Continent-arc collision in the Banda Arc imaged by ambient noise tomography

Robert W. Porritt^{1*}, Meghan S. Miller¹, Leland J. O'Driscoll¹, Cooper W. Harris¹, Nova Roosmawati¹, and Luis Teofilo da Costa²

¹Department of Earth Science, University of Southern California

²Institute of Petroleum and Geology – IPG, Timor Leste

*Corresponding author: rporritt@usc.edu, (906) 281-3675

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Abstract: The tectonic configuration of the Banda region in southeast Asia captures the spatial transition from subduction of Indian Ocean lithosphere to subduction and collision of the Australian continental lithosphere beneath the Banda Arc. An ongoing broadband seismic deployment funded by NSF is aimed at better understanding the mantle and lithospheric structure in the region and the relationship of the arc-continent collision to orogenesis. Here, we present results from ambient noise tomography in the region utilizing this temporary deployment of 30 broadband instruments and 39 permanent stations in Indonesia, Timor Leste, and Australia. We measure dispersion curves for over 21,000 inter-station paths resulting in good recovery of the velocity structure of the crust and upper mantle beneath the Savu Sea, Timor Leste, and the Nusa Tenggara Timur (NTT) region of Indonesia. The resulting three dimensional model indicates up to ~25% variation in shear velocity throughout the plate boundary region; first-order velocity anomalies are associated with the subducting oceanic lithosphere, subducted Australian continental lithosphere, obducted oceanic sediments forming the core of the island of Timor, and high velocity anomalies in the Savu Sea and Sumba. The structure in Sumba and the Savu Sea is consistent with an uplifting forearc sliver. Beneath the island of Timor, we confirm earlier inferences of pervasive crustal duplexing from surface mapping, and establish a

link to underlying structural features in the lowermost crust and uppermost mantle that drive upper crustal shortening. Finally, our images of the volcanic arc under Flores, Wetar, and Alor show high velocity structures of the Banda Terrane, but also a clear low velocity anomaly at the transition between subduction of oceanic and continental lithosphere. Given that the footprint of the Banda Terrane has previously been poorly defined, this model provides important constraints on tectonic reconstructions that formerly have lacked information on the lower crust and uppermost mantle.

1. Introduction

Coupling across convergent boundaries is fundamental to many plate boundary and plate tectonic processes. Most commonly, convergent boundaries take the form of subduction of oceanic plates where negatively buoyant oceanic lithosphere pulls the downgoing plate into the lower mantle. This style of convergence often results in a relatively narrow zone of deformation focused along the plate interface. However, when anomalously buoyant subducting lithosphere (e.g. continental or volcanic arc) encounters a convergent boundary, the downgoing plate resists subduction leading to a complex deformational environment. While the understanding of the process of oceanic subduction has advanced to reveal subtleties in the variations of subduction parameters, the process of continent-arc collision is not well-represented on the modern Earth (e.g. Moresi et al., 2014 and references therein). The southeast Asian archipelago is one such region of continent-arc collision and therefore provides a unique opportunity to investigate these processes (e.g. Audley-Charles, 1968; Harris, 2011; Hall and Spakman, 2015).

The Banda Arc is a collisional environment where various plate fragments in southeast Asia interact with the northwestern Australian plate (i.e. Hamilton, 1979). This collision is occurring at the eastern termination of the Java Trench where the Indian Ocean plate subducts under Eurasia forming the Indonesian Arc. The seafloor ages here increase to the east along the arc

reaching Jurassic ages south of the Island of Sumba (Müller et al., 2008) where the lithosphere transitions from oceanic to the Australian continental margin. At this transition, there is a significant bifurcation of the arc, as the volcanic arc goes from a single linear chain to a set of inner and outer arc islands, with the inner arc being volcanically active and the outer arc representing the collisional front. It is this inner and outer arc system that is collectively referred to as the Banda Arc. The collisional orogen is relatively young based on plate reconstructions (e.g. Harris, 1991; Hall, 2002). However, other data from onshore geological investigations offer a range of ages of initiation of the collision from ~16 Ma to ~1 Ma (Audley-Charles, 1968; Rutherford et al., 2001; Keep and Haig, 2010; Spakman and Hall, 2010; Harris, 2011; Hall and Spakman, 2015).

Oblique convergence along the plate boundary is evident up the orientation of the Timor Trough, plate motions in a no-net rotation frame (DeMets et al., 2010), and GPS velocity vectors (Bock et al., 2003; Calais et al., 2006; Simons et al., 2007; Nugroho et al., 2009, Koulali et al., 2016x) (See Fig. 1). In the no-net rotation frame, the Australian plate is moving to the northeast as the islands of Flores and Sumba move towards the east, but the GPS vectors relative to the Sunda block show north-northeast displacements. This suggests the collision progressed with initial collision along a promontory of the Australian plate and then continued southwestward along the Timor Trough. This effective time-space trade-off allows investigation into the temporal evolution of arc-continent collision by comparing different sections of the orogen (e.g. Harris, 2011). Past studies have either taken a large-scale view from teleseismic imaging (e.g. Widiyantoro et al., 2011; Hall and Spakman, 2015) or very small scale views from mapping outcrops (e.g. Audley-Charles, 1968; Rosidi et al., 1979; Harris et al., 1998, 2009; Roosmawati and Harris, 2009; Harris, 2011; Tate et al., 2015) and active source seismic images (e.g. Lüschen, et al., 2010; Rigg and Hall, 2012).

75 The region is among the most seismically active globally and has a clearly defined Wadati-
76 Benioff zone tracing out a northward dipping slab to ~600 km depth (Das, 2004). This structure
77 is confirmed by global P wave tomography (e.g. Widiyantoro et al., 2011) which indicates a high
78 velocity body coincident with the seismicity (e.g. Das, 2004; Ely and Sandiford, 2010) and
79 follows the trench continually from the Java Trench around the Timor Trough to southward
80 dipping subduction at the northeastern end of the Banda Sea. However, it is argued whether
81 there is a single slab or two slabs around this juncture at the eastern end of the Banda Sea (e.g.
82 Milsom, 2001; Das, 2004; Widiyantoro et al., 2011).

83 Recent imaging by Fichtner et al. (2010) using full waveform tomography indicates the
84 presence of high velocity material at 200 km depth under Timor and eastward, but not under the
85 western part of the Banda Arc. They infer that this high velocity anomaly is subducting
86 Australian continental lithosphere due to the correspondence with He^3/He^4 ratios of ~1 (Hilton et
87 al., 1992), which indicates continental material in the source of arc magmas. Furthermore, the
88 the imaged high velocity body is continuous with the Australian cratonic lithosphere, suggesting
89 a connected plate. This suggests that the Australian lithosphere is subducting, but their model
90 lacks the necessary resolution to determine what happened to the lower density continental
91 crust. Therefore, higher resolution images of the system are needed, such as those that are
92 achievable through the newly developed method of ambient noise tomography.

93 Crustal duplexing during the collision has largely been determined through structural
94 mapping of the islands (e.g. Audley-Charles, 1968; Rosidi et al., 1979; Harris et al., 1998, 2009;
95 Roosmawati and Harris, 2009; Harris, 2011; Tate et al., 2015) and offshore active source
96 seismic imaging (e.g. Lüschen, et al., 2010; Rigg and Hall, 2012). Major units of the bifurcated
97 arc include the Australian and Gondwana Passive Margin Sequence, the Banda Terrane and
98 synorogenic sediments and mélangé (e.g. Audley-Charles, 1968; Harris, 1998). The Australian
99 Passive Margin Sequence is the youngest sequence and consists primarily of carbonates that

100 formed on the continental slope and turbidites shed from the shelf to the south (Rigg and Hall,
101 2012). This unit is underlain by Permian to Jurassic pre- and syn-rift Gondwana sequence units,
102 which are primarily sedimentary deposits that formed as the Australian passive continental
103 margin was developing (e.g. Carter et al., 1976). The Banda Terrane (Audley-Charles and
104 Harris, 1990) is a forearc terrane with several different metamorphic and volcanic rocks, and
105 Asian affinity sedimentary units (Harris, 2006). Many of these units are found in an
106 autochthonous setting in Sumba (e.g. Rutherford et al., 2001), and as thrust nappes throughout
107 Timor (Carter et al., 1976). This terrane is the basement of the modern inner arc and has
108 undergone reworking through recent volcanic processes, but it can be documented throughout
109 the entire arc (e.g. Harris, 2006). Beneath the nappes of Banda Terrane is a thick tectonic
110 mélange that represents the original subduction channel that is now exhumed in Timor (Harris
111 et al., 1998). Synorogenic sediments deposited during collision include a transition from deep
112 water carbonates to turbidites and eventually non-marine conglomerates (Roosmawati and
113 Harris, 2009).

114 Significant recent structural work has been done to understand the relationships between
115 these units in the context of continent-arc collision. Balanced cross-sections for East Timor
116 (Timor Leste) presented by Tate et al. (2015) estimate that the collision has been
117 accommodated by 326-362 km of shortening inferring that 215-229 km of Australian lithosphere
118 has been subducted. Much of this shortening can be observed as Banda Terrane overthrust
119 from the north onto Timor and duplexing of the under-thrust material. This has resulted in a
120 complex system of folds and thrusts, positioning mafic and mélange material over largely
121 carbonate sequences of Australian affinity. For the other major outer arc island, Sumba, two
122 viable models have been proposed. The first model proposes that it is an autochthonous forearc
123 sliver of Banda Terrane, which is uplifting due to collision with the Scott Plateau, a promontory
124 of Australian lithosphere, from the south (e.g. Harris, 2006). An alternative model, proposed by

Rutherford et al. (2001), suggests Sumba is acting as a small scale block escaping due to collisional tectonics from an earlier position east of Flores to its current location southwest of Flores. In this alternative model, the collision with the Scott Plateau arrests the southwestward motion of the Sumba Block and initiated uplift.

The large scale views of the Banda Orogen provide a clear case of continental subduction under the Banda Arc (e.g. Hall and Spakman, 2015 and references therein) while structural mapping (e.g. Harris, 2011; Tate et al., 2015; Duffy et al., 2013) provides a view of collision between a passive margin sequence and a forearc terrane and possible extrusion tectonics. This leaves an important gap in our understanding at the meso-scale where the lower crust and uppermost mantle transfer the driving force of subduction into upper crustal deformation. We image that gap by use of ambient noise tomography to constrain the relative shear velocity structure within the upper 50 km.

2. Methods

The data for this study focuses on an NSF funded temporary deployment of 30 stations in the region installed and maintained by the University of Southern California (USC), with the support of the Portable Array Seismic Studies of the Continental Lithosphere (PASSCAL). Additionally, we use 16 stations from Geoscience Australia, 17 stations from GeoFon, 1 station from the Indonesia Tsunami Early Warning System (IA) and the preparatory commission for the Comprehensive nuclear-Test-Ban Treaty Organization (CTBTO), 1 station from the Global Seismograph Network (GSN), 1 station from the Malaysian National Seismic Network, and 3 stations from the Seismographs in Schools in Australia program to broaden the available path distribution (Fig. 1 inset). The temporary stations were initially deployed in March 2014 and were most recently serviced in August 2015, providing over one year of data. Data from the permanent stations are from January 2013 through August 2015.

Ambient seismic noise has recently been shown to be a useful tool for characterizing seismic velocity structure (e.g. Shapiro et al., 2005; Moschetti et al., 2007; Porritt et al., 2011). We follow the processing outlined in Bensen et al. (2007) to obtain inter-station empirical Green's Functions (EGFs) for all available station pairs of the 69 broadband stations in the region. This process begins with a time domain normalization via a running mean smoothing and frequency domain whitening to reduce the effect of large transient signals, such as local or teleseismic earthquakes, and then the EGFs are computed by cross-correlation. This processing step alters the absolute amplitude of the correlation, but maintains the phase information that is proportional to the actual Green's Function (Aki, 1957). The resulting EGFs show a clear surface wave signal (Fig. S1), which we measure with a Frequency Time Analysis (FTAN) (Dziewonski et al., 1969; Levshin et al., 1989) to obtain inter-station group and phase velocity dispersion curves. This method of measuring two-station dispersion curves has been repeatedly validated (e.g. Shapiro et al., 2005; Moschetti et al., 2007; Pawlak et al., 2012; Ward et al., 2013; Porritt et al., 2015) for arrays of similar design to what is available in the Banda Arc; alternative methods, such as Aki's method (Ekström et al., 2009) or Eikonal tomography (Lin et al., 2009), generally work better with higher station density on a regular grid, but have been shown to recover highly similar results (e.g. Ekström et al., 2009). The path-based measurements are quality controlled with a minimum signal-to-noise ratio of 4 and a minimum inter-station path length of 1.5 wavelengths. This quality control metric is less strict than some other studies, such as Bensen et al. (2007), where a minimum signal-to-noise ratio of 20 and minimum inter-station path of 3 wavelengths are used, but recent work by Luo et al. (2015) has shown that paths as short as 1 wavelength can be used and the low signal-to-noise threshold is used to account for the limited number of available paths. The path-based measurements are inverted for group and phase velocity maps via the method of Barmin et al. (2001). Examples of

the velocity maps are presented in Fig. 2 and discussed in the results section below. Maps at all measured periods are available in the supplementary information (Fig S2).

Group and phase velocity measurements at each spatial location for 8 to 35 seconds period are extracted from the maps and are inverted for the velocity of vertically polarized shear waves (V_{sv}) using Computer Programs for Seismology surf96 software (Herrmann, 2013). The inversion is parameterized as a 45-layer smooth model with 2 km thick layers in the upper 50 km, 5 km thick layers between 50 and 100 km depth, and 10 km thick layers below 100 km depth; this parameterization is chosen to provide relatively fine-scale layering where our depth resolution is best and broader-scale layering where our resolution is worse. The starting model is set as continental AK135 in the mantle, but with a constant velocity of ~ 4.5 km/s in the upper ~ 130 km of the model to avoid artificially inserting low velocity layers within the crust. The inversion is damped towards small, smooth changes in velocity and a total of 30 iterations are run at each latitude, longitude grid point. The upper 50 km of the inversion are used for analysis because, as shown in Fig. 3, the sensitivity depreciates below crustal depths.

Ambient noise tomography has emerged in recent years as a powerful imaging tool, with the above processing flow representing a relatively standard method of application. This tool has two main advantages as an imaging method. First, the illumination is primarily a function of array geometry, as every station acts as a virtual source for every other station. Therefore, ambient noise is able to simulate a source at each station location, whereas earthquake-based imaging relies on a heterogeneous distribution of sources and therefore is not always able to recover structure in desired regions. Secondly, the source of seismic noise is primarily thought to be oceanic microseisms, which excite surface wave energy at 10-20 seconds period for the primary microseism and 5-10 seconds period for the secondary microseism. Often, these periods are not strongly excited by local or teleseismic earthquakes and much of their energy

attenuates or scatters before reaching recording stations and therefore ambient noise tomography illuminates a structure often unavailable from other methods.

Ambient noise has been broadly adapted for imaging and monitoring since the method was shown as plausible by Lobkis and Weaver (2001). Their formulation, however, relies on assumptions which are only approximated in the real Earth. First, they assume an attenuation-free, homogeneous media between the two receivers. This is generally not the case in the real Earth, but is approximated by smoothly varying structures, such as those which are recovered by surface wave signals. Secondly, Lobkis and Weaver (2001) assumes perfectly homogeneously distributed surface sources. The actual distribution of background seismic energy is largely dependent on ocean weather systems and how the energy couples into the subsurface (e.g. Landès et al., 2010; Porritt et al., 2011; Traer et al., 2011). This leads to strong asymmetry in some EGFs depending on the relative orientation of the path to nearby coastlines and the deep ocean storms. Furthermore, recent work has shown that the sensitivity kernel of phase velocity to inter-station paths can be strongly dependent on asymmetry in the noise source (e.g. Fichtner, 2014, 2015). This effect is offset by shallow scattering in the subsurface acting as secondary sources. Additionally, processing steps, such as including more than 1 year of data to capture seasonal variation, running a time domain normalization and spectral whitening, and taking the symmetric version of the correlation, help to offset this asymmetry. Another challenge in the source distribution is the excitation of non-surface sources (e.g. Halliday and Curtis, 2008; Tsai, 2009; Froment et al., 2010; Kimman and Trampert, 2010). However, this can be utilized to extract higher mode information from noise correlations, leading to a more complete subsurface image (e.g. Halliday and Curtis, 2008).

3. Resolution

We present four proxies to assess the resolution of our model. The first two of these measurements are assessments of the lateral resolving power of the group and phase velocity

maps, which is fundamentally a measure of the path coverage (Figs 4, 5, S2). The first estimate we present is the resolution function described in Barmin et al. (2001). This estimate represents the minimum distance at which two delta functions could be separated and therefore smaller values indicate better resolution. We present this metric in Fig. 4d-f for the group velocity path coverage and Fig. S3d-f for phase velocity. These maps indicate lateral resolution of ~150 km in the Savu Sea and along the well-instrumented islands at the representative periods. However, resolution depreciates in the southwest and southeast where the crossing ray coverage decays. We also present checkerboard resolution tests (Fig. 5) for the phase (a-c) and group (d-f) velocity maps with checkers of size $2^\circ \times 2^\circ$. These tests, while less quantitative than the Barmin et al. (2001) estimate, confirm the ~150 km lateral resolution along the islands and in the Savu Sea. They also indicate reduced recovery of anomaly amplitudes for the group velocity (5d-f). This is likely due to the increased smoothing parameter used in the group velocity map inversion. The checkerboard tests also provide directionality to the resolution estimate because oversampled azimuths appear as local streaking of anomalies. Streaking artifacts appear most clearly in the southeastern and northeastern portions of our model. We expect the southeastern streaking reflects paths traveling along roughly the same azimuth between the island of Timor and Australia and the northeastern streaking reflects a lack of paths to stations in eastern Indonesia.

While lateral resolution is commonly assessed in surface wave studies, resolution in depth is less often investigated. The mean velocity model (Fig. 3a) does not contain a sharp velocity increase in depth, such as the crust-mantle interface (Moho). This is common in surface wave based inversions due to the smooth kernels for sensitivity to velocity with depth (Fig. 3b, c) and regularization in the inversion towards smooth, damped models. We present the RMS misfit between the measured dispersion curves and the calculated dispersion curves based on our final model in Fig. S4 as a map view. This shows the maximum misfit south of Timor, where we

lack ray coverage due to the placement of the stations and islands, and western Sumba near the edge of the model. Areas around the edge of the map with low misfit are likely a consequence of the damped model approaching an easily fit mean model. It is also worth considering how well the inversion recovers anomalous layers of high or low velocity. In Fig. S5 we present inversions of synthetic data with alternating high and low velocity layers in the crust. This test indicates that with our parameterization and dataset, the absolute amplitude of the anomalies is reduced and smoothed through multiple layers. However, the general shape is recovered for a high velocity mid-crustal layer and a low velocity mid-crustal layer, but not a low velocity lower crust. Tests with thinner layers were unable to recover the input alternating anomalies and therefore we approximate our vertical resolution to be ~ 10 km and note that layered structure will appear to be smoothly varying from the inversion.

4. Results

The group and phase velocity maps of Fig. 2 indicate some clear, first order structural trends. The island of Sumba is characterized by a northwest-southeast trending low velocity anomaly along its northeastern coastline at 10 seconds in both group and phase velocity maps. At 20 seconds, Sumba appears as a relatively high velocity feature and at 30 seconds it is a slightly low velocity anomaly. The Savu Sea appears as a significant low velocity anomaly at 10 seconds, but at 20 seconds it appears as low phase velocity, but high group velocity. At 30 seconds, the Savu Sea is largely a low velocity feature. The velocity structure beneath Timor is the highest amplitude velocity feature in these maps and appears relatively low at all presented periods. The crust and upper mantle of the islands of Flores, Alor, and Wetar, which comprise the inner arc, are represented in these maps as generally high velocity anomalies at all periods, but with a low velocity anomaly at $\sim 123^\circ\text{E}$ in the 10 second maps.

Fig. 6 shows maps of relative shear velocity at depths of 5-45 km. The velocity structure beneath Sumba appears as a relatively high velocity anomaly at the shallowest levels (5, 15

km), but in the mid- and lower-crust and uppermost mantle it is relatively low velocity. In the Savu Sea, the shallow structure is relatively low velocity, but with a high velocity body striking northeast-southwest in the upper 15 km between Alor and Sumba. At greater depth, the velocity structure is relatively low under the Savu Sea, with the highest amplitude low velocity anomaly at 25 km and slightly low velocities in the uppermost mantle. Beneath Timor, the significant low velocity anomalies seen in the group and phase velocity maps have resulted in large amplitude low velocity anomalies through the crust and uppermost mantle. However, at depths greater than ~35 km there is a high velocity anomaly striking northeast-southwest along the Timor Trough and it appears to dip towards the northwest. The inner arc islands show largely high velocity anomalies in the upper 25 km, but by 35 km depth the velocities are relatively low.

The cross-sectional views of the model in Fig. 7 further illustrate the structures of the region. Section A-A' depicts the high velocity anomaly in the upper to middle crust under Sumba and Flores. This also clearly shows the low velocity anomaly under Sumba and a relatively high velocity anomaly under Flores at 25-30 km depth. B-B' crosses the island of Savu, the Savu Sea, and Flores island. This section shows a slightly low velocity feature in the upper crust under Savu and a significantly low velocity lower crust or uppermost mantle. These low velocity anomalies interlayer with high velocity anomalies under the Savu Sea and Flores. The general structure changes from west to east in sections C-C' and D-D', which we interpret as the collision across the Timor Trough. The shallow structure under Timor in C-C' is a low velocity feature with a basal dip towards the south at ~10 km depth. Between 12 km depth and 30 km depth, the structure under Timor is a low velocity anomaly inclined towards the north and this feature overlies a high velocity anomaly at depths greater than 35 km. In the northern part of the section, under Pantar and Alor, there is a shallow low velocity veneer, but it is largely a high velocity crustal feature split into two distinct patches. The shallower patch is flat lying between 5 and 15 km and the deeper part dips towards the south. Further east, in section D-D', the low

velocities under Timor are again evident in the upper crust and this layer is separated from a deeper low velocity anomaly which lies on top of a high velocity uppermost mantle feature. Under Wetar, three high velocity anomalies are present, with one at ~10 km depth wedging under Timor and separating the two low velocity anomalies.

Much of our study region is below sea level and therefore a zero shear velocity water layer may have an effect on the inversion from group and phase velocity dispersion curves to shear velocity if the dispersion curves respond to the upper fluid layers rather than only the solid earth. Figs. S6 and S7 in the supplementary information show depth maps and cross-sections, similar to Figs. 6 and 7, through an inversion in which we've constrained the uppermost layer of nodes below sea level to a shear velocity of near 0 km/s. The effect is most clearly seen in the shallow ~5 km where this constraint is imposed and results in a signal of relatively low velocity features immediately under the sub-sea level portions of the model. However, this has little effect in the model below ~5-10 km and we therefore focus our discussion on the inversion that does not add the water layer as a prior constraint.

5. Discussion

In the following discussion we follow three guidelines for the interpretation of anomalies at various depth ranges. (1) Anomalies within the crust are considered to form largely from compositional differences such as between sedimentary packages and emplaced igneous rocks. (2) Anomalies in the mantle derive primarily from temperature contrasts within the lithosphere, although a significant portion of the anomalies may be due to compositional contrasts when comparing sub-continental mantle lithosphere to sub-oceanic mantle lithosphere. (3) Anomalies near the Moho (crust-mantle interface) may represent a comparison between lower crustal rocks and upper mantle rocks that likely have both compositional and temperature contrasts, and we therefore limit our interpretation of structures near this interface.

5.1 Sumba and the Savu Sea

We infer from our velocity model that Sumba is a forearc sliver that is actively being uplifted. This velocity structure is illustrated in Fig. 8 by a set of cross-sections. Cross-section E-E' shows that the velocity structure under the island of Sumba is characterized by high shear velocity crust in the upper 15-20 km. This velocity anomaly is similar to the high velocities of the inner arc and consistent with stratigraphic columns which show the basement of Sumba is composed of Cretaceous to Paleocene plutons, volcanoclastic deposits, and metamorphic units (Audley-Charles, 1985; Fortuin et al., 1994; Rutherford et al., 2001). High velocity anomalies are also seen under the center of the Savu Sea in section F-F' as both a very shallow layer (upper 5 km) and a mid-crustal layer at ~15 km depth. Sections G-G' and H-H' are oblique to the long axis of Sumba and both depict a continuous lower crustal high velocity anomaly from Flores and the eastern part of the Savu Sea to Sumba. This suggests that these structural units are continuous as would be expected for either the model of an uplifted forearc sliver or translation of the Sumba Block.

The model of Sumba as an uplifted forearc sliver reconciles some otherwise enigmatic features such as rapid uplift, GPS velocities indicating Sumba is not moving with Australia, relatively high shear velocity in the upper crust, and reconstructions inferring subduction of the Scott Plateau. In this model, Sumba was an ordinary piece of the forearc before collision, but it began to be uplifted as the relatively high buoyancy of the Scott Plateau entered the trench and pushed the forearc up from below. This is most clear in section H-H' (Fig. 8e) as the high velocity anomaly under most of the Savu Sea is at ~15-20 km depth, yet the high velocity feature is shallower under Sumba. The alternative model appeals to escape tectonics, suggesting that the Sumba Block, composed of Banda Terrane material, was extruded westward upon collision between the Australia and Banda plates (Rutherford et al., 2001). The continuity of the high velocity mid-crust through the Savu Sea in section H-H' may reflect the extruded track of Banda Terrane, but this model has significant shortcomings. First, this model

predicts a lithospheric scale fault to accommodate the motion of the Sumba Block and such a fault is not observed in our data. Second, the high velocity characteristic of this anomaly suggests the track is compositionally high rigidity (e.g. mafic or metamorphic units of the Banda Terrane) and this would make it relatively high viscosity and therefore resistant to flow in response to compression.

5.2 Timor

The collision across the Timor Trough is highly oblique with the most mature part of the collision (i.e. the most shortened) occurring at the northeastern edge of Timor and the youngest part of the collision at the southwestern end of the Island (e.g. Harris, 2011 and references therein). We present cross-sections along strike of this collision zone (Fig. 9) to show deeper structural levels of the inferred 'time-space trade-off'. The most recent version prior to this study is presented in Harris (2011) as his Figure 7.15 and allows direct comparison between velocity anomalies and mapped structural units. In the northeast of Timor, profile I-I', we observe a low velocity anomaly in the upper 5 km, which appears inclined upwards toward Wetar in the northwest. A similar dipping layer is also seen in Harris (2011) profile (a) as an imbricate stack of Australian Passive Margin Sequence abutted against the Banda Terrane and an antiform of duplexed Gondwana Sequence beneath the Banda Terrane. Profile I-I' has a high velocity anomaly under Wetar dipping towards the southeast, identified as the Banda Terrane by Harris (2011). Additionally, we observe a high velocity lower crust under Wetar and a low velocity lower crust under Australia. To the southwest, profile J-J', depicts a thicker low velocity anomaly under Timor and further separation of the upper and lower crust by a high velocity wedge. The shallow low velocity anomaly under Timor in J-J' has a similar inclination as in I-I', but the contact with the Banda Terrane of Alor is abrupt. The lower crust and mantle structure of J-J' is similar to I-I', in that we observe a northward dipping low velocity lower crust under Timor, which is underlain by a high velocity uppermost mantle. The last section crossing southwestern Timor,

K-K', depicts separation between the upper and lower crust at the southeast end of the profile, but it is far less pronounced than I-I' and J-J' and cannot be fully distinguished due to resolution limitations. However, there is a clear change in the structure as the high shear velocity Banda Terrane is not observed as a wedge which separates the crust of Timor, but rather is inclined upwards to the southeast. This structural inclination is consistent with Banda Terrane having been thrust over the Gondawana Sequence and Australian Passive Margin Sequence, but has been since largely eroded and preserved locally as klippen. The structure across Rote, L-L', is significantly different than the profiles in Timor, reflecting a less mature stage of collision. Here, the shallow crust is largely observed as a relatively low velocity, as would be expected for the mapped sedimentary sequences, but under the Savu Sea, the upper crust is also characterized as low velocity and the mid to lower crust is a significant high velocity anomaly. This suggests that in the early stages of collision, the forearc sequence of the Banda Terrane was resting at these mid-crustal depths and then it was forced to the surface during the collisional process. Under Savu, M-M', the structure is a mildly low velocity anomaly in the upper crust and a slightly high velocity anomaly across the Savu Sea. The velocity anomalies here are interlayered, but there is no clear evidence of collision. However, this region is anomalous in the lower crust and uppermost mantle under Savu, as it is a near zero low velocity anomaly rather than a high velocity anomaly as seen in the other profiles. From this we infer the relatively low velocity represents thickened oceanic crust on the downgoing plate, such as would be expected for subduction of the Scott Plateau. While this may alternatively be thickened continental crust of the overriding plate, there is no correlative evidence of thickened crust in the region aside from the Scott Plateau.

The high correlation between the surface mapped faults and terranes with our velocity model provides independent confirmation on each other, and our imaging results allow for downward extension of the collision zone identified from surface geology. The structure under

Timor and Rote at these depths clearly indicates a subduction sequence of relatively low velocity lower crust on top of higher velocity subducting mantle lithosphere. The higher density lithosphere is forcing the Australian margin crust into the collision where the lower density sedimentary sequences of the upper crust are being delaminated from the lower crust by thrusting against and beneath the higher density portions of the Banda Terrane. This decoupling of the crust allows the relatively denser lower crust to subduct with the lithospheric mantle. The decoupling varies along strike, with the northeastern-most section, I-I', showing the separation under the Timor-Wetar Strait and the central Timor section, J-J', showing the separation under Timor. The central and western Timor slices also show the high velocity backstop further to the southeast than I-I', suggesting that this feature is pushed north and upwards as collision evolves.

5.3 Inner Arc

The inner arc islands of Flores, Wetar, and Alor in figures 6-9 are largely characterized by a high velocity upper crust, low velocity middle crust, high velocity lower crust, and low velocity uppermost mantle. The high velocity upper crust likely reflects volcanoclastic sediments and shallow plutons from recent igneous processes. The low velocity middle crust could indicate a set of mid-crustal magma chambers with some percentage of partial melt and the high velocity lower crust could be an eclogitic zone undergoing alteration by rising heat and fluid from the slab (e.g. Austrheim, 1987). The low velocity uppermost mantle is inferred to be part of the mantle wedge where hydrous mantle rocks are upwelling to feed the arc system. While this general structure is largely consistent along strike, there are some anomalous regions. For instance, as noted above, there is a shallow low velocity anomaly near 123°E in the upper crust. This longitude corresponds to a similar change in shear velocity imaged at 200 km depth by Fichtner et al. (2010) and a change in the helium ratios (Hilton and Craig, 1989; Hilton et al., 1992).

Along the northern coasts of the inner arc islands, we observe clear high velocity dipping bodies in the lower crust to uppermost mantle (Fig. 7). These provide evidence of full crustal-scale thrust faulting with the inner arc islands acting as a backstop and possible early stages of a subduction reversal. The Flores Thrust and Wetar Thrust have long been recognized as important in accommodating collision in the inner arc (e.g. Silver et al., 1986), but the relatively planar geometries of the two thrust faults are newly observed in our model. Under Flores, mid to lower crustal high velocity anomaly is relatively horizontal over ~150-200 km (Fig 7c) whereas under Pantar, Alor, and Wetar it is inclined ~10 km over ~50 km lateral distance (Fig 7d). The ~10° dip of this feature, if confirmed with other methods with higher sensitivity to dipping structures, may indicate an incipient subduction polarity reversal where the Banda Sea microplate will subduct beneath the arc. Polarity reversals are recognized as an important component of the plate tectonic cycle (e.g. Yuan and Dueker, 2005; Porritt et al., 2014) but are generally only recognized in preserved lithospheric scale structures. Therefore, our model may provide rare evidence of an incipient polarity reversal at the meso-scale.

5.4 Uniqueness of the Bifurcated Arc

The Banda Arc is a complex orogenic system with interactions between the inner and outer arc systems. Fig. 10 depicts the major units in their relative spatial positions. Fig. 10a shows the upper Australian lithospheric mantle under southeastern Timor as a relatively high velocity anomaly, while Fig. 10b and 10c depict the upper and lower Timor crust respectively as low velocity anomalies. The broadly defined Banda Terrane is isolated in Fig. 10d and is observed beneath Sumba, the Savu Sea, Flores, Wetar, and Alor at ~10-20 km depths. It is notably absent near 123°E along the inner arc as described above. Infilling between the Banda Terrane and Timor, we see slightly low velocity anomalies in the middle of the collision (Fig 10e). These are likely Gondwana sequence sedimentary rocks, which may have been reworked during collision and tectonically buried. Removing the mid-crustal low velocity anomalies in the

3D model and viewing at an oblique angle (Fig. 10f) highlights the high velocity Australian lithosphere driving the low velocity crustal sequences into the higher velocities we interpret as the Banda Terrane of the overriding plate. In response to the collision across a northeast trending trench and driven by northward motion, we see Banda Terrane high shear velocity material elongated to the southwest.

6. Conclusions

We apply ambient noise tomography in the Banda Sea region using 69 temporary and permanent network stations. Analysis of the resulting 3D shear velocity model provides new insights into the collisional tectonics of the Banda Arc. We've reconciled competing hypotheses for the origin of Sumba showing its crustal component is most consistent with an uplifting forearc sliver due to collision with the Scott Plateau. Across Timor, the model confirms the tectonic-stratigraphic relationships previously inferred from surface mapping, but our model provides new, deeper images of how the subducting Australian lithosphere is colliding into the mid-crustal Banda Terrane and separating along an upper crust-lower crust interface. This has resulted in deposition of shelf-sedimentary units on top of Banda Terrane and subduction of the Australian lower crust with the lithosphere. Variations in the inner arc are subtler, but a low velocity anomaly is noted at the inferred transition between subduction of oceanic lithosphere and continental lithosphere. These observations suggest a highly complex structure in the Banda Arc resulting from collisional tectonics.

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648 **Figures**

649 **Figure 1:** Location map depicting seismic recording stations in the Banda Arc. Yellow inverted triangles indicate
650 stations, red triangles with bubbles indicate quaternary active volcanoes of the Banda Arc. Solid lines indicate plate
651 boundaries from the USGS; teeth indicate the overriding plate. GPS velocity vectors referenced to a stable Sunda
652 block and their uncertainties are shown in black arrows (Bock et al., 2003; Calais et al., 2006; Simons et al., 2007;
653 Nugroho et al., 2009) and blue arrows (Koulali et al., 2016) and the no net rotation plate motion is shown with white
654 arrowheads (DeMets et al., 2010). Inset map indicates all stations used in the study and a larger scale view of the
655 plate motion velocity vectors.

656 **Figure 2:** Rayleigh wave speed maps. (a-c) indicate phase velocity perturbation to a local mean. (d-f) indicate group
657 velocity relative to a local mean. Measurement period indicated in the southeast of each map and mean velocity
658 indicated in the northwest. Yellow inverted triangles indicate stations and grey triangles indicate quaternary active
659 volcanoes.

660 **Figure 3:** Mean Vsv and kernels of sensitivity to shear velocity as a function of depth at representative periods. (a)
661 1D local mean Vsv. (b) the phase velocity sensitivity and (c) is the group velocity sensitivity. Kernels are computed
662 with respect to a 1D local mean shear velocity model for the Banda Arc in the upper 50km (a) and AK135 below. X-
663 axis is scaled for presentation.

664 **Figure 4:** Group velocity resolution proxies. (a-c) path coverage in the region of interest. Each yellow line indicates a
665 path used in the inversion (minimum distance of 1.5 wavelengths and minimum SNR of 4). (d-f) resolution estimate of
666 Barmin et al. (2001) describing the minimum distance at which two delta functions could be separated. Lower values
667 indicate better resolution. Inverted triangles indicate stations and grey triangles indicate volcanoes. A similar figure is
668 available in the supplementary information for phase velocity.

669 **Figure 5:** Checkerboard resolution tests. Checkers are 2° by 2° and represent a perturbation of ±5% relative to a
670 mean of 5km/s. (a-c) indicates the phase velocity coverage. (d-f) indicates the group velocity coverage. Stations are
671 plotted as white inverted triangles and arc volcanoes are grey triangles.

672 **Figure 6:** Slices through the Vsv model at constant depth. Velocity is presented as perturbation to a local 1D mean
673 shear velocity (V0, northwest corner). Depths indicated in the southeast corner of each map. Inverted white triangles
674 indicate stations and grey triangles indicate volcanoes.

Figure 7: South to north cross-sections progressing west to east. (a) Indicates location of profiles in b-e. Each profile has the topography plotted in brown and sea level indicated by the gray background. Islands labeled in the topography are approximate. Horizontal axis in profiles shows distance along profile in arc degrees (1 degree ~ 111.2 km). Vertical exaggeration of cross-section is ~1.25 and vertical exaggeration of topography is ~10.

Figure 8: Cross-sections showing the structure of Sumba and the Savu Sea. (a) Map view indicating the location of the profiles. (b,c) Southwest to northeast cross-sections down the long axis of Sumba and offset, respectively. (d,e) Southwest to northeast cross-sections across the short axis of Sumba and down the Savu Sea.

Figure 9: Cross-sections of time-progressive structure along orogenic strike. Locations of the profiles indicated in (a). Earliest collision is shown in I-I' (b) and gets progressively older from (c-f). Velocity scale is presented as perturbation to a 1D local mean velocity. In the top portion of (b-e), topography/bathymetry is indicated with brown and grey shading in the background indicates sea level. Shaded and patterned fields based on Figure 7.15 of Harris, 2011. Horizontal axis in profiles shows distance along profile in arc degrees (1 degree ~ 111.2 km).

Figure 10: Isosurfaces of velocity anomalies discussed in the text. Each panel adds on top of the previous panels. (a) Subducting Australian lithosphere. (b) Sedimentary sequences on Timor of Australian affinity. (c) Subducting Australian crust. (d) Banda Terrane. (e) Gondwana and other Australian affinity lower crustal sediments. (f) The geometrical relationship of the subducting features and Banda Terrane as viewed from the southeast. Collisional features are labeled following our interpretation.

Figure 1

Figure 2

Figure 7

Figure 8

Figure 9