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Future accreted terranes: a compilation of island arcs, oceanic plateaus, submarine ridges, seamounts, and continental fragments

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Abstract

Allochthonous accreted terranes are exotic geologic units that originated from anomalous crustal regions on a subducting oceanic plate and were transferred to the overriding plate during subduction by accretionary processes. The geographical regions that eventually become accreted allochthonous terranes include island arcs, oceanic plateaus, submarine ridges, seamounts, continental fragments, and microcontinents. These future allochthonous terranes (FATs) contribute to continental crustal growth, subduction dynamics, and crustal recycling in the mantle. We present a review of modern FATs and their accreted counterparts based on available geological, seismic, and gravity studies and discuss their crustal structure, geological origin, and bulk crustal density. Island arcs have an average crustal thickness of 26 km, average bulk crustal density of 2.79 g cm^{-3}, and have 3 distinct crustal units overlying a crust-mantle transition zone. Oceanic plateaus and submarine ridges have an average crustal thickness of 21 km and average bulk crustal density of 2.84 g cm^{-3}. Continental fragments presently on the ocean floor have an average crustal thickness of 25 km and bulk crustal density of 2.81 g cm^{-3}. Accreted allochthonous terranes can be compared to these crustal compilations to better understand which units of crust are accreted or subducted. In general, most accreted terranes are thin crustal units sheared off of FATs and added onto the accretionary prism, with thicknesses on the order of hundreds of meters to a few kilometers. In addition many island arcs, oceanic plateaus, and submarine ridges were sheared off in the subduction interface and underplated onto the overlying continent. And other times we find evidence of collision leaving behind accreted terranes 25 to 40 km thick. We posit that rheologically weak crustal layers or shear zones that were formed when the FATs were produced can be activated as detachments during subduction, allowing parts of the FAT crust to accrete and others to accrete. In many modern FATs on the ocean floor, a sub-crustal layer of high seismic velocities, interpreted as ultramafic material, could serve as a detachment or delaminate during subduction.

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1 Introduction

Terrane accretion is considered to be one of the main contributors to the growth of continental crust (Stern and Scholl, 2010; Clift et al., 2009a; Cawood et al., 2009). Although continental crust is lost by erosion and/or recycled into the mantle at subduction zones, crust is also added to continents at subduction zones by accretion and magmatic events. Accreted terranes can be made of accreted crustal units of volcanic arcs, oceanic plateaus, continental fragments, seamounts, accretionary prisms, melanges, ophiolites, and flysch. The accretion of volcanic arcs, oceanic plateaus, and seamounts to continents adds mafic juvenile crust that eventually will mature into felsic compositional continental crust by progressive magmatism and lower crustal foundering (Stern and Scholl, 2010).

The concept of accreted terranes was first born in the 1970's and has evolved greatly since then (Monger et al., 1972; Irwin, 1972; Coney et al., 1980; Snoke and Barnes, 2006). Irwin (1972) was the first to introduce terranes into geologic lexicon as "an association of geologic features, such as stratigraphic formations, intrusive rocks, mineral deposits, and tectonic history, some or all of which lend a distinguishing character to a particular tract of rocks and which differ from those of an adjacent terrane." It was in the sutured rock belts of different affinities (oceanic crust and island arc) in the Klamath Mountains that Irwin (1972) first coined the term after recognizing that these tectonically juxtaposed rocks must have been scraped off in a subduction zone. In following years the attributes of "suspect" or "accreted" were added to specify terranes of allochthonous affinity which were juxtaposed tectonically to autochthonous deposits on continents, such as by accretionary processes at a subduction zone (Coney, 1978; Jones et al., 1982). The quest to identify and map accreted terranes led to the patchwork tapestry of terrane belts of Western North America (Coney et al., 1980) and the idea that continents grew from accretionary processes at subduction zones.

In addition to identifying suspect terranes on the continents, researchers sought to map out regions of the oceanic floor that could possibly become future accreted ter-

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ranes. The advancement of oceanic seismology in the 1980s led to catalogueing of anomalous crustal regions on oceanic plates that could eventually become accreted terranes (Carlson et al., 1980; Ben-Avraham et al., 1981; Nur and Ben-Avraham, 1982). These anomalous crustal regions were initially called oceanic plateaus; a term which encompassed every region of anomalously thick crust on the ocean plate. In this context, oceanic plateaus included large igneous provinces (LIPs), island arcs, hot spots, extinct mid-ocean ridges, seamounts, and submarine plateaus with continental crust (Ben-Avraham et al., 1981). Later compilations of anomalous crustal structures on the oceanic floor separated oceanic plateaus, thermal swells, and continental submarine plateaus (Schubert and Sandwell, 1989; Marks and Sandwell, 1991). Cloos (1993) designated basaltic oceanic plateaus, active spreading ridges, continental and island arc crust, continental passive margins, and seamounts as "future colliders". These compilations have focused on constraining the crustal thicknesses and volumes of oceanic plateaus, thermal swells, and continental submarine plateaus (Ben-Avraham et al., 1981; Sandwell and MacKenzie, 1989; Schubert and Sandwell, 1989; Marks and Sandwell, 1991). In the past decade, numerous and advanced marine geophysical and geochemical studies have been undertaken to characterize the crustal composition of oceanic LIPs, submarine ridges, island arcs, continental submarine plateaus, and seamounts.

Naturally the following question was posed: can we quantify the likelihood of accretion or subduction of these crustal features? Researchers used analytical studies of the buoyancy forces of oceanic plateaus, continental fragments, and island arcs preventing or allowing them to subduct or collide in a subduction zone (Molnar and Gray, 1979; Cloos, 1993; Moore and Wiltscko, 2004). Molnar and Gray (1979) and Moore and Wiltscko (2004) suggest the contrast between the external force of slab pull and the internal force produced by buoyant terrane crust will control the amount of terrane crust subducted vs. accreted. Molnar and Gray (1979) estimate that only 10 km of continental crust is subductable. Based on isostatic analyses of the subductability of oceanic plateaus, island arcs, and continental crust, Cloos (1993) calculated that collision would

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occur for oceanic plateaus with a crust > 17 km thick, continental crust > 20 km thick, and young, hot island arcs. Seno (2008) calculated the forces in the subduction zone necessary for a crustal block of continental affinity to shear off of a subducting plate, and concluded that accretion can only occur in a relatively dry (low pore pressure) subduction interface. More recently, analog and numerical geodynamic experiments have examined the subductability of oceanic LIPs, submarine ridges, island arcs, continental submarine plateaus, and microcontinents and the effects on the subduction zone dynamics after subduction (Ellis et al., 1999; van Hunen et al., 2002; Boutelier et al., 2003; Martinod et al., 2005; Espurt et al., 2008; De Franco et al., 2008; Mason et al., 2010; Afonso and Zlotnik, 2011; Tetreault and Buiter, 2012). Of course, observations of thick oceanic LIPs subducting (such as the Ontong Java and Hikurangi plateaus: Mann and Taira, 2004; Scherwath et al., 2010; Bassett et al., 2010) and the relative absence of entire island arc crusts in the geologic record (Condie and Kröner, 2013) indicate that the accretion, subduction, and collision of thick crustal regions might not always follow the analytical and geodynamic estimates. The addition of crustal material to continents at accretionary zones usually occurs by adding slivers of thrusted crustal units to the accretionary prism region (Coney et al., 1980; Snyder, 2002; Cawood et al., 2009), more often than collision and addition of the entire crustal thickness to the continent.

In the vein of earlier studies (Ben-Avraham et al., 1981; Sandwell and MacKenzie, 1989; Schubert and Sandwell, 1989; Marks and Sandwell, 1991), we catalog the regions of anomalous crust on the ocean floor and compare them to accreted terranes using new geophysical and geological studies from the last couple of decades. We group island arcs, oceanic LIPs, submarine ridges, seamounts, hot spots, submarine continental fragments, and microcontinents all as future allochthonous terranes (FATs). Although accreted terranes can also be units from accretionary prisms and melanges, these pre-accretion units are actually part of the subduction zone and are autochthonous to the convergent margin, and therefore are not covered in this study. In this paper we review the crustal compositions of modern and accreted examples of FATs and discuss the processes that lead to accretion, subduction, or collision for each of these anomalous crustal features on the ocean floor. Geophysical, geological, and geochemical studies provide us with new insight on the crustal layers and constraints on densities of FATs, and we will show in our summary that there are no significant differences between seismic velocity profiles from continental crust and mafic oceanic plateau crust. This compilation will summarize average crustal thicknesses, bulk crustal densities, and crustal structures of FATs. A better understanding of modern analogues of accreted allochthonous terranes will improve our understanding of the volume of crust accreted and subducted and the processes and kinematics affecting accretion and subduction, and collision. We hope therefore that this compilation will constrain future modelling studies of terrane accretion.

2 Accretionary processes

Accreted terranes are typically composed of units scraped off of FATs and mixed in with other subducting sediments or crust in melange or accretionary prism formations. The FAT also undergoes severe internal deformation while accreting/subducting. We observe four types of accretion processes in the Phanerozoic geologic record: incorporation into the accretionary complex, underplating to the overriding crust (sometimes termed subcretion), obduction over the overriding plate (or flake tectonics), and collision (Fig. 1).

Incorporation of FAT crust into the accretionary prism occurs through offscraping or underplating onto the prism (Cloos and Shreve, 1988). Offscraping of FAT crust into the accretionary wedge or imbricate thrusting onto the front of the accretionary wedge are observed often in the geologic record (Fig. 1a). In this type of accretion, the FAT crust does not subduct completely, but instead builds out the accretionary wedge seaward, as in an accretionary plate margin (Clift and Vannucchi, 2004). Landward-verging imbricate thrust faults typically shear off blocks of tens to hundreds of meters of FAT or oceanic crust (Kimura and Ludden, 1995). For example, the Oso Melange and Oso Igneous Complex in Costa Rica records the history of accreted oceanic plateaus,

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island arcs, and seamounts which were mixed in with accretionary prism sediments (Buchs et al., 2009).

Underplating of FAT crustal material onto the overriding plate during or after subduction, also called subcretion, is perhaps the most common type of terrane accretion. Crustal units can be offscraped and underplated onto the overriding plate by stacked thrust faults or they can be sheared and incorporated into the subduction channel (Fig. 1b) and later exhumed as part of the melange units. Moore (1989) suggests that temperature and strain rate control whether mass transfer of material by underplating or diffusive subcretion in the subduction channel is the primary accretion method. Active underplating in a modern subduction zone is clearly observed in seismic refraction studies of the Sagami trough in Japan (Kimura et al., 2010). Thrust slices of underplated FAT crust are often interlain with thrusted melange units, as seen in the imbricated intraoceanic arc and melange slices of the Klamath Mountains (Wright and Wyld, 1994). In addition, weak crustal layers can be activated as detachments that allow for shearing of crustal units (Zagorevski et al., 2009; Tetreault and Buiter, 2012).

Flake tectonics is the process of obduction of terranes during subduction/collision on top of an overriding strong wedge (Oxburgh, 1972). Accretion of FATs via flake tectonic mechanics is most notably evident in southwestern Canada where the Paleozoic Quesnellia, Stikinia, and Cache Creek terranes were thrust and subsequently transported hundreds of kilometers inland over a Proterozoic metasedimentary wedge (Snyder et al., 2009; van der Velden and Cook, 2005; Cook et al., 2004). Other notable examples of accretion via flake tectonics include the Alps (Oxburgh, 1972) and the Archean greenstone belts (Hoffman and Ranalli, 1988). The paucity of flake tectonic-mechanics in Phanerozoic terrane accretion is explained by the absence of a strong overriding wedge in most subduction zones (Ellis, 1988).

Intact accretion of FAT crusts by “docking” is technically a collisional process rather than an accretionary process, but is still a method of continental growth via addition of exotic crustal material (Fig. 1d). Continental fragments and composite terranes typically lead to collision. In this process, it is possible to preserve the whole crustal section of

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terranes. Many of the larger FATs such as oceanic plateaus and continental fragments are accreted by collision. A notable example of docking of major crustal units is in Canada, where lithospheric suture zones bounding major terranes are identified with seismic refraction lines (Clowes et al., 1995). Intraoceanic island arcs often are on the overriding plate, on the receiving end of accretion processes. Arc-continent collision in this configuration allows for the overriding island arc to be added as an intact unit to the subducting continent.

3 Island arcs

3.1 Island arcs: general setting

Island arcs are volcanic island chains that form on the overriding oceanic plate at subduction zones (Fig. 2). Extinct intra-oceanic island arcs, also called remnant arcs, back arc, or ridges, also can become accreted allochthonous terranes of island arc affinity. Continental volcanic arcs are defined as volcanic arcs built on the continental upper plate of a subduction zone, and therefore excluded from this compilation. However, some oceanic island arcs are built on fragments of continental crust, most notably is Japan, and can eventually become accreted terranes, and those special cases are included. Island arc chains are geographically curvilinear, spanning hundreds of kilometers along strike and about 100 km in width (Calvert, 2011). The topography of island arcs is quite striking, with the elevation rising from sea floor to sometimes a couple of kilometers above sea level over just 10 or 20 km distance. The locations of island arcs ($\sim$ 120 km from the trench in subduction zones (England et al., 2004)) are believed to be dictated by slab dip and melting in the mantle wedge (England and Katz, 2010) and/or fluid release from the downgoing slab (Grove et al., 2009). Remnant arcs are created by either backarc rifting of the forearc or abandonment due to changes in plate motion (Karig, 1972). The Izu-Bonin-Mariana arc system is one such example: it

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is composed of several active island arc chains with more than one remnant backarc produced by changing plate motions (Stern et al., 2003).

Island arcs are the most widely intuited contributor of continental crustal growth (Stern and Scholl, 2010), primarily because the crustal composition is believed to be most similar to the felsic continental crust. Using volume estimates from Condie and Kröner (2013), we project about 13 % of post-Archaen accreted terranes are oceanic island arcs and 55 % are continental arcs. Cloos (1993) estimated that island arcs greater than 15 km in thickness are buoyant enough to collide with continental crust, however the paucity of whole crustal sections of island arcs in the geologic record does not agree with this hypothesis (Condie and Kröner, 2013).

3.2 Island arcs: modern examples

There is a noticeable variation in crustal thickness and structure of modern island arcs between arc systems and even along strike within arc systems (Fig. 3) (Calvert, 2011), which can be attributed to the level of maturity in arc crustal evolution (Tatsumi et al., 2008), the amount of back arc extension (Nishizawa et al., 2007), and the magmatic production rate (Christeson et al., 2008). Mature island arc systems, such as the Izu-Bonin-Mariana system, have three crustal layers which were developed by partial melting of the initial immature basaltic arc crust (Tatsumi et al., 2008). The upper crustal layer often has a sharp velocity gradient and *P* wave velocities ranging from 3 to 6 $\mathrm{km\,s^{-1}}$ (Fig. 3), which are interpreted to be layers of volcaniclastics, volcanic flows, and sediments. The mid-crustal layer is characterized by seismic velocities of around 6 to 6.5 $\mathrm{km\,s^{-1}}$. This low velocity layer is often interpreted to be a layer of felsic to intermediate igneous rocks in many modern oceanic island arcs (South Sandwich: Leat et al., 2003; the Izu-Bonin-Mariana system: Kodaira et al., 2007a; Takahashi et al., 2007, 2009; Tonga Arc: Crawford et al., 2003). The felsic mid-crustal unit is produced by repetitive anatexis of the mafic lower crust (Tatsumi et al., 2008; Rioux et al., 2010). Juvenile island arcs are believed to lack this felsic middle layer, as in the cases of the Lesser Antilles and Leeward Antilles (Magnani et al., 2009; Christeson et al., 2008)

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and parts of the Kyushu-Palau Ridge (Nishizawa et al., 2007). The mid-crustal layer of the mature Aleutian arc, on the other hand, is inferred to be of a more mafic than intermediate composition, based on the higher seismic velocities at depths of 11 to 20 km (Shillington et al., 2004). The lower crustal unit of island arcs is typically characterized by seismic velocities ranging from 6.7 to 7.3 km s^{-1} (Fig. 3), and is interpreted to be gabbroic in composition underlain by mafic to ultramafic cumulates. The mafic and ultramafic cumulates are sometimes classified as a separate unit from the lower crust called the crust-mantle transition layer (CMTL) (Takahashi et al., 2007, 2009). The CMTL has typical seismic velocities around 7.0 to 7.6 km s^{-1} (Fig. 3). We include the CMTL as part of the crust because it is above the seismic moho in modern arcs and also found above mantle rocks in the accreted Talkeetna arc in Alaska (Rioux et al., 2007; Greene et al., 2006), Kohistan arc in Pakistan (Kono et al., 2009), and Guanajauto arc in Mexico (Lapierre et al., 1992). Seismic velocities ranging from 7.6 to 8.0 km s^{-1} are found below the lower crust of the Mariana arc and West Mariana rear arc in a thick layer, but the authors interpret the reflections between this layer and the lower crust as the Moho discontinuity and not the CMTL (Takahashi et al., 2007, 2008). Seismic reflections are also observed below this layer (Takahashi et al., 2007, 2008) and they are attributed to transformation of mafic materials during arc crustal generation rather than melt in the mantle (Takahashi et al., 2008; Tatsumi et al., 2008).

The average crustal thickness of island arcs (including remnant arcs), determined from the thickest regions in 26 seismic and gravity studies of island arcs, is $\sim 26 \pm 6$ km (Table 1). Bulk crustal densities were calculated from the *P* wave velocities of 17 seismic refraction studies, using the Nafe–Drake curve (Ludwig et al., 1970), the Christensen and Mooney (1995) relationships for all rocks at 10 km depth intervals, and the Christensen and Shaw (1970) curve based on mafic rocks from the mid-Atlantic ridge (Table 2). The densities calculated for the CMTL layer in this compilation range from 3.02 to 3.32 g cm^{-3}, using the Christensen and Mooney (1995) relationships. These values are within the range of, if not slightly lower than, the densities calculated based on mineral assemblages and sub-Moho conditions (> 0.8 MPa and 800–

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1000 °C) for the ultramafic pyroxenites from accreted island arcs (~ 3.25–3.40 g cm^{-3}) (Jull and Kelemen, 2001; Behn and Kelemen, 2006). Three seismic refraction studies constrained their crustal structure models with gravity modelling, and inferred a whole crustal density for the arcs which we compare to our calculated densities (Table 2). Coincidentally, the average crustal density (2.79 g cm^{-3}) calculated with the Christensen and Mooney (1995) relationship is identical to the average density calculated with the Nafe–Drake curve. The average crustal densities calculated from the three relationships are lower than the bulk density of average continental crust (2.83 g cm^{-3}: Christensen and Mooney, 1995) and the average density for oceanic crust (2.86 g cm^{-3}: Carlson and Herrick, 1990).

3.3 Island arcs: accreted examples

Accreted island arcs are mostly identified in the geologic record as calc-alkaline volcanic units. The amount of crustal thickness that is actually accreted varies significantly throughout the geologic record. It is not common to find the entire crustal section preserved in terranes of accreted island arcs. Only a few accreted island arc terranes (i.e. Talkeetna, Bonanza, Kohistan, Canyon Mountain, and El Paxtle arcs) contain parts of all of the original crustal layers, but these accreted layers are severely thinned. Geobarametric and geologic studies suggest original crustal thicknesses of 30–35 km for the Talkeetna arc (Greene et al., 2006; Hacker et al., 2008), 24 km for the Bonanza arc (Canil et al., 2010), 45 km for the Kohistan arc (Miller and Christensen, 1994), and about 30 km for the Canyon Mountain complex (Pearcy et al., 1990). The remaining preserved crustal thicknesses are 18 km thickness for the Talkeetna arc (Greene et al., 2006), 15 km for the Bonanza arc (Canil et al., 2010), and about 8.3 km for the Canyon Mountain complex (Pearcy et al., 1990). The Kohistan arc is believed to be entirely preserved in crustal thickness (Miller and Christensen, 1994; Petterson, 2010). Interestingly, the estimated original crustal thicknesses of these accreted terranes are significantly larger than the average thickness of modern island arcs, most likely be-

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cause of the large uncertainty and often lack of constraints in estimating the depth of crystallization. Truncated units from all crustal layers are also found in the accreted Alisitos-Teloloapan arc in Mexico (Lapierre et al., 1992) and the Alisitos Arc in Baja (Busby, 2004; Busby et al., 2006), but no estimates of original thickness has been made.

Based on the few terranes that contain units from the entire arc crust and even the upper mantle, accreted island arcs are composed of three crustal layers. The upper crust in accreted island arcs is mostly composed of volcaniclastics, basalt flows, tuffs, and sediments (Lapierre et al., 1992; Pearcy et al., 1990). The middle layers identified in accreted island arc suites are felsic to intermediate composition plutons; such as tonalities, diorites, and trondhjemites (Fig. 3) (Rioux et al., 2010; Greene et al., 2006). In the accreted Talkeetna arc, the middle crustal layer is composed of intermediate to felsic plutons that produce seismic velocities of 6 to 6.5 km s^{-1} (Rioux et al., 2010). The lower crust is typically mafic in composition, including garnet gabbros, layered gabbros, and pyroxene granulites (Debari and Sleep, January, 1991; Greene et al., 2006; Lapierre et al., 1992). Ultramafic cumulates such as pyroxenite gabbros and dolerites are best preserved in the accreted Kohistan arc (Kono et al., 2009; Miller and Christensen, 1994), but smaller units are also found in the El Paxtle arc in the Guerrero terrane (Lapierre et al., 1992), Talkeetna arc (Greene et al., 2006), Canyon Mountain complex (Pearcy et al., 1990), and Bonanza arc (Canil et al., 2010). Seismic velocities from the Tonsina pyroxenite unit of the accreted Talkeetna arc are 7.3 to 7.6 km s^{-1} (Behn and Kelemen, 2006) and from the Jijal garnet pyroxenites of the accreted Kohistan arc are 7.8 to 8.4 km s^{-1} (Kono et al., 2009), correlative to the CMTL in modern island arcs.

The preserved thicknesses of crustal units of island arcs in accreted terranes varies depending on the style of accretion and collision and the subduction polarity in an arc-continent convergence zone (Draut and Clift, 2013). Because island arcs form on the overriding plate at subduction zones, whole-arc accretion is most likely due to a continent entering the subduction zone on the downgoing plate, and then the arc is obducted or collided onto the continent. This type of arc-continent collision is currently observed at the Luzon Arc in Taiwan (Clift et al., 2009b). The mostly intact, accreted Kohistan arc in Pakistan is a notable example of arc-continent collision (Searle et al., 1999). But in this case the Kohistan arc is believed to have been on the subducting plate in a "backwards facing" arc-continent collision polarity (Draut and Clift, 2013). Besides arc-continent collision, island arcs collide/accrete to another FAT (such as an oceanic plateau) and create a large composite terrane that will collide and suture to continents and preserve remnants of the island arc crust. Accreted examples are the Talkeetna arc in Wrangellia Composite Terrane in Canada: (Greene et al., 2006) and the Stikine arc in Canda: (English and Johnston, 2005; Johnston and Borel, 2007). Quite possibly the modern-day Ontong Java Plateau-Solomons islands in the southwest Pacific: (Petterson et al., 1999) will be a future accreted composite terrane.

However, in most cases only the upper 2 to 5 km of arc crust are accreted onto continents through thin-skinned thrusting and preserved. This most likely occurs when island arcs are on the subducting plate and arc material is underplated and accreted onto the overriding plate. For example, in the eastern Klamath Mountains of North America, Devonian island arc units are 2.5 to 3.5 km in thickness and include mafic pillow basalts and a felsic upper unit, indicitive of upper to middle crustal layers (Dickinson, 2000). Island arc fragments in the Central Asia orogenic belt are also units bound by imbricate thrust faults (Windley et al., 2007). Detachment faults produced by thinning during back-arc extension or rheologically weak crustal layers can enable accretion of island arc crustal units. Zagorevski et al. (2009) suggest that Ordovician terranes of arc and back-arc origins in the Central Newfoundland Annieopsquotch accretionary tract, were accreted onto Laurentia because of low angle detachments within the arcs that were produced during back-arc extension. Also, the felsic middle crustal layer could be weakened by metasomatism from fluids released during subduction and act as a decollement layer to underplate arc crustal units onto the continent (van Staal et al., 2001).

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Another possible mechanism for accretion is the delamination of the CMTL and increased buoyancy of the remaining island arc crust. The CMTL, composed of ultramafic cumulates and peridotites, is often cited as a layer that delaminates either pre- or syn-accretion (Behn and Kelemen, 2006; Garrido et al., 2007). The delamination of the ultramafic CMTL will result in a more felsic overall composition for island arcs allowing the remaining arc crust to match better with the composition of continental crust (Takahashi et al., 2007, 2009). Densities calculated from mineral assemblages and in-situ conditions from gabbronites and pyroxenites of the CMTL in accreted island arcs are 0.05 - 0.25 $g\,cm^{-3}$ greater than that for mantle material at those conditions, therefore leading to a negative buoyancy instability (Jull and Kelemen, 2001; Behn and Kelemen, 2006). Evidence for CMTL delamination is cited in trench-parallel upper mantle anisotropy observed below modern island arcs (Behn et al., 2007). Greene et al. (2006) also find that the volume of pyroxenites in the Talkeetna arc is much less than needed to produce the arc's crustal composition, and infer that this discrepancy is due to either foundering of much of the CMTL or the missing pyroxenites were not accreted. On the other hand, the Tonsina pyroxenites of the Talkeetna arc are conformably underlain by upper mantle harzburgites (Rioux et al., 2007), suggesting the unlikelihood that volumes of the pyroxenite are removed. Furthermore, the depleted rare earth element (REE) signature of the ultramafic section of the Kohistan arc indicates that it did not form from crustal fractionation, but is a result of mantle and crust mixing (Garrido et al., 2007). The thickness of CMTLs cannot be clearly determined through crustal fractionation modeling, and the apparent missing thickness due to delamination may not be valid, at least for the Talkeetna arc.

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4 Oceanic plateaus, submarine ridges, and seamounts

4.1 Oceanic plateaus, submarine ridges, and seamounts: general setting

Mafic igneous regions with thicker crust than the surrounding ocean crust can be large regions of thick igneous crust such as oceanic plateaus and submarine ridges (Fig. 4) or smaller igneous regions such as seamounts (Fig. 5). Large igneous provinces (LIPs) are large, aseismic regions of mafic igneous crust on continental crust and ocean crust (Coffin and Eldholm, 1992, 1994; Bryan and Ernst, 2008). The reclassification and redefinition of categories within and of LIPs themselves have been the subject of debate through the years. Originally, oceanic LIPs (oceanic flood basalts) were termed oceanic plateaus (Kroenke, 1974). By the 1980s the term "oceanic plateau" was expanded to include any plateau-like features on the ocean floor, including seamounts, extinct ridges, and even continental plateaus and remnant island arcs (Ben-Avraham et al., 1981). Oceanic plateaus were redefined to exclude remnant island arcs and only include oceanic flood basalts, submarine ridges, rifted continental fragments, seamount chains, and hot spot tracks in the global compilations of Schubert and Sandwell (1989) and Marks and Sandwell (1991). Eventually, the term "LIP" was introduced by Coffin and Eldholm (1992, 1994) to define all regions of "voluminous emplacements of predominantly marine extrusive and intrusive rock whose origins lie in processes other than "normal" seafloor spreading." This classification included continental flood basalts and associated intrusive rocks, volcanic passive margins, oceanic plateaus, submarine ridges, ocean basin flood basalts, and seamount groups. In line with this volumetric definition, Sheth (2007) called all volcanic regions of large areal extent "LIPs", which led to the unusual addition of subduction-related plutons and seafloor-spreading-related volcanics in the LIP classification. This confusion led to the clarification of continental and oceanic LIPs by Bryan and Ernst (2008) as "magmatic provinces with areal extents $> 0.1\,\mathrm{M\,km^2}$, igneous volumes $> 0.1\,\mathrm{M\,km^3}$ and maximum lifespans of ~ 50 Myr that have intraplate tectonic settings or geochemical affinities, and are characterised by igneous pulse(s) of short duration ($\sim$ 1–5 Myr), during which a large proportion ($> 75\,\%$)

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of the total igneous volume has been emplaced." From the viewpoint of terrane accretion, the smaller volume submarine ridges and seamounts are also likely to accrete to continental crust and can be difficult to differentiate from accreted oceanic LIPs, so we discuss submarine ridges and seamounts with oceanic plateaus. Even though the seismic crustal structures of oceanic plateaus and thick-crustal submarine ridges appear similar, their origins are different (Bryan and Ernst, 2008). In this review, we follow the definition of oceanic plateaus as outlined by Kerr (2003); Kerr and Mahoney (2007) and Bryan and Ernst (2008) for differentiating between oceanic plateaus vs. submarine ridges. Kerr (2003) and Kerr and Mahoney (2007) state that oceanic plateaus are a type of LIPs formed on oceanic crust, and are vast, wide regions of anomalously thick oceanic crust: submarine analogues to continental flood basalts. Some submarine ridges, such as the Nazca Ridge, Cocos Ridge, and the Tuamotu Plateau have even been classified as oceanic plateaus in past studies. However, based on the definition of Bryan and Ernst (2008), these mafic regions are not voluminous enough nor formed due to rapid magmatism, and thus must be classified as submarine.

The origin of oceanic plateaus has been a point of vigorous discussion in the literature in terms of whether the feeder magmas originate from deep plumes or in the upper mantle (Richards et al., 1989; Foulger, 2007; Campbell and Kerr, 2007). Still, there are numerous geochemical, geophysical, and geodynamic studies that support the hypothesis of plume-formed oceanic flood basalts (Campbell, 2007; Hastie and Kerr, 2010; Hoernle et al., 2010). Interestingly, several modern-day oceanic plateaus were emplaced during the Cretaceous and were later rifted apart at triple junctions, such as the Kerguelen-Broken Ridge (Frey et al., 2000), Manihiki-Hikurangi-Ontong Java (Taylor, 2006; Davy et al., 2008), and Agulhas-Maud Rise-Northeast Georgia Rise plateaus (Parsiegla et al., 2008). The accreted Sorachi plateau is related to the Shatsky Rise oceanic plateau (Ichiyama et al., 2012), and could be another possible triple junction-related oceanic plateau (Sager, 2005).

Submarine ridges are the result of significant magmatism produced at hot spot tracks, leaky transforms, or now-extinct mid-ocean ridges. Recent seismic imaging

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and geochemical research has shown that the Nazca Ridge, Cocos Ridge, and the Tuamotu Plateau are not hotspot-related (Woods and Okal, 1994; Patriat et al., 2002; Sallares et al., 2003; Harpp et al., 2005), but rather formed due to leaky transform faults and mid-ocean ridges. Conversely, the Carnegie and Madeira Tore ridges formed from hotspot magmatism that intruded and underplated the oceanic plate, creating a new, thick mafic crust (Sallares et al., 2005; Geldmacher et al., 2006).

The third province of mafic anamolous crustal regions on the ocean floor that we include in this section are large seamounts and seamount chains (Coffin and Eldholm, 1994). In general, seamounts are submarine volcanoes, smaller in areal extent to oceanic plateaus and submarine ridges, with geochemical signatures that suggest different sources for different seamount chains. The number of seamounts > 1.5 km in height currently on the ocean floor is estimated to be more than 13 000 based on satellite altimetry (Wessel et al., 2010). For the review on crustal structure we focus only on large submarine volcanoes (> 3 km high) which are included in the list of LIPs by Coffin and Eldholm, 1994, (Fig. 5). Many of these large seamounts have heights of 3 to 5 km above the surrounding ocean floor. Seamounts form by various tectonic processes: they can be the result of plate extension and upper mantle mini-convection cells under mid-ocean ridges or transforms (Buck and Parmentier, 1986), deep mantle upwellings, short-lived hotspot volcanism, upper asthenospheric upwelling, and lithospheric cracking (Forsyth et al., 2006; Briais et al., 2009; Sandwell and Fialko, 2004). Koppers et al. (2003) theorize that the seamounts of the South Pacific are a result of discontinuous volcanism from a broad mantle upwelling that encompasses plumelets of distinct geochemical signatures. Whereas Sandwell and Fialko (2004) argue that flexural response to lithospheric thermal contraction produces the cracking and warping on the Pacific plate, and that small convection cells and mantle upwellings are not necessary. The different formation processes may produce thermal structures and weakness that will affect the ability of seamounts to accrete.

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4.2 Oceanic plateaus, submarine ridges, and seamounts: modern examples

Oceanic plateaus and submarine ridges have similar crustal thicknesses, and from 32 seismic and geophysical studies, their combined average crustal thickness is approximately 21 ± 4 km (Table 3). Even though the 33 km thick Ontong Java Plateau is commonly used to exemplify the typical crustal thickness of an oceanic plateau, it is anomalously thick for oceanic plateaus (Fig. 6). Oceanic plateau and submarine ridge bathymetry generally is 2 to 3 km above the surrounding ocean crust. Oceanic plateaus and submarine ridges typically have a sedimentary layer, upper crust, lower crust, and mafic underplating identified in seismic interpretations, although several oceanic plateaus and submarine ridges have an additional middle crustal layer (Fig. 6). Seismic refraction studies indicate an upper layer of 1–4 km thickness of low seismic velocities, correlated to limestones, pelagic sediments, and volcaniclastic sediments. Underlying that is the upper crust with P wave velocities of 4.5–6.0 km s^{-1}, commonly interpreted as mixed basaltic flows and pelagic material, altered basalts, and other submarine flows. The upper crust is sometimes correlated to oceanic layer 2 because of similar seismic velocities. In oceanic plateaus and submarine ridges where three crustal layers are identified, the upper crust has very low seismic velocites (3.5–4.5 km s^{-1}) and the middle crust has velocities typical of basalts (5.0–6.0 km s^{-1}). The lower crust typically has seismic velocities of 6.5–7.0 km s^{-1} in all oceanic plateaus and submarine ridges. Over-thickened lower crusts are common in this group of FATs, especially in submarine ridges. The lower crust is often interpreted to be gabbroic or correlative to oceanic crust layer 3. We caution against relating crustal units of this FAT to oceanic crust because oceanic plateaus and submarine ridges are formed differently from typical oceanic crust. Many oceanic plateaus and submarine ridges have a basal unit of high seismic velocities (7.0–7.9 km s^{-1}), also highlighted in a compilation by Ridley and Richards (2010). Grevemeyer and Flueh (2000) and Gupta et al. (2010) suggest that this mafic basal unit is underplated material due to plume magmatism. Early studies have suggested that the high seismic velocity lower crustal layer was representative of a ductile layer that occurs in crust greater than 15 km thick (Schubert and Sandwell, 1989). However, the theory that all large oceanic igneous provinces will have a ultramafic layer was debunked by the compilation of Ridley and Richards (2010).

The seismic crustal structure of seamounts consists of one or two layers and may contain a thick intrusive volcanic core. Seamounts are volcanoes build up on top of oceanic crust (Koppers and Watts, 2010). The upper crustal layers of seamounts and oceanic crust correlate with the seismic velocities of basalts. The lower crustal units are interpreted to be gabbros and sheeted dikes. Many seamounts, such as the those in the O'Higgins and Musician seamount chains, have two crustal layers similar to oceanic crust and no seismically discernable intrusive core (Kopp et al., 2003, 2004). Other submarine volcanics, such as Great Meteor seamount and Marcus-Wake seamount chain, have a thick layer that is seismically different from the surrounding oceanic crust and is interpreted as the volcanic core (Weigel and Grevemeyer, 1999; Kaneda et al., 2010). In some seamounts, such as the Hawaiian chain (Leahy et al., 2010) and La Reunion (Charvis et al., 1999), the oceanic crust is underplated by a seismically fast layer. Yet other submarine volcanics, including the Louisville hot spot track (Contreras-Reyes et al., 2010), Musician seamounts (Kopp et al., 2003), O'Higgins Seamount (Kopp et al., 2004), and Marcus-Wake seamount chain (Kaneda et al., 2010), do not have any seismic high-velocity layer below the crust. The high seismic velocities found in the Louisville and Marcus-Wake seamount chains are attributed to mafic intrusions in the lower crust (Contreras-Reyes et al., 2010; Kaneda et al., 2010). The subcrustal high-velocity layer in other seamounts is theorized to be from mafic dikes formed as a lithostatic response to loading (Hawaii: Leahy et al., 2010), hot spot material (La Reunion: Charvis et al., 1999), or hydrated lithosphere (O'Higgins seamount: Kopp et al., 2004).

We caculated an average crustal density from the *P* wave velocities from 23 seismic refraction studies of oceanic plateaus and submarine ridges (Table 4). The average crustal density is estimated to be 2.84 g cm^{-3} from the Christensen and Mooney (1995) depth-dependent relationship, 2.84 g cm^{-3} using the Nafe–Drake curve (Ludwig et al.,

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1970), and 2.82 g cm^{-3} with the Christensen and Shaw (1970) depth-dependent relationship (Table 4). Interestingly, these values are close to the densities of average continental crust (2.83 g cm^{-3}: Christensen and Mooney, 1995) and average oceanic crust (2.86 g cm^{-3}: Carlson and Herrick, 1990). Generally, the densities of oceanic plateaus and submarine plateaus calculated from the Nafe–Drake and Christensen–Mooney relationships are similar to the densities determined in combined seismic-gravity studies (Table 4).

4.3 Oceanic plateaus, submarine ridges, and seamounts: accreted examples

Accreted oceanic plateaus and submarine ridges are typically identified in the geologic record as mafic to ultramafic basalts unit in accreted terranes. Kerr (2003) presents a diagnostic criteria for identifying ancient oceanic plateaus in the geological record based on geology, petrology, and geochemistry. Oceanic plateaus are composed mainly of tholeiitic basalts with minor amounts of picrites and komatites, and are geochemically distinct from mid-ocean ridge basalt (MORB)-type and ocean-island basalt (OIB)-type mantle sources (Kerr, 2003; Hastie and Kerr, 2010). Depending on their origin, submarine ridge basalts can also have MORB or ocean-island basalt OIB signatures. It is quite likely that many greenstones and mafic accreted units, identified as accreted ophiolites or oceanic crust, may actually be oceanic plateaus (see Table 4 in Kerr et al., 2000). For example, the hotspot related greenstones of the Chugoku and Chichibu belts in Japan were re-interpreted as accreted oceanic plateau/submarine ridges rather than the earlier inference of mid-ocean ridge basalts, based on high Zr/Y ratios that are more similar to OIB geochemical signatures (Tatsumi et al., 2000).

The total amount of preserved crustal structure and thickness of oceanic plateaus varies in the observed geological record of accreted terranes. Sometimes the entire crustal thickness is preserved in accreted terranes, as in the Triassic Wrangellia terrane of North America, or only truncated units from all crustal layers are found, as in the accreted Gorgona and Columbia oceanic plateaus of South America. Seismic refrac-

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tion studies indicate that the total thickness of the Wrangellia composite terrane crust is about 25^{+} km in Vancouver (Ramachandran et al., 2006; Clowes et al., 1995) and 30 km in Alaska (Brennan et al., 2011). Approximately 6 km of exposed stratigraphic thickness, correlated to the sedimentary and upper crustal layers of the Wrangellia oceanic plateau, is found in Vancouver Island (Greene et al., 2010). Wrangellia's exposed units are composed of limestone and pelagic sediments, pillow lavas, massive flood basalts, subaerial and submarine flows, and olivine-rich basalts (Greene et al., 2009, 2010). In other accreted oceanic plateaus, the preserved crustal thicknesses can be as low as 2 to 7 km thick. The total reconstructed thickness of the accreted Columbia oceanic plateau is only 8 to 15 km, but units from all of the original crustal layers are found (Kerr et al., 1998). The accreted Colombian oceanic plateau also has preserved units of the ultramafic layer below the lower crust, which include olivine gabbronorites and pyroxenites (Kerr et al., 1998). In Ecuador, fragments of the Gorgona oceanic plateau include pillow basalts, dolerite sheets, and gabbros of the upper and mid crust, overlying the plume-derived magmas of the lower crust in thin-skinned thrust sheets (Kerr and Tarney, 2005; Kerr et al., 1998).

Accreted submarine ridges and seamounts are typically only truncated units of crustal layers. In Central America, various "ophiolitic" units are found with OIB geochemical signatures, which are interpreted as hotspot-related seamounts or submarine ridges (Hoernle et al., 2002; Geldmacher et al., 2008; Buchs et al., 2009). The enigmatic Siletz terrane of northern California and Oregon is composed of volcanics with OIB signatures that has been variously interpreted as a hot spot track, slab window, and mid-ocean ridge (Schmandt and Humphreys, 2011; McCrory and Wilson, 2013). Examples of accreted seamounts, identified primarily by their OIB signature, are the alkali basaltic units found in Japan (Isozaki et al., 1990). Typical seamount-derived terranes include thin-skinned units of radiolaran cherts, limestones, serpentinized peridotites, layered gabbros, and alkali basalts that are on the order of hundreds of meters thick (Geldmacher et al., 2008; Buchs et al., 2009). Accreted ocean-island basalts, interpreted to be remnants of seamounts, are often found within accretionary complexes

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(e.g. Cache Creek terrane: Johnston and Borel, 2007). Accreted seamounts are often "decapitated" in the accretionary prism, instead of underplated to the overriding plate. The seamount terranes of the Oso Igneous Complex in Costa Rica are within an accretionary prism complex, suggesting that the seamounts were decapitated within the prism and subsequently accreted to the Central American active margin (Buchs et al., 2009). Watts et al. (2010) suggest that even small seamounts can be accreted if the subduction channel is narrow, highly coupled, or if the seamount is regionally compensated by a thick, strong lithosphere.

Accretion of oceanic plateaus and large submarine ridges can occur as collision and whole crustal addition to a continent, or by underplating and accretion of sheared crustal units. Kerr et al. (2000) suggest that after mafic oceanic plateaus are accreted or collided, causing the subduction zone to jump, silicic magmas intrude and mature the accreted plateau lithology towards a more continental crust lithology. The basal cumulate layer may be a ductile layer that serves as a detachment to allow for underplating, an idea originally speculated by Schubert and Sandwell (1989) to develop in plateaus that exceed 15 km in thickness based on the rheological relationship of strength with depth. Even though this layer is not found in all LIPs and seamounts of great thicknesses (Ridley and Richards, 2010) (Fig. 6), the cumulate or underplated magma layer could definitely serve as a ductile layer to initiate detachment within the subduction zone. The Colombian oceanic plateau is the only documented accreted plateau that has accreted units of the basal ultramafic cumulate layer, most likely due to the onset of collision early after plateau formation (Kerr et al., 1998), leading us to hypothesize that this ultramafic basal layer commonly serves as a detachment layer, therefore it is not observed in other accreted oceanic plateaus.

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5 Continental fragments and microcontinents

5.1 Continental fragments and microcontinents: general setting

Continental fragments, microcontinents, and continental ribbons are submarine regions of continental crust on the oceanic plate (Fig. 7) that are the result of rifting events. Modern continental fragments on the ocean floor include the Rockall Bank, Hatton Bank, Campbell Plateau, Lord Howe Rise, and the Norfolk Rise (Fig. 7). These submarine plateaus of continental crust resulted from extensional episodes forming passive margins (Peron-Pinvidic and Manatschal, 2010). Continental fragments are bound by oceanic crust on one side and thick sedimentary basins overlying extremely thinned continental crust on the other. Microcontinents, such as Jan Mayen and the Seychelles, are surrounded by oceanic crust. Because continental fragments and microcontinents are formed during extensional processes, it is likely they are bound by deep crustal detachment faults and are thinned from normal faulting (Peron-Pinvidic and Manatschal, 2010; Reston, 2011). Continental fragments and microcontinents are theorized to form as a result of plume interaction with passive margins (Müller et al., 2001), localized thinning on the basins surrounding continental fragments (Peron-Pinvidic and Manatschal, 2010), or inherited structural grains from ancient sutures zones (Hitchen, 2004).

5.2 Continental fragments and microcontinents: crustal structure

Naturally, continental fragments and microcontinents have crustal compositions similar to that of typical continental crust. In general, seismic studies have identified two crustal layers with low seismic velocity values, representative of their continental affinity. However, the rifting processes that led to the formation of continental fragments and microcontinents most likely affects their layer and entire thicknesses (Morewood et al., 2005), as well as adding mafic intrusions to the crust. From 36 geophysical studies of continental fragments, we determine an average crustal thickness of $\sim 24.8 \pm 5.7$ km (Table 5). Continental fragments have a sediment layer that can be up to 5 km thick

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overlying 2 to 3 crustal layers, some of which are underplated with a mafic layer (Fig. 8). The thick sedimentary layer is generally devoid of volcanics, but some rift-related sills may intrude the sedimentary sequences of continental fragments in regions of high magmatism (Richardson et al., 1999; Davison et al., 2010). The upper crust has seismic velocities around 5.5 $\mathrm{km\,s^{-1}}$, most likely from rocks of granitic and gneissic composition. The seismic velocities of the mid-crustal layer range from 6.0 to 6.5 $\mathrm{km\,s^{-1}}$. The lower crust typically has velocities of 6.5 to 7.0 $\mathrm{km\,s^{-1}}$ and is inferred to be gabbroic. In only a few continental fragments, a basal layer with high seismic velocities (7.4 to 7.8 $\mathrm{km\,s^{-1}}$) is found above the seismic Moho (Fig. 8). The high velocity layer under the Faroe Bank is interpreted to be a layer of mafic sill intrusions in the crust related to the Iceland plume or convective upwellings (Harland et al., 2009). Under the Rockall Bank, this layer is believed to be serpentinized upper mantle (O'Reilly et al., 1996). For the continental fragments off the Australian margin, the high velocity lower layer is interpreted as mafic underplating (Grobys et al., 2009). Mostly, the high velocity seismic layer is found below the surrounding basins with oceanic or thinned continental crust. In these regions, the high velocity layer is also hypothesized to be either serpentinized mantle or mafic underplating (O'Reilly et al., 1996; Reston et al., 2001; Lundin and Doré, 2011).

The average crustal density of continental fragments and microcontinents, determined with the Christensen and Mooney (1995) depth-dependent relationship from seismic velocities from 20 studies, is $\sim 2.81\,\mathrm{g\,cm^{-3}}$ (Table 6). As we expect, the average crustal density of continental fragments and microcontinents is similar to that of the typical continental crust. Despite having thicknesses much lower than the average continental crust (25 km compared to 41 km) the lower densities calculated because of the smaller depths (< 25 km) are balanced by the mafic underplating contribution to several of the continental fragments. Interestingly, the average crustal density determined from the 8 seismic studies that constrained their models with gravity measurements, is a lower value of $2.79\,\mathrm{g\,cm^{-3}}$. The lower densities derived by gravity modelling are

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mainly from studies on continental fragments with no seismically-identified mafic basal layer.

5.3 Continental fragments and microcontinents: accreted examples

Because the classification and the identification of how such features form offshore of passive margins is relatively new (Peron-Pinvidic and Manatschal, 2010, see references therein), there has been little recognition of such features in the accretionary record. The most recognized accreted continental crustal units are found in the Alps. Many of the crustal units accreted in the Alps are believed to be rifted continent fragments (Manatschal, 2004), such as the Brianconnais terrane (Handy et al., 2010), gneiss units of the Piemonte units (Beltrando et al., 2010), and the Monte Rosa nappe (Froitzheim, 2001). In Newfoundland, the Dashwoods terrane is interpreted to be a rifted microcontinent block on the passive margin of Laurentia that was later reunited with Laurentia during the Taconic orogeny (Waldron and van Staal, 2001).

Accretionary and collisional processes could utilize the underlying detachment faults or surrounding exhumed and serpentinized mantle lithosphere. There is evidence for detachment faults that are inherited from initial rifting on the Brianconnais terrane and other accreted continental fragments (Reston, 2011). In western Norway mantle peridotite melange units, reinterpreted as hyperextended crust, underlie accreted microcontinent slivers of Gula, Jotunn, and Lindas nappes (Andersen et al., 2012). Precambrian terranes with continental affinities (gneisses) of the Central Asian Orogenic belt are bound by ophiolitic sutures and interpreted as microcontinents rifted off of the East Gondwana margin (Windley et al., 2007). It is possible that the ophiolites (characterized by sedimentary units, volcanics, and deep marine formations: Windley et al., 2007) bounding these continental terranes are hyperextended crust.

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6 Composite terranes

Often it is the case that FATs will combine before accreting onto a continent – such as oceanic plateau-island arc composite terranes. In general, the larger mass of these FATs makes it inevitable for accretion by collision. Modern examples of composite terranes include the Phillippines and the Ontong Java-Solomon Islands. The currently accreting Yakutat terrane in Alaska has been speculated to be a continental-oceanic composite terrane. Parts of the Yakutat subducting under Alaska involves oceanic basement or oceanic plateau crust, while the eastern region of the crust that is accreting is of continental composition (Bruhn et al., 2004).

In the geological record, large volumes of crustal accretion are carried out by the collision of composite terranes or continental fragments onto continents (Vink et al., 1984). In North America, the amalgamation of the Wrangellia and Stikinia terranes resulted in a ribbon continent (SABIYA) that was ~ 8000 km long and ~ 500 km wide (Johnston, 2001). During the collision of the superterrane to North America, the mantle lithosphere belonging to the microcontinent was also sutured to the continent, as evidenced by seismic reflection lines (Hammer et al., 2010) and mantle xenoliths from both regions (Johnston, 2008). Another notable accreted ribbon composite terranes is the Cimmerian superterrane which closed the Tethyan sea (Sengor, 1979).

7 Discussion

7.1 FAT similarities and differences

This review of the crustal composition of future accreted terranes highlights the variability in crustal thickness and structure between FAT groups as well as within each group. A comparison of modern FATs to their accreted versions can help us understand crustal composition of accreted units, the amount of crust lost during subduction, and the processes that allow for accretion and collision. Based on average crustal thick-

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ness and density, there appears to be no significant difference between FAT groups that would indicate that one particular group would be more subsceptible to subduction or accretion. The seismic velocity profiles from each of the three FAT groups show considerable overlap with the average continental crust given in Christensen and Mooney (1995) (Fig. 9). However, all three groups show considerable variablity in their crustal structure, depending on their formation and tectonic history, and this will play a part in terrane accretion.

The crustal structure of island arcs is composed of two to three layers which are commonly underlain by ultramafic cumulates (the CMTL). The main differences in arc crustal composition and thickness are products of maturation: juvenile arcs are more mafic, thinner and smaller, while mature island arcs have undergone repetitive anatexis to produce a felsic middle layer. The ultramafic cumulate layer found in most arcs could be formed during early anatexis of the inital basaltic island arc crust (Tatsumi et al., 2008). Foundering of this subcrustal ultramafic layer on mature island arcs would leave a crustal composition that is intermediate composition and a better contributer to the continental crust. However, many accreted terranes from island arcs do contain units from the ultramafic CMTL, so further modification needs to occur to produce a more compositionally similar crust to continents, post-arc accretion.

Oceanic plateaus and submarine ridges are quite varied in their crustal structure, and some are also underlain by a high seismic velocity layer. Moreover, recognized oceanic plateaus do not have unique seismic crustal structures or thicknesses which can be differentiated from submarine ridges (Fig. 6). In determining whether a large mafic igneous feature on the ocean floor is an oceanic plateau or submarine ridge, obviously the geochemical and geodynamic history is needed. Accreted mafic terranes, typically greenstone belts, represent oceanic plateaus, submarine ridges, and seamounts that have been added to continents by accretion or collision. The large terranes (> 30 km thick) of Wrangellia and Siletz in North America indicate that these mafic bodies are significant contributors to continental crust, despite their mafic composition. Indeed, Archeaen greenstone belts have led some researchers to suggest

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that accreted oceanic plateaus were the major crustal contributor in the Precambrian (e.g., Puchtel et al., 1998; Desrochers et al., 1993). However, more recent (Paleozoic) tectonic growth of continents is believed to be from felsic island arcs or modified post-accretion oceanic plateaus (Clift et al., 2009a, b; Stern and Scholl, 2010). There is observational evidence for modern day subduction of oceanic plateaus and submarine ridges: the Hikurangi oceanic plateau subducting seemingly intact to approximately 65 km depth under New Zealand (Reyners et al., 2006), the Ontong Java Plateau subducting under the Solomon Islands (Mann and Taira, 2004), and the Nazca Ridge under Peru (Hampel et al., 2004). In these instances, units from the sedimentary and upper crustal layers are being actively scraped off at the accretionary prism (Mann and Taira, 2004), or underplated at the plate interface (Contreras-Reyes and Carrizo, 2011) leaving behind evidence of the oceanic plateau's existence after subduction.

The accretion of continental fragments or microcontinents does not require post-accretion modification to achieve the average composition of continental crust. Being rifted off fragments of continental crust, continental fragments have crustal compositions identical to continental crust. Continental fragments are part of passive margin architecture, and therefore precursers to continents when entering a subduction zone. The ability of continental crust to subduct has been documented in the coesite found in exhumed ultrahigh pressure terranes (Chopin, 2003). However, continental fragments, because of their geographic relation to continents, will most likely lead continents into continent-continent collision.

In terms of seismic crustal structure, there is too much variation within and between groups to determine whether a crustal profile belongs to an island arc, oceanic plateau and submarine ridge, or continental fragment (Fig. 9e). While the seismic velocity profiles of continental fragments do appear to best match the average continental crust profile, there is significant overlap between the velocity profiles of continental fragments and oceanic plateaus/submarine ridges (Fig. 9). Clearly, seismic velocity profiles should not be the sole basis for determining the nature of crustal composition of an unclassified region of anomalous crust on the ocean floor. One example is the re-

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cent finding of granite in deep sea drilling of Rio Grande Rise, that would reclassify that feature as a continental fragment rather than a submarine ridge (Corfield, 2013). We would argue that combining gravity measurements with seismic models can narrow the origin of an undetermined FAT crust, as also suggested by Barton (1986) for calculating densities directly from seismic values. Many regions of anomalous crust on the Arctic ocean floor have been identified as both continental fragments and oceanic plateaus because of the low constraints provided by only using seismic velocities to determine the crustal composition (Dove et al., 2010; Lebedeva-Ivanova et al., 2006; Artyushkov, 2010). When determining the true crustal nature, seismic, gravity, and geochemical studies should also be reinforced with tectonic reconstructions to gain insight on the geological history of an unknown FAT.

7.2 From FAT to accreted terrane

Accretionary orogens are built of accreted terranes that are hundreds of meters thick, characterized by thin-skinned deformation, and suture bound. In terranes where units from the entire crust of island arcs and oceanic LIPs are preserved, the remaining crustal thickness has been severly sheared and thinned. Although buoyancy is an enabling factor in crustal accretion at subduction zones, it is likely that accretion can occur because weak layers in the FAT crust enable detachments and shear zones to develop within the subduction zone as the crust is subducting. Recent geodynamic experiments show that if a weak zone or detachment fault is present within the crust of the subducting crustal region, whether it is an island arc, oceanic plateau, or continental fragment, accretion will occur and leave a severly thinned terrane (Afonso and Zlotnik, 2011; Tetreault and Buiter, 2012). In island arcs, possible delamination units are the felsic middle crust and the CMTL. Pre-existing weaknesses in island arcs produced by backarc rifting can also serve as detachment faults during subduction. The depth of the weak layer or detachment determines the amount of crust and the layers of crust that can be underplated (Tetreault and Buiter, 2012). Continental fragments also may contain pre-existing faults from their earlier rifting stage that could serve as detachment

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faults during subduction. And while there is no observed evidence for delamination of the ultramafic layer underplating oceanic plateaus, we infer that this layer could also act similar to the ultramafic layer found in island arcs, and serve as a decollement during accretion.

The crustal deficit of most accreted island arcs, oceanic plateaus, submarine ridges, continental fragments, and even seamounts suggests that a significant amount of crustal material is recycled back into the mantle. Perhaps, the foundering of the lower crust and CMTL of oceanic plateaus and island arcs, which is considered to be a major mechanism of terrane accretion, can account for the volumetric loss of crustal material (Stern and Scholl, 2010). Whether the ultramafic unit below the lower crust in many FATs is dense enough to create instability and delamination can be determined from laboratory studies of accreted ultramafic units. The ultramafic cumulates of the CMTL in island arcs are inferred to have higher densities than upper mantle dunites when calculated with the expected temperatures and pressures at lower crustal depths (Behn and Kelemen, 2006). Results from seismic anisotropy studies and crystal fractionation modelling of arc crustal magma development support the theory that the ultramafic high velocity layer under island arcs is often delaminated before or during accretion. And in the accreted Wrangellia oceanic plateau, seismic refraction studies of the crust do not show any high P wave velocities (Brennan et al., 2011; Ramachandran et al., 2006), which can be interpreted as loss of the ultramafic subcrustal layer. However, interestingly enough, combined gravity and seismic studies of modern island arcs, oceanic plateaus, and submarine ridges do not involve a high density unit between the crust and mantle (Larter et al., 2003; Grow, 1973; Magnani et al., 2009; Christeson et al., 2008; Gohl and Uenzelmann-Neben, 2001; Sallares et al., 2003; Recq et al., 1998; Walther, 2003; Sinha et al., 1981; Peirce and Barton, 1991; Hampel et al., 2004; Shulgin et al., 2011; Patriat et al., 2002), contrary to the laboratory-derived densities of the arc CMTL rocks. In addition, the ultramafic units below the lower crust could be a rheologically weak layer that leads to decollement-related underplating during subduction.

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Besides the crustal features of FATs, other factors that may influence terrane accretion are the thickness of the subduction zone interface, whether the subduction zone is accretionary or erosive, and slab pull forces. Numerical experiments have shown that a thin subduction interface will promote shearing of the FAT crust and accretion of the upper crustal layers (De Franco et al., 2008). The nature of the accretionary prism region can be either erosive or accretionary depending on the sedimentary and erosive fluxes (Clift and Vannucchi, 2004; Scholl and von Huene, 2010), and this will factor into whether crust is recycled back into the mantle or not. And finally, the force of the subducting slab drives subduction and most likely can overcome the buoyancy of small crustal units (Molnar and Gray, 1979; Cloos, 1993). In addition, eclogitization of the oceanic lithosphere will increase the negative buoyancy of the slab and even allow continental crust to subduct (Afonso and Zlotnik, 2011).

8 Conclusions

Regions of high topography and anomalous crust on the oceanic floor that encounter an active subduction zone are likely to become accreted terranes. These future allochthonous terranes (FATs) include island arcs, oceanic plateaus, submarine ridges, seamounts, continental fragments, and microcontinents. By comparing modern FATs to examples of accreted terranes, we can better constrain the quantities of crust that are subducted and the material parameters that contribute to accretion. We find that modern island arcs have an average crustal thickness of 26 km, oceanic plateaus and submarine ridges have an average thickness of 21 km, and continental fragments and microcontinents have an average crustal thickness of 25 km. Yet most accreted terranes of island arc, oceanic plateau, submarine ridge, seamount, and continental fragment affinity are on the order of meters to kilometers thick. In the cases where collision rather than accretion by underplating or scraping into the accretionary prism, accreted terranes are interpreted to be 25 to 40 km thick. The average crustal densities for is-

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land arcs is 2.79 g cm^{-3}, 2.84 g cm^{-3} for oceanic plateaus and submarine ridges, and 2.81 g cm^{-3} for continental fragments and microcontinents.

The different crustal structures of these FATs and their rheological differences can lead to various processes of accretion, including accretionary prism thrusting, underplating, and collision. Crustal slivers of island arcs typically underplate and accrete to the overriding continent. Subduction of oceanic plateaus and submarine ridges often leads to accretion by collision. Seamounts and submarine volcanics subduct easily if they are not incorporated into the accretionary prism. Continental fragments likely lead to collision rather than accretion via underplating as they are connected to passive margins. In addition to the buoyancy of FAT crust, weak crustal layers and delamination of the lower crust and subcrustal layers lead to accretion and formation of accreted terranes.

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Table 1. Island arc crustal thicknesses, including the crust mantle transition layer (CMTL). All thicknesses are taken from seismic interpretations except for the Tonga Arc, which was derived by gravity modelling.

Island Arc	thickness (km)	reference
Aleutian Arc	35–37	Shillington et al. (2004)
Aves Ridge	26	Christeson et al. (2008)
Bonin Arc (S. Izu Active Arc)	25	Takahashi et al. (2009); Kodaira et al. (2007b)
Chikogu Arc (SW. Japan)	30	Ito et al. (2009)
Daito Ridge	20–25	Nishizawa et al. (2005)
N. Izu Arc	26–32	Kodaira et al. (2007a)
S. Izu Rear Arc	18	Takahashi et al. (2009)
Japan (Honshu Arc)	26	Arai et al. (2009)
Japan (Chikogu segment)	30	Ito et al. (2009)
Kuril Arc	33	Nakanishi et al. (2009)
Kyushu-Palau Ridge	20	Nishizawa et al. (2007)
Lau-Colville Ridge	15	Karig (1970)
Leeward Antilles Arc	27	Magnani et al. (2009)
Lesser Antilles Arc	24	Christeson et al. (2008)
Lesser Antilles at Montserrat	26–34	Sevilla et al. (2010)
Luzon Arc	25–30	Yumul et al. (2008); Dimalanta and Yumul (2004)
Mariana Arc	18	(Calvert et al., 2008)
Mariana Arc	20	Takahashi et al. (2007)
W. Mariana Ridge	17	Takahashi et al. (2007)
New Hebrides Arc (Vanuatu)	27–28	Coudert et al. (1984); Ibrahim et al. (1980)
Ogasawara Ridge (Bonin Ridge)	21	Takahashi et al. (2009)
N. Ryukyu Arc	23–27	Nakamura and Umedu (2009)
S. Ryukyu Arc	29–44	Nakamura and Umedu (2009)
Solomon Islands	27	Miura et al. (2004)
South Sandwich Arc	20	Larter et al. (2003)
Sunda Arc	20	Kopp et al. (2002)
Tonga Arc	22.2	gravity modelling: Bryan et al. (1972)
average	26 ± 6	

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Table 2. Bulk crustal densities (in g cm^{-3}) of modern island arcs, determined from seismic velocities using different seismic velocity-density relationships. Crustal densities include the density of the CMTL. Bulk densities are also reported from studies where the authors combined gravity and seismic data to determine crustal density. References are (1) Holbrook et al. (1999), (2) Lizarralde et al. (2002), (3) Shillington et al. (2004), (4) Christeson et al. (2008), (5) Takahashi et al. (2009), (6) Kodaira et al. (2007a), (7) Ito et al. (2009), (8) Nakanishi et al. (2009), (9) Nishizawa et al. (2007), (10) Magnani et al. (2009), (11) Takahashi et al. (2007), (12) Leat et al. (2003), and (13) Crawford et al. (2003).

Island Arcs	Nafe–Drake	Christensen–Mooney	Christensen–Shaw	reported in the study
Aleutians[1]	2.70	2.73	2.73	
Aleutians[2]	2.81	2.81	2.83	
Aleutians[3]	2.97	3.02	3.05	
Aves Ridge[4]	2.77	2.71	2.70	2.70
Bonin Arc (S. Izu Arc)[5]	2.86	2.86	2.85	
Izu-Bonin Arc[6]	2.81	2.82	2.79	
S. Izu Rear Arc[5]	2.83	2.82	2.80	
SW. Japan[7]	2.77	2.80	2.76	
Kuril Arc[8]	2.65	2.62	2.51	
Kyushu-Palau Ridge[9]	2.83	2.83	2.84	
Leeward Antilles Arc[10]	2.73	2.71	2.63	
Lesser Antilles Arc[4]	2.76	2.76	2.70	2.66
Mariana Arc[10]	2.78	2.77	2.73	
W. Mariana Ridge[11]	2.65	2.57	2.46	
Ogasawara Ridge[5]	2.89	2.91	2.91	
South Sandwich Arc[12]	2.76	2.73	2.68	2.89
Tonga Arc[13]	2.80	2.79	2.75	
average	2.79 ± 0.08	2.79 ± 0.10	2.75 ± 0.14	

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Table 3. Crustal thicknesses of oceanic plateaus and submarine ridges. Thicknesses are derived from seismic studies unless otherwise noted. [1] The crustal thickness was extrapolated in the original study because the Moho was not imaged.

Oceanic plateaus and submarine ridges	thickness (km)	reference
Agulhas Plateau	20	Parsiegla et al. (2008)
S. Agulhas Plateau	25	Gohl and Uenzelmann-Neben (2001)
Alpha Ridge	38	Dove et al. (2010)
Broken Ridge	20.5	Francis and Raitt (1967)
Caribbean Plateau	10–20	Mann and Taira (2004); White et al. (1998); Mauffret and Leroy (1997)
Carnegie Ridge	13–19	Sallares et al. (2005, 2003)
Cocos Ridge	21	Walther (2003)
Crozet Plateau	17	Recq et al. (1998)
Del Cano Rise	17.5	Goslin et al. (1981)
Eauripik Ridge	16[1]	Den et al. (1971)
Faroe-Iceland Ridge	23	Bohnhoff and Makris (2004)
Hikurangi Plateau	16–23	gravity modelling: Davy et al. (2008)
N. Kerguelen Plateau	17	Charvis et al. (1995)
S. Kerguelen Plateau	21–25	Operto and Charvis (1995)
Laccadive Ridge	24	Gupta et al. (2010)
Madagascar Ridge	25	Sinha et al. (1981)
Madeira-Tore Rise	17–18	Peirce and Barton (1991)
Maldive Ridge (Chagos Laccadive)	15	Francis and George G. Shor (1966)
Malpelo Ridge	21	Marcaillou et al. (2006)
Manihiki Plateau	21.4[1], 25	Hussong et al. (1979); gravity modelling: Viso et al. (2005)
Marquesas Island	15–17	Caress et al. (1995)
Maud Rise	11–14	Orsted Satellite data: Kim et al. (2005)
Mozambique Ridge	22–24	(Konig and Jokat, 2010; Hales and Nation, 1973)
Nazca Ridge	18–21	Hagen and Moberly (1994); Woods and Okal (1994); Hampel et al. (2004)
Nightyeast Ridge	24	Grevemeyer et al. (2000)
Ogasawara Plateau	> 20	Kaneda et al. (2005)
Ontong Java Plateau	33	Miura et al. (2004)
Rio Grande Rise	11–12	gravity modeling: Mohriak et al. (2010)
Roo Rise	12–18	Shulgin et al. (2011)
Shatsky Rise	26	Gettrust et al. (1980)
Tuamotu Plateau	21	Patriat et al. (2002)
Wallaby Plateau	18	Mihut and Muller (1998)
Walvis Ridge	12.5	Chave (1979)
Zenith Plateau	18	Mihut and Muller (1998)
average	21 ± 4	

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Table 4. Bulk crustal densities of oceanic plateaus and submarine ridges. Bulk crustal densities (in g cm^{-3}) are determined from seismic velocities using different seismic velocity-density relationships. Bulk densities are also reported from studies where the authors combined gravity and seismic data to determine crustal density. References are (1) Parsiegla et al. (2008), (2) Gohl and Uenzelmann-Neben (2001), (3) Francis and Raitt (1967), (4) Sallares et al. (2003), (5) Recq et al. (1998), (6) Walther (2003), (7) Bohnhoff and Makris (2004), (8) Charvis and Operto (1999), (9) Operto and Charvis (1995), (10) Gupta et al. (2010), (11) Sinha et al. (1981), (12) Peirce and Barton (1991), (13) Hussong et al. (1979), (14) Caress et al. (1995), (15) Hales and Nation (1973), (16) Hampel et al. (2004), (17) Grevemeyer et al. (2000), (18) Miura et al. (2004), (19) Shulgin et al. (2011), (20) Den et al. (1969), and (21) Patriat et al. (2002).

Oceanic plateaus and submarine ridges	Nafe–Drake	Chistensen-Mooney	Christensen–Shaw	reported in the study
Agulhas Plateau[1]	2.85	2.85	2.84	
S. Agulhas Plateau[2]	2.82	2.80	2.75	3.03
Broken Ridge[3]	2.82	2.82	2.80	
Carnegie Ridge[4]	2.85	2.85	2.83	2.89
Cocos Ridge[5]	2.91	2.93	2.94	2.93
Crozet Plateau[6]	2.70	2.63	2.53	2.62
Faroe-Iceland Ridge[7]	2.82	2.83	2.80	
N. Kerguelen[8]	2.90	2.92	2.92	
S. Kerguelen Plateau[9]	2.76	2.76	2.71	
Laccadive Island[10]	2.87	2.89	2.88	
Madagascar Ridge[11]	2.89	2.89	2.89	2.89
Madeira-Tore Rise[12]	2.77	2.74	2.68	2.90
Malpelo Ridge[4]	2.91	2.90	2.91	2.86
Manihiki Plateau[13]	2.79	2.80	2.77	
Marquesas Island[14]	2.91	2.87	2.87	
Mozambique Ridge[15]	2.70	2.70	2.62	
Nazca Ridge[16]	2.88	2.89	2.89	2.88
Ninetyeast Ridge[17]	3.01	3.04	3.08	
Ontong Java Plateau[13]	2.85	2.87	2.85	
Ontong Java Plateau[18]	2.88	2.91	2.90	
Roo Rise[19]	2.75	2.74	2.68	2.75
Shatsky Rise[20]	2.96	2.97	3.00	
Tuamotu Plateau[21]	2.80	2.79	2.74	2.74
average	2.84 ± 0.08	2.84 ± 0.09	2.82 ± 0.13	2.85 ± 0.12

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Table 5. Crustal thicknesses of continental fragments from seismic studies unless otherwise noted.

Continental fragments and microcontinents	thickness (km)	reference
Alpha-Mendeleev	26	Lebedeva-Ivanova et al. (2011)
Bill Bailey Bank	26	Funck et al. (2008)
Bounty Platform	23	Grobys et al. (2007)
Bower's Ridge	25	Cooper et al. (1981)
Campbell plateau	24	Grobys et al. (2009)
Chatham Rise	22	Grobys et al. (2007)
Chatham Rise	20	gravity modelling: Davy et al. (2008)
East Greenland Ridge	9–11	Døssing et al. (2008)
Elan Bank	16	Borissova et al. (2003)
Exmouth Plateau	20	magnetotellurics: Heinson (2005)
Fairway Ridge	23	Klingelhoefer et al. (2007)
Falkland plateau	25–30	gravity modeling: Kimbell and Richards (2008)
Faroe Bank	27.5	Funck et al. (2008)
Faroe Islands	35–40	Richardson et al. (1999)
Flemish Cap	33	Gerlings et al. (2011)
Flemish Cap	30	Funck (2003)
Galicia Bank	22	González et al. (1999)
Hatton Bank	26.5	Fowler et al. (1989)
Hatton Bank	23	White and Smith (2009)
Jan Mayen	16	Breivik et al. (2012)
Jan Mayen	19	Kodaira et al. (1998)
Lomonosov Ridge	26	Jackson et al. (2010)
Lomonosov Ridge	26	Poselov et al. (2003)
Lord Howe Rise	25	Klingelhoefer et al. (2007)
Lord Howe Rise	29	Shor et al. (1971)
Lousy Bank	24	Funck et al. (2008)
Lousy Bank	25	Klingelhoefer et al. (2005)
Mendeleev Ridge	32	Lebedeva-Ivanova et al. (2006)
Norfolk rise	205	Klingelhoefer et al. (2007)
Norfolk rise	21.6	Shor et al. (1971)
Porcupine Bank	28	Whitmarsh et al. (1974)
Porcupine Bank	25	Morewood et al. (2005)
Rockall Bank	30	Vogt et al. (1998)
Rockall Bank	28.5	Morewood et al. (2005)
Sao Paolo Plateau	12–16	gravity modelling: Scotchman et al. (2010)
Seychelles	39	Collier et al. (2009)
average	24.8 ± 5.7	

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Table 6. Bulk densities ($g\,cm^{-3}$) of continental fragments and microcontinents determined from seismic velocities using various velocity-density curves. Bulk densities are also reported from studies where the authors combined gravity and seismic data to determine crustal density. References are (1) Funck et al. (2008), (2) Grobys et al. (2007), (3) Cooper et al. (1981), (4) Grobys et al. (2009), (5) Borissova et al. (2003), (6) Klingelhoefer et al. (2007), (7) Funck (2003), (8) Gerlings et al. (2011), (9) Fowler et al. (1989), (10) Breivik et al. (2012), (11) Lebedeva-Ivanova et al. (2006), (12) Morewood et al. (2005), (13) Vogt et al. (1998), and (14) Collier et al. (2009).

Continental fragments and microcontinents	Nafe–Drake	Christensen–Mooney	Christensen–Shaw	reported in the study
Bill Bailey Bank[1]	2.80	2.81	2.78	2.79
Bounty Platform[2]	2.83	2.86	2.87	
Bower's Ridge[3]	2.90	2.92	2.93	
Campbell Plateau[4]	2.78	2.79	2.75	
Chatham Rise[2]	2.82	2.83	2.85	
Elan Bank[5]	2.82	2.85	2.84	
Fairway Rise[6]	2.78	2.77	2.72	2.74
Faroe Bank[1]	2.79	2.81	2.77	2.77
Flemish Cap[7]	2.82	2.85	2.83	
Flemish Cap[8]	2.81	2.83	2.85	
Hatton Bank[9]	2.92	2.96	2.98	
Jan Mayen[10]	2.75	2.69	2.74	
Lord Howe Rise[6]	2.81	2.82	2.79	2.77
Lousy Bank[1]	2.79	2.79	2.76	2.79
Mendeleev Ridge[11]	2.84	2.85	2.82	
Norfolk Rise[6]	2.74	2.71	2.64	2.77
Porcupine Bank[12]	2.76	2.75	2.79	
Rockall Bank[13]	2.85	2.88	2.89	2.83
Rockall Bank[12]	2.79	2.80	2.82	
Seychelles[14]	2.89	2.94	2.92	2.86
average	2.82 ± 0.05	2.81 ± 0.08	2.83 ± 0.06	2.79 ± 0.04

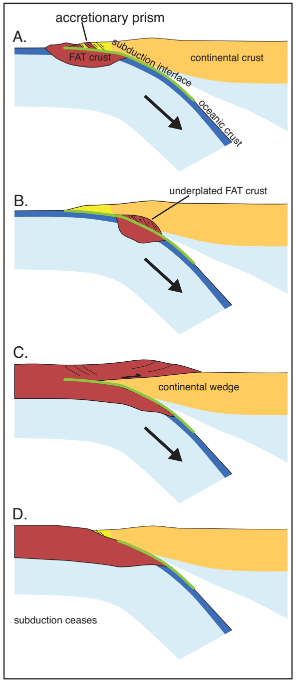

Figure 1. Cartoon schematics of FAT crust in subduction zones for four accretionary processes: **(A)** accretion in the accretionary prism, **(B)** subcretion, **(C)** flake tectonics, and **(D)** collision. In **(A)**, sediments and crustal units from the subducting oceanic plate and FAT are scraped off and accumulated in the accretionary prism in front of the forearc. The majority of the FAT crust is subducted. Subcretion **(B)** occurs below the accretionary prism, as crustal slices of the FAT are sheared and thrust onto the overriding continent. **(C)** Flake tectonics is the accretionary process where FAT crust is obducted onto the overriding continent, likely over a thick, strong prism of metasedimentary rocks in the overriding plate. **(D)** Collision will occur for large FATs, after some of the crust has subducted and accreted. The subducting slab will eventually detach.

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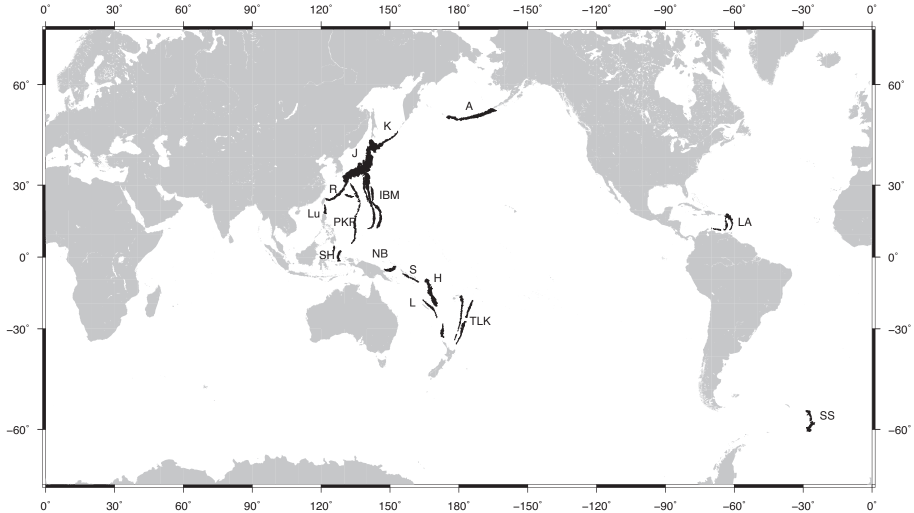

Figure 2. Location map of island arcs (shown in black) on the present day ocean floor. Arc systems labelled on the map are: A – Aleutians, H – New Hebrides, IBM – Izu-Bonin (Ogasawara)-Mariana arc system, J – Japan Arc, K – Kuril Arc, L – Loyalty Arc, LA – Lesser and Leeward Antilles, Lu – Luzon Arc, PKR – Palau-Kyushu Ridge, NB – New Britain Arc, R – Ryukyu Arc, S – Solomon Arc, SH – Sangihe-Halmera arc system, SS – South Sandwich Arc, and TKL – Tonga-Lau-Kermadec arc system.

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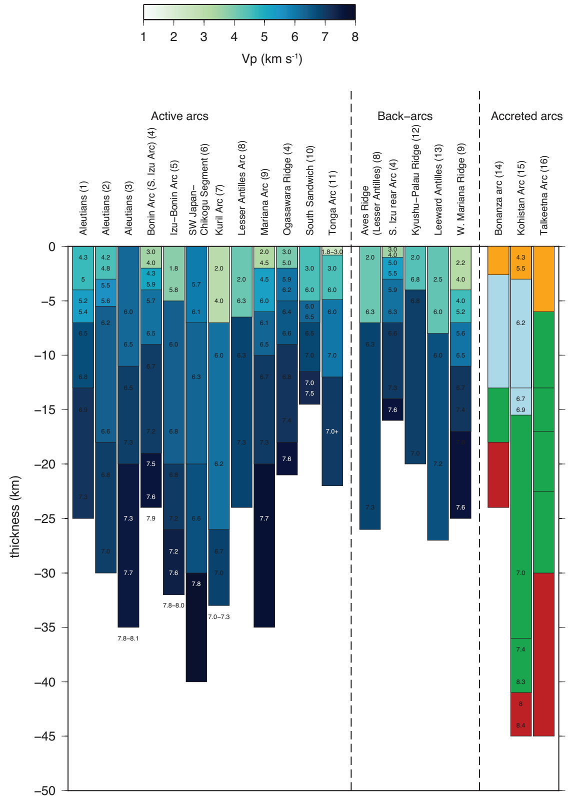

Figure 3. Seismic crustal structure of modern island arcs and calculated structure of accreted island arcs from previous studies. The thicknesses of units in the accreted arcs are calculated with geobarametric methods (Canil et al., 2010; Greene et al., 2006; Miller and Christensen, 1994) and the seismic velocities for the Kohistan units were measured in the lab (Miller and Christensen, 1994). For the accreted island arcs: orange represents upper crust, light blue represents middle crust, green is lower crust, and red represents the CMTL. References are (1) Holbrook et al. (1999), (2) Lizarralde et al. (2002), (3) Shillington et al. (2004), (4) Takahashi et al. (2009), (5) (Kodaira et al., 2007a), (6) Ito et al. (2009), (7) Nakanishi et al. (2009), (8) Christeson et al. (2008), (9) Takahashi et al. (2007), (10) Leat et al. (2003), (11) Crawford et al. (2003), (12) Nishizawa et al. (2007), (13) Magnani et al. (2009), (14) Canil et al. (2010), (15) Greene et al. (2006), and (16) Miller and Christensen (1994).

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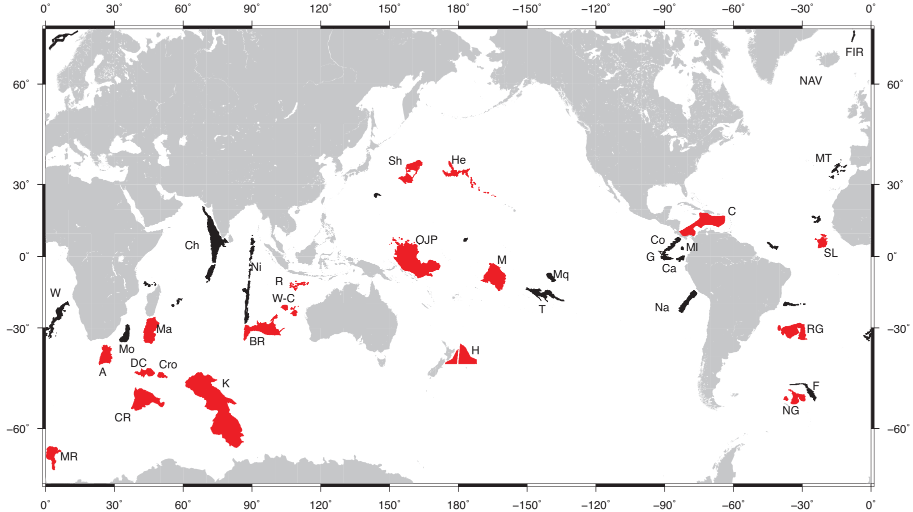

Figure 4. Location map of oceanic plateaus (shown in red) and submarine ridges (shown in black). Revised from LIP list of Coffin and Eldholm (1994) based on the definition of Bryan and Ernst (2008). Oceanic plateaus and submarine ridges labelled in the figure are: A – Agulhas Plateau, BR – Broken Ridge, C – Carribean Plateau, Ca – Carnegie Ridge, Ch – Chacos-Laccadive Ridge, Co – Cocos Ridge, CR – Conrad Rise, Cro – Crozet Bank, DC – Del Cano Rise, F – Faulkland Ridge, FIR – Faroe-Iceland Ridge, G – Galapagos Ridge, H – Hikurangi Plateau, He – Hess Rise, K – Kerguelen Plateau, M – Manihiki Plateau, Ma – Madagascar Ridge, ML – Malpelo Ridge, Mo – Mozambique Ridge, Mq – Marquesas Ridge, MR – Maud Rise, MT – Madeira-Tore Rise, Na – Nazca Ridge, Ni – Ninetyeast Ridge, NAV – North Atlantic Volcanic Province, NG – Northeast Georgia Rise, OJP – Ontong Java Plateau, R – Roo Rise, RG – Rio Grande Rise, Sh – Shatsky Rise, SL – Sierra Leone Rise, T – Tuamotu Plateau, W – Walvis Ridge, and W-C – Wallaby Plateau and Cuvier Plateau.

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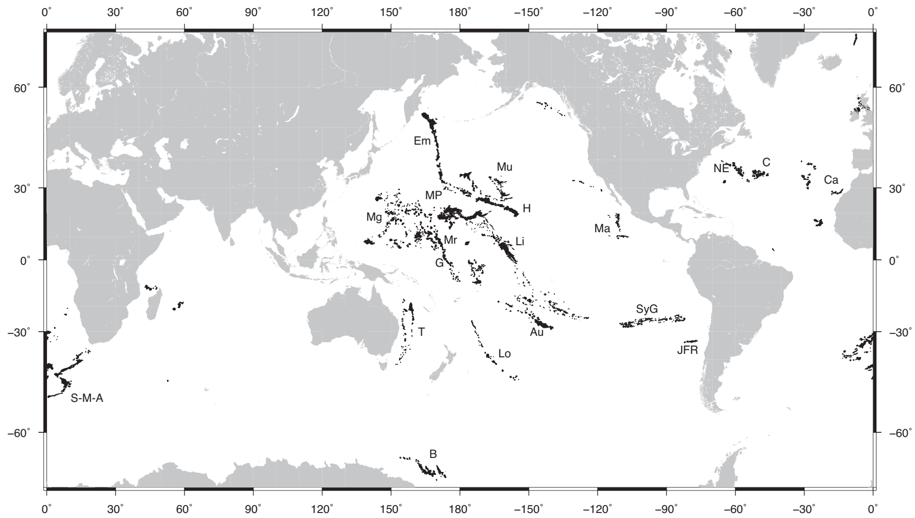

Figure 5. Location map of seamounts (shown in black). Revised from LIP list of Coffin and Eldholm (1994). Seamounts labelled are: Au – Austral Seamounts, B – Balleny Islands, C – Corner Seamounts, Ca – Canary Islands, Em – Emperor Seamounts, G – Gilbert Seamounts, H – Hawaii, JFR – Juan Fernandez Ridge, Li – Line Islands, Lo – Louisville Ridge, Ma – Mathematician Seamounts, Mg – Magellan Seamounts, Mr – Marshall Seamounts, Mu – Musician Seamounts, MP – Mid Pacific Mountains, NE – New England Seamounts, SyG – Sala y Gomez chain, S-M-A – Shona-Meteor-Agulhas Ridges, T – Tasmantid Seamounts.

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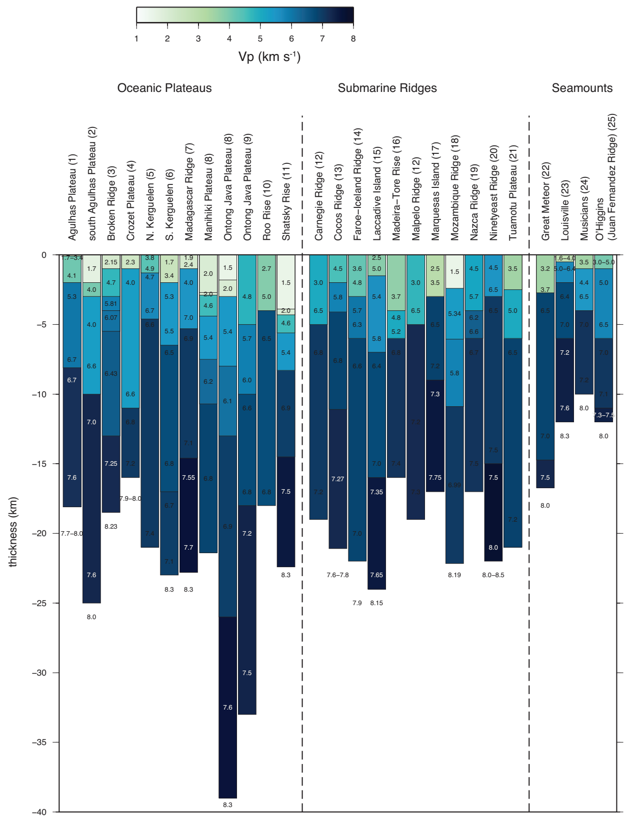

Figure 6. Crustal structures of modern oceanic plateaus and submarine ridges from seismic imaging studies. References are (1) Parsiegla et al. (2008), (2) Gohl and Uenzelmann-Neben (2001), (3) Francis and Raitt (1967), (4) Recq et al. (1998), (5) Charvis and Operto (1999), (6) Operto and Charvis (1995), (7) Sinha et al. (1981), (8) Hussong et al. (1979), (9) Miura et al. (2004), (10) Shulgin et al. (2011), (11) Den et al. (1969), (12) Sallares et al. (2003), (13) Walther (2003), (14) Bohnhoff and Makris (2004), (15) Gupta et al. (2010), (16) Peirce and Barton (1991), (17) Caress et al. (1995), (18) Hales and Nation (1973), (19) Hampel et al. (2004), (20) Grevemeyer et al. (2000), (21) Patriat et al. (2002), (22) (Weigel and Grevemeyer, 1999), (23) (Contreras-Reyes et al., 2010), 24 (Kopp et al., 2003), and (25) (Kopp et al., 2004).

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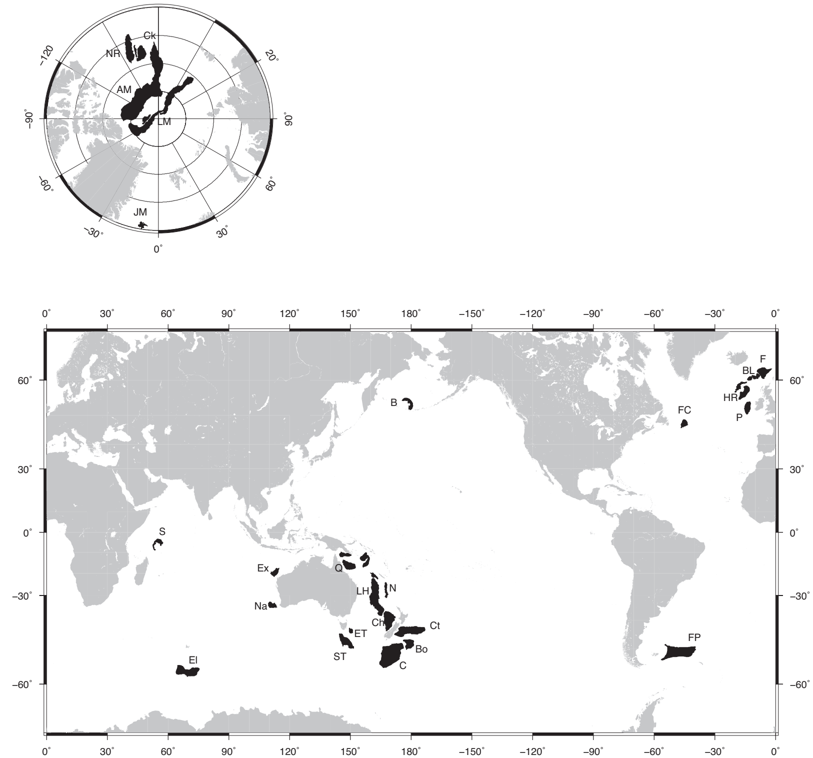

Figure 7. Location map of continental fragments (shown in black) compiled for this study. Continental fragments labelled are: AM – Alpha Mendelev Ridge, B – Bower's Ridge, Bo – Bounty Plateau, BL – Bill Bailey and Lousy banks, C – Campbell Plateau, Ch – Challenger Plateau, Ck – Chukchi Plateau, Ct – Chatham Rise, El – Elan Bank, Ex – Exmouth Plateau, ET – East Tasman Plateau, F – Faroe Bank, FC – Flemish Cap, FP – Falkland Plateau, HR – Hatton and Rockall Banks, JM – Jan Mayen, LH – Lord Howe Rise, LM – Lomonosov Ridge, Na – Naturaliste Plateau, N – Norfolk and Fairway Ridges, NR – Northwind Ridge, P – Porcupine Bank, Q – Queensland Plateau, S – Seychelles, and ST – South Tasman Plateau.

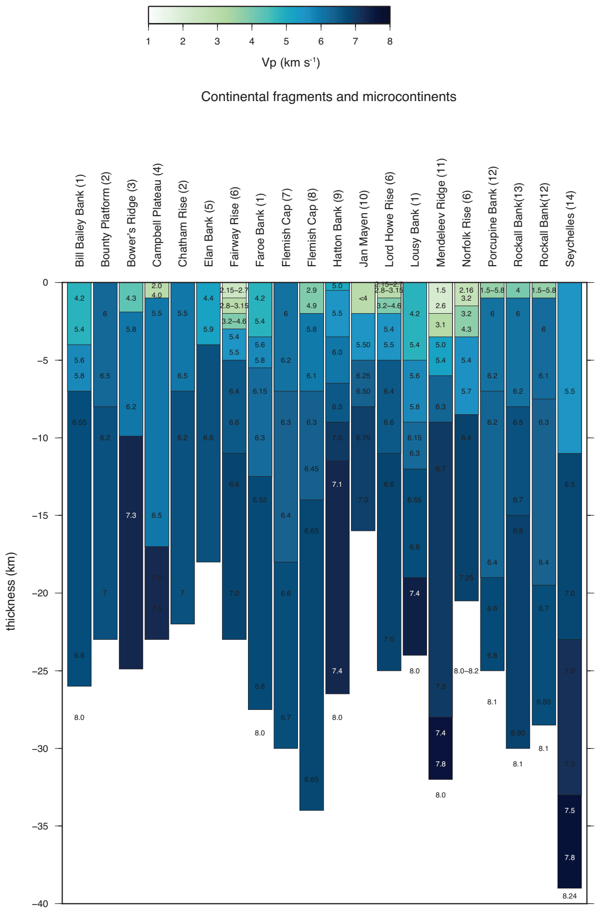

Figure 8. Seismic velocity profiles of modern continental fragments. References are (1) Funck et al. (2008), (2) Grobys et al. (2007), (3) Cooper et al. (1981), (4) Grobys et al. (2009), (5) Borissova et al. (2003), (6) Klingelhoefer et al. (2007), (7) Funck (2003), (8) Gerlings et al. (2011), (9) Fowler et al. (1989), (10) Breivik et al. (2012), (11) Lebedeva-Ivanova et al. (2006), (12) Morewood et al. (2005), (13) Vogt et al. (1998), and (14) Collier et al. (2009).

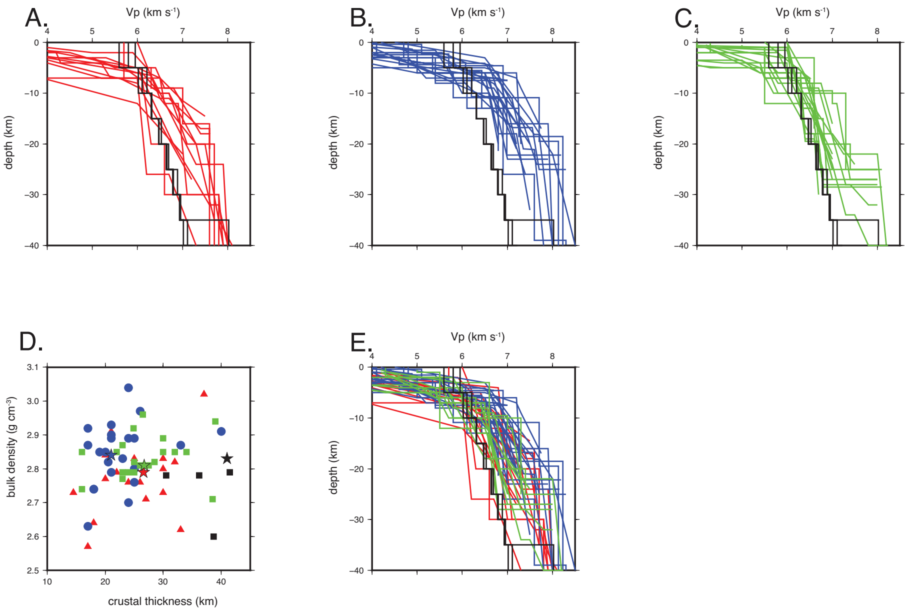

Figure 9. (A) Velocity profiles for island arcs (red), **(B)** oceanic LIPs (blue), **(C)** continental fragments (green), compared to the average velocity profiles of continental crust (black) from Christensen and Mooney (1995). **(D)** Bulk crustal density vs. crustal thickness for oceanic plateaus (blue circles), island arcs (red triangles), continental fragments (green squares) and continental crust (black squares). Average values for FATs and continental crust are plotted as stars. All densities are converted from seismic velocities using the relationships in Christensen and Mooney (1995). **(E)** Velocity profiles for all FATs plotted together.

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