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The southwestern edge of the Ryukyu subduction zone: A high *Q* mantle wedge

Yen-Ting Ko [a,b], Ban-Yuan Kuo [a,c,*], Kuo-Lung Wang [a], Shu-Chuan Lin [a], Shu-Huei Hung [b]

[a] Inst. of Earth Sciences, Academia Sinica, 128 Academia Rd., Sec. 2, Taipei 115, Taiwan
[b] Dept. of Geosciences, National Taiwan Univ., 1 Roosevelt Rd., Sec. 4, Taipei 106, Taiwan
[c] Dept. of Earth Sciences, National Central Univ., Jhongli, Taiwan

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ABSTRACT

The lateral edge of a subduction zone is usually depicted as an opening to the asthenosphere where invigorated dynamics and amplified magmatism take place. In this study we present evidence from seismic data suggesting the presence of a cold and dynamically sluggish edge environment at the southwest end of the Ryukyu subduction system. We measured attenuation, or $1/Q$, for P waves from subduction zone events at ~100 km depths received by OBSs in the Okinawa trough and land stations in NE Taiwan. In the Okinawa trough 100–200 km from the edge, Q values are lower than 100. In the vicinity of the edge, Q values increase from 100 to over 1000 towards Taiwan. To reconcile arguments from geophysical and geochemical observations, we propose that the mantle wedge near the edge has high Q values due to low temperatures and probably low water content. These may result from coupling of the slab laterally with the thick Eurasian lithosphere, which inhibits back-arc rifting, retards subduction, and reduces the water supply to the mantle wedge. The SW Ryukyu subduction system represents a subduction-zone edge type distinct from more commonly documented free or warm edges.

1. Introduction

Dynamically active slab edges are typified by the Kamchatka subduction zone where the pattern of anisotropy and adakite rock samples suggest a lateral flow around the NE edge of the Pacific slab and possible slab melting as a result of vigorous edge dynamics (Peyton et al., 2001; Yogodzinski et al., 2001). This free-edge type of slab with 3D flow and amplified magmatism has been observed in different subduction systems (Zandt and Humphreys, 2008; Civello and Margheriti, 2004; Baccheschi et al., 2007; Gvirtzman and Nur, 1999). Numerical models predict that additional flow is induced from outside into the overall 2D system through the open edge with or without a slab rollback (Kincaid and Griffiths, 2003; Jadamec and Billen, 2010; Kneller and van Keken, 2008). In this study we present a scenario different from the more commonly envisioned free and dynamically invigorated edge type. At the southwest Ryukyu subduction zone, the subduction is shallow and oblique such that the slab could be locked onto the adjacent Eurasian lithosphere (EL) laterally without significant rollback at this end of the subduction system. This and the associated effects of diminishing back-arc rifting and blocking of the edge of the mantle wedge with the EL all act to suppress vigorous 3D flow and lower the temperature. Here we present seismological evidence for a cold and dynamically sluggish edge environment.

2. SW Ryukyu subduction system

The Ryukyu subduction system terminates against the Eurasian continental margin in the vicinity of Taiwan, where several tectonic forces interact. The oblique convergence of the Philippine Sea plate with the Eurasian plate results in an arc–continent collision propagating southward (Suppe, 1984). Meanwhile, the oblique subduction of the Philippine Sea plate at the Ryukyu trench leads to a collision between the western edge of the slab and the EL at 50–90 km depths (Chou et al., 2006; Kao et al., 1998). Recent GPS models suggest that the Ryukyu arc is drifting away from the Eurasian margin at a rate of 50 mm/yr (Nakamura, 2004). In response, the Okinawa trough (OT) is opening as a back-arc basin. Because seafloor spreading has not begun (Iwasaki et al., 1990; Hirata et al., 1991), the submerged area north of the Ryukyu arc is still part of the Eurasian plate (Fig. 1).

The southwestern section of the arc–back-arc system also hosts several atypical features. The OT rapidly tapers towards

* Corresponding author at: Inst. of Earth Sciences, Academia Sinica, 128 Academia Rd., Sec. 2, Taipei 115, Taiwan. Tel.: +886 2 2783 9910x507; fax: +886 2 2783 9871.
E-mail address: byk@earth.sinica.edu.tw (B.-Y. Kuo).

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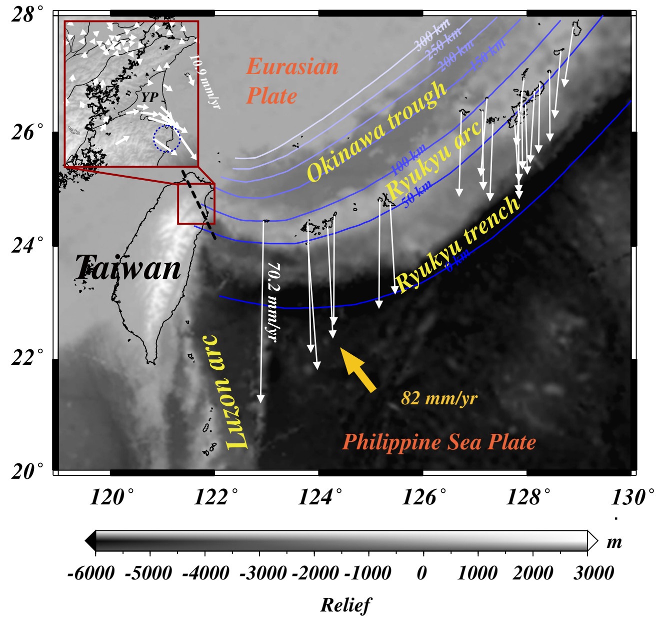

Fig. 1. Tectonic setting of the Taiwan–Ryukyu region. The Eurasian Plate occupies the west and north of the Philippine Sea plate. The oblique subduction results in a collision between the slab and the Eurasian lithosphere laterally, but the Philippine Sea plate is sliding down the Eurasian plate along the Ryukyu trench. The western boundary of the subducting slab defined by seismicity is marked as a dashed straight line. The inset shows the Yilan plain (YP), a triangular shape extensional structure, where the Okinawa trough (OT) terminates. GPS velocities at Ryukyu arc are highlighted to indicate retreat of the arc and equivalently the trench (Nakamura, 2004). The GPS velocities in NE Taiwan (inset) indicate a clockwise rotation and a progressively increasing southward movement of the southern Yilan plain (Rau et al., 2008). Both GPS data sets are referenced to Shanghai, China. Blue dotted circle (inset) marks the location of the serpentinized fore-arc mantle revealed by seismic tomography (Chou et al., 2009). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

west and ends at the Yilan plain, an extensional structure in NE Taiwan (e.g., Hou et al., 2009). The two westernmost islands of the Ryukyu arc, Yonaguni and Ishigaki, are composed of either sedimentary rocks or volcanic rocks dated back to 10–20 Ma, and are not present-day active volcanoes (Shinjo, 1999). Instead, the volcanic front of the Ryukyu subduction system can be delineated by connecting Kueishantao, an arc volcano islet 10 km offshore NE Taiwan, and several submerged volcanoes on the southern flank of the OT roughly above the 100 km depth contour of the subducting Philippine Sea plate (Chung et al., 2000). The axis of the arc roughly merges with the axis of the OT at Kueishantao.

The overriding plate of the subduction system, i.e., the Eurasian plate north of the Ryukyu trench may have been thinned thermally by an upwelling associated with rifting as numerical models generally suggest (e.g., Rey and Muller, 2010; Lin et al., 2010a). The arc-affiliation of the Kueishantao volcano suggests that it is underlain by a relatively thin EL above the mantle wedge. In contrast to the Kueishantao, the Tsaolingshan lava is located in NW Taiwan west of the western boundary of the underlying subducting slab (Fig. 2). The Tsaolingshan magma originates from melting of subcontinental lithospheric mantle metasomatized by subduction processes. Together with its ~2% melting (Wang et al., 2004), this implies a thick EL above the source of the magmatism. Geochemical characteristics of Tsaolingshan magma suggest its melting took place under garnet stability field, which indicates a thickness >70 km for the EL at this locality (Wang et al., 2004). Thus along the OT, we may envision a scenario in which the continental lithosphere thickens towards the continental interior in western Taiwan, although the details of thickening, whether abrupt or gradual, is unknown. In this study we examine lateral variation along the OT, with emphasis on the edge of the mantle wedge, by comparing seismic attenuation measurements in 3 groups of cross-arc profiles.

3. Data and measurements

Two broadband ocean-bottom seismometers (OBSs) were deployed in the northern flank of the OT (Lin et al., 2010b) to provide a reference model for the mantle wedge 100–200 km away from the plate edge. *F*-net stations (Okada et al., 2004) on Yonaguni and Ishigaki were used to pair with the OBSs. Several BATS (Broadband Array in Taiwan for Seismology; http://bats.earth.sinica.edu.tw) stations in northern Taiwan facilitate sampling of the western edge of the mantle wedge (Fig. 2). Events used in profile A–A′ and the other 2 profiles were located by Taiwan Central Weather Bureau network and Japan Meteorological Agency network, respectively. Events were selected in the subducted Philippine Sea slab with focal depths between 70 and 130 km (with one exception in profile C–C′; see below) and an epicentral distance to depth ratio < tangent 30°. Events at depths too deep and located far off along the strike of the slab from stations likely have a long refraction path within the slab before entering the mantle wedge. The data set is composed of paths preferably sampling the mantle wedge with little contamination of the slab (Fig. 2).

We follow the procedure of Ko et al. (2012) to obtain the spectra of *P* waves on vertical recordings. In this study, spectrum is measured in the frequency (f) band where signal to noise ratio > 5 with the maximum frequency at 8 Hz. For simplicity, Q is assumed to be frequency dependent as $Q=Q_0 f^{-\alpha}$ with α set at the commonly used experimental value of 0.27 (Jackson et al., 2002). Fixing α is part of our strategy of retaining as few degrees of freedom as possible in the inversion. Thereafter Q refers to Q_0 for *P* waves. In this study, raypaths sample similar portions of the mantle wedge relative to the wedge geometry (Fig. 2b), allowing a characterization of the lateral variation through comparing the path-averaged Q at different profiles across the OT.

We measure Q values for the OBS profiles using a two-station method (Roth et al., 1999) with the reference stations YNG and IGK at Yonaguni and Ishigaki islands, respectively. The *P* waves to these two *F*-net stations travel primarily within the slab and partly across the fore-arc mantle where attenuation is low (Stachnik et al., 2004; Wiens and Smith, 2003). We assume $Q_{ref}\rightarrow\infty$ for the two reference stations to recover the Q values for the mantle wedge. Each OBS profile consists of 3 two-station measurements and the error of the average Q is smaller than 10. One event at 50 km depth was used because its spectra at both stations are consistent with the respective spectra of other 2 events.

In northern Taiwan, the two-station method was not used because no stations with both suitable "slab" affiliation and acceptable waveforms qualified as reference stations. Because there are abundant event-station pairs for BATS, we applied the cluster-event method developed by Ko et al. (2012). The method reduces the number of degrees of freedom in the inverse problem to the point at which the tradeoff between Q and f_c is suppressed (Ko et al., 2012). A total of 34 events and 26 clusters were analyzed, with a cluster radius of 15 km and a minimum of 3 events in each cluster. The error in Q can be translated from the error of corner frequency (Ko et al., 2012). In this study the average error of corner frequency is 0.425 Hz (Supplementary Material), which corresponds to an error ranging from 20 for $Q<200$ to 100–150 for $Q=1000$. Examples of waveforms and measurements are shown in Fig. 3.