Crustal structure and evolution of Mariana intra-oceanic island arc

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ABSTRACT

A new high-resolution velocity model of the Mariana arc-backarc system obtained from active-source seismic profiling demonstrates velocity variations within the arc middle and lower crusts of intermediate and mafic compositions. The characteristics of oceanic island arc crust are a middle crust with velocity c. 6 km/s, laterally heterogeneous lower crust with velocities of about 7 km/s and unusually low mantle velocities. Petrologic modelling suggests that the volume of lower crust composed of restites and olivine cumulates after extraction of the middle-crust should be significantly larger than observed, suggesting that part of the lower crust, especially cumulates, is seismically part of the mantle.

Keywords: Seismic structure, Intra-oceanic island arc, Crustal growth, Izu-Bonin-Mariana Arc

INTRODUCTION

Oceanic island arcs are regarded as the prime location of growth of andesitic continental crust. However, this conjecture has largely been based on petrologic and geologic studies (e.g., Rudnick, 1995) and direct seismological detection of andesitic continental crust beneath the arcs has remained elusive. Although a thick middle crust with a tonalitic (intermediate to felsic) velocity of 6 km/s (Christensen and Mooney, 1995; Kitamura et al., 2003) was detected in the northern Izu arc (NIA) (Suyehiro et al., 1996), the across Aleutian arc has little or no middle crust with 6 km/s P-wave velocity (Holbrook et al., 1999; Shillington et al., 2004). Understanding crustal growth requires not only knowledge of the distribution of tonalitic middle crust but also imaging velocity variations and heterogeneity within each crustal layer. Particularly, the velocities of the middle and lower crusts and upper mantle should vary progressively during crustal growth. During growth, the initial basaltic crust differentiates to produce an intermediate crust having higher SiO_2 content, and the residual dense mafic component of the arc crust should be removed either during arc growth, or during subsequent arc accretion events (e.g., Kay and Kay, 1993; Jull and Kelemen, 2001). The nature of the crustal differentiation and the return of mafic components into the mantle are keys to understanding crustal growth.

The Izu-Ogasawara (Bonin)-Mariana (IBM) oceanic island arc developed as an oceanic arc within purely oceanic crust, without continental collision (e.g., Karig and Moore, 1975; Hall et al., 1995). Subduction initiated at 50-40 Ma and an early initial arc with no backarc rifting formed the nucleus of what has since become Kyushu-Palau ridge (KPR), the present Mariana arc-backarc system (MABS) and NIA (Stern et al., 2003) (Fig.1, inset). Backarc spreading rifted off part of this early arc to form KPR, west of the now-inactive Shikoku and Parece Vela basins (PVB) at 30 or 29 Ma, and ceased at 15 or 16 Ma (Okino et al., 1994; Okino et al., 1998). The Miocene Mariana arc has been split into the currently active Mariana arc (MA) as a non-volcanic margin and the inactive West Mariana ridge (WMR) by the Mariana trough and back-arc spreading center (MT) since 7 Ma (e.g., Stern et al., 2003), while NIA continues growth as an unrifted arc. For our seismic study of crustal growth, we carried out a deep seismic profile, with 106 ocean bottom seismographs (OBSs) and an airgun array with a total capacity of 196.6 litres (Takahashi et al., 2003) (Fig. 1).

P-WAVE VELOCITY MODELING TECHNIQUE AND RESULTS

Construction of a velocity model with high reliability and high resolution depends on the signal-to-noise ratio of OBS records. Because almost all our OBS records observe clear first phases through not only the crust but also the mantle to maximum offsets in some cases of over 150 km (Takahashi et al., 2003), our data have sufficient quality and depth penetration to image the full crustal thickness of MABS. Using these OBS records and our collinear reflection section, we constructed a P-wave velocity model from PVB, across WMR and MT, to MA using standard tomographic inversion and 2-D ray tracing techniques (Zelt and Barton, 1998; Zelt and Ellis, 1988; Zelt and Smith, 1992) (Fig. 2a). Interface locations and velocity contrasts were confirmed by minimizing traveltime residuals and comparison of waveforms between observed and synthetics.

Fig. 2b illustrates the good resolution of the final velocity model. Interface and velocity nodes were spaced at 5 km intervals, equivalent to the closest OBS spacing. Resolution values are a measure of model reliability and vary from 0 to 1, with values exceeding 0.5 conventionally regarded as reliable (Zelt and Smith, 1992). Based on this standard, almost all velocity nodes within the crust (except for the deepest parts of the middle and lower crusts, and the extreme ends of the profile) are reliable. Similarly, the interface nodes for the crucial boundaries between the middle and the lower crusts and the Moho exceed 0.5, meeting our reliability criterion and justifying our interpretation of the thickness of the middle and lower crusts and the shape of the Moho. Another measure of the quality of fit is that almost all residuals are below 150 msec, which is less than one period of the dominant frequencies of 5 Hz in the OBS data.

The Mariana arc (MA) consists of the active arc initiated in the Pliocene c. 30 km west of the Eocene-to-Miocene frontal arc and its forearc basin (Stern et al., 2003; Crawford et al., 1981). The Moho depth is c. 20 km beneath the volcanic front. The crust has four distinct layers separated by rapid increases and/or velocity contrasts in seismic velocity. We identify these as a sedimentary layer (2.0-4.5 km/s), upper crust (4.5-6.0 km/s), middle crust (6.1-6.5 km/s) and lower crust (6.7-7.3 km/s). Specifying the boundary between the arc and the back arc based on the shape of the Moho (Fig. 2a), their volumes are 311 km^3/km of arc length, 588 km^3/km, 591 km^3/km and 1,349 km^3/km, respectively, representing proportions of 11%, 21%, 21% and 48%, respectively (Table 1). These layers contain significant lateral heterogeneity: the average velocity of the middle crust is somewhat faster in the forearc than elsewhere, and is slowest adjacent to the backarc MT; and the velocity of the lower crust (7.1-7.3 km/s) beneath the forearc is significantly faster than that beneath the active arc (6.7-7.3 km/s). Sub-Moho velocities, at c. 7.7 km/s, are significantly less than the global average of 8.1 km/s.

WMR was a part of the Miocene arc developed before the Mariana trough opened in Late Miocene time (e.g., Crawford et al., 1981). The Moho depth is c. 17 km. The crust apparently consists of the same four layers as in the MA, a 2.2-4.0 km/s sedimentary layer, a 4.0-5.2 km/s upper crustal layer, a 5.6-6.5 km/s mid crustal layer and a 6.7-7.4 km/s lower crustal layer. Their volumes are 271 km^3/km, 358 km^3/km, 535 km^3/km and 883 km^3/km, and their relative proportions are 13%, 17%, 26% and 43% (Table 1). As in the MA, the lower crust exhibits its lowest seismic velocity adjacent to MT (6.7-7.2 km/s) in contrast to the western part with faster velocities of 7.2-7.4 km/s; and the sub-Moho velocity is again slow, with a velocity of 7.6 km/s.

DISCUSSION

Common structural characteristics of the northern Izu arc and the Mariana arc

The crusts of MA, WMR and NIA share important commonalities (Fig. 3), including the relatively laterally uniform upper crust with velocity 4.5-6.0 km/s, middle crust with velocity 6.0-6.5 km/s, and lower crust with velocity 6.8-7.3 km/s (Suyehiro et al., 1996). Based on petrologic studies (e.g., Kitamura et al., 2003; Christensen and Mooney, 1995), compositions of the upper, middle and lower crusts are interpreted to be basaltic, tonalitic and gabbroic, respectively. The volumes of the sedimentary, upper-crustal, middle-crustal and lower-crustal layers in NIA are 496 km^3/km, 732 km^3/km, 963 km^3/km and 2,455 km^3/km, and their proportions are 11%, 16%, 21% and 53%, respectively (Table 1). It might be expected that these proportions of the three arcs would be roughly similar, however, the proportion of the lower crust in WMR seems to be lower than that of NIA. Hence, it is important to consider how arc magmatic evolution might produce these structural characteristics.

The discovery of low velocities just beneath the Moho, now also recognized beneath the volcanic front of NIA (Kodaira et al., submitted), may be a key to understanding the crustal growth history. Candidates for the sub-Moho velocity in island arcs include melt, serpentinized mantle, crystallized basaltic magmas and transformation of crustal materials to upper mantle. These four scenarios all probably occur beneath the current active MA (Fig. 3). However, neither melts nor serpentinized mantle can cause the observed low mantle velocity on the backarc (eastern) side of the inactive WMR, away from the volcanic front: WMR cannot preserve a serpentinized mantle at present due to lack of a water source from a subducting slab and a higher mantle temperature at present than the MA (Hyndman and Peacock, 2003).

Crustal growth using the seismic structure and petrologic model

We next interpret the crustal growth represented by the above crustal structure and slow mantle velocity. It is widely accepted that underplating of mafic magmas produces fast velocities in the lower crust (e.g., Fountain, 1989). We assume that velocity variations within each layer represent petrologic and geological diversity and depend on the relative mafic component (Behn and Kelemen, 2003). One petrologic scenario for crustal growth (Tatsumi, 2000) is that tonalitic middle crust is produced by anatexis of the basaltic lower crust produced by differentiation from a primary, mantle-derived basaltic magma. The existence of an obducted tonalitic pluton at the northern tip of the Izu arc where the arc collides with the Japan arc supports the above mechanism, as the tonalite is thought to be produced by partial melting of a differentiated basaltic component of the Izu arc (Kawate and Arima, 1998). Based on this petrologic constraint, together with appropriate compositions of primary and differentiated arc basalt magmas, and experimentally inferred melting regimes of 25-35% partial melting (Beard and Lofgren, 1991), the volume of both restite and cumulate can be calculated from the observed volume of the upper and middle crusts.

The anatexis model used in our calculations is given in the Appendix. In most modern subduction zones, arc magmas and their solidified products (the arc crust) are derived from primary basalt magmas (PM) generated in the mantle wedge. At the juvenile stage of arc evolution, the mantle-derived basalt magma forms the initial arc crust, most simply composed of differentiated basalt (DB) and cumulate (CM) layers. This model regards the basaltic materials underplated at the bottom of the crust as a part of the "crust". The DB layer is then re-melted by basaltic underplating to produce tonalitic magmas (AM) (Tatsumi, 2000) that form the middle crust with a characteristic velocity of 6 km/s, and restites. Since the major product of surface volcanism at least in oceanic island arcs such as the IBM is basaltic (Tamura and Tatsumi, 2002), it may be reasonable to assume DB, which is produced from a mantle-derived primary basalt magma by crustal anatexis and subsequent fractionation of cumulates, erupts to form the upper crust.

Our anatexis model allows us to calculate the expected volumes of restites and cumulates (together forming the petrologic lower crust) from the seismologically measured volume of upper and middle crusts (Appendix; Table 1) A crucial observation in all three arcs is that the seismologically measured volume of lower crust is much smaller than the petrologically inferred combined volume of the restites and cumulates (Table 1). Our seismologically observed lower-crustal volumes fall within the range of our estimated restite volumes (MA and NIA), are somewhat smaller (WMR). In MA and NIA, this implies that dense cumulates are now seismologically part of the mantle and that the restites correspond to the lower crust as suggested by previous petrologic studies (e.g., Kay and Kay, 1993; Jull and Kelemen, 2001). In WMR, dense mixtures of restites and cumulates might also have become part of the upper mantle due to gravitational instability (e.g., Jull and Kelemen, 2001) possibly causing slow mantle velocity.

Table 1 also indicates a major compositional difference between oceanic island arcs and average continental crust (e.g., Zelt and Forsyth, 1994). Our observation of volume ratios of lower crust to middle crust in oceanic island arcs range from 1.65 (WMR) and to 2.54 (NIA), but that of the continental crust is less than 1.0. This transfer of the lower-crustal material into the mantle may be an important part of the crustal growth process.

Our crustal growth model begins by undeplating primitive basalts. Because the Eocene portion of the IBM arc includes high-Mg andesites (e.g., Kelemen et al., 2003), it is possible that a crust rich in high-Mg andesite rich existed there before underplating of primitive basaltic magmas. Even so, the same transformation of the crust during crustal growth process should occur and restites should still be present in the seismologically determnated mantle, because the high-Mg andesitic material should also be differentiated by heat supplied from underplating primitive basalts.

In the previous petrologic discussion we used the shape of the Moho to specify the boundary between the backarc and arc regions. A likely source of error in the estimation of volumes of each crustal layer, required to apply the above petrologic model, is the difficulty of specifying the arc-backarc boundary to within about 10 km, leading to uncertainties in lower-crustal volume of up to 50-100 km^3/km. Our robust conclusion that transformation of cumulate crust into seismic upper mantle is required in order to explain the observed middle-crustal volume is not affected by this relatively small uncertainty.

Our observations and calculations imply presence of cumulates and restites, produced during crustal growth, below the Moho. However, the lateral lower-crustal heterogeneity remains as an issue to be understood. This velocity variation could be strongly related to crustal growth, and relevant to the formation of continental crust, because the velocities are close to that of typical lower continental crust. Particularly, it seems important that the lower-velocity lower crust with less than 7 km/s is located beneath the current volcanic fronts of both MA and NIA, beneath the frontal arc within MA, and beneath regions within MA and WMR adjacent to MT. This lateral velocity variation within the lower crust is relatively reliable as shown by the high-resolution values in our model (Fig. 2b). Our current geochemical model is too simple to model this lateral variation, which therefore remains a target of future geochemical modelling.

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APPENDIX: Petrologic crustal growth model

For our petrologic model, we use a basalt composition that typifies IBM volcanism, as differentiated basalt (DB). Compositions of the primary basalt (PM) and the fraction of cumulate materials (CM) (22%) can then be obtained by assuming a simple olivine maximum fractionation from PM to DB, and Fe-Mg exchange partitioning between magma and olivine with $Fe^{2+}/(Fe^{2+}+ Fe^{3+})_{magma}=0.9$, and $Mg/(Mg+Fe)_{mantle\ olivine}=0.9$. Fractionation of olivine only, and subsequent formation of dunitic cumulate is confirmed by MELTS modeling (Ghiorso and Sack, 1995) for PM and DB compositions. An andesitic melt (AM) is produced by 25-35% partial melting of a basaltic component (~DB) (Beard and Lofgren, 1991). Major-element compositions of all these magmas are shown in Supplemental Material Table S1.

The petrologic model we use here is expressed by the formulae:

$V_{uc}=(1-R_{db})*V_b$,

$V_{mc}=R_{dm}*(1-R_{db})*V_a$,

$V_{res}=(1-R_{dm})*(1-R_{db})*V_a$,

$V_{cm}=R_{db}*(V_a+V_b)$

where V_a and V_b are the volume of primary basalt magmas (PM) intruded at the initial and anatexis stages; V_{uc}, V_{mc}, V_{res}, and V_{cm} are the volumes of basaltic upper crust, andesitic middle crust (AM), restites and cumulates (CM); and , R_{db} and R_{dm} are the fractional degree of crystallisation to produce CM (22%) and the degree of partial melting to produce AM

(25-35%), respectively. Re-writing these equations in terms of parameters V_{uc} and V_{mc}, obtained from our seismic profile, we obtain:

$V_{res}=(1-R_{dm})*V_{mc}/R_{dm}$

and

$V_{cm}=(V_{mc}/R_{dm}+V_{uc})*R_{db}/(1-R_{dm})$.

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FIGURE CAPTIONS

Figure 1 Bathymetry of the Mariana arc-backarc system around our wide-angle seismic profile. The thick black line and open circles indicate locations of the airgun shooting line and OBSs, respectively. Numerals denote the site numbers of OBSs. Contour interval is 1,000 m. Inset shows our profile superimposed on bathymetry of whole of IBM arc with 2,000 m contour lines, and locates the Kyushu Palau Ridge (KPR), Parece Vela Basin (PVB), West Mariana Ridge (WMR), Mariana Trough (MT), Mariana Arc (MA) and Mariana Trench.

Figure 2 Final velocity models of MABS and its resolution. (a) Final model. Solid circles indicate OBS locations. Numerals denote P-wave velocity (Vp) (km/s). Open arrows indicate boundaries between the arc and backarc areas used to calculate crustal volumes. UC, OL2 and OL3 are upper crust and oceanic layers 2 and 3, respectively. Dark shading shows region of seismic ray penetration. (b) Resolution of the final model. Resolutions of velocity nodes are indicated by the color scale with contour spacing of 0.1, and those of the interface nodes are shown by the size of the open circles. Red circles indicate OBS locations.

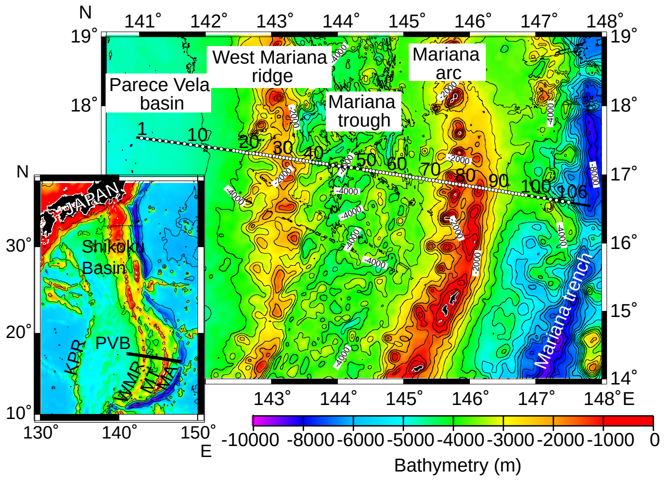
Figure 3 Schematic interpretations of crustal growth of MABS. MC and LC: middle and lower crust. (a) Miocene. (b) Current.

Table 1 Comparison of the seismologically measured crustal volumes (km^3/km) of MA, WMR and NIA. The Petrological estimate of the volumes of differentiated restites and cumulates, and their sum ("estimated lower crust") assume 25-35% partial melting in order to produce an andesitic layer equivalent to the observed middle crust.

Table 1 Comparison of crustal volume

	Layer	MA (km^3/km)	WMR (km^3/km)	NIA (km^3/km)	CC (km)
Seismological estimate	Sediments	311	271	496	
	Upper crust	588	358	732	7
	Middle crust	591	535	963	19
	Lower crust	1,349	883	2,455	18
Petrologic estimate	Estimated restites	1,773-1,098	1,605-994	2,889-1,788	
	Estimated cumulates	833-642	705-532	1,293-983	
	Estimated lower crust (restites+cumulates)	2,606-1,740	2,310-1,526	4,182-2,771	

Each value in CC (continental crust) column indicates thickness of each layer (Rudnick and Fountain, 1995; Zelt and Forsyth, 1994).



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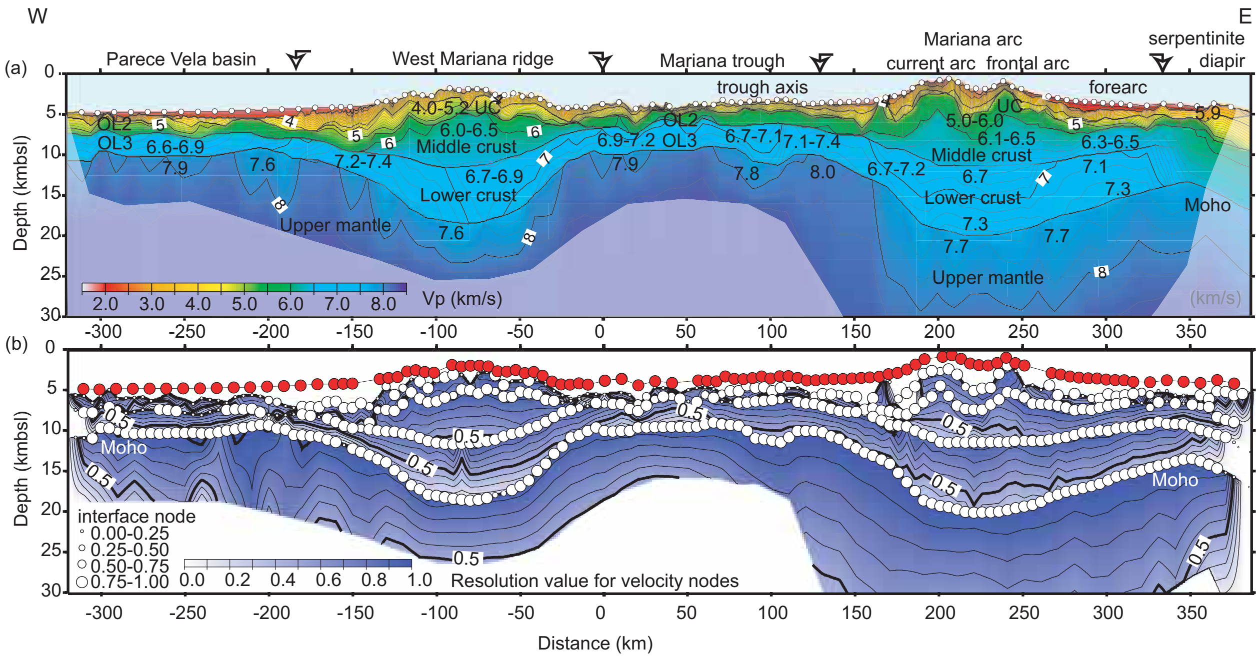

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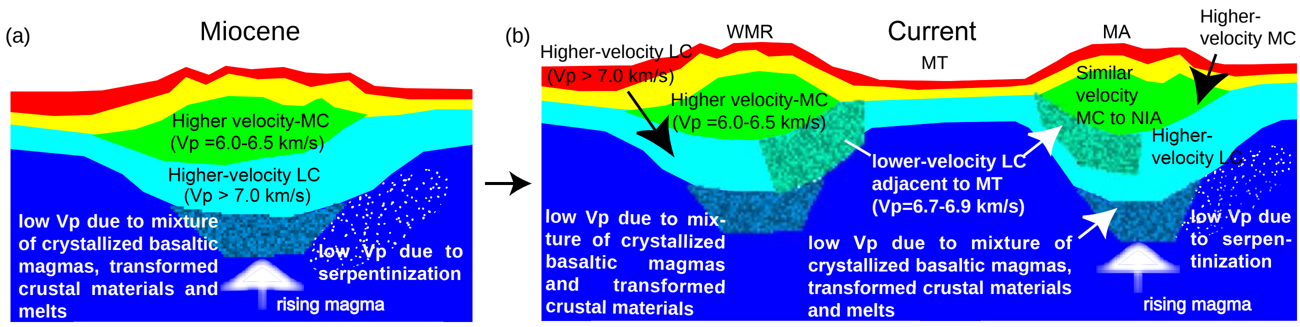

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