*Ludwig et al.*, 1970; *Barton*, 1986], ophiolitic rocks and drill holes in oceanic [*Becker et al.*, 1989; Alt et al., 1993] and volcanic margin [*Planke*, 1994] crust. Lower crust has the same velocity, 7.1 km/s, as the 500 m thick gabbro at ODP Site 735 where average core density is 2950 kg/m^3 [*Iturrino et al.*, 1991]. Thickness of the OJP crust suggests that comparison should be made with other oceanic plateaus or regions with excessively thick igneous crust where both velocity and density measurements are available. Icelandic layer 3 gabbros have ~7.0 km/s velocities and 3000-3100 kg/m^3 densities [*Genshaft et al.*, 1992], while *Christensen and Salisbury* [1989], *Mengel and Kern* [1992], and *Kern* [1993] report gabbro values of 6.7-7.4 km/s and 2900-3100 kg/m^3. On the other hand, generation of thick igneous crust requires increased melt volumes, generally associated with an increase in asthenospheric temperature. As the potential temperature increases, the percentage of MgO in the melt increases, as do seismic velocities. *White and McKenzie* [1989] predicted values of 6.9-7.2 km/s and 2990-3070 kg/m^3 for high MgO basaltic crust. Hence a 3050 kg/m^3 density is arbitrarily assigned to the lower crust, above a 3300 kg/m^3 mantle following *Sandwell and Renkin* [1988]. The 6.9 and 7.6 km/s layers of *Hussong et al.* [1979] are similarly assigned densities of 3000 and 3150 kg/m^3, respectively (Figure 4).

The modeled and observed fields correspond regionally, except in the Solomon collisional zone. Moho determined from gravity inversion [*Cordell and Henderson*, 1968] was consistent with that of the forward model. No attempt was made to account for the dynamic effects of subduction or to match local differences which we largely attribute to upper crustal density contrasts from structural relief and bathymetry.

The >36 km thick crust determined by *Furomoto et al.* [1976] on the basis of seismic refraction data (Figures 4 and 6) would result in a ~50 mGal lower regional level over the plateau along transect OJP-Nauru, and the ~43 km deep crustal root in transect OJP-Lyra would contribute a further lowering of ~60 mGal, yielding a total anomaly of ~110 mGal. In fact, a 3250-3300 kg/m^3 lower crust density, i.e., a very small or no crust/mantle density contrast, would be required to reproduce the observed regional field.

Figure 6 shows oceanic crust thicker than normal in the Lyra Basin. We ascribe this finding to thick, low-density basalts and sediments overlying thick oceanic crust. The Lyra Basin crust was modeled using the same crustal layers and thicknesses as in the Nauru Basin, except an increased sediment thickness, ~0.6 km, as suggested by seismic reflection data. The crustal thickness, excluding sediment, is nevertheless ~11 km, with a ~6 km thick layer 3B. A ~10 km thick igneous crust is also reported in the Venezuela Basin where Cretaceous volcanics overlie assumed older crust [*Diebold et al.*, 1981]; however, most of the thickness increase is in the upper crust. The thickness of the modeled layer 3B could be reduced by assuming a thicker cover of low-density basalts than in the Nauru or Venezuela Basins. A thick layer 3B may, however, be explained either by an increase of layer 3 during accretion as suggested by *Mutter and Mutter* [1993] or as a result of underplating during an intraplate magmatic event. Velocity data are necessary to resolve this question.

Horst and graben structures along the plateau-basin transition [*Erlandson et al.*, 1976] and the graben-like Lyra Trough led *Kroenke* [1972] to suggest that the Lyra Basin is a foundered part of OJP. The few single-channel seismic profiles available [e.g., *Ewing et al.*, 1968; *Winterer et al.*, 1971] suggest a much more continuous smooth basement reflector compared with the variable Nauru Basin extrusive cover [*Shipley et al.*, 1993]. Thus the entire Lyra Basin may be covered with postemplacement extrusives and thus forms a contiguous part of the Greater Ontong Java LIP.

The main difference between our three-layer OJP igneous crust and the four-layer crust of *Hussong et al.* [1979] is that ours has linear velocity-depth relations in the sediments and upper crust and lacks a 7.7 km/s lower crust (Figures 2 and 3 and Table 2). As a result, the average crustal column for the central plateau (Figure 4), i.e., water depths <3 km, is ~10 km thinner than the crust of *Hussong et al.* [1979]. The crustal model in Figure 6 has a thick 7.1 km/s lower crust and a Moho depth of 32 km beneath the central plateau. It is 5-10 km deeper than the homogenous, 2800 kg/m^3, crust of *Sandwell and Renkin* [1988]. The difference is due to the intracrustal density distribution and a higher average crustal density, 2860 kg/m^3, in the layered model. On the other hand, the *Furumoto et al.* [1976] crustal model is 5-15 km thicker (Figures 4 and 6). Using data from a seismic array in Micronesia, *Richardson and Okal* [1996] calculate the Moho depth benath OJP to be ~35 km. A similar depth has been estimated from OBS data on the SW part of the plateau [*Miura et al.*, 1996; also personal communication, 1996].

Despite the apparent continental thickness of OJP, velocities similar to those of Manihiki Plateau, Shatsky Rise, and oceanic crust led *Hussong et al.* [1979] to suggest "expanded" oceanic crust and an OJP "expansion" factor of 5.0 relative to oceanic crust was indicated. Although the factor is reduced to 3.75 by the thinner crust, linear expansion of oceanic crust does not reproduce the OJP velocity structure (Figure 4). *Mutter and Mutter* [1993] have argued that variations in total thickness of "normal" oceanic crust result primarily from changes in thickness of layer 3. They suggested that the zero-age depth to the layer 2/3 boundary is relatively invariant. Our results imply that the upper and middle crust of OJP are equivalent to layer 2 (11 km), and the lower crust is equivalent to layer 3 (15 km). This suggests that layer 2 is 5 times thicker, while layer 3 is only 3 times thicker than normal oceanic crust. Thus OJP crust does not appear to be expanded oceanic crust.

Studies of regional velocity structure and seismic properties of drilled lavas on thick oceanic crust on North Atlantic volcanic margins and Kerguelen Plateau have led to a three-layer crust of which the upper layer consists mainly of extrusives [*Eldholm and Grue*, 1994; *Planke and Eldholm*, 1994; *Charvis et al.*, 1995; *Operto and Charvis*, 1995]. OJP drilling results and velocity structure, together with similarities to the Nauru Basin, Kerguelen Plateau, and North Atlantic margins, support an extrusive upper crust. Furthermore, the 4.5 km thick upper layer is directly comparable to the ~4 km thick exposed extrusive (basement) basalts on Malaita Island [*Petterson*, 1995; also personal communication, 1996]. The middle, 6.1 km/s, crust is more ambiguous. In the North Atlantic, intrabasement reflectors have been associated with velocities as high as 6.0-6.5 km/s but only in the deepest seaward dipping wedges where a high proportion of dikes is inferred [*Eldholm and Grue*, 1994]. ODP Site 504B [*Becker et al.*, 1989; *Shipboard Scientific Party*, 1993] recorded 6.0-6.5 km/s velocities in layer 2C sheeted dikes, whereas the extrusive unit is underlain by a 5.2-6.4 km/s layer 2 in the Nauru Basin [*Wipperman et al.*, 1981]. On the northern Kerguelen Plateau, a 6.2-6.3 km/s layer was interpreted as a seaward extension of a plutonic complex observed on the Kerguelen Islands [*Charvis et al.*, 1995]. Other refraction profiles, both on the north and south Kerguelen Plateau, show a velocity discontinuity between a 4.8-5.8 km/s layer, interpreted as extrusive, and a 6.4-7.1 km/s layer, interpreted as a layer 3 equivalent [*Charvis et al.*, 1995; *Operto and Charvis*, 1995]. Thus a predominantly intrusive nature for the middle crust on the OJP is suggested.

The 7.1 km/s lower crust at OJP differs from the lower crust interpreted on Kerguelen Plateau. The northern Kerguelen Plateau features a thick layer 3-like lower crust with velocities of ~7.4 km/s overlying Moho [*Charvis et al.*, 1995]. The southern plateau is characterized by a low-velocity zone, 6.7 km/s, which has led *Operto and Charvis* [1995] to suggest a continental affinity for this part of the plateau. Also the lower crust on volcanic margins, e.g., North Atlantic and U.S. East Coast margins, is characterized by 7.2-7.7 km/s velocities [*Eldholm and Grue*, 1994; *Holbrook et al.*, 1994a, b]. Thus the OJP lower crust appears intermediate between a low-velocity (~6.7 km/s) "possibly continental" and high-velocity (7.2-7.7 km/s) "oceanic" nature, and its crustal velocity structure is not directly comparable to either normal oceanic or continental crust.

4. LIP Emplacement

Crustal composition depends on plate tectonic setting and lithospheric configuration during melting and migration of mantle material in the asthenosphere and lithosphere. This leads to emplacement end-member models for flood basalts, e.g., plate boundary or intraplate settings. An oceanic intraplate LIP intrudes preexisting oceanic crust and presumably sandwiches it between extrusive and underplated layers emplaced during LIP construction. In contrast, the entire crust is emplaced during the transient event in a plate boundary LIP [cf. *Saunders et al.*, 1996].

Compared to other oceanic LIPs, where lavas commonly are emplaced subaerially or in shallow water [*Coffin and Eldholm*, 1994], drilled OJP lavas appear to have been extruded in deep water. *Sliter and Leckie* [1993] infer that Aptian limestone just above basement at Site 807 was deposited in 1.5-3 km water depth. A Cretaceous carbonate compensation depth (CCD) of 2.5 km [e.g., *Thierstein*, 1979] suggests that most sites on the rim of the plateau may encounter basalts emplaced below the CCD. Indeed, lack of carbonates at Site 803 [*Shipboard Scientific Party*, 1991a] may imply water depths >2.5-3 km soon after cessation of magmatism. Assuming a plate boundary setting, the age-depth analysis of *Coffin* [1992] implies emplacement of Site 289 in ~300 m water depth. Alternatively, emplacement on 0-40 Ma oceanic crust implies that the lavas cover a preexisting, basement relief of ~2.2 km [*Parsons and Sclater*, 1977]. Despite these equivocal boundary conditions, summarized by *Coffin and Gahagan* [1995], we infer that most of OJP was emplaced below sea level, large parts probably at 1.5-3 km depth.

The other giant oceanic plateau, Kerguelen, was formed in a narrow, young ocean basin [*Royer and Coffin*, 1992] and may in part be a volcanic continental margin fragment [*Operto and Charvis*, 1995]. Shatsky and Hess rises in the Pacific were emplaced at triple junctions [*Vallier et al.*, 1983; *Sager and Han*, 1993]. Geochemical evidence from basalts [*Alibert*, 1991; *Mahoney et al.*, 1995] and deep crustal seismic velocities [*Operto and Charvis*, 1995] suggest that the southernmost Kerguelen Plateau has affinities with continental lithosphere, unlike either OJP or adjacent ocean basin flood basalts [*Storey et al.*, 1988; *Mahoney et al.*, 1995]. Plate reconstructions and geochemical studies indicate that OJP was emplaced away from continental areas, although the pre-Aptian plate tectonic history is poorly constrained [*Nakanishi et al.*, 1992; *Mahoney et al.*, 1993; *Yan and Kroenke*, 1993]. The ~40 m.y. age difference between the older OJP basalts (~122 Ma, anomaly M1) and oceanic basement of adjacent basins indicates a probable intraplate setting with the M1 plate boundary near the present southern rim of the plateau. A plate boundary setting would imply that most of the conjugate LIP has been subducted. However, *Cloos* [1993] argues that oceanic plateaus with crustal thickness >30 km are not subductable, a concept consistent with the apparent early Miocene shutdown of the Solomon subduction complex [*Kroenke*, 1972, also personal communication, 1996]. Furthermore, similarity of basement basalts on Malaita, NE Santa Isabel, Ulawa, and parts of Makira islands (Figure 1) with those drilled on OJP suggests that these islands are obducted splinters of OJP [*Hughes and Turner*, 1977; *Kroenke et al.*, 1986; *Mahoney et al.*, 1993; *Petterson*, 1995; *Tejada et al.*, 1996]. Thus part of the original OJP exists within the collision zone.

An intraplate setting is also compatible with plateau elongation along Early Cretaceous fracture zone trends. *McNutt et al.* [1989] suggested that fracture zones control the surface manifestations of weak plumes such as the Marquesas plume. Therefore one or more fracture zones may have acted as a catchment for mantle melts and led to asymmetric emplacement of OJP (Figure 7). Lithospheric extension amplifies melt production [*White and McKenzie*, 1989; *Farentani and Richards*, 1994], and an oceanic plateau may form by a surfacing plume head which would quickly link to a spreading center through its potential influence on rift propagation [*Mahoney and Spencer*, 1991; *Small*, 1995]. Thus the LIP may be emplaced by combined on- and off-axis magmatism [*Gladczenko*, 1994; *Winterer and Nakanishi*, 1995]. However, the distance to the closest spreading axis was probably of the order of 1000 km (Figure 7), i.e., about one half plume diameter after lithospheric impingement [*White and McKenzie*, 1989; *Hill*, 1991].

We suggest that OJP was emplaced in a near-ridge setting. Part of the melt migrated to thinned lithosphere along the plate boundary (Figure 7). Thus, the oceanic plateau occupies an asymmetric position relative to the paleoplate boundary without requiring a major conjugate feature. The emplacement model comprises (1) plume impact on oceanic lithosphere near a plate boundary, links to a spreading center [e.g., *Mahoney and Spencer*, 1991; *Small*, 1995; *Larson*, 1997], (2) transient, voluminous emplacement of mantle melts through a multiple feeder system, including the contemporaneous plate boundary and fracture zones [cf. *McNutt et al.*, 1989], (3) construction of an oceanic plateau, mainly on one side of the plate boundary, i.e., part of the plateau is underlain by preexisting oceanic crust, probably 0-40 m.y. old, (4) contemporaneous, but less voluminous, flood basalt emplacement as sills, dikes and flows in adjacent ocean basins [e.g., *Shipley et al.*, 1993; *Castillo et al.*, 1994], (5) migration of plate boundary (ridge jumps) to southern plateau flank during waning stage of the transient magmatic pulse, (6) resumption of normal seafloor spreading as suggested by symmetric M series magnetic anomalies east of the OJP [*Nakanishi et al.*, 1992], and (7) secondary plume head volcanism at ~90 Ma [*Bercovici and Mahoney*, 1994]. The model produces a thick crust, laterally changing from entirely new Early Cretaceous crust to intruded Late Jurassic crust capped by Lower Cretaceous rocks. Sophisticated normal incidence and wide-angle seismic experiments would be required to test this hypothesis.

5. Nature of Lower Crust

The thickness and velocity of OJP's lower crust are difficult to explain by invoking only those processes associated with the accretion of normal oceanic crust. Factors such as preexisting oceanic crust and magmatic underplating; increased asthenospheric potential temperature and variations in composition; and synemplacement and postemplacement hydrothermal and metamorphic processes must all be considered.

Magmatic underplating below an intruded Late Jurassic oceanic crust is difficult to reconcile with the 7.1 km/s lower crust, because an underplated body should have a higher velocity/density than the lowermost normal oceanic crust, ~7.5 km/s [e.g., *Kempner and Gettrust*, 1982; *White et al.*, 1992]. The small density contrast between layer 3 cumulate gabbro and primitive MORB magma may not be a sufficient density filter for mantle melts to be underplated [*Herzberg et al.*, 1983]. Moreover, a velocity >7.1 km/s is expected due to high melting pressures caused by the crustal lid [*Kelemen and Holbrook*, 1995; *Farnetani et al.*, 1996]. Thus the observed 7.1-7.3 km/s velocities appear more appropriate for a plate boundary setting, where no preexisting crust would provide a density filter.

Voluminous melts were necessary to produce OJP [e.g., *Coffin and Eldholm*, 1994]. The existence of high-MgO picrites and komatiites in some LIPs [*Cox*, 1972; *Storey et al.*, 1991; *Larsen*

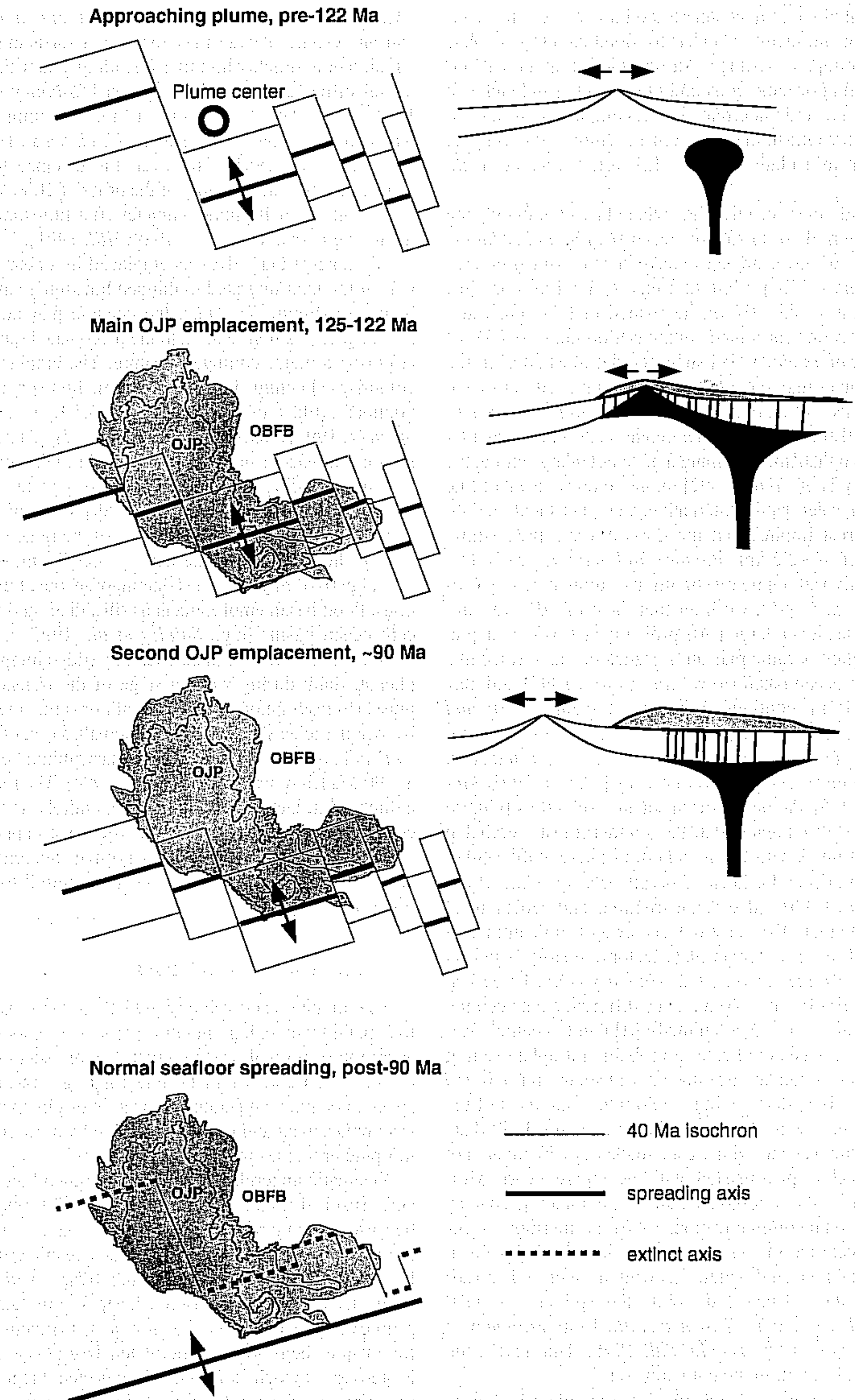

Figure 7. Sketch of the OJP LIP emplacement. Orientation of fracture zones and spreading axes is based on work by *Nakanishi et al.* [1992]. Note that the ages are approximate and refer to the 122 Ma $^{40}Ar/^{39}Ar$ basalt age at Sites 289/807 [*Mahoney et al.*, 1993] and Malaita and Santa Isabel islands [*Petterson*, 1995] and the ~3 m.y. emplacement period of *Tarduno et al.* [1991]. OJP, Ontong Java Plateau; OBFB, ocean basin flood basalts.

et al., 1994] implies elevated asthenospheric potential temperature. Furthermore, *White and McKenzie* [1989, 1995] showed that high MgO melts should generate a thick crust with a high average velocity. However, petrologic modeling shows that a temperature anomaly in excess of 300°C would be required to account for 7.5-7.6 km/s velocities in some LIPs, and thus we also need to consider fractionation processes during migration of a picritic liquid from mantle to surface [*Farnetani et al.*, 1996; *Neal et al.*, 1997]. If the picritic liquid path intersects a magma chamber containing already fractionated liquid, it will, depending on its density, either spread beneath the chamber with minimal mixing [*Sparks et al.*, 1980], or be trapped by the chamber roof [*Stolper and Walker*, 1980]. For the OJP, *Farnetani et al.* [1996] predict the presence of fractionated cumulates of mela-gabbro yielding 7.4-7.7 km/s velocities, in the lower crust, while in the middle crust, intrusions of olivine gabbro have seismic velocities of ~7.0 km/s. *Neal et al.* [1997] suggest that the lower crust contains a wehrlitic to pyroxenitic cumulate of density ~3250 kg/m^3. This might imply that the 8.0 km/s velocity upper mantle is actually a part of the lower crust. If this is the case, the gravity does not resolve the crust-mantle transition.

Dick et al. [1991] suggested that oxide-rich gabbro layers in ODP Hole 735B could be seismic reflectors. The very high densities of oxides, 4400-5200 kg/m^3, will increase density and velocity of host rocks, even though their content is limited to a few percent. Moreover, laterally continuous or even semicontinous thin lenses of high-velocity rock may act as seismic energy channels yielding refraction waves masking refractors from the lower velocity host rock [e.g., *Fountain et al.*, 1994].

An alternative mineral phase to olivine-rich cumulates or oxide-rich layers that would yield high velocities in mafic rocks is garnet. Garnet is not a stable phase in high Mg/Fe melts, e.g., basaltic to ultramafic bulk compositions, at pressures <~10 MPa and temperatures >700°C [*Kelemen and Holbrook*, 1995] and thus will probably not form in abundance during conductive cooling of anhydrous igneous rocks [*Ahrens and Schubert*, 1975; *Kelemen and Holbrook*, 1995]. However, effects of synemplacement and postemplacement hydrothermal and metamorphic processes on lower crustal velocities must also be considered. For example, depth to the lower crustal layer on North Atlantic margins may indicate a metamorphic facies boundary [*Eldholm and Grue*, 1994], and the nonvolcanic Galicia margin includes a layer of serpentinized peridote of 7.6 km/s velocity and 3260 kg/m^3 density [*Whitmarsh et al.*, 1993]. Increased serpentinization, which is fluid dependent, will decrease these values [*Kern and Tubia*, 1993]. The discovery of garnet-bearing mafic granulite xenoliths from the Kerguelen Islands suggests that the PT conditions in the lower crust of the plateau are appropriate for the formation of granulite-facies minerals and that the lower crust of the Kerguelen Plateau is composed of mafic granulites [*Gregoire et al.*, 1994, 1995]. Metamorphism of gabbro yields granulites and garnet granulites [e.g., *Bucher and Frey*, 1994] and OJP's lower crust lies within the garnet window corresponding to a 7.2 km/s velocity [*Furlong and Fountain*, 1986] (Figure 8). Metamorphism will only take place if the environment of the rocks changes, e.g., a temperature increase, introduction of fluids, or deformation. A temperature increase requires a magmatic event following emplacement of igneous crust. Such an event appears to have happened on the OJP, as suggested by the ~90 Ma basalts from Site 803 and Santa Isabel Island. The second magmatic event may have added enough heat to the preexisting gabbros to cause recrystallization. However,

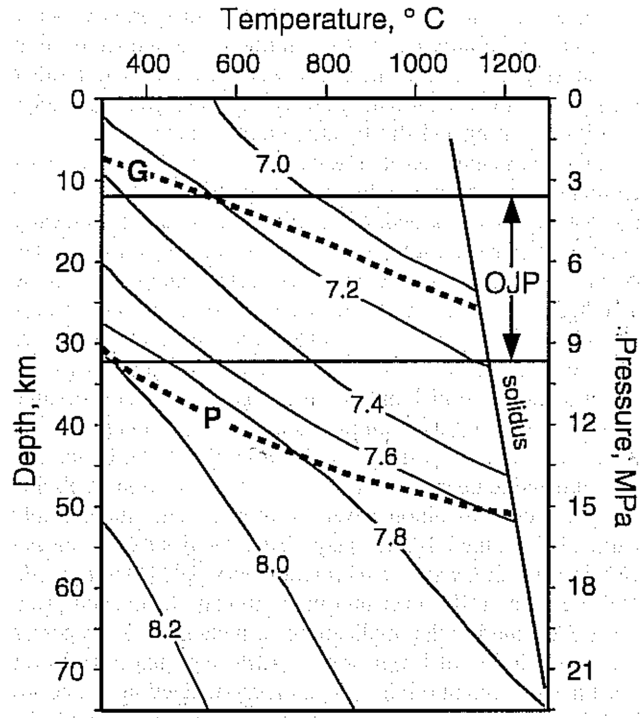

Figure 8. Gabbro-granulite-eclogite diagram for olivine gabbro composition [*Furlong and Fountain*, 1986]. Granulite and eclogite facies are indicated by the garnet-in curve (G) and plagioclase-out curve (P), respectively [*Wood*, 1984]. Light lines indicate seismic velocities (in km/s). The depth range of the OJP lower crust is shown, indicating that a granulite facies lower crust may explain the observed velocities.

for the gabbro→granulite transition to proceed at temperatures <~800°C, which might be expected after ~30 m.y. of cooling, fluid must be present [*Ahrens and Schubert*, 1975]. The fluid present or introduced must be CO_2-rich as water alone will form amphibolite [*Austrheim*, 1987]. This also suggests that metastability factors are of greater importance than compositional variations (e.g., high Mg/Fe basaltic bulk compositions). The presence of fluids in the lower crust is poorly constrained. If the greenschist facies is the limit [*Frost and Bucher*, 1994], the gabbros in the lower crust will remain anhydrous. On the other hand, if serpentinization takes place, seawater appears to reach the upper mantle. Evidence for fluids in the lower crust exists, especially in relation to shear zones [e.g., *Austrheim*, 1987; *Mével*, 1988; *Newton*, 1990] but also as a result of thermal contraction [*Dick et al.*, 1991]. Thus fluid-controlled metamorphism may be partly responsible for the observed velocities in the OJP lower crust.

Metamorphism resulting in seismic velocity changes may also take place due to deformation related to spreading, transform movements, and rift tectonics in the vicinity of spreading centers [*Mével*, 1988]. This process will facilitate fluid transport, although metamorphism also takes place in anhydrous conditions [e.g., *Mével*, 1988; *Cannat et al.*, 1991]. Oceanic gabbros tend to be strongly deformed [e.g., *Sinton and Detrick*, 1992]. Deformation begins in the crystal mush stage and appears to continue well below the solidus, when solid-state off-axis deformation takes place under granulite-facies metamorphic conditions [*Nicolas*, 1989; *Cannat et al.*, 1991]. In normal oceanic gabbro, temperature decreases to the stability condition for amphibolite-facies assemblages, while much thicker LIP crust appears to experience granulite-facies conditions within the garnet window [*Furlong and Fountain*, 1986], conditions maintained even at present (Figure 8). Thus the OJP lower crust, and that in other LIPs, may reflect deformation processes under granulite facies conditions during and shortly after emplacement.

If OJP was emplaced at or near a plate boundary, a crustal accretion process similar to that associated with a plate boundary hotspot may be envisaged. Moreover, synaccretionary and early postaccretionary deformation processes in the lower crust may occur. The main difference with normal spreading is the volume and thickness of the new crust. Hence the lower crust of an oceanic plateau will be exposed to higher pressure and temperature and therefore fractionation process yielding higher-density cumulates than at normal mid-oceanic ridges. Lowermost oceanic crust is composed of granulite-facies rocks, although garnets have not been observed [e.g., *Cannat et al.*, 1991]. Increasing pressure and temperature will favor recrystallization of olivine and plagioclase to clinopyroxene, orthopyroxene, and possibly garnet. In fact, the 7.1-2 km/s layer 3B velocities [*Kempner and Gettrust*, 1982; *Christensen and Salisbury*, 1989; *White et al.*, 1992] and the OJP lower unit may originate from such processes. Considering the implications of metamorphic processes for the evolution of lower crust in LIPs, e.g., that OJP lower crust may be rich in olivine, in part recrystallized into zones of high-velocity rock, we suggest that such processes should be considered in future modeling work.

6. LIP Dimensions

Parts of the OJP LIP along its southwestern flank have been obducted into and probably have been subducted beneath the Solomon collisional zone. OJP was thus originally larger than at present. Nonetheless, most of OJP probably did not subduct. Our OJP emplacement model (Figure 7) suggests that the LIP volume may have been larger than at present but not doubled as implied by a conjugate feature [*Larson*, 1991].

Coffin and Eldholm [1994] calculated the OJP crustal volume within the 4000 m contour. Although Figure 6 shows that the lower crust may be ~10 km thick beneath 4000 m water depth, the gravity model is fitted to a regional level on the plateau flank, and thus the crustal thickness is relatively poorly constrained. The calculated volume (Table 3) reflects the intermediate crustal thickness used in this study. Based on the integrated velocity/gravity model (Figure 6) of the OJP, we infer that the extrusive thickness varies from 4.5 km on the central plateau to ~2 km in the Nauru Basin. Thus the extrusive volume above 4000 m is the same as that of *Eldholm and Grue* [1994], while the total crustal volumes for off- and on-ridge setting, 44.4 $\times 10^6$ km^3 and 56.7 $\times 10^6$ km^3, respectively, are intermediate between the maximum and minimum values of *Coffin and Eldholm* [1994] (Table 3).

Coeval West Pacific extrusives in the Nauru, East Mariana, and probably the Lyra Basins, may all, or in part, belong to the greater OJP LIP. Therefore the crustal volumes are considered minimum values. However, no corrections have been made for coeval rocks within the Solomon subduction complex and possible OJP secondary head and tail volcanism [*Bercovici and Mahoney*, 1994], factors that would add and subtract to the volumes in Table 3, respectively.

7. Conclusions

Previous studies based on gravity modeling and refraction profiles yield a difference in maximum OJP crustal thicknesses of ~18 km. An evaluation of the available refraction data, gravity field, drilling results, and data from other LIPs yields a three-layer igneous crust. The main difference with the four-layer crust of *Hussong et al.* [1979], based on work by *Furumoto et al.* [1976], is that a relatively thick 7.1 km/s lower crust satisfies more constraints than their 6.9-7.6 km/s lower crust. Our model is consistent with the observed gravity. We interpret the OJP as composed of thick oceanic crust extending to a depth of ~32 km on the central plateau. The OJP crust is, however, not a linearly expanded normal oceanic crust nor does it contain only an expanded lower crust (layer 3) [*Mutter and Mutter*, 1993] but contains three distinct crustal units. Crustal volumes of 44.4 $\times 10^6$ km^3 and 56.7 $\times 10^6$ km^3 are calculated (Table 3) for off- and on-ridge emplacement settings, respectively.

In light of the anomalously thick Lyra Basin crust suggested by our modeling and the continuous smooth basement reflector we suggest that the basin has been covered by postemplacement extrusives and thus forms a contiguous part of the Greater Ontong Java LIP which also encompasses the Pigafetta, East Mariana, and Nauru Basins. We further suggest that the Greater Ontong Java LIP was emplaced by funnelling of melts to regions of thin lithosphere along the existing plate boundary [cf. *Small*, 1995; *Sleep*, 1996], and through preexisting, young oceanic lithosphere, possibly along fracture zones like the Lyra Fracture Zone (FZ). At the end of the transient event, the main injection zone migrated southwest and normal seafloor spreading resumed south of the plateau.

The emplacement conditions and geophysical parameters suggest that the LIP lower crust consists of ponded and fractionated primary melts. However, metamorphic granulite stability conditions within the OJP lower crust suggest that recrystallization processes may be responsible for the observed lower crustal velocities. Thus the OJP crustal composition should be considered in view of both primary magmatic and secondary hydrothermal and metamorphic processes.

Table 3. OJP Dimensions

	Area, $\times 10^6$ km^2	Extrusive Volume, $\times 10^6$ km^3	Total Crustal Volume, $\times 10^6$ km^3 Off-Ridge	On-Ridge
Main plateau (this study)	1.86	8.4	44.4	56.7
Eldholm and Coffin [1994]				
Refraction Moho[a]	1.86		49.0	61.3
Gravity Moho[b]	1.86		26.9	39.2
Eldholm and Grue [1994]	1.86	8.4		

Main plateau refers to water depths >4000 m.
[a]Igneous crustal thickness from *Furumoto et al.* [1976].
[b]Igneous crustal thickness from *Sandwell and Renkin* [1988].

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