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Abstract

Island arcs are geologically active and important structures. From a short-term perspective, they are a major source of seismic and volcanic hazards. From a longer-term perspective, arc processes are most likely a key component in the production of continental lithosphere. They are also the focus of numerous Geoscience investigations. In this thesis I investigate the seismic structure of island arcs at a regional (hundreds of kilometer) and a local (10’s of km) scale. My goal in this work is to contribute to our efforts to understand the origin and evolution of these geologically important structures. I focus seismic imaging methods on two regions, the Philippine Island region and the northern Lesser Antilles island of Montserrat.

The Philippine Island Arc (PIA) is commonly regarded as a complex structure in which subduction zones border its sides and the intra-arc, sinistral Philippine Fault System transects throughout its length. The arc is seismically active and volcanic activity spans almost the entire arc. While several studies provide a wealth of information on the tectonic and the geodynamic settings of PIA, few have looked carefully into the subsurface because they were limited by the availability of digital seismic data. For this reason, important data gaps exist, in particular the details of the subsurface seismic velocity structure. The recent deployments of relatively dense digital seismic stations offer an opportunity to conduct a detailed study on the arc’s velocity structure. Data from this new seismic network are used to determine the three-dimensional (3–D) velocity structure of the PIA by applying the $P$–wave travel time tomography.

A broad distribution of source depths and the arc-wide distribution of seismic stations allow tomographic imaging of structures down to 450 km depth with spatial resolution of about $\sim$50 km resolution. The prominent features of the tomographic images include the low velocity zones correlating with the overlying volcanic structures and high velocity zones that more or less coincide with the Wadati-Benioff zones of the subduction zones. The slabs are imaged as 2–6%
faster than the mantle velocity values of the IASP91 model. They commonly extend deeper than the seismicity suggesting that they penetrate aseismically to greater depths. Shallow low-velocity anomalies correlate with the fore–arc and intra–arc basins, in the mantle wedge at the top of the slabs, deep in the upper mantle, as well as the segments of the Philippine Fault Zones.

Images of island arc structure are generally limited to the high, narrow frequency bands employed in active-source seismic experiments or smoothed over substantially in low-frequency global and regional tomography. A second focus of the work in this thesis is the imaging the regional- and local-scale structure of an active arc, the island of Montserrat, located in the northern Lesser Antilles arc. I use receiver functions and local-earthquake generated P-wave arrival times to estimate the first-order subsurface seismic structure in the vicinity of the island. The receiver functions provide information on the first-order shear velocity and Poisson’s ratio of the crust. Although the precise sharpness of the crust-mantle transition is not resolved, this study can limit its thickness to less than $\sim 4$ km. The estimated mean crustal thickness of $\sim 30$ km and the observations suggest that the crust may be slightly thinner northwest of the island ($\sim 26-30$ km) than it is to the south ($\sim 30-34$ km). The high P-wave speed and Poisson’s ratio indicate a generally mafic lower crust, with rocks of intermediate composition not precluded in the upper part of the lower crust. We did not find evidence for a thick high-speed lower crust ($> 7.4$ km/s) as has been inferred in some other arcs. P-wave travel times are used to image the upper crust of the island, which includes the actively erupting Soufrière Hills Volcano. As with many studies of the island, imaging deeper than about 5-to-6 km beneath the volcano is very difficult because of limited ray coverage at depth. The earthquake-produced travel times primarily sample and illuminate the low-velocity upper regions of the island. The results are generally consistent with the results from recent active-source work imaging beneath the island.
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MT–Manila Trench, CZ–Collision Zone, NT–Negros Trench, CT–Cotabato Trench, ELT–East Luzon Trough, PT–Philippine Trench.

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MT–Manila Trench, CZ–Collision Zone, NT–Negros Trench, CT–Cotabato Trench, ELT–East Luzon Trough, PT–Philippine Trench.

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3.26 See Figure 3.25 for the general description of the figure. MT– Manila Trench, NLT– North Luzon Trough, SCS– South China Sea, ELT– East Luzon Trough.

3.27 See Figure 3.25 for the general description of the figure. MT– Manila Trench, VH– Vigan High, WLT– West Luzon Trough, SCS– South China Sea, ELT– East Luzon Trough, B.R.– Benham Rise, WPB– West Philippine Sea Basin.

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3.32 See Figure 3.25 for the general description of the figure. SCS– South China Sea, PT– Philippine Trench, WPB– West Philippine Sea Basin, PFZ– Philippine Fault Zone.

3.33 See Figure 3.25 for the general description of the figure. SS– Sulu Sea, PT– Philippine Trench, WPB– West Philippine Sea Basin, PFZ– Philippine Fault Zone.
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B.6 See Figure B.2 for the general description of the figure. MT– Manila Trench, SCS– South China Sea, M.B.– Marinduque Basin, ELT– East Luzon Trough, PT– Philippine Trench, WPB– West Philippine Sea Basin, PFZ– Philippine Fault Zone.

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B.11 See Figure B.2 for the general description of the figure. (see Figure B.8 for the slice locations). SS– Sulu Sea, NT– Negros Trench, PT– Philippine Trench, WPB– West Philippine Sea Basin, PFZ– Philippine Fault Zone. 

B.12 See Figure B.2 for the general description of the figure. (see Figure B.8 for the slice locations). ST– Sulu Trench, SS– Sulu Sea, PT– Philippine Trench, WPB– West Philippine Sea Basin, PFZ– Philippine Fault Zone, CMA– Central Mindanao Arc. 

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C.6  Results of checkerboard resolution tests. (Top) Map of Montserrat Island showing the locations of the vertical profiles (Bottom). The middle plots are the topographic profiles along the slices. SVH - Silver Hills, CH - Centre Hill, SHV - Soufrière Hills, SSH - South Soufrière Hills. The scale bar shows the velocity perturbation.  

C.7  Results of checkerboard resolution tests. (Top) Map of Montserrat Island showing the locations of the vertical profiles (Bottom). The middle plots are the topographic profiles along the slices. SVH - Silver Hills, CH - Centre Hill, SHV - Soufrière Hills, SSH - South Soufrière Hills. The scale bar shows the velocity perturbation.
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Dedication

This humble piece of work is dedicated to:

Chuck Ammon, His kindness, unwavering support, and extraordinary patience made me what I am today.

My beloved Philippines, the Pearl of the Orient Seas. May this research unravel a fraction of your rich geodynamic setting!

Our Heavenly Father, Geophysicist Extraordinaire. For your greater glory and honor!
Chapter 1

Introduction

In this thesis, I investigated the seismic velocity structure beneath two island arcs, the Philippine Island Arc (PIA) and a smaller region of northern Lesser Antilles. I begin with a review of my work to image the three-dimensional $P$-wave velocity structure of the PIA using $P$-wave traveltime tomography. The PIA is one of the series of arc systems in the Western Pacific region (Figure 1.1) that features spectacular volcanic and seismic activity and showcases a complex geologic evolution. For these reasons alone, understanding the tectonic structures of the arc is important in deciphering its part in the formation of the Western Pacific region and also in the context of seismic hazard analysis.

I begin focus on the PIA with my work on estimating the bulk crustal shear-wave velocity structure of the PIA (Chapter 2). Observations from three broadband seismic stations located in different geologic settings were used in receiver function analyses and joint inversion of receiver function and surface-wave dispersion measurements. The receiver function analysis is performed to estimate the crustal thickness and Poisson’s ratio under the stations. These two parameters are important in the study of arc growth rate of the PIA as well as in determining the lower crust composition. The observed receiver functions are quite complex as a result of the location in regions of complicated near-surface geology and the existence of a subducting (dipping) slab at depth. The calculated receiver functions and the fundamental-mode surface wave dispersion data from independent study are jointly inverted to construct the initial model for the tomography study. Jointly inverting these two data sets reinforce the advantages of each method and
Figure 1.1. Simplified tectonic map of the Western Pacific region. The PIA is one of the series of arc systems in the region from Japan to the north down to New Zealand to the south (Yumul et al., 2003). Dots are earthquakes from Engdahl et al. (1998) catalog color coded by hypocentral depths. Also shown are the tectonic plates from model PB2002 (Bird, 2003). Amur (AM), Australia (AU), Banda Sea (BS), Caroline (CL), Mariana (MA), Molucca Sea (MS), Birds Head (BH), Pacific (PA), Philippine Sea (PS), Sunda (SU), Yangtze (YA).
at the same time bridges the resolution gaps between them. Thus, the joint method can provide a better constrained shear–wave velocity model for the PIA. Of course the structure of the PIA is quite complicated to hope that one could unravel many of the details using observations from three locations. The broader region is the focus of work described in Chapter 3.

Chapter 3 documents the detailed three-dimensional, $P$-wave velocity structure of the PIA using local, regional, and teleseismic earthquakes (determined for the first time), which provides insight into the geodynamic characteristics of the arc. The tomographic procedure used in this study has illuminated the structure of the slabs bordering the PIA showing variations in the geometry and length within and among the subduction zones. The slabs are imaged to extend deeper than the maximum depth of seismicity, which may imply that the deeper part of the slabs are subducting aseismically. Distinctive low velocity zones are commonly observed in the mantle wedge above the slab, which often connect to the surface beneath the volcanic regions. Significant low velocity bodies are also observed deeper in the upper mantle and far from the slabs of the present-day subduction zones. These results provide evidence for the occurrence of slab dehydration processes as well as possible deep dehydration process of the remnant slab. Some other features revealed by the tomography are the possible migrating partial melts from the vertically dipping slab and extending upward below the active, Quaternary volcanoes in southwestern PIA, the flat–slab geometry observed in central Luzon Island that features subduction of a buoyant extinct ridge, and the possible detachment of portions of South China Sea slab in northern Luzon that correlate spatially to the volcanoes with enriched mantle source. Based from the minimum length of the imaged slab and from the GPS–derived convergence rate data, the age of the initiation of Manila Trench is Early Miocene, the Negros Trench is Late Miocene and the Philippine Trench is Pliocene. The apparent simplicity of the tomographic image of the PIA, at least the upper 400 km, suggests that the arc may have formed on or near its current location. The coherency of the high velocity zone structures and the spatial relationships of the low velocity zones can easily be explained by the in situ formation of the majority of the PIA at least during the Early Miocene rather than the arc forming from amalgamation of different terranes that were derived from different locations.
In chapters 4 and 5 I document my work applied to image the crustal structure beneath and in the vicinity of the Island of Montserrat, home of the active Soufrière Hills Volcano. Similar with the work I did for the PIA, receiver function analysis was conducted to estimate the crustal thickness and Poisson’s ratio of the island, which is presented in Chapter 4. Teleseismic waveforms recorded by 8 seismic stations of the Montserrat Volcano Observatory were used in the analysis and I used forward modeling to model the first-order subsurface seismic velocity structure. Chapter 4 was published as a journal article in Geochemistry, Geophysics, Geosystem (Sevilla, W. I., C. J. Ammon, B. Voight, and S. De Angelis (2010), Crustal structure beneath the Montserrat region of the Lesser Antilles island arc, Geochem. Geophys. Geosyst., 11, Q06013, doi:10.1029/2010GC003048.). Part of the results from Chapter 4 were used to construct an initial model for the tomography study of the island presented in Chapter 5. The main objective of this study is to map the depth of the magma source region of the Soufrière Hills Volcano using the tomography. Unfortunately, as with many similar studies, imaging deep enough to see the magma source remains an elusive goal. The depth of the earthquakes (or in the case of other studies, the depth penetration of the first-arriving seismic waves) do not sample the region of suspected magma. The work provides a valuable image of the shallow structure, which like many volcanic edifices is extremely heterogeneous. Variations in the shallow structure may lead to insights into the hydrologic and stress interactions in the uppermost crust beneath the island as they are applied to earthquake location and faulting geometry estimations in future studies.
Chapter 2

Lithospheric Structure of the Philippine Island Arc From Receiver Function Analysis.

2.1 Abstract

Teleseismic waveforms recorded from three broadband, seismic stations in the Philippine Island Arc are analyzed to investigate the first-order crustal structure beneath the island regions. The H-K stacking procedure is performed on the waveforms to obtain local estimates of crustal thickness and Poisson’s ratio. Although the stations are located in regions that underwent different tectonic history, the results of the analysis suggest that the PIA (at least on those regions sampled by the receiver functions) is underlain by relatively thick crust (∼35 km) with intermediate bulk crustal composition (Poisson’s ratio ∼0.26). These results are consistent with the intermediate to silicic composition of the magmatic deposits from nearby volcanic fields based from geochemical studies. The shear wave velocity model is also estimated by jointly inverting the receiver functions with surface wave dispersion data that were derived from an independent study. The obtained model placed bounds on the thickness of the overlying sediment from 5 to 10 km thick. Intra-crustal low velocity zones are also resolved in the inversion, which may indicate partially melted region corresponding to the source of magmatic materi-
als. But a firm conclusion on this is difficult given the complexity of the signals. The complexity of the waveforms limited the resolution of the inversion down to about 60–km depth.

2.2 Introduction

Fundamental in understanding the crustal generation and evolution of the Philippine Island Arc (PIA) system is the knowledge of the first-order crustal structure and bulk crustal composition of the arc. These parameters hold imprints of the geological processes that created the arc system, so sufficient information on these parameters is important for deciphering the tectonic history of the PIA. While several geological studies provided an important framework towards the understanding of the arc crust, information on the crustal thickness of PIA is rather limited to specific areas sampled with receiver function analysis (Besana et al., 1995). In some regions, estimates of the thickness were derived from geochemical relationships, seismicity depth, and gravity data (Dimalanta and Yumul, 2003, and references therein). Recent compilations on the crustal thickness suggested that the PIA crust is 15 to 65 km thick, with median values from 24 to 40 km (Dimalanta and Yumul, 2003). While the median values are consistent with the average thickness of mature island arc crusts (e.g. Condie, 2005), values as much as 65–km thick are probably overestimated; these large values were estimated from the arc geochemistry–crustal thickness systematics of Plank and Langmuir (1988) and from hypocentral depths. Such systematics are subject to considerable uncertainties in the calculation of the crustal thickness that includes the constant depth assumption for the top of the slab corresponding to the location of arc volcanoes (Plank and Langmuir, 1988). Recent studies proposed that these depths vary between and within arcs segments although the depths may correlate with the slab dip (Syracuse and Abers, 2006). The variation of these depths might have significant implications on the extent of the melting process beneath the arcs, thus affecting the geochemically-derived crustal estimate thickness (Syracuse and Abers, 2006). In global crustal models such as CRUST2.0 (Bassin et al., 2000), a 2 x 2 degree grid size global model, the PIA crustal thickness ranges from 19 to 33 km thick.
Information on the arc growth processes of PIA is not only important in terms of specific local or regional tectonic reconstructions, but also can provide insight on the formation and evolution of continental crust in general. At least during the Phanerozoic, a significant proportion of continental crust is suggested to be derived from island arcs through arc magmatic events and subsequent accretion of arc–derived materials to continents by subduction-related convergence (e.g. Davidson and Arculus, 2005). The PIA is in the process of colliding with the continental Eurasian margin and therefore it presents a good venue in understanding the development of an arc complex and before its accretion to a continental margin (e.g. Encarnación, 2004).

The majority of the PIA was built on oceanic lithosphere which was subsequently modified by episodes of volcanism, accretion, rifting, and collision of microcontinental blocks (e.g. Hamilton, 1979; Holloway, 1982; McCabe et al., 1982; Taylor and Hayes, 1983; Karig, 1983; McCabe et al., 1985; Karig et al., 1986; Mitchell et al., 1986; Geary et al., 1988; Rangin, 1991). Formation in an oceanic setting probably started in the Late Jurassic (Geary et al., 1988) with modification as the original lithosphere collided with rifted fragments of Eurasia during the Miocene (McCabe et al., 1982). The oceanic basement is manifested by widespread occurrences of upthrust or uplifted belts of ophiolitic terranes throughout the arc, and these are commonly capped by associated marine sediments (e.g. Karig, 1983; Geary et al., 1988; Rammlmair, 1993). The nature of these ophiolithic terranes is a controversial topic in Philippine geology. Several workers suggested that these terranes are allochthonous put together by convergence processes and wrench tectonics (e.g. Hamilton, 1979; Holloway, 1982; McCabe et al., 1982; Taylor and Hayes, 1983; Karig, 1983; McCabe et al., 1985; Karig et al., 1986; Mitchell et al., 1986; Geary et al., 1988; Rangin, 1991). Others suggested that at least in the Northern Philippines, the terranes are autochthonous and were generated by episodic generation of oceanic crust within the arc complex (Encarnación, 2004). These two different scenarios on the formation of the basement units have significant implications on the formation and growth of the PIA.

In this study, teleseismic waveforms recorded by three broadband stations in PIA are analyzed using receiver function technique to estimate the structure and composition of the crust. Receiver function analysis is a simple yet powerful tool to
map crustal thickness and to estimate Poisson’s ratio, an elastic parameter that is directly equivalent to the ratio of the $P$- and $S$-wave velocities. The combination of Poisson’s ratio and one of the seismic wave speeds can be used to characterize the bulk crust composition better than one of the speeds alone (e.g. Christensen, 1996). In conjunction with the information on upper crustal rocks (usually from direct field mapping of outcrops), the bulk-crustal estimates of Poisson’s ratio can also place constraints on the compositions of lower crust rocks, which are often inaccessible to direct sampling (Zandt and Ammon, 1995). The geological processes that operated on the arc can then be better deciphered with improved knowledge of the rock compositional variations. I begin with a brief overview of the receiver function methods in the next section followed by the results of the analysis for the three stations used in this study. Results from this study will add receiver function–derived estimates of crustal thickness to the current database for the PIA arc.

2.3 Receiver Functions

Detailed descriptions of receiver function analysis can be found in Langston (1979), Owens et al. (1984), Ammon (1991), Cassidy (1992), or Cassidy (1995). Receiver functions are time series consisting of a direct $P$-wave pulse, locally generated $P$-to-$S$ ($Ps$) phases, and crustal multiples generated at the boundaries beneath the recording station (Fig. 2.1). $P$-receiver function estimation begins with seismograms recorded at teleseismic distance ($30^0$ – $90^0$) from three-component, broadband seismic station. At this distance range, the $P$-wave comes in at a steep angle of incidence through the crust beneath the recording station resulting to polarized particle motion: the $P$-waves dominate the vertical component and the $S$-waves the horizontal component (e.g. Cassidy, 1992). Thus the generated $Ps$ phase and associated multiples are almost exclusively horizontal and dominate the waveform of the radial component of the station. Deconvolving the vertical component (dominated by source and path effects) from the radial component isolates the $Ps$ phases and suppresses the $P$-multiples yielding the radial receiver function (e.g. Cassidy and Ellis, 1991). The same process can also be applied to the transverse component to produce the transverse receiver function. The transverse
receiver function, although used rarely because of its instability in the presence of signal and processing noise (Langston, 1981), is useful in exploring for the presence of laterally, inhomogenous structure (e.g. Julia and Mejia, 2004). A laterally homogenous, nearly horizontal stratification will yield a small amplitude transverse receiver function while the laterally varying structure will introduce tangential component of motion (e.g. Julia and Mejia, 2004; Bostock, 2007).

Figure 2.1. (Top) Ray diagram showing the legs of the major phase arrivals that consist the receiver function using a single layer over a half-space model. (Bottom) The corresponding waveforms calculated using the model above. The figures are modified from Ammon et al. (1990).

In the deconvolution process, the information on the absolute time is lost and
all time information is now relative to the direct \( P \)-wave that arrives at zero time (e.g. Owens and Crosson, 1988). Provided the average \( P \)-velocity is known, the relative travel times of direct \( P \) with the \( Ps \) phase and multiples can supply constraints on the crustal thickness below or near the seismic station (e.g. Zandt et al., 1995). The relative travel time however is sensitive to the average velocity and depth to the interface (Ammon et al., 1990). This depth–velocity trade-off can be minimized by supplying \textit{a priori} information such as \( P \)-velocity from refraction studies, \( S \)-velocity models from surface wave analysis, and estimates of Poisson’s ratio (Ammon et al., 1990). The amplitudes of the \( Ps \) phases and multiples are dependent to the \( S \)-velocity contrast across the boundary. The stronger the contrast, the more distinct are the amplitudes of the phases. Low-velocity zones can be identified from the polarity of the \( Ps \) phases, \textit{i.e.} negative and positive polarity indicate the top and bottom of low \( S \)-wave velocity region.

The receiver function analysis is also useful in identifying dipping interfaces below the recording station. Dipping layers introduce transverse component of motion associated with the \( P \) and \( Ps \) arrivals. They also generate pronounced azimuthal variations in both the amplitude and arrival time of the \( Ps \) conversion. These azimuthal variations can be used to constrain the dip direction and angle of the interfaces like the subducting slab in subduction zone environments (Langston, 1979; Owens et al., 1988; Cassidy, 1992, 1995).

Calculating receiver function from observed seismograms is a linear inverse problem that can be addressed in either time or frequency domain. In this study, I used the iterative time–domain deconvolution of Ligorría and Ammon (1999). This iterative deconvolution is a least–squares minimization of the difference between the Gaussian filtered observed radial (or transverse) seismogram and the predicted signal produced from the convolution of an iteratively updated spike train with the vertical component seismogram. The method has several advantages that includes a causal receiver function, a reduction in side-lobes due to the use of Gaussian pulses, and it does not require an optimal stabilization parameter like the water-level or damping value. Detailed descriptions on the advantages of the time–domain iterative approach are discussed in Ligorría and Ammon (1999) and Kosarian (2006).
2.4 Receiver Function $H-\kappa$ Stacking Analysis

I used teleseismic waveforms recorded from three broadband stations in the Philippines, two located in Luzon Island and one in Mindanao Island (see Figures 2.2 and 2.13), during the period 1995 to 2009. The waveforms are from earthquakes with $M_w \geq 5.0$ to ensure good signal-to-noise ratios. The horizontal north and east-component seismograms are rotated to radial and transverse components. Then the vertical component is deconvolved from the horizontals. Two frequency bands corresponding to Gaussian width factors 2.5 ($f < 1.25$ Hz) and 1.0 ($f < 0.5$ Hz) were used in the deconvolution to produce two receiver functions with high and low frequency bands. The low frequency bands reduce the effects of scattering on the receiver function but only provide signals sensitive to the simpler features of the structures. The high frequency band signals can provide higher resolution of the structure but the resulting receiver functions may be complicated if the data are of low-quality or the structure is complicated (Julia et al., 2003).

After obtaining the receiver functions, I used the program $hkstack$ (Julia and Mejia, 2004) to estimate the crustal thickness ($H$) and Poisson’s ratio ($\kappa$). Program $hkstack$ performs the stacking procedure of Zhu and Kanamori (2000) with the inclusion of bootstrap approach for uncertainties estimation (Julia and Mejia, 2004). Using different $H$ and Poisson’s ratio or equivalently $V_p/V_s$ ratio, where $V_p$ and $V_s$ are $P$- and $S$-wave velocities respectively, the stacking procedure sums the amplitudes of receiver function at the predicted arrival time of the $Ps$ phase and the two crustal multiples $PpPms$ and $PpSms + PsPms$. The quality of the phases amplitudes varies such that the stacking procedure uses a weighted sum of the receiver function amplitudes, $P(H,\kappa)$,

$$P(H,\kappa) = W_{Ps}R(t_{Ps}) + W_{PpPms}R(t_{PpPms}) - W_{PpSms}R(t_{PpSms}) \quad (2.1)$$

where $H$ is thickness, $\kappa$ is the $V_p/V_s$ ratio, $W$’s are the a priori weights ($\Sigma W_i = 1$), and $R(t)$’s are the receiver functions. The choice for $W$’s is subjective and I used trial and error to come up with the combinations of $W$’s where the phases will be most consistent in the stacks. The preferred estimates for the $H$ and $V_p/V_s$ are when the three phases stack coherently producing the maximum amplitude.
Inherent in the procedure is the assumption of $V_p$ for the crust. Although the dependence of crustal thickness on crustal velocity is small (Zhu and Kanamori, 2000), (about 0.5 km change in thickness for a 0.1 km/s uncertainties in $V_p$), I tried $V_p$ values from CRUST 2.0 model, average $V_p$'s for island arc crusts (Condie, 2005), and $V_p$ estimated from the geology of the study region to facilitate comparisons with the global model as well as to determine the range of thickness values under the station.

The advantage of the stacking procedure is that the receiver function transforms directly into the $H$-$\kappa$ domain, which avoids the need to identify the phases and to pick their arrival times (Zhu and Kanamori, 2000). Implicit in the procedure is the assumptions that the underlying Moho surface is planar and the layers are laterally homogenous (e.g. Zandt et al., 1995; Julia and Mejia, 2004). In regions with dipping interfaces or with a gradational contact between the crust and the Moho, the procedure may give inaccurate results. Therefore large sets of azimuthally, well-distributed data are necessary to overcome the limitations of the stacking procedure (Julia and Mejia, 2004).

## 2.5 Shear-Wave Velocity Structure Estimation

Detailed modeling of the Earth’s subsurface requires wide-ranging frequency signals sensitive to the range of heterogeneity associated with the subsurface structures. To estimate the shear wave velocity structure of the PIA, I used receiver functions and fundamental-mode surface wave dispersion data. Both data sets are sensitive to $S$-wave velocity structure. However, both also have inherent resolution limits. The receiver function is sensitive to the velocity contrasts and relative vertical travel times but not to the absolute shear-wave velocity beneath the station (Ammon et al., 1990). The dispersion data are sensitive to the vertical average of the absolute $S$-wave velocity at different depth ranges but not to the details of the velocity layer (e.g. Özalaybey et al., 1997; Julia et al., 2000). Several studies jointly inverted these two data sets to bridge these resolution gaps obtaining highly constrained shear velocity structure than by using each data set individually (e.g. Özalaybey et al., 1997; Du and Fougler, 1999; Julia et al., 2000, 2003). In this study, I used the joint inversion scheme of Julia et al. (2000, 2003). This scheme is
an iterative least-squares approach where the final inverted model is a compromise among three parameters: the influence factor that controls the trade-off between fitting the receiver function and surface wave dispersion data, the smoothness parameter that affects the data fitting and model smoothness, and the importance of \textit{a priori} velocity weights relative to model fits and roughness (e.g. Julia et al., 2000, 2003; Kosarian, 2006). I performed several inversions using a range of values for the three parameters to obtain the final model for the PIA that will give the optimum (in the sense of fitting the data with a relatively simple model) compromise among the parameters. The inversion process assumes that the underlying velocity is nearly flat structure. A thorough description of the methodology can be found in Julia et al. (2000).

For the dispersion data, this study uses the low-frequency model of Harvard (Larson and Ekström, 2001), which is the only dispersion model available that includes the PIA region. The model contains smooth variations as a function of period (30–180 s) for both group and phase velocities of Love and Rayleigh waves. As this model has no information below the 30 s period, the upper crustal structure may not be adequately constrained. To evaluate the effects on the final model, I examine the influence of varying the weights on the receiver function and the dispersion data in the inversion.

The joint inversion requires an initial model but since there is no published result for crustal structure from refraction profiles in the region, I used the isotropic Preliminary Reference Earth Model (PREM) as the \textit{a priori} model (Dziewonski and Anderson, 1981). I started the inversion using PREM with different smoothing and influence factors. After several iterations, I examined visually the robustness of the fit between the receiver function and synthetics calculated from the model inversion results as well as the fit between the observed and predicted dispersion data. Assessing the uncertainty in the solutions from simple, straightforward computation of standard errors is a complicated process, which was discussed extensively by Julia et al. (2000, 2003). Simply put, judging exclusively the robustness of the results based on the calculated errors may not provide clear understanding on the accuracy of the uncertainties or the non-uniqueness of the solutions. Therefore several inversions were performed systematically using a range of weighting parameters. I then compare the resulting final models and selectively remove those
models with inadequate fits between the observations and predictions. I also adjusted accordingly the PREM model in the succeeding inversions based on the available geology information (e.g. putting a thin, low velocity layer for the surface layers to account for sedimentary basins) and the character of the receiver function.

2.6 Results

2.6.1 Stations BAG and TGY, Luzon Island, Philippines

Luzon, the largest island in the Philippines, is bounded on its eastern and western sides by two-opposing subduction zones (Figure 2.2). The eastern boundary is the site of westward subduction of West Philippine Sea basin along the East Luzon Trough (northeastern side) and Philippine trench (southeastern side) (e.g. Hamburger et al., 1983). The Manila Trench bounds the western side where the subduction of the South China Sea crust is taking place (e.g. Cardwell et al., 1980; Hamburger et al., 1983; Hayes and Lewis, 1984). Subduction along the trench generated the volcanic centers along the central and the western part of the island. These centers are part of the 1200 km belt of Miocene and Quaternary tholeiitic to calc-alkaline volcanic complexes from Taiwan (24°N) to Mindoro Island (13°N) (Defant et al., 1989). The island is also traversed by the sinistral Philippine Fault system, which might have been active since the Miocene (Karig, 1983). Forming the basement of the arc are ophiolitic bodies, which are widespread throughout the island and are often overlain by volcanioclastics and marine sediments such as carbonates and shale (e.g. Karig, 1983; Geary et al., 1988).

Seismograms from two broadband stations on the northern (BAG) and southwestern (TGY) Luzon were analyzed to estimate the S-velocity structure and the bulk crustal Poisson’s ratio using receiver functions. Results from the analysis can provide constraints on the crust and upper mantle structure of the island and to improve our understanding of the geological evolution of Luzon arc.
Figure 2.2. Map of Luzon Island showing the two stations (BAG,TGY) used in receiver function analysis and the main tectonic structures in the region. Barb lines outline the subduction zones with the barbs located on the overriding plate. The solid lines trace the approximate location of segments of the Philippine Fault Zone (PFZ) and the arrows show the relative motion directions on either side of the fault. The dots and cross marks are epicenters from Engdahl et al. (1998) (EHB) and National Earthquake Information Center (NEIC) catalog respectively. Dashed lines A–A’ and B–B’ are locations of seismicity profiles in Figures 2.3 and 2.9. The solid lines radiating from the stations are the approximate distances sampled by the $Ps$ multiples used in estimating the $Vp/Vs$ values. The inset figure is the geographic location of the study area. The Luzon Island is shaded in black. WPB – West Philippine Sea Basin, PhT – Philippine Trench.
2.6.1.1 Station BAG

Station BAG (16.41°N, 120.58°E) in the northwestern part of Luzon (Fig. 2.2) sits on Pliocene–Pleistocene Limestone formation, the youngest of the series of deep and shallow marine deposits that overlies the metamorphosed ophiolithic basement rock units (e.g. Yumul et al., 2003). This series of marine deposits reflect major tectonic upheaval in Northern Philippines from pre-Miocene subduction–related marginal basin setting to island arc setting and a shift in subduction from east (proto-East Luzon trough) to the present-day Manila trench in the west (e.g. Yumul et al., 2003). The Wadati-Benioff zone of the South China Sea crust subducting in the Manila Trench is about 50 km below BAG (Figure 2.3).

Figure 2.3. Plots showing the topographic profile (Top) and the vertical distribution of seismicity with the approximate trace of the subducted South China Sea crust (Bottom). Hypocentral depths are from Engdahl et al. (1998) (circle) and NEIC (cross) catalogs. The earthquakes are within the 30–km perpendicular distance on either sides of A–A’ line (see Figure 2.2). The inverted triangle marks the location of BAG.
Seventy-nine receiver functions with high signal–to–noise ratio were calculated from seismograms recorded between 1998 to 2009. Figure 2.4 shows the distribution of earthquakes used in the analysis. There are no data from south to west back azimuths due to the paucity of earthquakes with good quality teleseismic seismograms. Plots of low-pass filtered ($f < 1.25$ Hz) radial and transverse receiver functions by back azimuth and distance are shown in Figure 2.5. Although the transverse receiver functions are usually unstable in the presence of noise (Langston, 1981), these were included in the analysis to aid in identifying dipping layers (Cassidy, 1992). The receiver functions are stacked by back azimuth and by great circle arc distance to examine for azimuthal variations in the lithospheric structure. The receiver functions are also stacked over all back azimuths to suppress spurious arrivals and to enhance those arrivals associated with horizontal or near-horizontal interface (e.g. Cassidy, 1995).

Figure 2.4. Locations of earthquakes (dots) used in receiver function analysis recorded by BAG from 1998 to 2009. Concentric circles mark the distance in degrees from BAG.

In general, BAG appears to be underlain by complex structures that result in varying radial receiver functions with back azimuth and the significant energy on the transverse component receiver functions. Some of this transverse energy might have been introduced by the dipping subducted slab but the signals start relatively early and that suggests near-surface complexity is contributing as well.
The variations in waveforms by back azimuth necessitate the grouping of receiver functions into northwest (NW), northeast (NE), and southeast (SE) cluster for the analysis. All back azimuths appear to be dominated by shallow structure effects based from the large amplitude arrivals on the tangential components. Receiver functions from events in the NW show phase lag (∼0.5 s) in the direct $P$-arrivals. A shift in the arrivals suggests that the events sampled a shallow, low velocity layer (e.g. Sheehan et al., 1995; Zelt and Ellis, 1999). The signature of the shallow layer is also apparent on the large amplitude of the corresponding transverse receiver functions. The SE phase lags show coherent arrivals around 1.3 s but these are absent in the NW and not as distinct and appear to arrive earlier (∼0.9 s) for the NE. The corresponding transverse components show positive polarity from the SE and NE but negative polarity in the NW. These characteristics could indicate an intracrustal structure dipping to the south-southwest direction of the station or sources of crustal scattering. The $Ps$ conversion from the Moho arrives around 4.5 s. The arrivals are low amplitude suggesting the Moho is not a sharp feature, although such a signal could be masked by severe near-surface structure. Another notable set of arrivals in the data sets are the negative and positive polarity arrivals near 5.5 s and 7.5 s. The corresponding phase arrivals on the transverse component show positive polarity from the NW while negative polarity from the NE and SE. These are interpreted as $Ps$ phases generated at the top of the east–dipping, subducted South China Sea crust and oceanic Moho, respectively. Of course, the reader should keep the complexity of the signals in mind when assessing these tentative interpretations.

Taking into account the differences in the receiver functions with directions, the H-$\kappa$ stacking was performed by back azimuths to estimate the crustal thickness and Poisson’s ratio. Despite the signal complexity, the stacking procedure seems to produce reasonable results, suggesting that the inherent averaging is effective in extracting some value from the radial receiver functions. Figure 2.6 shows the results for the three back azimuths for a $Vp$ of 6.6 km s$^{-1}$. The values $w1$, $w2$, and $w3$ are the weights applied to the $Ps$ phase and multiples ($PpPms$, $PpSms + PsPms$). Trade off between depth and velocity exists between the initial input velocity and $Vp/Vs$ ratio in the stacking analysis therefore a strong negative correlation should be expected, which is given by the $corr$ value. The crust
Figure 2.5. Plots of radial and transverse receiver functions stacked by back azimuth (BAZ) and great circle arc (Garc) distance. The bottommost receiver functions on both panels are weighted stacks from all back azimuths bounded by ±1 standard deviation (dashed lines). Vertical dashed lines marked the approximate arrivals of important phases discussed in the text. $P_{sc}$ – intracrustal converted phase, $P_{sm}$ – Moho phase, $P_{SCS}$ – converted phase from the slab, $P_{pPms}$ – $P$-wave multiple.
is thickest in the SE (39 km) and slightly thinner in the other directions (average of 33 km in the NW and NE). The Vp/Vs ratio (or equivalently $\sigma$) also varies with direction; mafic compositions in the NW back azimuth (1.85 or $\sigma=0.294$), intermediate compositions in the NE (1.76 or $\sigma=0.262$), to felsic compositions in the SE (1.72 or $\sigma=0.245$). The crustal thickness values are comparable from all directions and estimates of $\sigma$ suggests that the bulk crustal composition of areas sampled by receiver functions varies with directions. The mafic crustal composition in the NW is consistent to the area sampled by $Ps$ multiples, which is the fore-arc basin 52 km from BAG. In the NE and SE back azimuths, the multiples sampled the calc-alkaline and shoshonitic volcanic centers (e.g. de Boer et al., 1980; Defant et al., 1989) respectively, in agreement with the low values of Poisson’s ratio obtained in this study. The across the arc variations in the crustal composition is consistent with the geochemical data from volcanic centers in Luzon island (e.g. de Boer et al., 1980; Defant et al., 1989).

The 1-D shear wave velocity structure under BAG was estimated by jointly inverting the receiver functions and the surface wave dispersion data. I focused on receiver functions from the SE back azimuth because of the larger number of events and the smaller amplitude of the transverse component compared to the other two back azimuths. The receiver functions were stacked within narrow range of great circle arc distance to account for amplitude variations from different distances. Two frequency bands for receiver functions corresponding to Gaussian width factors 2.5 ($f < 1.25$ Hz) and 1.0 ($f < 0.5$ Hz) were included in the inversion to ensure that the low frequency response, which is less sensitive to small-scale structural heterogeneity, is doubled in importance (Julia et al., 2003). For the dispersion data, the inversion used the Rayleigh wave group and phase velocities. Because of the inherent complexity in the data and the presence of the dipping slab underneath BAG, the inversion was carried down to 120 km depth. Below that depth, the shear velocities are constrained to smoothly converge to the velocities of the initial PREM model during the inversion. At best the results with a laterally uniform earth model can only approximate the true structure, or provide some insight into interesting features worthy of focus in future studies with more sophisticated (i.e. 3-D approaches).

Figure 2.7 is a summary of the joint inversion results. The results show remark-
Figure 2.6. H–κ stacking results for BAG by back azimuth. w1, w2, and w3 are the weighting values for the Ps and multiples. corr is the measure of the correlations between the thickness (h) and Vp/Vs ratio, which should be large and negative in value due to the depth–velocity trade-off. The confidence ellipses on the grid-search panel show the combinations of h and Vp/Vs values where the arrivals of Ps and multiples add constructively among the receiver functions. The plots of receiver functions show the phase matching marked by vertical lines.

Figure 2.6. H–κ stacking results for BAG by back azimuth. w1, w2, and w3 are the weighting values for the Ps and multiples. corr is the measure of the correlations between the thickness (h) and Vp/Vs ratio, which should be large and negative in value due to the depth–velocity trade-off. The confidence ellipses on the grid-search panel show the combinations of h and Vp/Vs values where the arrivals of Ps and multiples add constructively among the receiver functions. The plots of receiver functions show the phase matching marked by vertical lines.

ably good agreement between the predicted and observed dispersion values even though the initial values were obtained from independent source. The predicted receiver functions are well matched visually to the observed data in particular the important arrivals (e.g. Ps, Ps, PsSCS) in the data waveforms. The final inverted model shows a shallow, low velocity layer about 10–km thick, which may reflect the alluvial and limestone deposits mapped (Yumul et al., 2003) in the study area. Below 10 km, the S-wave velocity increases gradationally to around 30–km depth. Velocities at this depth range are lower compared with those from PREM or the values in the average crustal model of Christensen and Mooney (1995). These observations in the velocity trend can be correlated to the series of shallow and deep
marine environment deposits underneath BAG. The inverted model also shows the Moho at $\sim 40$ km depth in good agreement with the H–κ stacking results. A low velocity zone exists below the Moho around 50 km depth, which increases to values higher than PREM below 55 km. These zones are tentatively interpreted to be the slab and oceanic Moho respectively. The results are consistent with the depth of the estimated trace of the slab from seismicity data.

Figure 2.7. Summary of joint inversion results for BAG. The three panels show the receiver function fits (Upper left), the Rayleigh wave dispersion fits (Bottom), and the initial (flattened PREM) and the inverted model (Upper right). The arrows on receiver function plots marked the phases discussed in the text. The gray, shaded region in the model panel is the range of crustal S-wave velocities from Christensen and Mooney (1995) plotted for comparison. $P_{sc}$ – P-to-s intra-crustal phase, $P_s$ – Moho phase, $P_{SCS}$ – $P_s$ converted phase from the slab.
2.6.1.2 Station TGY

The station TGY (14.10°N, 120.94°E) located in the southwestern part of Luzon (Figure 2.2) sits along the outer ridge of Taal caldera underlain by base-surge volcanic deposits (FDSN.org, 2009b). The caldera is about 25 km x 26 km wide depression that contains a small volcano island (Taal Volcano) surrounded by 300–km² crater lake (Miklius et al., 1991). This volcanic complex is part of the Macolod Corridor, a 40–km wide zone of still active Quaternary volcanism and extensional faulting that crosses southwestern Luzon island in a N-S and NE-SW directions (Defant et al., 1988; Förster et al., 1990). To the west of TGY is the east–vergent Manila trench with Wadati–Benioff zone dipping almost vertical down to ∼250 km in the vicinity of the study area (Figure 2.8). Early study by Besana et al. (1995) suggested a crustal thickness of 34 km beneath TGY based on five receiver functions with back-azimuth between 30° – 45°. Their shear-wave velocity model suggests a low-velocity zone at 18 km depth and alternating low– and high– velocity zones below, which they attributed to the complex lower crust.

Seven receiver functions with high signal–to–noise ratio calculated from the 15 available teleseismic events are used in the receiver function analysis to complement the results of Besana et al. (1995) (Figure 2.9). The seismograms were recorded from 1997 to 1998. The transverse receiver functions are included in the analysis to test for the signature of laterally varying structure under the station. Figure 2.10 shows receiver functions plotted as a function of time with back azimuth. The transverse component shows large amplitude arrivals indicative of the complexity of the structure underlying the station. The radial waveforms have small P–wave amplitude and “ringing” like appearance beyond 5 s. These features are signatures of shallow, low-velocity layers, which were absent or were not resolved on the results of Besana et al. (1995). Around 1.5 s is a large, Ps converted phase with an associated large negative amplitude in the transverse receiver function. This phase may indicate a dipping structure or a flat anisotropic layer (e.g. Levin and Park, 1997; Savage, 1998). However, the absence of data from other back azimuths makes it hard to distinguish the cause of this phase. A large arrival with negative polarity around 2.7 s may indicate the presence of low velocity zone (LVZ) below the station. The phase most-easily associate with the crust-mantle transition (Ps) arrived at around 4 seconds.
To estimate the crustal thickness and the average Poisson’s ratio under TGY, the H–κ stacking procedure is performed on receiver functions smoothed with Gaussian filter and with cut–off frequencies of 1.25 Hz (α=2.5) and 0.5 Hz (α=1.0). The two frequency bands are used to evaluate the effects of the complex characters of the receiver functions on the estimates of crustal thickness and σ. Figure 2.11 shows the results of the H–κ stacking procedure for Vp of 6.6 km/s. The crustal thickness under TGY is 34 km consistent with the result of Besana et al. (1995) and Vp/Vs ratio with average of about 1.77 (≡ 0.266), which suggests that the
crust is on average intermediate in composition. The $P$-multiples sampled the region from about 42 km SE of TGY (see Figure 2.2), which from geochemical studies, the volcanic rocks in the region are calc-alkaline to silicic in composition (e.g. Defant et al., 1988; Miklius et al., 1991; Vogel et al., 2006).

The receiver functions were inverted with Rayleigh and Love wave dispersion data to estimate the S-wave velocity structure under TGY. Figure 2.12 shows the summary of the inversion results. The inversion was performed using Rayleigh, Love, and both Rayleigh and Love wave to evaluate the effects of the dispersion data on the inversion. In general, all the results show comparable final velocity model. The visual fit of the predicted receiver functions, in particular the phase arrivals (e.g. $P_{sc}$, $P_{sm}$), with the data are remarkably good. Also, the predicted dispersion data matched well with the observed dispersion. All the three models show a 5–km thick slow velocity surface materials and two low-velocity zones, one below 20 km and the other below 40 km. The inferred Moho depth is around 34 km depth, which is consistent with the H–κ stacking result.
2.6.2 Station DAV, south Mindanao Island, Philippines

Mindanao island is the largest of the islands in the southern portion of the PIA (Figure 2.13). It is composed of two terranes; the eastern Mindanao terrane characterized by rocks of island arc affinity and the western Mindanao terrane with rocks of continental origin (Pubellier et al., 1991; Rangin and Silver, 1991; Hawkins et al., 1985; Sajona et al., 1994). Several studies suggested that these terranes are the northern extension of the arc–arc collision occurring currently south in the Molucca sea region (Cardwell et al., 1980; Moore and Silver, 1983; Nichols et al., 1990; Pubellier et al., 1991; Lallemand et al., 1998; Hall, 2002). The collision and suturing of these two terranes occurred ca. 4–5 Ma ago, which was followed by the
Figure 2.11. H–κ stacking results for receiver functions smoothed with α=1.00 (Left panel) and α=2.5 (Right Panel). w1, w2, and w3 are the weighting values for the Ps and multiples. corr is the measure of the correlations between the thickness (h) and Vp/Vs ratio, which should be large and negative in value due to the depth–velocity trade-off. The confidence ellipses on the grid-search panel show the combinations of h and Vp/Vs values where the arrivals of Ps and multiples add constructively among the receiver functions. The plots of receiver functions show the phase matching marked by vertical lines. The crustal thickness and σ values are consistent from both plots.

initiations (ca. 3 to 4 Ma) of the subduction zones presently bordering the island (Pubellier et al., 1991; Sajona et al., 1993, 1994). These subduction zones, the Sulu trench on the west, the Cotabato trench to the south, and the Philippine trench in the east are the sites of subduction of Early to Middle Miocene Sulu basin, the Eocene Celebes Sea basin, and the Eocene Philippine Sea basin respectively (Acharya and Aggarwal, 1980; Hilde and Lee, 1984; Cardwell et al., 1980; Rangin and Silver, 1991; Müller, 1991). The subduction process produced extensive Pliocene–Quaternary volcanoes with lava compositions ranging from calc-alkaline to shoshonitic (Sajona et al., 1993, 2000a).

2.6.2.1 Station DAV

Station DAV (7.07°N, 125.579°E) sits on the edge of the eastern escarpment of an anticline that is made of metamorphosed coralline limestone (FDSN.org, 2009a). Nearby the station is the Davao basin, a N–S trending sedimentary basin bor-
Figure 2.12. (Top) Plots showing the receiver function (black line) in two frequency bands (f < 1.25 Hz and f < 0.5 Hz) and the fit of the synthetics (light line) using Rayleigh, Love, and both Rayleigh and Love dispersion data in the inversion. (Middle) Plots of the fit to the dispersion data. (Bottom) Initial (flattened PREM) and inverted shear-wave velocity model under TGY. Apparent on all plots are the 5–km thick shallow, low velocity layer and two low-velocity zones (LVZ). The Moho depth is around 34 km, consistent with the H-κ stacking result. The gray, shaded region is the range of crustal S-wave velocities from Christensen and Mooney (1995) plotted for comparison.
Figure 2.13. Geographic map of Mindanao Island showing the location of station DAV as well as the main tectonic structures in the region. Barb lines outline the subduction zones with the barbs located on the overriding plate. The solid lines trace the approximate location of the Philippine Fault Zone (PFZ) and the arrows show the relative motion directions on either side of the fault. The dots and cross marks are epicenters from Engdahl et al. (1998) (EHB) and National Earthquake Information Center (NEIC) catalog respectively. The dashed line A–A’ is the location of seismicity profiles in Figure 2.14. The solid line radiating from the station is the approximate distance sampled by the $Ps$ multiples used in estimating the Vp/Vs value. The inset figure is the geographic location of the study area. Mindanao is shaded in black. ST – Sulu Trench.
dered on the east and west by the Central and Pacific Cordilleras respectively (e.g. Pubellier et al., 1991; Sajona et al., 1997). West of the station is the Mt. Apo (the highest peak in PIA) volcanic center characterized by Pliocene–Quaternary pyroclastics flow deposits and Quaternary lavas (e.g. Sajona et al., 2000a) (Fig. 2.13). To the east of DAV, the slab associated with the subduction along the Philippine trench is \(\sim 100\) km away and can be traced down to 250 km depth (Figure 2.14).

![Figure 2.14](image)

Figure 2.14. Plots of topographic profile (Top) and vertical distribution of seismicity with approximate trace of the subducted West Philippine Sea crust (Bottom). The hypocentral depths are from Engdahl et al. (1998). The earthquakes are within 25–km perpendicular distance on either sides of line A–A’ in Figure 2.13. The inverted triangle marks the location of DAV.

Sixteen seismograms with high signal–to–noise ratio selected from 90 teleseismic events recorded from 1995 to 2009 were used in the receiver function analysis.
Figure 2.15 shows the distribution of source events. The receiver function stacks are plotted in Figure 2.16. In general, the receiver functions from all back azimuths are dominated by large reverberations from the basins. Receiver functions from the NE sampled the Davao basin while those from the SE traversed the Davao gulf. These reverberations have masked the later arriving $Ps$ phase from the Moho resulting to its small amplitude.

![Figure 2.15. Locations of earthquakes (dots) recorded at DAV from 1995–2009. Concentric circles mark the distance in degrees from DAV.](image)

The receiver functions from the SE back azimuth are used in the stacking analysis because of the greater number of observations compared with the other back azimuths. The summary of the stacking analysis is in Figure 2.17 for an initial $Vp$ of 6.6 km $s^{-1}$. The calculated crustal thickness is 24 km and $Vp/Vs$ of 1.82 ($\sigma =0.284$). The value of Poisson’s ratio is consistent with the geology of the area in which the $Ps$ multiples sampled the mafic crust in the Davao gulf (see Figure 2.13).

The shear velocity structure under DAV was estimated by jointly inverting the stack of the receiver functions from the SE back azimuth and the group and phase velocities of the Rayleigh and Love waves. Because of the complicated site response of the waveforms and the large reverberations beyond 5 s, the $S$-velocity is constrained to approach smoothly to the PREM values below 60 km depth. The
cut-off depth values were estimated from the series of inversions performed in this study. Inverting for the velocities deeper than 60 km resulted to unreasonably high \( S \)-velocity values because the inversion is trying to fit the huge amplitudes of the receiver functions beyond 5 s, which are due to the trap waveforms from the basin and not due to the structure underneath. At least before 5 s, the joint inversion results show remarkable fit between the receiver functions and synthetics as well as the very good match of the observed and predicted dispersions values. This is in spite of the inherent complexity of the waveforms. The final model is consistent
Figure 2.17. H–κ stacking result for receiver functions smoothed with α=2.5. w1, w2, and w3 are the weighting values for the Ps and multiples. corr is the measure of the correlations between the thickness (h) and Vp/Vs ratio, which should be large and negative in value due to the depth–velocity trade-off. The confidence ellipses on the grid-search panel show the combinations of h and Vp/Vs values where the arrivals of Ps and multiples add constructively among the receiver functions. The plots of receiver functions show the phase matching marked by vertical lines.

with the geology of the area having a very thick (10 km), shallow, relatively–low velocity layer, indicative of sedimentary basin. More reliable details could be extracted for the uppermost crust with shorter-period dispersion measurements. The shallowest velocities are quite low, and the smoothness constraint smears the strong low-velocity structure into the mid crust. The inferred Moho depth (∼28 km) from the shear wave model agrees with the H–κ stacking results, but must be
considered tentative given the complexity of the signals.

2.7 Discussion and Conclusion

The structure and bulk composition of the crust beneath the three broadband stations in the PIA were estimated using receiver function analysis and joint inversion of receiver function and surface-wave dispersion. The stations are located in regions that underwent different tectonic history, thereby the structures sampled in the analysis have different geologic signatures. Nevertheless, several common characteristics can be derived on the receiver function waveforms. All waveforms show complicated site responses and the important arrivals are often masked by conversions and reverberations in the shallow crusts. These observations imply that the effects of the sedimentary basins are prevalent on the seismic records. Results from the joint inversion provided bounds on the thickness of the basins from 5 km to perhaps as thick as 10 km in the Davao gulf, in the southern PIA. But in spite of the complexity of the waveforms, this study was able to resolve well the crustal thickness (sampled by receiver functions) of several regions within the PIA (average of 35 km) and the provide some Poisson’s ratio constraints on the bulk composition of the crust. In Luzon Island in the northern Philippines, the estimated Poisson’s ratio (∼0.26) suggests an intermediate bulk crustal composition and is consistent with the intermediate to silicic geochemical signatures of nearby volcanic arc complexes (e.g. de Boer et al., 1980; Defant et al., 1988, 1989; Miklius et al., 1991; Vogel et al., 2006). This result is consistent with the idea that the PIA underwent substantial modification of the crust since its generation from initially mafic oceanic crust (e.g. Encarnación, 2004). The crustal thickness and the presence of intra-crustal low velocity zone (LVZ) under TGY agree with the receiver function result of Besana et al. (1995) except that this study prefers a deeper LVZ (∼25 km) than the suggested depth of Besana et al. (1995) (∼18 km). The LVZ depth is also consistent with the model proposed by Vogel et al. (2006) explaining the source of abundant silicic volcanic deposits in the region. In their model the source of the silicic rocks comes from the differentiated magma that ponded in the mid-crust. In DAV, the high Poisson’s ratio indicates a mafic composition of the oceanic crust in Davao gulf. This result may reflect the original composition of
the crust of the nearby region before being modified by the subduction events that occurred in the last 5 Ma (e.g. Sajona et al., 1994).
Figure 2.18. (Top Left) Plots showing the receiver function (black line) in two frequency bands ($f < 1.25 \text{ Hz}$ and $f < 0.5 \text{ Hz}$) and the fit of the synthetics (light line). (Bottom) Plots of the fit to the dispersion data (Group and Phase velocities of both Rayleigh and Love waves). (Top Right) Initial (flattened PREM) and inverted shear-wave velocity model under TGY. Apparent on the plot is the very low, upper shallow layer that reflects the rock deposits of Davao basin sampled by the receiver functions from the SE. The high velocity below 50 km are due to the large reverberations beyond 5 s of the receiver functions and do not necessarily reflect the structure at that depth under DAV. The Moho depth is around 28 km depth, consistent with the $H-\kappa$ stacking result. The gray, shaded region is the range of crustal S-wave velocities from Christensen and Mooney (1995) plotted for comparison.
Chapter 3

Tomographic Imaging of The P–Wave Velocity Structure Beneath the Philippine Island Arc

3.1 Abstract

A total of 4040 local and regional earthquakes and 510 teleseismic earthquakes recorded by 73 stations of the regional seismic networks of the Philippines are jointly inverted to estimate the 3-D $P$-wave velocity structure beneath the Philippine Island Arc. Both hypocentral locations and velocity structure are estimated in the inversion. The tomographic inversion was able to resolve the structures down to 450 km depth with spatial resolution of about 40 km based on checkerboard resolution tests. The prominent features of the tomographic images include the low velocity zones correlating to the overlying volcanic structures and the high velocity zone that more or less coincide with the Wadati-Benioff zones of the subduction zones. The shape of the slab is obscure in some areas primarily because of the limitations in the resolving capabilities of the inversion given the distribution of the events and the stations as well as the parameterizations used in the inversions. Nevertheless, even without a priori information on the upper boundary of the subducted slabs in the initial model, this study was able to image the slabs, which appear 2–6% faster than the mantle velocity values of the IASP91 model.
Slabs commonly extend deeper than the seismicity suggesting they penetrate aseismically at greater depths. The shallow low-velocity anomalies correlate with the fore–arc and intra–arc basins, in the mantle wedge at the top of the slabs, deep in the upper mantle, as well as the segments of the Philippine Fault Zones.

3.2 Introduction

The Philippine Island arc (PIA) is an ideal place to perform a tomographic study. The arc is located in a unique tectonic setting that features diverse structures bearing imprints of a rich geodynamic history (see Figure 3.6). Subduction zones, with Wadati-Benioff zones down to $\sim 250$ km depth, border its sides generating abundant seismicity. Between the subduction zones formed the 1200–km long Philippine Fault system, a NW–SE trending active sinistral fault that essentially traverses the whole PIA (e.g. Barrier et al., 1991). The Philippine Fault is suggested to have been active since 15 Ma in its northern segment and around 4 Ma in its central segment (Karig, 1983; Barrier et al., 1991; Aurelio, 2000). The Philippine Trench and at least the central and southern segments of the Philippine Fault are suggested to exhibit slip partitioning mechanism (Fitch, 1972; McCaffrey, 1992; Aurelio, 2000). Under this mechanism, the Philippine Fault accommodates a third of the oblique convergence occurring at the Philippine Trench and the rest of the displacement is being taken up along the trench (Aurelio, 2000). Subsequently, the Philippine Trench is suggested to have formed due to the oblique convergence (Barrier et al., 1991; Aurelio, 2000). Finally, there are deep ($>300$ km) seismic events occurring in the southern PIA that are unrelated with the present day subduction zones. These events are often associated with the detached slab of the Molucca Sea Plate, which are currently being consumed by the ongoing arc–arc collision in the Molucca Sea region, south of Mindanao (e.g. Acharya and Aggarwal, 1980; McCaffrey et al., 1980).

Voluminous volcanic fields, both active and inactive, are scattered throughout the arc and mostly associated with the subduction zones. Their geochemical signatures and isotopic compositions vary along and across the arc (e.g. Knittel and Defant, 1988; McDermott et al., 1993; Mukasa et al., 1994; Sajona et al., 1997; Castillo, 1996; Sajona et al., 2000b; Castillo and Newhall, 2004). The sources of
the variations are controversial (Castillo and Newhall, 2004), which are either attributed to the amount of melt from the subducted basaltic crust (Sajona et al., 2000b) or to the variations in the amount and type of sediment melts that metamorphized the geochemically depleted mantle source (e.g. Defant et al., 1989; McDermott et al., 1993; Castillo and Newhall, 2004). The spatial distributions of the volcanoes also offer interesting insights to the source underneath. In northern PIA, for example, a series of volcanoes that started as a single chain in southern Taiwan, diverged into two chains around 17.8°N that are distinct in age and geochemistry. Studies suggested that the formations of these two chains and the observed differences are due to the tear in the subducted slab of the South China Sea crust (Yang et al., 1996; Bautista et al., 2001). Another example is the voluminous Pliocene–Quaternary volcanic field in central Mindanao island, south of PIA. The source of volcanisms cannot easily be correlated with the nearby subduction zones because the corresponding trace of the slabs do not extend at depths beneath the volcanic structures (Sajona et al., 1994, 2000a). Geochemical studies suggested that the structures were the magmatic response to the collision event that occurred 4–5 Ma (Pubellier et al., 1991) between the western and eastern part of Mindanao (e.g. Sajona et al., 2000a). This event is suggested to be the northern extension of the ongoing collision between the Halmahera and Sangihe island arcs in the Molucca Sea region.

Seismic tomography can image most of the features of the PIA described above. The intense seismic activity in the arc offers abundant earthquake data with wide range of depth from shallow to very deep (∼650 km). The number and depth distribution of the earthquakes can ensure enough crisscrossing rays to provide detailed image of the crust and upper mantle. In addition, this study incorporates roughly vertically propagating teleseismically recorded earthquakes with the local and regional events in the inversion simultaneously. In doing so, the inversion can provide images of the deeper structure and can further improve the resolution of the shallow structures as have been observed from similar studies in other subduction zones (e.g. Zhao et al., 1994, 1995, 1997; Gorbatov et al., 1999; Wang et al., 2009).

A fundamental issue in the seismotectonic study of the PIA is the paucity in the detailed information of the subsurface structures. Previous tomographic studies that covered the PIA are mostly of global and of wide, regional scales (e.g.
Dziewonski and Woodhouse, 1987; Fukao et al., 1992; Puspito et al., 1993; Bijwaard et al., 1998; Gorbatov and Kennett, 2003; Li and van der Hilst, 2010). The regional tomography studies largely focused on the velocity structure of the upper mantle and deeper structures (Fukao et al., 1992; Puspito et al., 1993; Lallemand et al., 2001; Gorbatov and Kennett, 2003; Li and van der Hilst, 2010) and have poor resolutions at the shallow regions. Fukao et al. (1992) imaged the P-wave structure beneath the Western Pacific region with horizontal spatial resolution of 150 km and vertical resolution of 50 km. Their results show that the southwestern PIA together with Japan, Ryukyu, and Taiwan arcs are underlain by intensive low-velocity anomalies shallower than 350 km (Figure 3.1). However because of poor resolution under the PIA, the arc is not further discussed in the paper. Lallemand et al. (2001) have interpreted some of the tomography results for northern Luzon island (Figure 3.2) using the global tomography model of Bijwaard et al. (1998). On their vertical slice along ∼20.0°N latitude, they suggested the South China Sea basin slab, which is subducting along Manila Trench, is dipping vertically and may have been overturned beneath the basin at the transition zone depth. They estimated the slab to be 880 km long and perhaps may reach 1180 km in length if the overturned feature is indeed the slab itself. They also pointed out the slab appears disconnected to the surface at around 100 km depth. South of this slice (∼18.5°N) shows a significantly different tomographic image. The slab is dipping shallowly to the east and is 660 km long (Lallemand et al., 2001). Similar with the previous slice, it is disconnected to the surface around 70 km depth. The regional body wave tomography of Gorbatov and Kennett (2003), also in the Western Pacific region, showed the image on the eastern side of the PIA. The positive velocity perturbation feature from 200 km down to 1500 km depth under central PIA (Figure 3.3) is correlated to the slab of the Philippine plate subducting along the Philippine Trench. The southern part of the PIA is included in the tomography study of Puspito et al. (1993) in the Indonesia region. The spatial resolution of their study is 0.5° and 33–50 km for the vertical. They were able to image the slabs under the western subduction zones of the PIA down to 300 km depth and the slab under the Philippine trench in the east to 200 km, which may extend to 450 km depth (Figure 3.4). The shallow subsurface structures of the PIA were not resolved due to poor resolution. The recent study by Li and
van der Hilst (2010) on the regional tomography study of East Asia includes the northern part of the PIA. Their tomographic image showed fast velocity down to \( \sim 200 \) km under the northern Luzon island, which is underlain by pronounced low velocity region down to about 410 km depth (Figure 3.5).

![Figure 3.1](image1.png)

**Figure 3.1.** Vertical cross section of the slowness anomalies from Fukao et al. (1992) discussed in the text. The profile cuts across the central Philippines.

![Figure 3.2](image2.png)

**Figure 3.2.** Vertical cross section of \( P \)–wave anomalies under northern Luzon from the global model of Bijwaard et al. (1998). The figures are modified from Lallemand et al. (2001). M.T.– Manila Trench.

Clearly the previous tomographic studies have discrepancies on the actual length of the slabs under the PIA and all of them obtained poor resolution of the shallow subsurface structure of the arc. In this study, local, regional, and teleseismic events are inverted simultaneously in the tomographic inversion to model
Figure 3.3. Location of the vertical cross sections (Left) of bulk–sound speed (Middle) and shear wave–speed (Right) model for the Philippines by Gorbatov and Kennett (2003). The figures are modified from Gorbatov and Kennett (2003). The arrow points to the image of the Philippine Slab (PS).

Figure 3.4. (Left) Map of the Indonesian region and vicinity showing the seismicity and location of the cross-section of the anomalies under the central Philippine island (Right). The figures are modified from Puspito et al. (1993). (Right) Velocity anomalies along line AA’. The circles denote the earthquake hypocenters.

in more detail the 3-D P-wave velocity structure of the PIA. The resulting tomographic images are integral in fully understanding the subsurface structures like the morphology of the slabs, the source and locations of deep volatiles, and the relationships between crustal structures and crustal earthquake distribution. Resolving the detailed subsurface structure can provide insights in the formation and evolution of the PIA as well as its relationship with the other island arcs in the
Western Pacific region.

### 3.3 Tectonic framework of the PIA

The PIA has long been recognized as one of the best laboratories to study the evolution of arc systems and its contribution to continental growth. The arc is a mature complex and is in the process of being accreted to the continental Eurasian margin (e.g. Encarnación, 2004). At present, the PIA serves as a complex boundary between the Sunda Plate and the Philippine Sea Plate (Figure 3.6) (e.g. Rangin et al., 1999; Bird, 2003). Convergence between the two plates partitions within the PIA resulting to intense seismic activity and volcanism. The arc features diverse tectonic structures such as subduction zones of opposite polarity, intra-arc sinistral fault zones traversing its entire length, volcanic chains, continent-island arc collision zones, and sedimentary basins (e.g. Holloway, 1982; Karig, 1983; Cardwell et al., 1980; Barrier et al., 1991; Bischke et al., 1990; Aurelio, 2000). In terms of seismic activity, the PIA can be divided into aseismic North Palawan block and the seismically active Philippine Mobile belt (e.g. McCabe et al., 1985) (Figure 3.7). The North Palawan block is suggested to be a rifted fragment of Eurasian plate that collided with the western PIA during the Miocene (e.g. McCabe et al., 1982).
Figure 3.6. Regional tectonic setting of the Philippine Island arc. Dots are earthquakes from Engdahl et al. (1998) catalog and are color coded by hypocentral depths. The longer arrows show the motion (cm/yr) of the Philippine Sea plate relative to the fixed Sunda plate (SU) while the shorter arrows show the motion of SU relative to the fixed Eurasian plate. The plate motion are from the global plate motion model, MORVEL (DeMets et al., 2010). Solid lines delineate the plate boundaries (Bird, 2003). Country names are in italics. MS – Molucca Sea Plate, BS – Banda Sea Plate, BH – Birds Head Plate.

3.3.1 Subduction zones, marginal basins, and volcanoes

Oppositely dipping subduction zones bound the western and eastern sides of the PIA. On the western side, the Manila, Negros, Sulu, and Cotabato trenches are the sites of subduction of the South China Sea, Sulu Sea, and Celebes Sea marginal basins. The East Luzon Trough and the Philippine trench border the eastern side, which mark the northwestward subduction of the West Philippine Sea basin. These subduction zones have generated volcanic structures that erupted materials with
Figure 3.7. Generalized seismicity (circle) map of the PIA from Engdahl et al. (1998) and Pacheco and Sykes (1992) catalog. Sizes of the circles show the rupture area (in km$^2$) of the earthquakes (if seismic moment, Mo, available) assuming a circular rupture and static stress drop of 30 bars. Subduction zones are marked by lines with sawtooth symbols that point to the overlying block. The thin lines trace the Philippine Fault system. Labeled are the subduction zones, marginal basins, and structures discussed in the text.
wide range in compositions from tholeiitic to calc-alkaline to shoshonitic. The wide spectrum of composition reflects the variations in the amount of contributions from the slab and subducting sediments in modifying the mantle source regions.

3.3.1.1 East Luzon trough

The East Luzon Trough (ELT) marks the eastern boundary of northern Philippine region (Figure 3.8). The trough is suggested to be at the very early stage of westward subduction of West Philippine Sea basin that is taking place at an older subduction zone, the proto-East Luzon Trough (p-ELT) (e.g. Fitch, 1972; Hamburger et al., 1983; Lewis and Hayes, 1983). There is not enough information on the p-ELT other than it was implied to explain the Eocene volcanics and plutons on the northeastern part of the PIA as well as the relict subduction complex offshore that cannot be attributed to the present day ELT (e.g. Lewis and Hayes, 1983; Yumul et al., 2003). The Early Miocene was proposed to mark the waning stage of p-ELT and the initiation of subduction of the South China Sea basin along the Manila Trench on the western side of the arc (Yumul et al., 2003). The mechanism that caused the subduction polarity reversal from p-ELT to the Manila Trench is poorly constrained. Some workers proposed the collision of the buoyant Benham Rise along p-ELT while others suggested the collision of North Palawan micro-continental block to central Philippines (Yumul et al., 2003).

3.3.1.2 Manila trench

Bordering the western side of the PIA is the Manila trench, which is the site of the eastward subduction of the late Oligocene–early Miocene South China Sea (SCS) basin (Figure 3.8) (e.g. Cardwell et al., 1980; Hayes and Lewis, 1984; Encarnación, 2004). The Wadati–Benioff zone can be traced down to ∼250 km. Estimates of average, long term convergence rate along the trench range from 1 cm/yr in the northern part to about 7 cm/yr in the south (Rangin et al., 1999). The northern and southern terminations of the trench are marked by collision zones between the PIA and the eastern margin of continental Sunda plate and the Palawan microcontinental block respectively (e.g. Hamilton, 1979; McCabe et al., 1982, 1987; Karig, 1983; Stephan et al., 1986). The subduction process, which probably started in
Figure 3.8. Map showing the different structures in the northern part of the PIA discussed in the text. Sizes of the circles show the rupture area of the earthquakes (color coded by hypocentral depths) assuming a circular fault and a static stress drop of 30 bars. The volcanoes with the associated subduction zones are labeled (see LEGENDS). The dotted lines mark the approximate location of the collision zones on the northern and southern ends of Manila trench. Dashed lines marked the approximate area coverage of the West Volcanic chain (WVC) and the East Volcanic chain (EVC) of Yang et al. (1996). Inset figure shows the geographic location of the area relative to the PIA.
the Late Oligocene to Early Miocene \cite{Hayes1984, Encarnacion1993} have generated the Miocene (10 Ma) to Recent Luzon arc \cite{deBoer1980, Richard1986, Defant1989, Defant1990, Yang1996}. The Luzon arc is a 1200–km chain of volcanoes from the Coastal Range of Taiwan (24$^\circ$N) to Mindoro island (13$^\circ$N) \cite{Defant1990}. It is suggested to consist of five groups or segments that are geochemically distinct from each other \cite{Defant1989, Defant1990}. The geochemical signatures of the volcanic rocks vary from tholeiitic to calc-alkaline reflecting the roles of different factors like degree of partial melting, mantle source composition, and the involvement of subducted sediments in the magmatic generation processes \cite{deBoer1980, Richard1986, Defant1989, Defant1990}.

The northern segment of Luzon arc is recognized as double arc structures consisting of the Western Volcanic Chain (WVC) and the Eastern Volcanic chain (EVC) \cite{Yang1996}. These two chains are 50 km apart north of 18$^\circ$N and converged together near 20$^\circ$N \cite{Yang1996}. The WVC is older (Miocene to Pliocene) and structurally eroded while the younger EVC consists of active Quaternary volcanoes, with well-developed cone shape structures, and enriched mantle sources \cite{Yang1996}. Based on the different age and geochemistry between the two chains, Yang et al. \cite{Yang1996} suggested that the volcanic center may had shifted eastward from under the WVC to EVC. The eastward shift is suggested to be the result of the change in the dip of the SCS slab due to the collision and eventual subduction of the Scarborough seamount, an extinct ridge, to Manila trench \cite{Yang1996}. The subduction of this buoyant ridge resulted to the temporary cessation of subduction and a tear in the slab around 19–20$^\circ$N after the resumption of subduction. Bautista et al. \cite{Bautista2001} however argued that the eastward shift is caused by the subduction of the buoyant plateau around 20$^\circ$N. The inferred tear in the slab is located further south ($\sim$16$^\circ$N) right at the site of ridge subduction.

3.3.1.3 Negros trench

The Negros trench is located on the west central part of the PIA (Figure 3.9) where the Southeast (SE) Sulu Sea basin is subducting in southeastward direction \cite{Hamilton1979, Sajona1993, Schlueter1996}. The SE sub–basin is a
back-arc basin that formed by seafloor spreading either during the early Oligocene (30–35 Ma) to late Miocene (10 Ma) (Roeser, 1991) or during the early Miocene (20 Ma) (Müller, 1991). The initiation of subduction is inferred to start in the late Miocene or early Pliocene (e.g. Hall, 1996; Sajona et al., 2000b). The subduction process generated ∼300 km long arc, which consists of six Pliocene to Quaternary stratovolcanoes (Solidum et al., 2003). These volcanoes erupted wide range of lava types from calc-alkaline to slightly shoshonitic basalts to dacites reflecting variable proportions of sediment contributions and oceanic–derived fluid fluxes from the subducting slab (Solidum et al., 2003). At present, the Wadati–Benioff zones associated with the slab can be traced to ∼150 km depth. Convergence rate along the trench ranges from 3.6–5.4 cm/yr (Rangin et al., 1999).

3.3.1.4 Sulu trench

The SE Sulu sub-basin is also subducting southward along the Sulu trench, which generated the Pliocene to Quaternary volcanoes of the Sulu arc and the Zamboanga peninsula (Figure 3.9) (Sajona et al., 1996; Castillo et al., 2002). Geochemical studies on the arc lavas showed enrichment in high field strength elements (Sajona et al., 1996; Castillo et al., 2002). The enrichment is either attributed to the melting of the subducted Sulu Sea basaltic crust (Sajona et al., 1996) or to the melting of the enriched component in the mantle wedge (Castillo et al., 2002). The subduction probably started from 15 to 10 Ma (Rangin and Silver, 1991). Currently, the Sulu trench is characterized by sparse seismicity and GPS data suggest a convergence rate of 2.8 cm/yr (Rangin et al., 1999).

3.3.1.5 Cotabato trench

The eastward subduction of the Eocene (43 Ma) Celebes basin occurs along the Cotabato Trench with a convergence rate of 3.5 cm/yr (Figure 3.9) (Shyu et al., 1991; Rangin et al., 1999). The associated Wadati–Benioff zone is poorly defined, which possibly extends down to ∼100 km depth.
Figure 3.9. Generalized tectonic map of the southwestern region of the PIA. Sizes of the circles show the rupture area of the earthquakes (color coded by hypocentral depths) assuming a circular fault and a static stress drop of 30 bars. The volcanoes with the associated subduction zones are labeled (see LEGENDS). Inset figure shows the geographic location of the area relative to the PIA.

3.3.1.6 Philippine trench

The Philippine trench marks the westward, oblique subduction of the Eocene (52–60 Ma) West Philippine Sea basin between 16°N and 2°N (Figure 3.10) (Seno, 1977; Cardwell et al., 1980; Hamburger et al., 1983; Hilde and Lee, 1984). At its northern end, it is connected to the East Luzon Trough via a left-lateral transform
fault (Lewis and Hayes, 1983; Hamburger et al., 1983). The trench is suggested to be a young feature because of its sharp morphology, very narrow forearc, shallow Wadati–Benioff zone (∼250 km), and poorly developed accretionary prism that diminishes abruptly to the south (Cardwell et al., 1980; Hamburger et al., 1983; Bloomer and Fisher, 1988). The associated volcanic centers lined the eastern margin of the PIA from Bicol in the north to Leyte Island in its central segment down to northeastern Mindanao in the south but are absent towards southeastern Mindanao (Hamilton, 1979; Cardwell et al., 1980; Hamburger et al., 1983; Sajona et al., 1994, 1997; Ozawa et al., 2004). Based on the aforementioned characteristics, the trench is said to be propagating southward towards the Molucca Sea region (Cardwell et al., 1980; Hamburger et al., 1983; Hall, 1987; Ozawa et al., 2004).

The timing of the initiation of subduction along the Philippine Trench is still controversial. Some studies suggest the trench is <4 Ma based on kinematic modeling, length of the subducted slab (∼250 km), and current estimates of the convergence rate (6–8 cm/yr) (Karig and Sharman, 1975; Cardwell et al., 1980; Aurelio, 2000). Recent study on the K–Ar ages of the associated volcanoes on the northern part in Bicol region however gave an older initiation age of ∼8 Ma (Ozawa et al., 2004).

3.4 Data

This study uses two sets of data to perform the joint tomographic inversion; the arrival times of local and regional events and the relative travel time residuals of teleseismic events. Local and regional events can provide images of the crust and upper mantle structures, the mantle wedge, and the region of the upper boundary of the slab but only down to the depth of the hypocenter or to the depth of propagation of the diving rays (Zhao et al., 1994). The teleseismic data can image the deeper structure but usually can not resolve the shallower structure because the rays propagate in nearly vertical direction and usually do not crisscross well at the shallow surface. Figure 3.11 shows the ray paths for the local and regional events and for the teleseismic events used in this study. Each data set features distinctive ray path geometries and characteristic sampling depths. The ray paths of the local and regional events crisscross well at regions above the hypocenter
Figure 3.10. Generalized tectonic map of the eastern part of the PIA. Sizes of the circles show the rupture area of the earthquakes (color coded by hypocentral depths) assuming a circular fault and a static stress drop of 30 bars. The volcanoes associated with the Philippine trench are labeled. Inset figure shows the geographic location of the area relative to the PIA.
while the crisscrossing teleseismic rays occur at much deeper depth. Thus, jointly inverting the data sets preserves the advantages each data offers and at the same time bridges the resolution gaps inherent among them if used individually (Zhao et al., 1994).

Figure 3.11. Perspective view of $P$-ray paths of local and regional (A) and teleseismic (B) earthquakes. Triangles marked the locations of the 73 seismic stations.

The events used in this study were recorded by the Philippine Institute of Volcanology and Seismology seismic networks from 1998 to 2009. The network consists of 73 seismic stations that are distributed throughout the island (Figure 3.11). All arrival times were selected from the International Seismological Catalog (ISC) (International Seismological Centre, 2001).

3.4.1 Local and regional earthquake data

To select the local and regional travel time observations, which are often quite noisy, I applied a preliminary selection process based on the following criteria:

1. Earthquakes that occurred in the area 5–21°N, 116–130°E
2. Distance between event and receivers is ≤ 1600 km
3. A given event should be recorded by minimum of eight stations
4. Offshore events with hypocentral depth < 5 km are excluded

The selection process eliminates events that are poorly recorded or outside the region of interest. The data are subjected further to a more rigorous elimination process, which include plotting the arrival times with distance to examine the scatter in the arrivals. Those arrivals that are widely scattered (severe outliers) in the plot were removed from the set. Figure 3.12 shows the 4040 events selected in the process yielding 55,878 arrival times. On the depth distribution plot of the events, majority of the earthquakes occur shallower than 250 km but there are also relatively small number of deep events (>350 km), which possibly originated from the inferred remnant slab under the southern PIA (e.g. Acharya and Aggarwal, 1980; McCaffrey et al., 1980).

### 3.4.2 Teleseismic earthquake data

The following criteria were used to select the events for the teleseismic data set:

1. Earthquakes with epicentral distance (great circle arc) between $27^0$ to $90^0$ from the seismic network. Using this distance range can avoid the effects on the travel times of complex structures of the upper mantle and the core–mantle boundary.

2. Minimum magnitude threshold of $M_b$ 5.5

3. Each event recorded by at least eight stations

Figure 3.13 shows the geographic location of the 510 selected events. The azimuthal coverage is fairly good except in the east azimuth. This study uses the relative travel times of teleseismic arrivals in order to minimize the effects of uncertainties in origin times and hypocentral locations as well as the velocity heterogeneity outside the modeling space. The relative residuals are calculated from the observed travel times following the method of Zhao et al. (1994). Initially, the theoretical teleseismic travel time of the ray between the hypocenter and station receiver is calculated using the IASP91 1-D earth model (Kennett and Engdahl, 1991). The calculation also finds the intersection of the ray with the bottom plane of the model space. The travel times are corrected for the ellipticity of the
Figure 3.12. (A) Hypocenters of local and regional events color coded by depths. The inverted triangles are the seismic stations that recorded the events. (B) Travel time plots as a function of epicentral distance (cross marks) color coded based on hypocentral depths. Superimposed for comparison are the theoretical travel times (circles) for events with depths of 10, 100, 200, 400, and 600 km calculated using the 1-D model of the tomography. Clearly the 3-D heterogeneity of the structures beneath PIA has strong influence on the arrival times. (C) Histograms of the depth distributions of the earthquakes

Earth (Dziewonski and Gilbert, 1976). Once the intersection of the ray and the bottom plane is located, the ray path between the intersection and the receiver is determined using a 3-D ray tracer (Zhao et al., 1994). The travel time residual $t_{ij}$, from the $j$–th earthquake to the $i$–th station is

$$t_{ij} = T_{ij}^{obs} - T_{ij}^{pred}$$ (3.1)
Figure 3.13. Epicenters of teleseismic events used in this study color coded by hypocentral depth. The location information are either from the catalog of Engdahl et al. (1998) or from the ISC (International Seismological Centre, 2001).

where $T_{ij}^{\text{obs}}$ and $T_{ij}^{\text{pred}}$ are the observed and predicted travel times, respectively. From equation 3.1, the relative travel time residual, $T_{ij}^{\text{rel}}$, is calculated from each station residuals by removing the mean residual, $T_{ij}^{\text{ave}}$, (average over all stations for each event): 

$$T_{ij}^{\text{rel}} = t_{ij} - T_{ij}^{\text{ave}}$$  \hspace{1cm} (3.2)

where the mean residual is,

$$T_{ij}^{\text{ave}} = \frac{1}{n_j} \sum_{i=1}^{n_j} t_{ij}$$  \hspace{1cm} (3.3)

and $n_j$ is the number of observed arrivals for the $j$th event.
Figure 3.14 shows the average relative residuals of all teleseismic events for each of the stations. Plots of active and inactive volcanoes as well as mapped areas of exposed ophiolitic terranes are included in the figure. 52% stations showed a delay in average residual travel times. The patterns of the delayed and early arrivals are, in general, consistent with the major features of the structures in the PIA. Commonly, stations with delayed arrivals coincide with the locations of nearby volcanic fields especially those in Luzon and Mindanao islands. In Mindanao, except for few stations in the east coast area, the stations showed delay arrivals. The delays can be correlated with those rays that passed through the low velocity regions associated with volcanic activity. Delayed arrivals are also observed on Palawan island, which is an exotic (terrane) to the Philippine arc (e.g. Encarnación, 2004). Early arrivals are observed on some stations near the coast in western Luzon and southeastern Mindanao. This early arrivals can be attributed to the high velocity, subducted slab under the area. The stations nearby the ophiolitic terranes or at a distance from the volcanoes also exhibit early arrivals. A few stations however do not show simpler patterns, which may reflect the localized effects of nearby structures.

3.5 Methodology

The TOMOG3D code of Zhao et al. (1994, 1992) was used to do the joint tomographic inversion of absolute travel times of local and regional events and the residuals of the travel time of teleseismic events. This method is popular in tomographic studies of subduction zone regions because the code can incorporate seismic discontinuity structures like the Moho surface or the upper boundary of the slab in the initial model. Detailed description of the method can be found in Zhao et al. (1994, 1992). The following are summaries of the main features of TOMOG3D.

1. A 3-D inhomogeneous model with seismic velocity discontinuities (SVD) can be included as a priori information in the inversion. The SVD includes Conrad (mid-crust), Moho discontinuities, and the top of the subducted slab.

2. The code relies on a pseudo–bending (Um and Thurber, 1987) and Snell’s
Figure 3.14. Average of relative travel time residuals distributions of teleseismic events from all azimuths at a given station. Shaded areas are ophiolitic terranes adapted from Encarnación (2004); Yumul Jr (2007).

3. At any point in the model, the velocity perturbation is calculated by linearly interpolating the perturbations at the eight grid nodes surrounding that point.

4. The LSQR algorithm of Paige and Saunders (1982) is used to solve the large
and sparse system of equations, which include damping (minimum length) and smoothing constraints.

5. The epicenters of local and regional events can be relocated prior the inversion or in an iterative sense, alternating location and velocity estimation.

A 3-D grid of nodes was set up throughout the PIA (Figure 3.15). A checkerboard resolution test (CRT) was employed to choose the optimum grid spacing that would provide adequate ray coverage and that would better resolve the structures in the region. Several resolution tests with different node separations (horizontal and vertical) were conducted following the method of \cite{Zhao and Hasegawa, 1993}.

In the CRT, velocity perturbations of ± 6% were assigned alternately at every grid point to construct the checkerboard model. Afterwards, the theoretical travel times of rays passing through this model are calculated to get the synthetic data. Gaussian errors with 0 s mean and 0.1 s standard deviation were added to the synthetic travel time data to account for possible errors embedded on the data. The quality of the resolution test is evaluated based on how well the inversion recovered the original checkerboard pattern (Figure 3.15). Examples of CRT results for grid spacing of 0.33°, 0.4°, and 1° are presented in Figures 3.16, 3.17, and 3.18. In this study, the horizontal grid spacing of 0.4° and vertical spacings of 20–50 km are the minimum spacings that produced the best recovered checkerboard pattern. The pattern is well-resolved down to depth of 450 km. I also performed other resolution tests by adding uniformly distributed random errors with standard deviation of 0.1 s on the synthetic data. Including non-Gaussian errors generally results in less resolution than inferred from CRT’s with added Gaussian errors. The resulting CRT patterns only differ slightly from the tests with added Gaussian noise but the general overall resolvable regions are similar.

3.5.1 Inversion of the Observations

To date, there are no regionally applicable published, $P$–wave velocity models for the PIA. To construct the initial 1-D model, I used the velocity models obtained from the receiver function analysis to constrain the crustal layer. The mantle layers were derived from the global IASP91 model \cite{Kennett and Engdahl, 1991}. This combined receiver function–derived crustal model and the unperturbed mantle
Figure 3.15. Horizontal (top left) and vertical (bottom left) grid node configurations used in this study. The horizontal and vertical spacings are 0.4° and 20–50 km respectively. The right panels are samples of the input checkerboard pattern using ±6% perturbation.
Figure 3.16. Checkerboard resolution test results for horizontal grid spacing of 0.33° at different depth slices.
Figure 3.17. Checkerboard resolution test results for horizontal grid spacing of 0.4° at different depth slices.
Figure 3.18. Checkerboard resolution test results for horizontal grid spacing of 1.0° at different depth slices.
layers, here referred to as PHMOD (Figure 3.19), can provide a better basis for the structure under the PIA. The crustal model is averaged to come up with a simple, three-layer model; the upper layer corresponding to the sedimentary basin, the middle layer for the upper crust, and the bottom layer for the lower crust. The P-wave velocity ($V_p$) of the upper layer is 3.5 km/s. For the upper and lower crust velocities, the $V_p$ values are 6.0 and 6.7 km/s respectively. Two seismic velocity discontinuities, corresponding to the top of the lower crust (13 km) and the Moho (30 km) are included in PHMOD.

Tomographic studies in subduction zones that included these discontinuities and also the upper boundary of the subducted slab in their initial model suggested that the hypocenters are better located and the reduction in the final travel time root-mean square (RMS) residuals are larger than without incorporating these discontinuities (e.g. Zhao et al., 1994; Gorbatov et al., 1999). The PHMOD however features a flat, instead of a complex shape discontinuities because of the absence of information on the lateral depth variations of the discontinuities. Some studies used and compared both flat and varying depths in crustal discontinuities and observed only minor difference in the RMS and some differences in amplitudes of the velocity anomalies but the general patterns of velocity distributions are the same (e.g. Serrano et al., 2002; Sun et al., 2008). The upper boundary of slabs are not included in PHMOD because of the complex tectonic characteristics of PIA. Subduction zones occur on either side of the arc while the corresponding Wadati-Benioff zones are only well defined in some areas. Not including the slabs in the initial model however will allow to evaluate whether or not the tomographic inversion can resolve these features. In general, the velocity and layer thickness of PHMOD are comparable with the 1-D velocity models used in other subduction zone regions like Japan (e.g. Zhao et al., 1994), Kamchatka (Gorbatov et al., 1999), Alaska (Zhao et al., 1995), and Taiwan (Wang et al., 2009).

TOMOG3D uses damping parameter and smoothing regularization to stabilize the inversion (Zhao, 2001). To choose the appropriate weights for these associated regularization parameters, I conducted an empirical approach to obtain the optimum damping factor for the least-squares procedure. Using the data and the initial velocity model, a suite of inversions were run with different damping factors and smoothing values to find a good compromise between the reduction in travel
Figure 3.19. (Left) Initial 1-D P-wave velocity model (PHMOD) used in the tomography inversion. The crustal layer is averaged from the shear velocity model obtained from the joint inversion of receiver function and surface wave dispersion conducted by this study. The mantle layers are adopted from the IASP91 model (Kennett and Engdahl, 1991). (Right) The upper 50-km layers of PHMOD.

time residuals and the smoothness of the obtained 3-D velocity model. Figure 3.20 shows the results of the tests. For the smoothing, the value at the approximate junction where a change in the slope of the curve occurs (~0.005), is chosen for this study. Using this preferred smoothing, several inversions were run with different damping parameters to come up with a trade-off curve between the norm of the solution and the root-mean-square (RMS) travel time residual. The preferred damping value was chosen based on the balance between the reduction of the residual and the simplicity of the obtained velocity model, which is 10.0. The RMS of travel time residuals before the inversion is 0.75 s. After the inversion, where the local and regional earthquakes are relocated and using the preferred smoothing and damping values, the RMS is reduced to 0.57 s. The variance reduction in travel time residual is 41%.
Figure 3.20. (Left) Plot of smoothness versus travel time residuals for different damping values. The arrow points to the preferred smoothing value used in this study. (Right) Trade-off curve for several damping values, as indicated on top of the circles, using the smoothing chosen from the left panel. The arrow marks the damping value used in this study.

3.6 Results and Discussions

Prior to this study, tomography results that covered the PIA region by far were only able to image structures deeper than 200 km. The shallower structures are resolved poorly because of the inherent limitations in the resolution of the previous tomography studies. To obtain a better resolution of the 3-D structures of the PIA, local, regional, and teleseismic events are inverted simultaneously in this study. In the following sections, the tomographic images of the subsurface structures of the PIA are presented in Figures 3.21 to 3.23 (depth slices).

The shallower structures of the arc (depth <40 km) exhibit large regional variations in $P$–wave velocity ($V_p$) perturbations but the main anomalies correspond roughly to the known structures at the surface. At 10–km depth, low-velocity (low–$V$) zones, indicated by negative $V_p$ perturbations, correlate with the fore-arc and intra-arc basins as well as with the regions at or near the volcanic structures. The low–$V$ zones are also observed along segments of the Philippine Fault zones (PFZ). In fact, a prominent low velocity is confined around 13°N right where the fault bifurcates. The presence of low–$V$ beneath the intra-arc fault zone has been observed in other regions like the Median Tectonic Line, which is the largest fault
zone in Japan (Sun et al., 2008). The low–V is attributed to the weak materials in the fault zone (Sun et al., 2008).

A similar trend in the distribution of the anomalies can be observed in the 25–km depth slice. At this depth, the low–V zones along segments of the PFZ seemed to merge and to follow the trace of the fault specifically the segment in central PIA. A linear high velocity (high–V) zone feature, which is essentially parallel to the Philippine trench is also observable east of this linear low–V zone structure. North of the PFZ around 15°N is a high–V zone with significant amplitude. It should be noted that the epicenter of M 7.7 1990 Luzon earthquake (earthquake.usgs.gov, 2010) is located right at this high velocity region.

The 45–km depth slice features wider low–V zone and more heterogeneous anomaly distributions compared with the shallower slices. This depth roughly corresponds to the mantle wedge region under the PIA and the heterogeneity may reflect the complexity of the structures at this depth. The amplitude of the low–V zone is stronger in particular under active volcanic fields. The existence of these low–V regions is commonly associated with the magmatic centers, the upwelling of hot mantle materials, and the fluids liberated from the dehydration of the subducting slab (e.g. Zhao et al., 1994; Gorbatov et al., 1999; Wang et al., 2009). Clearly, these processes are reflected on the depth slices of tomography images of the PIA.

The image at 65–km, which unlike the 45–km depth slice, is dominated by linear high velocity structures and again trend parallel with the subduction zones. The structure along the collision zone (~12.0°N), which was discussed in the previous sections, is complex. It appears that a low–V anomaly, most prominent in 100–km depth slice, separates the slabs subducting along Manila Trench (MT) and Negros Trench (NT) (Figure 3.22). It is not clear however if this anomaly is related to the hot mantle materials disrupted by the collision event. The strike of the trace of the slab along NT seems to deviate towards the northeast similar with the orientations of some of the islands in the central PIA. Depth slices from 100 km and deeper show clear traces of linear high–V zones corresponding to the slabs. The slab signatures can be traced down to 450 km, which is the depth limit of the resolution of this study based on the checkerboard tests. Some portions of the PIA show a distinct but incoherent signature of slabs. The incoherency may be due to the imperfect
crisscrossing of ray paths, which are dictated by the distribution of the stations and events used in the inversion. Nevertheless, an important result of this study is the subducted slabs are imaged successfully even though these are not included in the initial model.

### 3.6.1 Luzon Island, northern Philippines

A series of vertical cross sections perpendicular to Luzon Island are presented in Figures 3.25 – 3.30 and the geographic locations of these sections are shown in Figure 3.24. For the sole purpose of clarity, discussions on the structures under the central and southern Philippines are presented in another subsection, Central and Southern Philippines, following this subsection. The results of the resolution tests are presented in Appendix C. In general the checkerboard patterns are well recovered on the tests down to 450 km.

Vertical cross section A–A’ (Figure 3.25) extends across the submerged part of Luzon ridge where the strike of Manila Trench curved outward to the right. The slice cuts through the North Luzon trough, a forearc basin filled by up to 4 km of sediments (Hayes and Lewis, 1984). The feature under this region is shown in A–A’. The South China Sea slab is imaged as a linear, eastward dipping high velocity anomaly reaching a depth of ∼350 km. The slab dips shallowly under the ridge but at a much steeper angle to the east side. The seismicity does not extend deeper than 250 km, which suggests that the slab is aseismic below that depth. The thickness of the anomaly is about 50–70 km. A simple, but rough, approximation on the thickness of subducted SCS lithosphere, \( L \), can be made using the half–space cooling model (equation 3.4) (Fowler, 1990).

\[
L \equiv 2\sqrt{\kappa t} 
\]  

(3.4)

\( \kappa \) is thermal diffusivity and \( t \) is the age. Based on magnetic anomaly, the opening of the SCS basin occurred ca. 32 to 17 Ma (e.g. Taylor and Hayes, 1980). Using this information and assuming \( \kappa = 10^{-6} \text{ m}^2\text{s}^{-1} \), the theoretical thicknesses of SCS lithosphere are 64–46 km, which are well within the estimated thickness of the anomaly. Slightly south of this slice, Lallemand et al. (2001) interpreted the high velocity zone from the global tomography model of Bijwaard et al. (1998) as the
Figure 3.21. Tomographic images of the PIA at different depth slices. Squares marked the seismic station locations. The saw–toothed lines trace the subduction zones while the lines traversing the arc mark the Philippine fault zone systems. The scale shows the velocity perturbation relative to PHMOD. The dashed line outlines the inferred landward extent of the collision zone after Yumul et al. (2003). MT–Manila Trench, CZ–Collision Zone, NT–Negros Trench, CT–Cotabato Trench, ELT–East Luzon Trough, PT–Philippine Trench.
Figure 3.22. Tomographic images of the PIA at different depth slices. Squares marked the seismic station locations. The saw–toothed lines trace the subduction zones while the lines traversing the arc mark the Philippine fault zone systems. The scale shows the velocity perturbation relative to PHMOD. The dashed line outlines the inferred landward extent of the collision zone after Yumul et al. (2003). MT–Manila Trench, CZ–Collision Zone, NT–Negros Trench, CT–Cotabato Trench, ELT–East Luzon Trough, PT–Philippine Trench.
Figure 3.23. Tomographic images of the PIA at different depth slices. Squares marked the seismic station locations. The saw–toothed lines trace the subduction zones while the lines traversing the arc mark the Philippine fault zone systems. The scale shows the velocity perturbation relative to PHMOD. The dashed line outlines the inferred landward extent of the collision zone after Yumul et al. (2003). MT–Manila Trench, CZ–Collision Zone, NT–Negros Trench, CT–Cotabato Trench, ELT–East Luzon Trough, PT–Philippine Trench.
Figure 3.24. Map of Northern Luzon region showing the locations of the vertical sections of the tomography images. Also labeled are the structures discussed in the text. The square symbols mark the seismic stations used in this study. NLT– North Luzon Trough, VH– Vigan High, WLT– West Luzon Trough, BNPP– Bataan Nuclear Power Plant, NPCT– North Palawan Continental Terrane, M.B.– Marinduque Basin. Inset figure shows the geographic location of the area relative to the PIA.
SCS slab, which dips vertically and appears to be disconnected beneath the Luzon ridge around 100 km depth. They estimated the slab to be at least 880 km long and can be up to 1180 km if it is the same high velocity anomaly that extends to the transition zone. They also suggested the slab appears overturned beneath the SCS itself (see Figure 3.2). The tomography results of Li and van der Hilst (2010) also showed an almost vertical high velocity zone but only extend to depth shallower than 300 km (see Figure 3.5), which is in agreement with the apparent length of the slab obtained by this study. Both tomography models of Bijwaard et al. (1998) and Li and van der Hilst (2010) show a flat lying high velocity zone above the 660-km transition zone under the region. The relatively low velocity perturbation (∼2%) under the forearc region (<50 km depth) may be due to the absence of offshore stations used in the inversion. As expected this low velocity feature is apparent on tomography results using local and regional events only (Figure B.2). Resolution test suggests that ray paths crisscross in this forearc region (see profile A–A’ in Figure A.3). However, the region is outside of the seismic network, which are typically difficult to resolve in conventional local tomography methods (e.g. Zhao, 2007). Zhao (2007) suggested that including the depth phases (e.g. pP, sP) in relocating the offshore events prior the inversion process can overcome the limitations brought about by the absence of offshore stations. The depth phases (e.g. pP, sP) can give more accurate epicentral parameters in relocating the events than by using the first P– or S–wave arrival times only (e.g. Forsyth, 1982; Engdahl and Billington, 1986; Zhao, 2007). In this study, the depth phases were not included. Despite of this, the structures under the seismic networks are well image under the PIA, which are critical in understanding the seisntectonics of the PIA.

Section B–B’ (Figure 3.25) cuts across the area between the Calayan island, an extinct Miocene volcano, and the Babuyan Island, which is an active Quaternary volcano. As discussed in the previous section (3.3.1.2), the volcanoes are suggested to be part of the Western Volcanic Chain (WVC) and Eastern Volcanic Chain (EVC) respectively. Together, these chains were proposed to be a double arc structure that formed due to a shift in the position of the magma generation depth eastward from under the WVC to below the EVC (Yang et al., 1996). The eastward shift is proposed to occur when the dip of the SCS slab became
shallower due to buoyancy (Yang et al., 1996; Bautista et al., 2001). So far, two mechanisms were put forward to explain the cause of the buoyancy. These are either from the subduction of Scarborough Seamount chains, an extinct mid-ocean ridge, around 16°N–17°N (Yang et al., 1996) or the partial subduction of buoyant plateau around 20°N (Bautista et al., 2001). Both studies suggested a tear in the SCS slab (although the location differ between the two mechanisms) inferred from the observed gap in strain energy release and abrupt change in the slab dip south of 18°N (Bautista et al., 2001) or in order to explain the geochemically enriched lava of some volcanoes of the EVC (Yang et al., 1996). Slices D–D’ and E–E’ are constructed to image the fate of the subducted part of the Scarborough Seamounts, which will be discussed later. The image along B–B’ shows a different feature compared to A–A’, which is about 90 km to the north. Two prominent high-V zones at the upper 90 km and below 200 km depths separated by a low-V zone (between 100–200 km depths) are observable in the image. B–B’ is within the region where Yang et al. (1996) suggested the detachment of the deeper part of the slab that occurred 4–5 Ma during the time when the Scarborough Seamount reached and accreted to the Manila trench. The subduction process was temporarily halted due to the resistance of the seamounts to subduct. Subduction finally resumed around 2 Ma forming the EVC. Yang et al. (1996) speculated that the subducted slab was torn near the boundary between the oceanic crust of the SCS and the submerged continental margin. This is to explain the unusual Sr–Nd isotopic signatures of some EVC lavas (Defant et al., 1989, 1990; McDermott et al., 1993) and ultramafic xenoliths (Vidal et al., 1989) that plot below the mantle array (after White and Hofmann, 1982) on the Nd–Sr isotope diagram. The tear served as a pathway for the enriched mantle component suggested to reside in the mantle of the subducting slab to reach the magma source of the EVC. The tomographic image appears to agree with the speculated detached slab scenario. The high velocity zone below 200 km depth most likely correspond to the detached slab and the low velocity zone may represent the upwelling molten materials through the gap, which are emanating most likely from the mantle of the subducted SCS lithosphere. The low velocity zone around 50 km depth appears to be located beneath the nearby Babuyan Island, which is one of the volcanic centers with rocks that exhibit enriched mantle component.
Figure 3.25. Vertical slices of tomography results (see Figure 3.24 for the slice locations). (Top) Plots of topography profiles (vertical exaggeration, V.E. = 15x). (Bottom) Open circles are best located earthquakes (1964–2006) selected from Engdahl et al. (1998) catalog. Labeled are the major structures described in the text. MT– Manila Trench, NLT– North Luzon Trough, SCS– South China Sea. The scale bar shows the velocity perturbation relative to PHMOD.

Slices C1–C1’ and C2–C2’ (Figure 3.26) traverses the Northern Luzon segment of Luzon arc. The segment consists of Pliocene to Pleistocene volcanic necks and plugs dispersed along the western section of the arc (Defant et al., 1989). Both slices ended offshore of East Luzon Trough (ELT) where the accretionary prism and seismicity associated to ELT is suggested to vanish rather abruptly (Lewis and Hayes, 1983; Hamburger et al., 1983). The slices are presented to determine the structure at the inferred northern end of the ELT. The prominent feature in both slices is the vertically dipping, SCS slab. The thickness (∼65 km) is consistent with the anomaly in A–A’. Its depth extent (∼350 km) and location (middle of Luzon arc) is also consistent with the results of Lallemand et al. (2001) and Li and van der Hilst (2010). Similar in slice A–A’, the shallow, low-V zones offshore
and west of the slices are due to artifacts from poor resolution. The trace of ELT, however, is not resolved on both slices.

Figure 3.26. See Figure 3.25 for the general description of the figure. MT– Manila Trench, NLT– North Luzon Trough, SCS– South China Sea, ELT– East Luzon Trough.

Slices D–D’ and E–E’ (Figure 3.27), as mentioned previously, are constructed to image the fate of the subducted part of the Scarborough Seamounts under Luzon Island. D–D’ cuts along latitude 17.0°N traversing the Vigan High and ends offshore of ELT. The Vigan High is a broad structural high suggested to be the expression of the underplated ridge (Scarborough Seamounts) (Hayes and Lewis, 1984; Pautot and Rangin, 1989). The tomographic image reveals an eastward dipping high–V zone that flattens below Luzon Island. The flattened segment is thicker than the previous slices, which may be due to the vertical smearing of the image. The plot of seismicity (west side) extends only to ∼150 km depth. This study speculates that this high–V zone represents the flat–subduction segment of the SCS basin, assuming the subducted seamounts are located in this latitude. It appears that the SCS basin subducts down to ∼100–120 km then flattens and
resumes its descent horizontally into the mantle for about 300 km distance. This flat subduction style has been observed in other subduction zone regions like in central Peru and central Chile, both featuring the subduction of Nazca Ridge and Juan Fernandez Ridge respectively (Gutscher, 2002; Ramos and Folguera, 2009; Alvarado et al., 2009). Gutscher (2002) suggested that the buoyancy of the thickened oceanic crust of moderate to young age (<30 Ma) can adequately explain the occurrences of flat-slab subduction. These features are present in this part of Luzon island, i.e. the thickened crust (Scarborough Seamounts) of the young subducting SCS basin (32–17 Ma). In addition Bautista et al. (2001) suggested that the seamounts are still hot and deforming plastically based from heat flow data and from the gap in strain energy release of intermediate depth earthquakes. The segment of the slab oriented horizontally is aseismic. The shallow seismicity (<100 km) on the eastern side of the slice may correspond to the subduction of West Philippine Sea basin (WPB) along the ELT. The associated slab is poorly resolved in the image. The E–E’ slice on the other hand shows the ”normal” eastward dipping subduction of the SCS basin. It appears that in this part of Luzon Island, for a short amount of along-arc distance between slices D–D’ and E–E’, the SCS slab exhibits a dipping to flat to dipping style of subduction. This is similar with the subduction style in the Chilean margin where the ridge subduction is taking place (Gutscher, 2002). The dipping SCS slab is traceable down to ~350 km but is aseismic below 200 km. On the eastern side, the possible slab associated with the ELT is shown by a dipping high-V zone visible down to <100 km depth. The plot of seismicity coincided with this feature. Another prominent feature of the slice is the low velocity zone at mid-depth and seaward of the ELT. The location is approximately where the Benham Rise is suggested to be colliding or subducting along the ELT (e.g. Hamburger et al., 1983). The Benham Rise is usually interpreted to be a zone of thickened crust resulting from excess of volcanism at a spreading ridge (e.g. Karig, 1975; Hilde and Lee, 1984; Deschamps and Lallemand, 2002). It is uncertain however if this low velocity zone is associated with the SCS slab because its location is outside the seismic network or with the Benham Rise because previous studies suggested that this structure formed farther west (Hilde and Lee, 1984) or south (Deschamps and Lallemand, 2002) of its current location.
F–F’ (Figure 3.28) cuts across the Mt. Pinatubo area and extends eastward towards the junction of the ELT and the left lateral transform fault. The fault connects the ELT with the Philippine Trench (e.g. Hamburger et al., 1983). Mt Pinatubo, which started to erupt on June 12, 1991, produced the second largest eruption in this century causing widespread damage in the Philippines and the global temperatures to drop temporarily by 0.5°C (Newhall et al., 1997). The tomography result shows an eastward–dipping SCS slab that extends to ∼350 km depth and essentially underlies the whole width of Luzon arc at least at this latitude. Two low–V bodies possibly related to the slab are observable, one at the top of the slab at intermediate depth (150–250 km) and the other at the mantle wedge depth that extends to the surface under the volcano. The existence of these low–V bodies may indicate the dehydration of fluids from the SCS slab. Also visible in the figure is the shallow high velocity zone that is aseismic and located beneath the mountain range in eastern Luzon. It is uncertain however if this is
related to the ELT. The possible slab associated with the ELT is manifested by a
dipping high–velocity zone that coincides with the seismicity plots.

Figure 3.28. See Figure 3.25 for the general description of the figure. MT– Manila
Trench, WLT– West Luzon Trough, SCS– South China Sea, ELT– East Luzon Trough,
PT– Philippine Trench, WPB– West Philippine Sea Basin.

Section G–G’ (see Figure 3.28) is constructed to image the structure under
Mt. Natib, a late Pliocene to Pleistocene strato–volcano and one of the series of
volcanic structures comprising the Bataan segment of Luzon arc that includes Mt.
Pinatubo (Ruaya and Panem, 1991; Defant et al., 1988, 1991). The volcano has
no historical record of eruption but recent study by Cabato et al. (2005) suggested
that the latest eruption occurred in the last 11.3 to 18 ka. Field mapping revealed
the volcano had erupted violently in the past producing tephra deposits, pyroclas-
tic flows, and lahars (Wolfe and Self, 1983; Volentik et al., 2009). The geochemical
compositions of the eruptive products range from tholeiitic, calc-alkaline to alkaline
suites (de Boer et al., 1980; Defant et al., 1988, 1991). Currently, thermal springs
are observed within the Mt. Natib caldera suggesting the presence of active hy-
drothermal systems (Ruaya and Panem, 1991). Imaging the structure under Mt.
Natib is important in the study of hazard analysis in the Philippines because ~15
km southwest of the volcano located the Bataan Nuclear Power Plant (BNPP) (
*D’Amato and Engel*, 1988; *Volentik et al.*, 2009). BNPP is a nuclear power plant, that was built but was not put in operation because of highly controversial issues that includes the integrity and safety of the plant in light of its location nearby active volcanoes and faults (e.g. *D’Amato and Engel*, 1988). Although it is not the scope of this study to assess the controversy surrounding the plant, the tomography results can be used as another independent source of information in evaluating the eruption potential of this dormant volcano. The tomographic image shows an east–dipping SCS slab that may extend to ∼300 km depth. Similar with the previous slices, the slab is aseismic deeper than 200 km. Perhaps the structures that are critical in assessing the risk of Mt. Natib are shown by the the low–V bodies around 100–125 km depth and around the mantle wedge depth (≤50 km). The checkerboard test results (see Appendix C) show that these bodies are well–resolved in the inversion. The presence of these low–V bodies suggests that slab dehydration is taking place at this region, which may indicate that Mt. Natib is a potentially active volcano. In fact the shallower, horizontally oriented, low–V zone appears to extend to the surface. The west–dipping high–V zone on the eastern side of G–G’ is interpreted to be the WPB subducting along the Philippine Trench. The slab extends down to ∼100 km and unlike the SCS slab, the seismicity coincides throughout its length.

The slice H–H’ (Figure 3.29) traverses the region characterized by extensive Pleistocene to Holocene volcanisms in southwestern Luzon called the Macolod Corridor (MC) (e.g. *Defant et al.*, 1988). The Corridor is ∼40 km wide zone of NE–SW and N–S oriented extensional faulting (e.g. *Defant et al.*, 1988; *Förster et al.*, 1990; *Pubellier et al.*, 2000). Several studies have proposed various explanations on the formation of MC but the general consensus is that it is a pull–apart rift zone (e.g. *Karig*, 1983; *Defant et al.*, 1988; *Förster et al.*, 1990). Some studies attributed the source of volcanism to rifting (e.g. *Wolfe and Self*, 1983; *Defant et al.*, 1988; *Knittel et al.*, 1997) rather than the subduction process primarily because based on seismicity studies (e.g. *Cardwell et al.*, 1980; *Hamburger et al.*, 1983; *Bautista et al.*, 2001), the Wadati–Benioff zone of the SCS slab is steeply dipping and does not underlie the volcanic fields of MC. Isotope geochemistry studies on volcanic eruptive products indicate that the mantle wedge below the
volcanoes are contaminated by slab–derived melts or slab–derived fluids or both (e.g. Knittel et al., 1997; Defant et al., 1989; Mukasa et al., 1994). In slice H–H’, the SCS slab is traceable down to ∼350–400 km depth. The seismicity coincided with the slab only down to ∼280 km depth. Low–V zones exist above the slab from around 250 km depth and that extend upward to the surface underneath MC. These low–V anomalies can be attributed to the dehydration of the slab that feeds the magma source region of the volcanoes in MC. The presence of low–V at crustal depth agrees with the receiver function analysis results of this study and consistent with the model of Vogel et al. (2006) where they suggested that the partial melting of ponded, crystallized magma at crustal level depth may explain the generation of silicic volcanic products in the MC. The features on the tomographic image suggest that although the volcanoes are far from the presumed magma generation depth, the magmatic structures in MC is related to subduction through the migration of fluids or melts arising from the SCS slab in agreement with the geochemical characteristics. The WPB subducting at the Philippine trench is not well–imaged at this region.

Figure 3.29. See Figure 3.25 for the general description of the figure. MT– Manila Trench, SCS– South China Sea, M.B.– Marinduque Basin, ELT– East Luzon Trough, PT– Philippine Trench, WPB– West Philippine Sea Basin, PFZ– Philippine Fault Zone.
I–I’ cuts between the southern end of Manila Trench and the northern end of the site of collision between the North Palawan Continental Terrane (NPCT) and the central PIA (Figure 3.29). The NPCT is a microcontinental fragment that was rifted from southeast China during the opening of the Early Oligocene to Early Miocene SCS basin (McCabe et al., 1982; Holloway, 1982; Taylor and Hayes, 1983; Sarewitz and Karig, 1986; Encarnación et al., 1995; Almasco et al., 2000). The collision event is suggested to occur ca. 8–9 Ma (Marchadier and Rangin, 1990) or ca. 10–15 Ma (McCabe et al., 1982). At the culmination of this event, the subduction along the Manila Trench in this region is inferred to halt and the convergence was transferred eastward manifested by the internal deformation in the central PIA (e.g. Barrier et al., 1991). The locking of convergence due to the collision is proposed to result in a flip in subduction zone to the east forming the Philippine Trench (Barrier et al., 1991; Ozawa et al., 2004). This collision model for the initiation of the Philippine Trench suggests the trench is younger than 8–9 Ma (Barrier et al., 1991). Ozawa et al. (2004) used K–Ar dating to determine the age of volcanisms associated with the Philippine Trench in the Bicol region, which is in the northeastern part of Luzon. They obtained ∼6.6 Ma for the age in the north region and ∼3.5 Ma in the central region of the trench (Ozawa et al., 2004). Using these ages, Ozawa et al. (2004) suggested that the initiation of the Philippine Trench occurred right after the collision event. Barrier et al. (1991) also supports the collision model but they suggested that the central PIA underwent internal deformation first and that the trench initiated around 2.5–3 Ma. A study on the Philippine Fault system using GPS data however argues for the formation of the trench due to the change in the movement of the Philippine Sea Plate relative to Eurasia from northward to northwestward motion around 4 Ma (Aurelio, 2000). Aurelio (2000) cited the paleo-stress trajectory conducted by other studies in Taiwan and Japan as well as studies on the patterns of volcanisms in Japan and Ryukyu arcs that support the change in the plate motion of the Philippine Sea Plate around 4 Ma. Based on these information and the apparent synchronism in the formation of the Philippine Trench and Fault system (Barrier et al., 1991; McCaffrey, 1992), Aurelio (2000) proposed that the change in the plate movement resulted to the formation of the Philippine Trench and Philippine Fault system through shear partitioning mechanism. The image obtained in I–I’
shows a well-defined SCS slab. The slab is characterized by steeply-dipping high-$V$ zone down to $\sim 350-400$ km depth. As in previous slices, the slab is aseismic at depth deeper than $\sim 250$ km. A notable shallow, low-$V$ anomaly is observable beneath the Marinduque Basin. The Marinduque Basin is an intra-arc basin floored by oceanic crust and formed by sea floor spreading type processes (Sarewitz and Lewis, 1991; Encarnación, 2004). The basin is also truncated by a segment of the Philippine fault. Similar in slice H–H', the trace of the slab associated with the Philippine trench is not image coherently but can be traced down to 100 km coinciding roughly with the seismicity plots. The slab probably is not long enough such that its surface projection is outside the seismic network, which result in poorly crisscrossing of rays.

The trace of the SCS slab started to vanish south of $12.8^\circ$N (see slices J1–J1' & J2–J2', Figure 3.30). At this latitude, the slab is dipping almost vertical but is aseismic from 100 km depth down. A shallow, low-$V$ zone around 100 km depth indicates possible presence of partial melts originating from the slab. The trace of the slab subducting in the Philippine trench is imaged as westward dipping ($\sim 30^\circ$), high-$V$ zone down to $\sim 250$ km depth.

The timing of the initiation of subduction along Manila trench is poorly constrained (e.g. Encarnación, 2004; Yumul et al., 2003). The trench is probably been active since Late Oligocene – Middle Miocene (Hayes and Lewis, 1984) or pre–Late Oligocene (Encarnación et al., 1993). Defant et al. (1989) inferred that the subduction immediately followed the opening of the SCS basin (ca.32–17 Ma) based on radiometric dating of the volcanoes (approx. 10 Ma for the oldest eruption) and biostratigraphic correlations. In this study, the SCS slab is imaged down to $\sim 300$ and possibly to 400 km depths. Assuming the long term average subduction rate of 2 cm/yr as estimated by Hayes and Lewis (1984) and using the length of the imaged slab resulted to a 15–20 Ma initiation of subduction, in agreement with the early studies (e.g. McCabe et al., 1982, 1987; Hayes and Lewis, 1984; Defant et al., 1989). The trace of the slab underneath the zone of collision between the continental NPCT and the central PIA also extends down to $\sim 300-400$ km depth. If the collision event occurred ca.10–15 Ma (McCabe et al., 1982), the convergence rate along the Manila Trench at this region should be at least 4 cm/yr. This rate should account for the length of the slab that had consumed along the trench prior
3.6.2 Central and Southern Philippines

Subduction zone systems border the eastern and western sides of central and southern PIA (Figure 3.31). The western side is marked by the eastward and southward subduction of Sulu Sea basin along the Negros Trench and Sulu Trench as well as the subduction of the Celebes Sea basin along the Cotabato Trench. On the east side is the site of northwestward subduction of the West Philippine Sea Basin along the Philippine Trench. The region is also traversed by the Philippine Fault that trends essentially parallel to the Philippine Trench. Several slices (see Figures 3.32–3.36) are constructed across this part of the PIA to image the structures underneath the region.

Two slices (K1–K1’ and K2–K2’) are made that crosses the collision site between North Palawan and the Central PIA (Figure 3.32). The slices show complicated subsurface structures, which may reflect the consequence of the collision.

Figure 3.30. See Figure 3.25 for the general description of the figure. MT– Manila Trench, SCS– South China Sea, M.B.– Marinduque Basin, PT– Philippine Trench, WPB– West Philippine Sea Basin, PFZ– Philippine Fault Zone.

to the culmination of the collision.
Figure 3.31. Map of Central and Southern Philippines showing the locations of the vertical sections of the tomography images. Also labeled are the structures discussed in the text. The square symbols mark the seismic stations used in this study. Inset figure shows the geographic location of the area relative to the PIA. CMA - Central Mindanao Arc, WPB – West Philippine Sea basin.

On the west side of both slices, at the inferred collision site, is what appear to be a detached SCS slab. The seismicity is shallow <70 km and coinciding with the eastward dipping, high–$V$ zones. Below these zones are low–$V$ bodies that reached down to <200 km depth, which is underlain by another eastward dipping high–$V$ zones reaching perhaps 300–km depth. Although these structures are well–resolved in the resolution tests, this study only speculates the detached slab scenario because there exists no corroborating evidence for the possible detachment of the SCS slab. The low–$V$ zones do not extend to the surface, which may possibly explain the absence of associated volcanic structures. The WPB subducting along the Philippine Trench on the east side is shown as a coherent, westward–dipping high–$V$ zone. On slice K2–K2’, a shallow, low–$V$ zone extends
to the surface under Mt. Biliran, an active volcano. Those shallow, low-$V$ zones offshore are probably due to artifact because of poor resolution and the absence of offshore seismic stations.

Figure 3.32. See Figure 3.25 for the general description of the figure. SCS– South China Sea, PT– Philippine Trench, WPB– West Philippine Sea Basin, PFZ– Philippine Fault Zone.

Slices L–L’ and M–M’ (Figure 3.33) cut through Palawan island in the west, the Negros Trench, and ends offshore of the Philippine Trench in the east. The prominent features on both slices are the two oppositely dipping high-$V$ zones. The westward dipping zone is associated with the subducted Sulu Sea basin along the Negros trench. The slab is traceable down $\sim 250$ km that extends deeper than the maximum seismicity depth ($<100$ km). Assuming the minimum slab length of 250 km and convergence rates of 3.6–5.4 cm/yr, the age of the possible initiation of subduction ranges from 7.0–4.7 Ma (Late Miocene to Early Pliocene) consistent with the age from earlier studies (Hall, 1996; Sajona et al., 2000b). The west-dipping WPB associated with the Philippine Trench can be traced down to $\sim 150$ km and maybe deeper ($<300$ km).

The Sulu Sea slab in slices N–N’ and O–O’ (Figure 3.34) is also a coherent feature similar from the previous slices. The seismicity is sparse in both slices.
Figure 3.33. See Figure 3.25 for the general description of the figure. SS– Sulu Sea, PT– Philippine Trench, WPB– West Philippine Sea Basin, PFZ– Philippine Fault Zone.

and plots shallower than 80 km. N–N’ slice continues through Camiguin, a small volcanic island, and ends offshore of the Philippine Trench. Camiguin Island is consisted of four volcanoes, that includes the active Mt. Hibok-Hibok, that erupted basaltic to rhyolitic, calc-alkaline lavas in the last ca. 400 ka. (Castillo et al., 1999). The Island is said to be the northern extension of the Central Mindanao Arc (CMA) characterized by the most voluminous volcanic field in the PIA (Sajona et al., 1993, 1994; Castillo et al., 1999; Sajona et al., 2000a). The source of volcanisms of the CMA cannot easily be related to the surrounding, present–day subduction zones because the Wadati–Benioff zones do not extend beneath the arc (e.g. Sajona et al., 1993, 2000a). Instead the CMA is associated to the detached slab of the Molucca Sea Plate (e.g. Sajona et al., 1993, 2000a; Solidum et al., 2003). This detached slab is inferred from deep seismicity (>300 km) under Mindanao. South of Mindanao, the Molucca Plate is pinched between two converging island arcs, the Halmahera and the Sangihe arc in the Molucca Sea region (Lallemand et al., 1998). The possible feature of this remnant slab is imaged by the tomography study of Bijwaard et al. (1998) as a high–V zone from ~300 km depth down beneath Mindanao Island. The tomography image under Camiguin (N–N’) reveals
a low amplitude, negative perturbation around the mantle depth. There is no coherent feature of the detached slab resolved in the slice. The tomography model of Bijwaard et al. (1998) shows a clear image of this detached slab under N–N’ around 450 km depth, which is deeper than the maximum resolution depth of this study. Beneath the basin offshore east of Camiguin shows a large amplitude, negative velocity perturbation. The boundary between the negative and positive velocity perturbation is sharp. It is interesting to note that the Philippine Fault cuts right through this sharp boundary. The slab related to the Philippine Trench can be traced down to 200 km and roughly coincide with the seismicity plot.

**Figure 3.34.** See Figure 3.25 for the general description of the figure. SS– Sulu Sea, NT– Negros Trench, PT– Philippine Trench, WPB– West Philippine Sea Basin, PFZ– Philippine Fault Zone.

To image the magma source region of CMA, slices P–P’ and Q–Q’ are constructed from Sulu Sea cutting across the Sulu Trench and the heart of CMA until offshore of Philippine Trench (Figure 3.35). The possible slab associated with the Sulu Trench is imaged as a positive velocity perturbation (≤2.0%) on the west side of the slice in P–P’. The slab may reached a depth of 100 km but the seismicity only extend ≤50 km. The prominent feature in both slices is the large-scale low-\(V\)
anomaly (∼100 km) emanating in the upper mantle (∼200 km) and extending up to the CMA area at the surface. This low–V zone may indicate partial melts or hot upwelling materials from the mantle. The size of this anomaly seems to be consistent with the huge volume of eruptive products in the CMA region. Again the inferred remnant slab of the Molucca Plate is not image because it is located deeper than the depth resolution of this study. The earthquake hypocenters related to this remnant slab are ∼400 km and deeper. What is apparent however is that the low–V anomaly is at a distance from the slabs under the Sulu Trench and the Philippine Trench. The WPB is clearly observed in both slices and the depth extents (∼200–225 km) are consistent with the previous slices.

Figure 3.35. See Figure 3.25 for the general description of the figure. ST– Sulu Trench, SS– Sulu Sea, PT– Philippine Trench, WPB– West Philippine Sea Basin, PFZ– Philippine Fault Zone, CMA– Central Mindanao Arc.

Sections R–R’ and S–S’ (Figure 3.36) are constructed to see the Celebes Sea slab under the Cotabato Trench. The possible slab is imaged as low amplitude, positive velocity perturbation in S–S’ (∼2%) that dips to the east and traceable down to 100 km. This feature is not observed on the tomography results of Bijwaard et al. (1998) (see also Figure 3.2). The prominent feature in the image are the two
high amplitude, negative velocity perturbations at depths 250 km and 350 km. The shallower feature appears to be located beneath the Celebes Sea slab. This study can only speculate the origin of these two features from the upwelling of hot mantle materials associated with the deep Molucca Sea slab subduction in the lower mantle. These features are also similarly observed in the deep structures from tomography studies under Japan (Zhao et al., 1994; Zhao, 2004; Abdelwahed and Zhao, 2007). A high amplitude, positive velocity perturbation is located east of Davao Gulf and under the Pujada Peninsula. This high velocity may correspond to the ophiolitic terranes comprising the Pujada Peninsula. The other high velocity zone that extends down to $\sim$200 km corresponds to the West Philippine Sea slab under the Philippine Trench. Seismicity plots coincide with this feature.

![Figure 3.36](image)

**Figure 3.36.** See Figure 3.25 for the general description of the figure. CT– Cotabato Trench, CS– Celebes Sea, PT– Philippine Trench, WPB– West Philippine Sea Basin, PFZ– Philippine Fault Zone.

### 3.6.3 Subducted slabs and volcanisms in the PIA

The joint inversion of local, regional, and teleseismic events used in this study is successful in determining the shallow and deep $P$-wave velocity structures be-
neath the PIA. In general, the tomographic images reveal high velocity zones corresponding to the subduction zones along the western and eastern sides of the Philippines. Commonly, these high velocity zones are deeper than the maximum seismicity depths, which indicate that slabs in the PIA subduct aseismically at deeper depths and that caution should be taken when using the Wadati–Benioff zone in determining the slab lengths. The lack of seismic stations offshore and the imperfect crisscrossing of rays along the slab region resulted in some slices with incoherent structures. Nevertheless, this study was able to image successfully the slabs even without a priori information on the initial model.

The South China Sea slab (32–17 Ma) is the prominent subsurface structure in north Luzon Island. For the most parts, the slab is imaged as eastward dipping, high velocity zone. Its depth extent (down to ~350 km) is consistent throughout and the deeper portion is subducting aseismically. Slab detachment is speculated in the northern part of Manila Trench (between 19°N–20°N), which may have facilitated the introduction of enriched mantle components on the erupted lavas of volcanoes in that area. The subsurface structure is complicated at the latitude ~17.0°N where the subducted seamounts are suggested to be located (Bautista et al., 2001). Here, the SCS basin subducting along the Manila Trench is suggested to exhibit flat–slab subduction similar to the flat slab segments in Peru and Central Chile (Gutscher, 2002; Alvarado et al., 2009; Ramos and Folguera, 2009). At a relatively short along–arc distance, the style of subduction along the Manila Trench changes from dipping to flat to dipping perhaps due to the buoyancy of the subducting, hot seamounts. The SCS slab is imaged as vertically dipping structure in the collision zone region at the southern end of Manila Trench. The tomography results confirm the early seismicity studies showing the slab located far from the active volcanic field (Macolod Corridor) in southern Luzon. This study further shows that even though the supposed ideal magma generation depth is at a distance from the volcanoes, materials (e.g. slab–derived aqueous fluids, melts from subducted sediments) are being supplied to the volcanoes through the migration of melts. The migration of melts can also explain the enriched subduction components of the lavas in that region (e.g. Castillo and Newhall, 2004). With regards to the timing of initiation of Manila Trench, this study suggests that the initiation occurred during the Early Miocene based on the minimum length of the slab (~350
km) and convergence rate of 2 cm/yr.

The West Philippine Sea basin, inferred to be subducting along the East Luzon Trough in the northeastern PIA, is not well resolved in this study. The slab is imaged possibly between latitudes 18.00°N and 15.20°N. Structures at the northern and southern ends of the slab are not resolved. Bautista et al. (2001) proposed that the Benham Rise collided with ELT during the Pliocene based on their estimated displacement (240 km) along the E–W transform fault that connects the ELT with the Philippine Trench. The image of ELT is incoherent essentially throughout its entire trace. Perhaps the best image of the slab is observed around 15.80°N where it can be traced down to ~100 km depth, but well short of the supposed displacement along the transform fault.

The subduction of the Sulu Sea Basin along Negros Trench is also clearly imaged as a west-dipping, high velocity zone starting from latitude ~10.80°N. The slab is traceable down to ~250 km depth but the plots of seismicity only extend down to ~150 km. Low velocity zones are found to exist at the mantle wedge depth above the slab with the large amplitude low velocity anomalies located beneath the active volcano in Negros Island. Based on the minimum length of the slab, subduction along the Negros Trench may have initiated during the Late Miocene.

Slabs from both Sulu Trench and Cotabato Trench are not well-resolved in this study possibly due to the distribution of the seismic stations and the imperfect crisscrossing of rays. Nevertheless, the probable Sulu Sea Basin slab is imaged down to 100 km while the Celebes Sea basin slab under the Cotabato Trench may extend down to 100–150 km depths.

On the east side of the PIA, the tomography sections reveal the subducted part of the West Philippine Sea basin along the Philippine Trench reaching depths down to 200–300 km. Seismicity plots extend to ~250 km depth. In some tomography sections, the slab appears discontinuous. This is rather due to the steep dip of the slab making it essentially outside the seismic network which results in inadequate crisscrossing ray coverage for the whole slab structure. Slow anomalies are observed above the slab in particular under the region with known volcanoes. These slow anomaly bodies may be attributed with the dehydration of the slab, similar with what have been tomographically image from other subduction zones.

Some studies suggested the formation of the Philippine Trench to the collision
of Palawan micro–continental block with the central Philippines (e.g. Barrier et al., 1991; Ozawa et al., 2004). The collision event essentially locked the subduction process occurring along the collision site in the west and induced the initiation of subduction to the east. Ozawa et al. (2004) measured K–Ar ages of 37 volcanic eruptive products from the Bicol Peninsula in southeastern Luzon to constrain the initiation of subduction of the Philippine Trench. They obtained ages mostly within 0–7 Ma and two outliers of 27 and 43 Ma (Figure 3.37). They also noted possible age gaps in magmatic activity around 5.0–6.0 Ma and 1.0–1.5 Ma. Samples younger than 7 Ma are considered as associated with the subduction along the Philippine Trench while the two older samples are suggested to be related with older magmatic events in the PIA. Based on these younger volcanic ages, Ozawa et al. (2004) estimated that the subduction along the Philippine Trench was initiated at the latest at 8 Ma right after the collision event in 8–9 Ma (age from Marchadier and Rangin (1990)). It should be noted however that excluding the two outliers, only four samples have age range from 6.6–4.38 Ma while the remaining 31 samples are 3.38 Ma and younger.

Another study extrapolating GPS data suggested that the initiation of the Philippine Trench occurred much later, about 4 Ma (Aurelio, 2000). Around that time, the relative movement of the Philippine Sea Plate with respect to Eurasia changed from northward to northwestward motion (Aurelio, 2000). This change in direction of motion favored the formation of the Philippine Trench and the Philippine Fault system through shear partitioning mechanism. Under this mechanism, the Philippine Fault (at least in central and southern PIA) accommodates the lateral component of oblique convergence between the Philippine Sea Plate and Sunda Plate while the Philippine Trench absorbs the perpendicular component of convergence (Fitch, 1972; Pubellier et al., 2000; Aurelio, 2000). The measured displacement rates along the Philippine Fault and the convergence rate along the Philippine Trench from GPS data are suggested to strongly support the partitioning of the slip of oblique convergence between the two structures (Duquesnay et al., 1994; Aurelio, 2000). Geologic data in central Philippines suggested a 2–4 Ma age for the Philippine Fault thus supporting the synchronism of the Philippine Fault and Philippine Trench (Barrier et al., 1991).

Puspito et al. (1993) concluded in their tomography study that the slab of the
West Philippine Sea basin extends below 200 km and further suggested that it may reach a depth of 450 km. The average convergence rate along the Philippine Trench is 6–8 cm/yr (Barrier et al., 1991). Ozawa et al. (2004) cited these information to support their model on the timing of the initiation of the Philippine Trench. The results of this study favor the 4 Ma initiation of the Philippine Trench. Similar with the result of Puspito et al. (1993), the slab of the West Philippine Sea basin was well imaged down to 200 km but the possible maximum depth of penetration of the slab is shallower (∼ 300 km) than suggested by Puspito et al. (1993) (450 km). Results of checkerboard resolution test of this study showed that the structures can be resolved down to 450 km. Puspito et al. (1993) used 11 stations located in the Philippines while this study had 52 stations for the same regional coverage.

Figure 3.37. Map of sample locations from Ozawa et al. (2004). The numbers in parentheses are the K–Ar ages obtained by Ozawa et al. (2004). See also the Bicol region in Figure 3.24 for the geographic location of the study area relative to the PIA
The tomographic model of this study also confirmed that the slabs surrounding the Mindanao Island do not extend below the CMA, which is suggested to be the locus of the most voluminous volcanic field in the PIA. The tomographic image shows that the CMA is underlain by slow anomaly bodies emanating from a deep upper mantle region and extending upward to the surface. These bodies may indicate upwelling of hot mantle materials possibly associated with the remnant slab of the Molucca Sea Plate. This remnant slab however is not resolved because it is located deeper than the depth resolution of this study. These deep upwelling features have also been observed in Central and Southwest Japan where the low velocity zones were attributed to the effect of deep dehydration of the Pacific Plate slab (Abdelwahed and Zhao, 2007). The occurrences of deep dehydration process have been suggested by experimental mineral physics study (e.g. Ohtani et al., 2004).

The Philippine Island Arc is commonly regarded as a collage of exotic terranes assembled by various tectonic episodes resulting to its present configuration (e.g. Hamilton, 1979; Karig, 1983; McCabe et al., 1985; Karig et al., 1986; McCabe et al., 1986; Geary et al., 1988; Rangin, 1991; Billedo et al., 1996). In fact this concept is always considered in the interpretation and kinematic reconstruction of the paleotectonic history of the PIA. A recent review of these terranes, in particular the terranes in northern Philippines however suggested that they are autochthonous and were formed by in situ sea floor spreading with respect to the PIA region (Encarnación, 2004).

In general, the detailed tomography conducted by this study reveals relatively simple subsurface structures of the PIA. The subducted part of the marginal basins on the western side of the PIA along Manila Trench and Negros Trench is well imaged as high velocity zones dipping to the east. Although these two subduction zones are considered to form independently of each other (e.g. Hamilton, 1979; Cardwell et al., 1980; Hayes and Lewis, 1984; Encarnación, 2004) the apparent similarity in the thickness and the depth extent of the slabs may also point to the possibility that these two are once connected prior to the collision event in the central PIA. Similar with the result of Puspito et al. (1993), the Cotabato Trench is not well resolved and may extend no deeper than 100 km depth. The Sulu Trench and East Luzon Trough are also poorly imaged most likely due to the
inadequate distribution of stations at those regions. The slab along the Philippine Trench is a consistent feature on the east side of the PIA dipping to the west down to $\sim$200–300 km. Although this study does not have sufficient resolution for structures deeper than $\sim$400 km, the current configurations of the slabs as imaged in the tomography procedure are consistent with the simple kinematic model of Encarnación (2004) for the formation of the PIA.

Similar with tomography studies in other subduction zone environments (Zhao et al., 1997; Gorbatov et al., 1999; Abdelwahed and Zhao, 2007; Wang et al., 2009), the subducted slabs under the Philippines were imaged as high velocity anomalies with average perturbation values 2%–6% faster than the normal mantle velocities. Consistent also with the other tomography studies are the slow velocity anomalies that are clearly visible beneath volcanic regions. These anomalies commonly extend down to $\sim$100 km depths and often are connected to the surface in particular under active volcanic regions. The characters of these anomalies vary with the different sections but the perturbation amplitudes (-2% to -6%) are within the range of values obtained from other subduction zone environments. These distinctive low-velocity anomalies are commonly related to the dehydration processes from the subducting slabs. Some deep (> 200 km), low velocity bodies are also observed in some sections. The resolution tests confirmed the reliability of these anomalies. These anomalies are similar with what have been observed in Japan (Abdelwahed and Zhao, 2007) and in Tonga Arc (Zhao et al., 1997), which were suggested to be due to the deep dehydration or to the upwelling of hot mantle materials. The remarkable coherency of the slab features, which considering that these features are not included a priori in the initial model, and the consistency of the results with other studies suggest the tomography provided reliable images of the subsurface of the PIA.

3.7 Conclusion

A high resolution tomography study with spatial resolution of 40 km and down to $\sim$400 km is determined in the Philippine Island Arc. A large number of travel times from local and regional events and residuals of the travel times of teleseismic events are jointly inverted to image a detailed three-dimensional velocity structure
of the Arc. In most cases, the slabs are imaged clearly as dipping high velocity zones, 2–6% faster than the normal IASP91 mantle, and are consistent with the perturbation values from other tomographic studies of subduction zones. The extent of the slab is often deeper than the deepest seismicity, which implied that the slabs penetrate the greater depths aseismically. The tomographic images reveal east-dipping high velocity zones on the western side of the PIA that corresponds to the Manila, Negros, Sulu, and Cotabato subduction zones. The eastern side of the PIA is characterized by west-dipping high velocity zones along the East Luzon Trough and Philippine Trench. While the slabs are well resolved under the Manila, Negros, and Philippine subduction zones, the Sulu Trench, Cotabato Trench, and East Luzon Trough are poorly defined primarily because of imperfect resolutions and parameterization used in the tomography model. Nevertheless, this study is successful in imaging the slabs even though the slabs are not included \textit{a priori} in the initial model. Based on the estimated minimum length of the slabs and convergence rates from other studies, subduction along the Manila Trench occurred in Early Miocene, the Negros Trench in the Late Miocene, and the Philippine Trench during the Pliocene.

Slow velocity perturbations are apparent in the mantle wedge at the top of the slabs and often extend to the surface under the volcanic arc fronts. These low velocity zones may indicate the dehydration reactions in the slab feeding the source materials for magma generations. The magmatic process in Macolod Corridor in southeastern Luzon can be related to the migration of fluids from the steeply-dipping subducted slab of the South China Sea Basin. Deep dehydration reactions are also observed under Central Mindanao Arc of Mindanao island, which may be caused by upwelling of hot mantle materials associated with the subducted remnant slab. These features are observed where the slabs associated with the surrounding subduction zones do not extend under the Arc.
4.1 Abstract

We use receiver function analysis to estimate the first-order subsurface seismic structure in the vicinity of the island of Montserrat, located in the northern Lesser Antilles arc and home of the active Soufrière Hills Volcano. Near-surface complexity in the island structure inhibits our ability to resolve lateral variations in seismic structure, so we average the signals from eight stations in two directions to produce a first-order regional seismic velocity model. Although we are unable to resolve a precise sharpness of the crust-mantle transition, we can limit its thickness to less than $\sim 4$ km. Lateral variations in shallow structure complicate a receiver-function based estimate of Poisson's ratio, but the data indicate that in both directions sampled by the receiver functions, the lower crustal Poisson's ratio is greater than 0.27 and more likely in the range 0.29—0.30. We estimate a mean crustal thickness of $\sim 30$ km and the observations suggest that the crust may be slightly thinner northwest of the island ($\sim 26$-30 km) than it is to the south ($\sim 30$-34 km). The high P-wave speed and Poisson's ratio indicate a generally mafic lower crust, with rocks of intermediate composition not precluded in the upper part of the lower
crust. We did not find evidence for a thick high-speed lower crust (> 7.4 km/s) as has been inferred in some other arcs.

4.2 Introduction

The island of Montserrat is located in the northern segment of the 700-800 km long Lesser Antilles Volcanic Arc, about 150 km east of the trench, and about 120-140 km above the subducting North American lithosphere (Figure 4.1). The Lesser Antilles arc broadens north of Martinique and Dominica where the current volcanic arc (Guadeloupe to Saba), active since the early Miocene, is located west of an older Eocene to mid-Oligocene arc (eastern Guadeloupe, Antigua, Barbuda, St. Maarten) (Macdonald et al., 2000). The entire region has a relatively shallow sea floor with depths in the range 0.5–1.0 km (Figure 4.1). Limestone outcrops on the eastern arc and observed as far west as northern Montserrat and St. Eustatius suggest that the region has experienced some uplift in the Quaternary (Macdonald et al., 2000).

The crust of the Caribbean plate has a thickness intermediate between that of typical ocean and continental lithosphere (Officer et al., 1957, 1959). Thus Montserrat’s volcanic complexes likely formed on modified oceanic crust thickened early in the history of the Caribbean plate, before ~85 Ma and prior to the emplacement of the plate between North and South America (Donnelly, 1989; Pindell and Barrett, 1990). The Montserrat crust may have been further modified during activity associated with the Eocene to mid-Oligocene age arc now located east. The entire region from the trench to at least the Aves ridge (an older late Cretaceous–to–Paleogene age arc structure) appears to contain modified oceanic crust with thickness estimates ranging from < 10 to 40 km in places along the arc (Officer et al., 1959; Boynton et al., 1979; Whitmarsh et al., 1983). Early estimates from sparse seismic and gravity observations suggested a crustal thickness of 30-35 km for the northern Lesser Antilles arc (Boynton et al., 1979). More recent work in the southernmost region of the Lesser Antilles arc suggested a crustal thickness of 20-25 km, i.e. 5 to 10 km thinner than the northern part of the arc; this result derives partly from lower velocities, and densities, than were assumed in previous models (Christeson et al., 2008). The island of Montserrat extends ~22 km in the
north-south direction and $\sim$13 km in the east-west direction, and about one-third of this area lies just below sea level (Figure 4.2). The highest elevation is associated with the Soufrière Hills Volcano (SHV), and is $\sim$1 km above sea level. The exposed island is a composite of three volcanic complexes decreasing in age from north to south. In the north, the Silver Hills complex was active $\sim$1.2-2.6 Ma, south of that is the Centre Hills complex, which was active $\sim$600 ka, and still further south is the currently active Soufrière Hills complex that has been active for about the last 170 ka (Harford et al., 2002). The Soufrière Hills Volcano (Figure 4.2) is a composite of a group of andesitic domes flanked by pyroclastic and collapse deposits, and began its most recent eruption in 1995 (Young et al., 1998; Kokelaar,
2002; Luckett et al., 2008). This recent activity has been characterized by four cycles of lava dome building with occasional explosions and dome collapse, interspersed by periods of relative inactivity (Kokelaar, 2002; Voight et al., 2006). The Montserrat Volcano Observatory (MVO) has operated seismic networks mainly covering the southern two-thirds of the island (surrounding the SHV) since 1995. Data from the network has been used for a number of investigations of volcano-related seismicity patterns (e.g. Aspinall et al., 1998; Rowe et al., 2004; Roman et al., 2006; Luckett et al., 2008). By 2000, the network was a mix of broadband and short-period instruments. In 2005, the original short-period seismic network was upgraded to include eight Guralp CMG-40T three-component seismic stations (Figure 4.2). We use teleseismic $P$ waves recorded on the broadband stations to investigate crustal structure beneath the island region. Estimating crustal properties beneath active volcanic regions is a challenge due to the intrinsic structural complexity of such regions. Strong near-surface variations in rock properties are produced by the complex processes of extrusion, dome collapse, subsurface intrusion, weathering, geothermal system evolution, thermal metamorphism, and the alteration of the volcanic edifice and its foundation. In our analysis, near-surface complexity is handled by averaging the responses across the entire island network to isolate information on the island-wide structure. The area of the active volcanic edifice is small compared with the geometry of receiver function analyses, but our effort provides reliable information on the island and regional structure. We find that the region surrounding Montserrat has a crust thickness of roughly 30 km, with a higher than average Poissons ratio in the lower crust. The crust-to-mantle transition generates substantial $P$-to-$S$ converted waves, which is consistent with a relatively sharp (less than \( \sim 4 \) km thick) crust-to-mantle transition.

4.3 Observations

Most imaging of intraoceanic island arcs has been performed using narrowband, relatively high frequency controlled source methods (Holbrook et al., 1999; Shillingston et al., 2004; Kodaira et al., 2007; Christeson et al., 2008). We use the P wave receiver function method (e.g. Langston, 1979; Owens et al., 1987; Ammon, 1991; Bostock, 2007) that samples the structure more locally with lower-frequency sig-
Figure 4.2. Montserrat surface relief and bathymetry with a 100 m contour interval, MVO station locations (inverted triangles) and Soufrière Hills Volcanic center (star). Stations with Guralp 40T sensors were used in the analysis. The dotted circles represent the area covered by crust-mantle transition Fresnel zones for the Ps converted phase originating near 30 km depth and the back azimuths of the events used in the analysis.

...als, which are less sensitive to rapid changes in structure but more sensitive to gradational boundaries (Owens and Zandt, 1985). We computed teleseismic P wave receiver functions using the time domain iterative deconvolution approach described by Ligorría and Ammon (1999) and selected P waveforms for events between 30 and 90 distance to minimize interference with upper mantle and core-mantle boundary structure. As with all small island sites, Montserrat has a fair amount of surf-generated seismic background motions, so many teleseisms had low signal-to-noise ratios within the primary receiver function analysis period range...
(about 30 s to 1 s period). We obtained a total of about 60 stable receiver functions that naturally clustered into two back azimuth directions (northwest and south) by examining available teleseismics recorded between 2000 and 2008.

Individual observed receiver functions are quite complicated and have large tangential (off-azimuth) motions, indicative of strong lateral heterogeneity. Since the complex motions begin immediately after the P arrival, we believe much of the complexity is associated with structure in the upper few kilometers of the crust. We also believe that response of the shallow structure beneath each station is uncorrelated with the response at other stations. We can do little to model the near-surface complexities, but the availability of observations from eight relatively nearby stations can allow us to extract information on the bulk crustal properties and the regional crustal thickness. Our assumption is that the deeper structure is common to all the signals and can be enhanced by averaging signals from all the stations to reduce the influence of the near-surface heterogeneity. Toward this end, we split the observed receiver functions at each station into two back azimuthal groups and averaged the results in each group. For completeness, we also averaged all the observations to create a single azimuthally averaged response.

The estimated receiver functions are shown in Figure 4.3. Figure 4.3 (top) shows the results of stacking all the observations with equal weights to each back azimuth cluster. Figure 4.3 (middle) shows the results for the northwest back azimuthal cluster, and Figure 4.3 (bottom) shows results for the southern back azimuthal cluster. In roughly laterally homogeneous structures, the transverse receiver function should be small, and in our case the tangential amplitudes of the receiver function stacks are substantially smaller than the individual station signals. We have labeled the arrival times of the waves that we interpret as the converted phase from the crust-mantle boundary ($P_s$) and the multiple arrival ($PpPmS$), both of which are defined with ray diagrams in Figure 4.4. Since the receiver functions include a low-pass Gaussian filter (width factor of 1.0), the arrival time of the phase is the peak. Our selection of the wave arrival times was aided using lower-frequency versions of the receiver functions (shown later). Although no time for the multiple is identified for the southern back azimuth receiver function, that arrival is clear when the signal is low-pass filtered to enhance periods longer than about 3 s. The averaged responses are relatively simple for an island station
(e.g. Leahy and Park, 2005). In fact, the tangential receiver functions are typical of those observed at many sites and are relatively small at the times when the converted phase and the crustal multiple arrive. The amplitudes of the $P_s$ and $PpPmS$ phases are sensitive to the seismic velocity contrast across the boundary generating them, and the arrival times are sensitive to the bulk properties of the crust (Langston, 1979; Ammon et al., 1990; Zandt and Ammon, 1995).

Figure 4.3. Receiver function stacks of (top) all observations, (middle) signals derived from $P$ waves arriving from the northwest, and (bottom) south back azimuths. The symbol R identifies radial, and T identifies tangential receiver functions. The receiver function low-pass filter Gaussian width factor is 2.5. The relatively small tangential motions associated with the converted $P_s$ phase and the multiples indicate that the multistation stacking has produced a reasonable average of the structure.
Figure 4.4. Ray diagram for the main phases modeled in the receiver functions. All the rays are generated by the incident P wave that sweeps across the region from left to right. The multiple (PpPmS) and the converted phase, Ps, share the last ray leg. The multiple bounces approximately 15–20 km offshore.

Before moving on to the interpretation of the signals, we note that the area sampled by these receiver functions is substantially larger than the island of Montserrat (Figure 4.4). Obviously one of the most interesting regions of the lower crust to sample is that directly beneath the SHV, but geometry limits our ability to isolate this region. Fresnel zones associated with the receiver function converted phase from a depth near 30 km are on the order of 510 km diameter for the shorter periods in the receiver functions. In Figure 4.2 we show the areas for the two back azimuths that are sampled by the Ps converted waves from the eight stations. Individual Fresnel zones for a depth of 30 km were estimated and then the approximate areas spanned by the zones from all stations were outlined with the circles shown. Vertical resolution is a few km, and is addressed later when we model receiver function waveforms. As a result of the geometry of the incoming P waves, none of the conversions occur beneath SHV. Assuming a roughly 30 km thick crust, the multiples (PpPmS) average the crustal and crust-mantle boundary structure roughly 30 km along the direction of the incident P waves (which
extends off the map in Figure 4.2). Thus our results are applicable to the structure in the region surrounding Montserrat, and are not a specific sampling of the region directly beneath the active SHV.

4.3.1 Crustal Thickness and Properties of the Lower Crust

Earlier studies of the shallow seismic structure on the island focused on simple layered models suitable for estimating hypocentral locations. The models of Rowe et al. (2004) and those in use by the Montserrat Volcano Observatory (MVO) include strong $P$ velocity transitions ($\sim 2–3$ km/s) where the shallow surface structure meets the basement, which has a $P$ wave velocity in the 5 to 6 km/s range, at a depth of about 3 to 4 km below sea level. These models support our inference that a strong shallow velocity transition plays the dominant role in producing complicated individual receiver functions. Shalev et al. (2008, 2010) used traveltimes of signals generated by a controlled marine source as part of the onshore-offshore SEA-CALIPSO experiment to produce a three-dimensional (3-D) model of the upper crust beneath the island. They also constructed an island-wide average one-dimensional (1-D) model as a starting point for the 3-D tomography. We use this 1-D island-wide model to simulate the average structure of the uppermost crust to a depth of 7 km. For simplicity, we assumed a Poisson’s ratio (or equivalently the $V_p/V_s$ ratio) for the upper crust and focused our modeling on the average thickness, speed, and Poisson’s ratio of the middle to lower crust. To constrain the shallow structure’s shear wave speed, we adopted the Poisson’s ratio relationship from Ludwig et al. (1970) as presented by Brocher (2005). Poisson’s ratio is high in the upper kilometer or two, but averages about 0.26 in the upper 9 km of the model.

We fixed the $P$ velocity down to a depth of 7 km with the 1-D tomography model, then increased the speed to a value in the range of 6.5–7.0 km/s at a depth of about 9 km. The receiver functions contain no indication of a strong sharp velocity transition within the lower crust, so we modeled the lower part of the crust as a single uniform layer. A smooth gradient will not affect the receiver functions, so the constant speed could be also viewed as the average of a structure with a gradient. Although the receiver functions provide some information on the
velocity contrast at the base of the crust, the information on the absolute speed of the lower crust is limited (Langston, 1979; Ammon et al., 1990). To circumvent the lack of information, we tested \( P \) velocity average values in the range of 6.7–7.0 km/s for the mean of the lower crust. To simplify our discussion, we focus on a model with a lower crust \( P \) velocity of 6.9 km/s, but incorporate results from values of 6.7 and 7.0 km/s in our estimates of uncertainties for our model’s lower crustal properties. For completeness, we also varied the strength of the crust-mantle transition contrast, varying the mantle \( P \) wave speed from 7.8 to 8.2 km/s.

A receiver function response depends on the initial \( P \) wave incidence angle or equivalently the wave’s horizontal slowness, which is a function of distance to and depth of the teleseismic source. Variations in the horizontal slowness correspond to variations in wave amplitudes (from variations in the reflection and conversion coefficients) and the wave traveltimes, particularly the multiples. A broad sampling of slowness values is ideal for receiver function analyses. The average incident \( P \) wave horizontal slowness values of the two azimuthal clusters is \( \sim0.06 \text{ s/km} \), which corresponds to \( P \) wave sources at about 60° distance (as determined from global seismic traveltime tables). For signals from the southern back azimuthal cluster, although the average was 0.06 s/km, we also included several signals from greater distance that correspond to a lower incident \( P \) wave slowness. To account for the incidence angle variability, when comparing the signals with the observed receiver function stacks, we weighted and combined the responses for the different ray parameters to match the variability that existed in the observations. We computed the one-dimensional Earth model responses using the reflection matrix method of Kennett (1983).

The preferred fit associated with a lower crust \( P \) velocity of 6.9 km/s is shown in Figure 4.5, which contains three plots, one for the whole-island stack, and one each for the two azimuthal clusters. In each plot, the observed receiver function (thick line) is compared with the predictions from models with crust thicknesses between 25 and 35 km. The best visual fits are identified with stars, but the results on either side of the preferred thicknesses are acceptable. Inspection of results for other lower crust speeds suggests that an uncertainty of about 23 km is a reasonable estimate. Poisson’s ratio in the lower crust of this model is relatively high, \( \sim0.29 \), and we can fit the observations reasonably well assuming a Poisson’s ratio value.
between about 0.28 to 0.31. Assuming this level of variation in the Poisson’s ratios of the lower crust does not appreciably change the crustal thickness estimates beyond the 23 km uncertainty estimates.

Figure 4.5. Radial receiver function stacks compared with simple Earth model responses for a range of crust thickness. The upper crust is fixed to resemble the one-dimensional active source $P$ wave travelt ime based model of Shalev et al. (2008) with an assigned Poisson’s ratio of 0.26. The receiver function low-pass Gaussian filter width factor is 2.5, but they have been filtered using an acausal two-pole Butterworth filter with a corner period of 4 s. Stars identify the preferred thicknesses. The lower crust Poisson’s ratio in the model is 0.29.

Poisson’s ratios estimated using flat-lying surfaces are obviously approximate in island analyses where the multiple reflection point is offshore. In this case, the $PpPmS$ multiple bounces up to $\sim 1$ km beneath the station elevation (Figure
The result is that the multiples will arrive about 0.4 s earlier than they would if the structure was laterally uniform. We can test the consequences of the bias, and explore the uncertainty in the estimated lower crustal structure more rigorously. Using a fixed horizontal slowness, we model the observed arrival times of the converted phase ($Ps$) and multiple ($PpPmS$) with a simple model with multiples that reflect off the seafloor about 0.5–1.0 km beneath the station. We performed a grid search over crustal thickness and Poisson’s ratio. We assumed a difference in elevation between the multiple bounce point and the station elevation of 1.2 km (an average ocean floor depth of 1 km and an average station elevation of 200 m), although the results are not overly sensitive to this number. We changed the velocity model for the first two legs of the multiples by removing the top layers of the model used in the 1-D modeling.

The results for three different values of lower crustal $V_p$ centered on 6.9 km/s are shown in Figures 4.6 and 4.7. The contoured misfit is one half (2 observations) the sum of the variance-weighted traveltimes. We assumed a standard deviation for the arrivals of 0.125 s for the $Ps$ phase and 0.25 s for the $PpPmS$ phase. We can pick the peak of the arrivals at least that well, but the simplifying assumptions of lateral homogeneity led us to these conservative estimates. The match to the times is very good, minimum misfit values are about one tenth of the assumed uncertainty. The ranges of crustal thickness and $V_p/V_s$ ratios that match the observations illuminate the tradeoff between Poisson’s ratio and the crustal thickness (as constrained by traveltimes). The crustal thickness is relatively well constrained and Poisson’s ratio is greater than 0.27 for all cases (compared with the often assumed standard value of 0.25) and the best fits are associated with higher ratio values. Consistent with the waveform modeling, we observe a difference in crustal thickness; the estimated thickness south of the island is a few kilometers more than that to the northnorthwest. Our preferred results are for the $P$ wave speed of 6.9 km/s. The results for the $V_p = 6.6$ km/s and $V_p = 7.2$ km/s are shown to illustrate the relative insensitivity of the thickness and Poisson’s ratio estimates to the assumed $P$ wave speed. Models with high values of $V_p$ begin to reduce the amplitudes of the receiver function converted phases, which also leads us to favor the results for a $V_p$ in the lower crust of 6.9 km/s. We prefer the higher value because it is more consistent with the refraction-based velocity estimate for the
lower crust in the arc region between Guadaloupe and Martinique (Boynton et al., 1979). Assuming a one standard deviation equivalent for the assumed uncertainty, the Poisson’s ratio range lies within the range of 0.28–0.33 for both azimuths. The crustal thickness averages about 28±2 km northwest of the island and 32±2 km south of the island. The uncertainties would be slightly larger if we included uncertainty in the assumption of the lower crustal $P$ velocity.

![Figure 4.6](image.png)

**Figure 4.6.** Contoured normalized traveltime misfit for the northwest back azimuth radial receiver function. Dark shaded regions show the best fitting values of Poisson’s ratio (or $V_p/V_s$ ratio) and crustal thickness. The misfit values are normalized such that a value of one indicates a misfit to the $Ps$ by 0.125 s and the $PpPmS$ by a value of 0.25 s (conservative estimates of how well we can pick the peaks of these arrivals). The dotted lines outline the range of models that fit each observation within the picking criteria. (left) The misfit assuming a mean crustal $V_p$ of 6.6 km/s for the lower crust and a mean lower crustal $V_p$ of (middle) 6.9 and (right) 7.2 km/s. Previous seismic studies and the receiver function amplitudes prefer the higher lower crust $P$ wave velocities; the result at 6.6 km/s is shown to illustrate the traveltime sensitivity.

In summary, the receiver functions indicate that the crustal thickness in the Montserrat region averages roughly 30 km, with slightly thinner crust to the northwest of the island. The Poisson’s ratio of the lower crust is at least 0.28 and perhaps slightly larger, a value which is high relative to global continental crustal averages (Zandt and Ammon, 1995; Christensen and Mooney, 1995). Our observations are consistent with an average lower crust velocity of about 6.9 km/s. The high Poisson’s ratio and high $P$ wave speed suggest a chiefly mafic lower crust. Because the constant 6.9 km/s speed can also be viewed as the average of a layer structure with a gradient, the results are also consistent with a lower layer exhibiting a gradation
of rock types, from intermediate types such as diorite near the top, to dominantly mafic types such as gabbro near the base. This is particularly likely under SHV.

4.4 Discussion

Since island arcs are thought to play an important role in continental growth (at least during the Phanerozoic), the first-order seismic structure of the arcs provides important constraints on how continents evolve. The general assumption is that the bulk composition of material extracted from the mantle beneath most island arcs is basaltic (e.g. Davidson and Arculus, 2005). If one assumes that most intraoceanic arcs retain an overall basaltic composition, one challenge to the idea of building continents from arcs is that average continental composition is andesitic, not basaltic (McClennan et al., 2005). Substantial modification of initially mafic melts can occur due to fractionation and assimilation, leading to residual melts and rocks of intermediate rock compositions, but also generating cumulate assemblages.
But unless arcs do not originate with a basaltic composition (Kelemen, 1995), at least one process is needed to allow the arc crust to lose substantial mafic material if it is to play a major role in continent growth. A leading hypothesis to explain the change in composition is the delamination of a dense mafic and ultramafic cumulate from the base of arc crust (Nelson, 1991; Holbrook et al., 1999; Davidson and Arculus, 2005). Evidence from exposed island arc sections such in Talkeetna, Alaska and Kohistan, Pakistan indicate that the most mafic components of the arc-forming melt systems are in fact absent (Behn and Kelemen, 2006; Garrido et al., 2007). Whether that process occurs continuously or during times of strong tectonic activity such as arc-involved collisions remains to be determined. Early studies of intraoceanic arcs suggested that the lower crust had high $P$ velocities, consistent with dominantly mafic material. Recent work has confirmed the early results to some extent, but also shown that structure between arcs, and within large segments of arcs, can vary significantly. Controlled seismic source data from the eastern Aleutian arc, for example, suggest that the Aleutian arc is underlain by a thick, high-speed lower crust ($7.0 \leq V_p \leq 7.7$ km/s) interpreted to be mafic to ultramafic cumulates with a crustal thickness varying between 35 and 37 km (Holbrook et al., 1999; Shillington et al., 2004). Kodaira et al. (2007) used active source data to image the subsurface beneath the northern Izu intraoceanic arc. They inferred a variable structure with crustal thickness ranging from 26 to 32 km (estimated uncertainty of $\pm 2$ km) along the length of the arc, with alternating crustal thickness changes varying on a length scale of $\sim 50–100$ km. The average middle and lower crustal velocity varied along the arc length ($\sim 80$ km wavelength) with regions of slower average velocity ($\sim 6.7$ km/s) alternating with regions of faster average $P$ wave speed ($\sim 7.1$ km/s). With the exception of the Aleutians, most arc lower crusts seem to have $P$ wave speeds less than about 7.4 km/s, which Behn and Kelemen (2006) suggest is roughly the range for which density instability begins in rocks such as gabbronorite, which are found deep within the exposed sections of intraoceanic arcs.

A simplified comparison of the $P$ velocity structures for the Lesser Antilles region is presented in Figure 4.8. Early work by Officer et al. (1959) around the Caribbean region suggested that a variable, perhaps modified oceanic crust underlies the entire region. We show an average representation of their results
for the Venezuelan Basin, west of Aves Ridge in Figure 4.8. In general, the basin crust is roughly 50% thicker than typical oceanic crust. We believe this structure may resemble the primitive crust upon which the Lesser Antilles arc was built, but the regional tectonic history of the Caribbean plate is complex (Pindell and Barrett, 1990; Macdonald et al., 2000; Christeson et al., 2008), and the assumption could be incorrect. Boynton et al. (1979) analyzed explosion-generated refracted wave travel times along the central Lesser Antilles (Martinique to Guadeloupe) and estimated a variable thickness 6.0–6.2 km/s middle crust overlying a 6.9±0.2 km/s lower crust along the arc. Upper mantle arrivals were absent in the line parallel to the arc, so direct observations of crustal thickness were unavailable. Using shots from the survey line perpendicular to the arc that had a favorable geometry for propagation along the arc south of Martinique, they estimated an average crustal thickness of about 30 km (Figure 4.8). Combining seismic data from the profile perpendicular to the arc with gravity observations, they estimated a slightly thicker crust of ~34 km beneath St. Vincent and a faster lower crustal layer (7.4 km/s) (Figure 4.8). Models derived using estimated traveltime observations from a single explosion near Barbuda observed on stations from throughout the northern Lesser Antilles during the Lesser Antilles Deep Lithospheric Experiment (LADLE) suggested a slightly slower lower crust (6.6 km/s) and a slightly thinner crust (27 km) (Whitmarsh et al., 1983). Christeson et al. (2008) estimated a Lesser Antilles crustal thickness of 24 km. The lower crust in their model has a slightly higher $P$ velocity than typical continental material, and they inferred that to indicate that it is slightly more mafic than the average continent. An average of the model’s lower crust contains a smooth gradient with a maximum speed less than 7.3 km/s, and thus no evidence for mafic or ultramafic cumulates with $V_p > 7.4$ km/s at the base of the crust.

Although the variability in the seismic structure at intraoceanic arcs complicates generalizations, a typical intraoceanic arc structure consists of an upper crust of intrusive and extrusive igneous rocks, sedimentary and volcaniclastic rocks that are altered, fractured, and porous and have substantially slow $P$ wave speeds ($\leq 5.0$ km/s). The middle crust varies in thickness between and within arcs, and generally has wave speeds in the 6.0–6.5 km/s range. Slower speeds and thin (~5 km thick) middle crusts have been interpreted as original oceanic crust on which the arc was
built. Alternate views suggest that this layer is composed of intrusive rocks of mafic and intermediate composition developed by differentiation of deeper mafic magmas (e.g. Holbrook et al., 1999). Thus in the vicinity of Montserrat, observed seismic velocities of \( \sim 6.2 \) km/s at \( \sim 5–10 \) km depth seem compatible with an intermediate, dioritic composition (Paulatto et al., 2010; Christensen and Mooney, 1995), and this is consistent also with the observed eruptions of andesite at SHV. In most arcs the lower crust \( P \) wave speeds range from the high sixes to the high sevens of km/s. The thickness of this deeper layer ranges from a few kilometers beneath some arcs to 10–20 km beneath others. This layer is interpreted to represent mafic rocks, and mafic and ultramafic cumulates left behind by magmatic differentiation processes.

Our estimated range of crustal thickness, 26–34 km, for the Montserrat region is consistent with earlier estimates of the thickness beneath the northern Lesser Antilles arc (Boynton et al., 1979; Whitmarsh et al., 1983) and slightly thicker than the Christeson et al. (2008) estimate for the southern Lesser Antilles. The
assumed $P$ wave speed for the lower crust is compatible with the older refraction profile estimates (e.g. Boynton et al., 1979; Whitmarsh et al., 1983) and with the average speed in the model of Christeson et al. (2008). We have also shown that the Poisson's ratio for the lower crust is relatively high (>0.27). The receiver function observations allow values as high as 0.33–0.34, but we think that it is more likely that the value is in the range of 0.28–0.30 since few common rocks have ratios higher than 0.30 under laboratory conditions (Christensen, 1996). The high Vp value and the relatively high Poisson's ratio are consistent with a mafic composition such as gabbroic rocks (Christensen, 1996; Brocher, 2005; Behn and Kelemen, 2006). We interpret the lower crust (greater than ~10 km thick) in our model to be modified Caribbean crust (originally similar to the Venezuelan basement) that has been intruded and thickened with differentiated mafic material added to the crust as part of the subduction process. We see little evidence for a deep layer of mafic to ultramafic cumulates (Vp > 7.4 km/s) although our modeling of the $Ps$ and $PpPmS$ amplitudes suggests that a layer a few kilometers thick could exist undetected near the crust-mantle boundary.

One possible explanation for the absence of cumulates is that they are removed by subduction processes as they form. This may be typical of most arcs, since the Aleutians is the only arc with a region containing rocks with Vp > 7.4 km/s. An alternate view is that what we commonly identify as the crust-mantle transition is in fact a seismic velocity increase associated with an increase in garnet within the cumulates beneath the arc. However, Montserrat andesite displays no geochemical signature of a garnet influence in crystal fractionation (R. S. J. Sparks, personal communication, 2008). This observation argues strongly against a role of garnet in explaining the 30 km deep velocity increase, and is at the same time consistent with our crustal thickness estimate of ~30 km, rather than ~35 km. Other supportive arguments include the models of Behn and Kelemen (2006), that suggest garnet would form in gabbroic rocks at >35–40 km depths or deeper, and the models of Furlong and Fountain (1986), that predict lower seismic wave speed increases from garnet additions to crustal rocks than required by our data to match the amplitude of the receiver function converted phases. Thus we prefer to interpret the $Ps$ and $PpPmS$ arrivals as originating from the crust-mantle boundary beneath the Montserrat region.
Wadge (1984) used the volume of observed surface volcanic rocks to estimate that during the last 0.1 Ma, the Lesser Antilles arc volcanic production rate per unit kilometer of arc length was $\sim 4.0 \text{ km}^3 \text{ Ma}^{-1} \text{ km}^{-1}$. If we assume that the original Lesser Antilles crust resembled the Venezuelan Basin crust, then the thickness has tripled from roughly 10 km to roughly 30 km. To produce the observed thickening of 20 km in 50 Ma across a 50 to 100 km wide arc (Figure 4.1) requires a growth rate of 20 to 40 km$^3$ Ma$^{-1}$ km$^{-1}$, which is lower than Reymer and Schubert (1986)’s estimate of 70 km$^3$ Ma$^{-1}$ km$^{-1}$ for the long-term growth rate of the Lesser Antilles, but comparable to the range (30–50 km$^3$ Ma$^{-1}$ km$^{-1}$) of more recent results from numerical modeling of subduction processes (Nikolaeva et al., 2008). Finally, we note that to convert a 30 km thick Lesser Antilles arc structure to average continental crust from the Lesser Antilles arc, would require thickening the crust by about 10 km and slightly reducing the mean speed of the lower crust (e.g. Christensen and Mooney, 1995). The point here is not that crust evolves that way in any specific case, but appears to do so in bulk. The average Poisson’s ratio for our Montserrat region crust is 0.28, which is slightly above continental average of 0.27, but well within the ranges observed on the continents Zandt and Ammon (1995).

4.5 Conclusions

The observations from this work include the timing and amplitude of $P$ to $S$ converted phases propagating between a relatively strong and less than 4 km thick crust-mantle transition located roughly 26–34 km beneath the region surrounding Montserrat. We estimate that the average thickness of the crust beneath the Montserrat region is $\sim 30$ km, and slightly thinner to the northwest than the south. If we assume that the Lesser Antilles arc formed on top of modified oceanic crust similar to that found in the Venezuelan basement, then the crustal thickness has increased by at least 10–20 km in the last 40 Ma. This value is reduced if the current arc formed on a thick accretionary prism of the older Aves Ridge arc or on a Mesozoic protoarc structure (Bouysse et al., 1990; Macdonald et al., 2000). Tomography results (Shalev et al., 2008, 2010; Paulatto et al., 2010) show that the upper crust has a strong gradient in P wave speed starting from very slow
velocities. Our resolution of the seismic wave speeds in the middle and lower crust speed is limited. We assumed that the mean speed in the lower crust is roughly 6.9 km/s, and that for reasonable perturbations around that value, the lower crust has a Poisson’s ratio greater than about 0.27 ($V_p/V_s > 1.78$) and more likely is closer to 0.29–0.30. The high wave speed and Poisson’s ratio strongly favor a mafic (gabbro) composition for the lower crust, but intermediate compositions in the upper part are not precluded, especially under SHV and the extinct volcanic centers. We see little evidence of a substantial cumulate layer ($V_p > 7.4$ km/s) in the lower crust, suggesting that cumulates remaining from deep fractionation were removed earlier (or perhaps continuously), consistent with other observations of active intraoceanic
5.1 Abstract

The three-dimensional $P$-wave velocity structure beneath Montserrat is imaged tomographically down to $\sim 8$ km depth. The tomographic inversions used 3116 arrival times from 294 local events and recorded by 47 permanent and temporary seismic stations to image the first-order crustal structure. Key results include a velocity anomaly extending to $\sim 6$ km depth with $P$ velocity ranging from 3.5–5.6 km/s corresponding to volcaniclastics and andesitic intrusive bodies. Discrete magma storage regions were not specifically resolved by the tomographic inversion and consequently any shallow magma pockets are $< 2$ km in diameter and the main magmatic reservoir is most likely deeper than 6–8 km.

5.2 Introduction

Montserrat is located on the northern segment of the Lesser Antilles island arc (Figure 5.1), which formed from the westward subduction of the lithosphere of the Atlantic Ocean beneath the Caribbean plate (Macdonald et al., 2000). Arc seismicity is characterized by shallow to intermediate depth earthquakes, with the majority of events occurring east of the arc axis (Feuillet et al., 2002). The
Caribbean crust has a thickness intermediate between that of typical ocean and continental lithosphere, and near Montserrat a mean crustal thickness of 30 km has been established from receiver function analysis (Sevilla et al., 2010).

Figure 5.1. Topographic map of Montserrat Island. The inverted triangles are the seismic stations and the dots are one example of grid node configurations used in the tomographic inversion. The main physiographic features are indicated by the labels; SVH - Silver Hills, CH - Centre Hill, SHV - Soufrière Hills, SSH - South Soufrière Hills.

The crustal structure underlying Montserrat is further investigated, with the main aim to improve understanding of its magmatic system (cf. Druitt and Koke-
Laar, 2002). Presented here are the tomographic results of 3D velocity structure using travel-time picks from selected local events. Although most of the local observations sample the crust above presumptive magma storage targets (Voight et al., 2010), the plan was that improved resolution of the shallow structure would, in a further iteration, allow enhanced imaging of deeper structure. Unfortunately, late-discovered timing issues on more recently-deployed seismic recording systems interfered with the inclusion of recent (post-2000) teleseismic observations into the tomographic inversion. The resulting model has no resolution in the deep middle and lower portions of the crust, but still has the potential to help image deep structure at a later time, and may provide a useful framework for improved studies of island seismicity.

5.2.1 Background on the Montserrat Magmatic System

Montserrat is comprised of three main volcanic centers with non-overlapping activity: Silver Hills (c.2600 to 1200 ka), Centre Hills (c.950 to 550 ka) and Soufrière–South Soufrière Hills (c.170 ka to present) (Figure 5.1) (Harford et al., 2002). The current eruption of Soufrière Hills volcano (SHV), 1995–present, is attributed to an andesitic body at shallow to mid-crustal levels reheated periodically by an influx of hot mafic magma from a deeper source (e.g. Murphy et al., 2000; Harford and Sparks, 2001). Constraints on the top of shallow magma storage at ∼5 km depth are inferred from the distributions of local earthquakes (Aspinall et al., 1998; Miller et al., 2010), petrology and geochemistry (Barclay et al., 1998), and ground deformation measurements (Mattioli et al., 2010; Voight et al., 2010). Recent shallow-crust tomographic results of the SEA-CALIPSO experiment are consistent with this result (Paulatto et al., 2010; Shalev et al., 2010). Further details on storage geometry are controversial (Voight et al., 2010). Deformation data suggest sources at mid-crustal levels (Mattioli et al., 2010; Voight et al., 2010), and upper storage zones may connect to a deeper reservoir >10 km depth, and to the ponded mafic magma near the Moho by vertical conduit systems (Devine et al., 2003; Zellmer et al., 2003; Elsworth et al., 2008). The aim of this study is to further explore the aspects of magmatic system structure at crustal level depth.
5.3 Seismic Data and Tomographic Methods

The TOMOG3D code of Zhao et al. (1994, 1992) is used to invert the P-wave travel time data. TOMOG3D adopts a pseudo-bending (Um and Thurber, 1987) and Snell’s law approach to ray tracing, and a damped least squares method with additional smoothing constraints [e.g., Aki and Richards, 1980] to invert the travel time data to velocities at grid nodes. At any point in the model, the velocity perturbation is calculated by linearly interpolating the perturbations at the eight grid nodes surrounding that point. The code also allows 3-D inhomogeneous velocity model with velocity discontinuities of complex shapes as an *a priori* information.

For the initial model, the crustal 1-D velocity model of Shalev et al. (2008, 2010) was adapted and averaged every 1-km depth for the upper 9-km crustal layer. The mid and lower crust initial model layers are obtained from the receiver function analysis of broadband stations in Montserrat by Sevilla et al. (2010) and the IASP91 (Kennett and Engdahl, 1991) Earth model for the mantle layers. Only about one-third of the island of Montserrat is above sea-level, which is about 16 km long and 10 km wide. Several checkerboard resolution tests with different node separations were conducted to choose the optimum grid spacing that would match the available ray coverage to image well resolved structures in the region. A velocity perturbation of ± 6% was assigned alternately at every grid point on the initial model. Figure 5.2 shows an example of one resolution test with horizontal node separation of ~2 km and vertical (depth) separation of 2 km. The checkerboard pattern is well-resolved to about 6 km to 8 km depths, which is within or at the top of the presumed shallow magma chamber. The minimum horizontal node separation may be larger than the scale length of heterogeneity therefore four different grid configurations were used to mapped-out the horizontal locations of the perturbations.

The travel time data set consists of travel times from earthquakes that occurred in 1995 during the initial stages of volcanic activity in Montserrat and in 2007 during the SEA-CALIPSO seismic experiment. The arrival times of the earthquakes were manually picked with accuracy within 1–2 ms and the hypocenters were re-located using precise relocation techniques (Miller et al., 2010). A total of 3116 arrival times from 294 local events and recorded by 47 stations were used in the
Figure 5.2. An example of checkerboard resolution test results for a grid spacing of ~2 km. Random errors were added on the synthetic travel time data with 0s mean and 0.1s standard deviation to simulate the possible errors on the original data.

5.4 Results

An empirical approach was used to obtain an optimum damping factor for the inversion. Using real data set and the initial velocity model, a number of inversions were run with different damping factors to find a good compromise between data and solution variances. The preferred damping parameter value is 6.0 that gives the best compromise between the model roughness and stability of the results.

After performing the tomographic inversion, the weighted RMS residual for all four grid configurations is reduced from about 0.32 s at the initial inversion to 0.26 s after the inversion. Figure 5.3 shows an example of the final tomographic images of the Island of Montserrat. Other slices and corresponding checkerboard
resolution test results are presented in Figures C.1 to C.7 in Appendix C. The key features of the results are the near-surface, relatively high velocity zone, but slower than the 1-D model of Shalev et al. (2008, 2010) beneath the Soufrière Hills. The velocity anomaly extends down to \( \sim 6\text{–}8 \text{ km depth} \) with \( P \) velocities ranging from 3.5–5.6 km/s. The velocity values likely correspond to volcaniclastics and andesitic intrusive bodies and are consistent with the results of Paulatto et al. (2010) from their refraction/reflection tomography results.

![Tomographic results showing the absolute \( P \)-wave velocity from the inversions using arrival times of local earthquakes (dots). See Figure 5.1 for the slice locations.](image)

**Figure 5.3.** Tomographic results showing the absolute \( P \)-wave velocity from the inversions using arrival times of local earthquakes (dots). See Figure 5.1 for the slice locations.

### 5.5 Discussion

The 3-D \( P \)-wave velocity structure of the Island of Montserrat is presented in detail up to \( \sim 8 \text{ km depth} \). South of the island and within the Soufrière Hills area is a relatively high velocity region where the clustering of local earthquakes
occur. On either side of this region are lower velocity zones 2–4 km below the surface and may be related to assemblages of prior volcanic eruption deposits. The island is underlain by simple crustal velocity structure with P-wave velocity slower than the average velocity obtained by Shalev et al. (2008, 2010) at similar depths. These relatively low P–velocity values may reflect the elevated temperature of the region due to volcanic activities. The tomographic image results do not reveal anomalously low velocity zones that may indicate magma storage region on any part of the island. This observation may suggest that any shallow storage regions are too small to be resolved by the tomographic inversion or there are too few deep events to illuminate the structure at depth. With limited first-arrival energy penetrating to depth, all we can conclude is that any magmatic features in the shallow crust are likely smaller than 2 km.

5.6 Conclusion

The subsurface P–wave velocity structure beneath Montserrat was imaged using travel time tomography of local earthquakes. The tomographic inversions used 3116 arrival times from 294 local events and recorded by 47 permanent and temporary seismic stations. Based on checkerboard resolution tests, the inversion was able to resolve structures as deep as \( \sim 6–8 \) km. Key results include a velocity anomaly extending to \( \sim 6 \) km depth with \( P \) velocity ranging from 3.5–5.6 km/s corresponding to volcaniclastics and andesitic intrusive bodies. Discrete magma storage regions were not specifically resolved by the tomographic inversion and consequently are <2 km diameter and most likely perhaps deeper than 5 km.
Appendix A

Checkerboard Resolution Tests of The Vertical Sections Along The PIA
Figure A.1. Histogram of the randomly generated noise with 0 s mean and 0.1 s standard deviation added to the synthetic travel times used in the checkerboard resolution tests.
Figure A.2. Map of Northern Luzon region showing the locations of the vertical sections of the checkerboard images. Also labeled are the structures discussed in the text. The square symbols mark the seismic stations used in this study. NLT– North Luzon Trough, VH– Vigan High, WLT– West Luzon Trough, BNPP– Bataan Nuclear Power Plant, NPCT– North Palawan Continental Terrane, M.B.– Marinduque Basin. Inset figure shows the geographic location of the area relative to the PIA.
Figure A.3. Results of checkerboard resolution tests (see Figure A.2 for the slice locations). (Top) Plots of topography profiles (vertical exaggeration, V.E. = 15x). The major structures described in the text are traced on top of the patterns. MT– Manila Trench, NLT– North Luzon Trough, SCS– South China Sea. The scale bar shows the velocity perturbation.
Figure A.4. See Figure A.3 for the general description of the figure. MT– Manila Trench, NLT– North Luzon Trough, SCS– South China Sea, ELT– East Luzon Trough.
Figure A.5. See Figure A.3 for the general description of the figure. MT– Manila Trench, VH– Vigan High, WLT– West Luzon Trough, SCS– South China Sea, ELT– East Luzon Trough, B.R.– Benham Rise, WPB– West Philippine Sea Basin.
Figure A.6. See Figure A.3 for the general description of the figure. MT– Manila Trench, WLT– West Luzon Trough, SCS– South China Sea, ELT– East Luzon Trough, PT– Philippine Trench, WPB– West Philippine Sea Basin.
Figure A.7. See Figure A.3 for the general description of the figure. MT– Manila Trench, SCS– South China Sea, M.B.– Marinduque Basin, ELT– East Luzon Trough, PT– Philippine Trench, WPB– West Philippine Sea Basin, PFZ– Philippine Fault Zone.

Figure A.8. See Figure A.3 for the general description of the figure. MT– Manila Trench, SCS– South China Sea, M.B.– Marinduque Basin, PT– Philippine Trench, WPB– West Philippine Sea Basin, PFZ– Philippine Fault Zone.
Figure A.9. Map of Central and Southern Philippines showing the locations of the vertical sections of the checkerboard resolution tests results. Also labeled are the structures discussed in the text. The square symbols mark the seismic stations used in this study. Inset figure shows the geographic location of the area relative to the PIA. CMA - Central Mindanao Arc, WPB– West Philippine Sea basin
Figure A.10. See Figure A.3 for the general description of the figure. (see Figure A.9 for the slice locations). SCS– South China Sea, PT– Philippine Trench, WPB– West Philippine Sea Basin, PFZ– Philippine Fault Zone.

Figure A.11. See Figure A.3 for the general description of the figure. (see Figure A.9 for the slice locations). SS– Sulu Sea, PT– Philippine Trench, WPB– West Philippine Sea Basin, PFZ– Philippine Fault Zone.
Figure A.12. See Figure A.3 for the general description of the figure. (see Figure A.9 for the slice locations). SS– Sulu Sea, NT– Negros Trench, PT– Philippine Trench, WPB– West Philippine Sea Basin, PFZ– Philippine Fault Zone.

Figure A.13. See Figure A.3 for the general description of the figure. (see Figure A.9 for the slice locations). ST– Sulu Trench, SS– Sulu Sea, PT– Philippine Trench, WPB– West Philippine Sea Basin, PFZ– Philippine Fault Zone, CMA– Central Mindanao Arc.
Figure A.14. See Figure A.3 for the general description of the figure. (see Figure A.9 for the slice locations). CT– Cotabato Trench, CS– Celebes Sea, PT– Philippine Trench, WPB– West Philippine Sea Basin, PFZ– Philippine Fault Zone.
Appendix B

Tomography Results Using Local and Regional Events Only
Figure B.1. Map of Northern Luzon region showing the locations of the vertical sections of the images. Also labeled are the structures discussed in the text. The square symbols mark the seismic stations used in this study. NLT– North Luzon Trough, VH– Vigan High, WLT– West Luzon Trough, BNPP– Bataan Nuclear Power Plant, NPCT– North Palawan Continental Terrane, M.B.– Marinduque Basin. Inset figure shows the geographic location of the area relative to the PIA.
**Figure B.2.** Tomography results using local and regional events only (see Figure B.1 for the slice locations). (Top) Plots of topography profiles (vertical exaggeration, V.E. = 15x). The major structures described in the text are traced on top of the patterns. MT–Manila Trench, NLT– North Luzon Trough, SCS– South China Sea. The open circles are the earthquake data that I used in the inversion. The scale bar shows the velocity perturbation.
Figure B.3. See Figure B.2 for the general description of the figure. MT– Manila Trench, NLT– North Luzon Trough, SCS– South China Sea, ELT– East Luzon Trough.
Figure B.4. See Figure B.2 for the general description of the figure. MT– Manila Trench, VH– Vigan High, WLT– West Luzon Trough, SCS– South China Sea, ELT– East Luzon Trough, B.R.– Benham Rise, WPB– West Philippine Sea Basin.
Figure B.5. See Figure B.2 for the general description of the figure. MT– Manila Trench, WLT– West Luzon Trough, SCS– South China Sea, ELT– East Luzon Trough, PT– Philippine Trench, WPB– West Philippine Sea Basin.
Figure B.6. See Figure B.2 for the general description of the figure. MT– Manila Trench, SCS– South China Sea, M.B.– Marinduque Basin, ELT– East Luzon Trough, PT– Philippine Trench, WPB– West Philippine Sea Basin, PFZ– Philippine Fault Zone.

Figure B.7. See Figure B.2 for the general description of the figure. MT– Manila Trench, SCS– South China Sea, M.B.– Marinduque Basin, PT– Philippine Trench, WPB– West Philippine Sea Basin, PFZ– Philippine Fault Zone.
Figure B.8. Map of Central and Southern Philippines showing the locations of the vertical sections of the tomography images. Also labeled are the structures discussed in the text. The square symbols mark the seismic stations used in this study. Inset figure shows the geographic location of the area relative to the PIA. CMA - Central Mindanao Arc, WPB– West Philippine Sea basin.
Figure B.9. See Figure B.2 for the general description of the figure. (see Figure B.8 for the slice locations). SCS– South China Sea, PT– Philippine Trench, WPB– West Philippine Sea Basin, PFZ– Philippine Fault Zone.

Figure B.10. See Figure B.2 for the general description of the figure. (see Figure B.8 for the slice locations). SS– Sulu Sea, PT– Philippine Trench, WPB– West Philippine Sea Basin, PFZ– Philippine Fault Zone.
Figure B.11. See Figure B.2 for the general description of the figure. (see Figure B.8 for the slice locations). SS– Sulu Sea, NT– Negros Trench, PT– Philippine Trench, WPB– West Philippine Sea Basin, PFZ– Philippine Fault Zone.

Figure B.12. See Figure B.2 for the general description of the figure. (see Figure B.8 for the slice locations). ST– Sulu Trench, SS– Sulu Sea, PT– Philippine Trench, WPB– West Philippine Sea Basin, PFZ– Philippine Fault Zone, CMA– Central Mindanao Arc.
Figure B.13. See Figure B.2 for the general description of the figure. (see Figure B.8 for the slice locations). CT– Cotabato Trench, CS– Celebes Sea, PT– Philippine Trench, WPB– West Philippine Sea Basin, PFZ– Philippine Fault Zone.
Appendix C

Tomographic and Checkerboard Resolution Tests Results for Montserrat Island
Figure C.1. (Top) Map of Montserrat Island showing the locations of the vertical profiles (Bottom). The dots are the local earthquakes within the 2-km radius of the slices. The middle plots are the topographic profiles along the slices. SVH - Silver Hills, CH - Centre Hill, SHV - Soufrière Hills, SSH - South Soufrière Hills.
Figure C.2. (Top) Map of Montserrat Island showing the locations of the vertical profiles (Bottom). The dots are the local earthquakes within the 2–km radius of the slices. The middle plots are the topographic profiles along the slices. SVH - Silver Hills, CH - Centre Hill, SHV - Soufrière Hills, SSH - South Soufrière Hills.
Figure C.3. (Top) Map of Montserrat Island showing the locations of the vertical profiles (Bottom). The dots are the local earthquakes within the 2–km radius of the slices. The middle plots are the topographic profiles along the slices. SVH - Silver Hills, CH - Centre Hill, SHV - Soufrière Hills, SSH - South Soufrière Hills.
Figure C.4. Histogram of the randomly generated noise with 0 s mean and 0.1 s standard deviation added to the synthetic travel times used in the checkerboard resolution tests.
Figure C.5. Results of checkerboard resolution tests. (Top) Map of Montserrat Island showing the locations of the vertical profiles. (Bottom). The middle plots are the topographic profiles along the slices. SVH - Silver Hills, CH - Centre Hill, SHV - Soufrière Hills, SSH - South Soufrière Hills. The scale bar shows the velocity perturbation.
Figure C.6. Results of checkerboard resolution tests. (Top) Map of Montserrat Island showing the locations of the vertical profiles (Bottom). The middle plots are the topographic profiles along the slices. SVH - Silver Hills, CH - Centre Hill, SHV - Soufrière Hills, SSH - South Soufrière Hills. The scale bar shows the velocity perturbation.
Figure C.7. Results of checkerboard resolution tests. (Top) Map of Montserrat Island showing the locations of the vertical profiles (Bottom). The middle plots are the topographic profiles along the slices. SVH - Silver Hills, CH - Centre Hill, SHV - Soufrière Hills, SSH - South Soufrière Hills. The scale bar shows the velocity perturbation.


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