Continental crust is basically a mosaic of orogenic belts, and these in turn are largely nests of island arcs, welded and melded together. Arcs are what is produced in the overriding plate of a convergent margin when subduction is sufficiently rapid (faster than $c.$ 2 cm a$^{-1}$) for long enough that the subducted plate reaches magmagenetic depths (100–150 km) and causes the mantle to melt. This configuration must exist for long enough that melts generated in the overlying asthenospheric wedge not only reach the surface but also persist until a stable magmatic conduit system is established. When this happens, magma produced by hydrox fluxing of the mantle will rise towards the surface to form volcanoes and plutons of a magmatic arc. This magmatic locus and its often spectacul vacanoes are important parts of an arc, but neovolcanic zones make up a relatively minor part of any arc. Arcs can be built on continental or oceanic lithosphere. Here the focus is on those arcs that are built on oceanic crust, which may be called intra-oceanic arc systems (IOASs). These can be distinguished from arcs built on continental crust, also known as ‘Andean-type arcs’.

IOASs are constructed on thin, mostly mafic crust, and consequently these magmas are not as contaminated by easily fusible felsic crust as are magmas from Andean-type margins, which are built on much thicker and more felsic continental crust (Fig. 1). IOASs are widely acknowledged as sites where thickened welts of juvenile (i.e. derived from melting of the mantle) crust is produced. In spite of this significance, IOASs are significantly less well studied than Andean-type arcs. The main reason for this is that the vast bulk of IOASs are submerged below sea level and difficult to study. Nevertheless, our understanding of IOAs has advanced greatly since these were first reviewed a quarter of a century ago by Hawkins et al. (1984); an overview of arcs in general was also prepared about that time (Hamilton 1988). New developments in marine technology [global positioning system (GPS), sonar swathmapping, deep-sea drilling, manned submersibles, remotely operated vehicles (ROVs) and autonomous underwater vehicles (AUVs)] are allowing the study of IOASs to advance rapidly, and we now have a good grasp of their most important features. The purpose of this paper is to summarize our understanding of IOASs for a broad geoscientific audience, with the hope that this understanding will allow geologists studying ancient crustal terranes to better identify fossil IOASs in orogens and cratons. This review draws heavily on 30 years of studying the Izu–Bonin–Mariana arc system, especially the Marianas, and examples are drawn heavily from this IOAS. The general tectonic and magmatic relationships observed there are mostly
applicable to the general case of modern IOAS formation and evolution.

What is an intra-oceanic arc system?

Arc–trench systems comprise the lithosphere (crust and uppermost mantle) between the trench and the back-arc region. IOASs are one endmember in the spectrum of arc–trench systems, the opposite end of the spectrum from Andean-type arcs. This spectrum reflects differences in what the arc is constructed on and thus its crustal composition, thickness, and elevation. IOASs are built on oceanic crust, whereas Andean-type arcs are built on pre-existing continental crust. Correspondingly, IOAS crust is thinner and more mafic than that beneath Andean-type arcs, which is thicker and more felsic (Fig. 1). This reduces the opportunity for primitive, mantle-derived magmas to interact with the crust en route to the surface; consequently, IOAS lavas are less differentiated, more mafic, and less contaminated than are lavas erupted from Andean-type arcs. The more primitive nature of IOAS igneous rocks better preserves evidence of the subduction-modified mantle-derived melts that formed them than do igneous rocks from Andean-type arcs, and this ‘window on subduction-zone processes’ is a paramount reason that studies of IOAS arcs are advancing rapidly (although it must be noted that all arc lavas tend to be more fractionated than are mid-ocean ridge basalt (MORB) or ‘hotspot’ lavas).

IOASs are the fundamental building blocks that are assembled over geological time into orogens, cratons, and continents. If the continental crust can be likened to a brick wall, then an IOAS plays the role of a single brick. This analogy has limited utility, because bricks in a wall and IOASs look and behave very differently. First, bricks are homogeneous, whereas IOASs are heterogenous, with significant vertical and transverse compositional variations, both in the crust and in the mantle lithosphere. Second, bricks are annealed solid aggregates, whereas IOASs form from melts and continue through their lives to interact with mantle-derived melts as well as generating eutectic crustal melts. Continued magmatic additions to the base of IOAS occurs across a ‘transparent Moho’, with some mantle-derived melts permeating or traversing the crust whereas others are underplated to the base of the crust, at the same time that dense cumulates and residues ‘delaminate’ or sink back into the convecting asthenosphere. Third, bricks are rectangular parallelepipeds, or cuboids, whereas IOAS crust has the relative dimensions of flattened noodles:

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![Fig. 1. Comparison of Andean-type arc (a) and intra-oceanic arc system (b), greatly simplified.](image-url)
wider (c. 250 km) than thick (15–35 km), much longer (hundreds to thousands of kilometres) than wide. Finally, bricks cemented into a wall are inert and immutable, whereas juvenile crust composed of accreted IOASs continues to deform as well as interact with and generate melts long after these are accreted. These melts, continuing deformation, and attendant metamorphism serve as the cement that welds different IOASs together.

IOASs are built on oceanic crust, but this crust generally forms when subduction begins, as discussed below. Crustal thicknesses are typically 20–35 km beneath the magmatic arc, thinning beneath the back-arc region and tapering trenchward. Detailed seismic studies of arcs are needed to actually measure crustal thicknesses, and these are relatively uncommon (the western Aleutian and Izu–Bonin–Mariana arcs are the IOASs with the best-imaged crustal structure; see Stern et al. 2003). However, the relative paucity of geophysical soundings of arcs does not seriously impede our ability to identify these among Earth’s inventory of convergent margins. Because crustal thickness typically is reflected in elevation, intra-oceanic arcs are generally mostly below sea level. This serves as a useful (and simple) criterion for identifying intra-oceanic arcs: if much of the volcanic front of a given arc system lies below sea level, it can usefully be described as an intra-oceanic arc. Similarly, if the volcanic front of a given arc mostly lies above sea level, this is probably an Andean-type arc. This twofold classification differs from the threefold subdivision of de Ronde et al. (2003), who identified: (1) intra-oceanic arcs (those with oceanic crust on either side); (2) transitional or island arcs (those along the margins of island chains with a basement of young continental crust); (3) continental arcs (those developed along the margins of continents). De Ronde et al.’s (2003) intermediate category of ‘island arc’ is the principal difference, and although I agree that arcs built on crust that is transitional between true oceanic and continental crusts exist (e.g. the Philippines), assignment to this intermediate group requires more information about the nature of crustal substrate than is commonly available. Figure 2 shows the distribution of IOASs as they are identified here.

IOASs tend to form where relatively old oceanic crust, generally Cretaceous or older, is subducted. Such lithosphere is negatively buoyant, which causes the subducting sea floor to sink vertically as well as subduct down-dip, in turn causing the trench (and the associated arc system) to ‘rollback’ (Garfunkel et al. 1986; Hamilton 2007). Such a situation favours development of convergent plate margins within the oceanic realm and thus IOASs. In contrast, subduction of young oceanic lithosphere engenders a strongly compressional convergent margin so that the arc system migrates away from the ocean basin and onto any flanking continent. Subduction of older oceanic lithosphere favours development of an overall extensional strain regime in the hanging wall of the associated convergent margin and is an important reason why IOASs tend to be associated with back-arc basins, discussed further below.

It is essential to distinguish between the early stages in the formation of an IOAS and its subsequent evolution. The latter stages are usefully referred to as ‘mature’, and all of the IOASs that

Fig. 2. Location of convergent plate margins and distribution of intra-oceanic arc systems.
are active today are in this stage. An IOAS changes somewhat once it becomes mature, but these changes are episodic as well as progressive. Episodic changes include the formation of back-arc basins and tectonic erosion of the forearc. Progressive changes include thickening of the crust beneath the volcanic–magmatic front, extent of serpentinitization of subforearc mantle, and the thickness of sediments that are accumulated. The early stages in the life of an IOAS are called ‘infant’, ‘nascent’, or ‘immature’ and few of the characteristics of mature IOASs pertain to this stage. The infant arc stage is relatively brief, lasting c. 5–10 Ma. Because none of the currently active IOASs are in this stage, our understanding of infant arcs is reconstructed from mostly early Cenozoic examples in the Western Pacific, especially the Izu–Bonin–Mariana arc. Except for the final section ‘IOAS forearc structure preserves its early history’, this review concentrates on mature IOASs.

How do we know about intra-oceanic arc systems?

Because arcs are large and heterogeneous geological entities, their study involves the full range of geoscientific perspectives: geochemistry, sedimentology, geophysics, geodynamics, structure, metamorphism, palaeontology, etc. Such studies of IOASs are more difficult in many ways than those of Andean-type arcs, simply because the former are largely below and the latter largely above sea level. Consequently, submarine geological studies are more expensive than studies on land. It is only when studying the trench and deeper parts of the forearc (which are submerged for both Andean-type and intra-oceanic arc systems) that the same marine geological approaches must be used. Also, Andean-type arcs lie near many population centres and pose serious volcanic, landslide, and seismic hazards (also benefits such as geothermal energy and mineral deposits), whereas IOASs are isolated, sparsely populated, and any mineral deposits are difficult to exploit. Because traditional land-based geoscientific approaches can be used to study Andean-type arcs, and because many nations rightfully have concerns for the public good, studies of such systems and our understanding of them are relatively advanced. Some geoscientific work on IOASs can be done above sea level but this is limited to the upper slopes of the tallest volcanoes and isolated structural highs. Understanding IOASs requires using research vessels and marine geotechnology, which are expensive. Our understanding of IOASs naturally lags behind that of Andean-type margins but is advancing rapidly.

Intra-oceanic arc-itecture

At broad scale, IOASs consist of four components, as shown in Figure 3: trench, forearc, volcanic–magmatic arc, and back-arc. These same components are just as useful for subdividing the transverse structure of Andean-type arc-trench systems. IOASs are often associated with back-arc basins, where sea-floor spreading occurs, or with narrower rift zones. Other IOASs show no evidence of extension, but no modern IOAS is associated with back-arc shortening. Further details about these four components are provided below.

Trench

The trench marks where the two converging plates meet. Because one plate bends down to slide beneath the other, trenches are global bathymetric lows, several thousands of metres deeper than sea floor away from the trench. A trench can be filled with sediments or contain very little sediment, depending on how much is supplied to it. Sediment flux reflects proximity to continents. Trenches associated with Andean-type arcs can receive large volumes of sediments delivered by rivers or glaciers and thus are often filled. In contrast, very few of the trenches associated with IOASs contain significant sediment. The only IOAS trenches with significant sediment fill are found in the southernmost Lesser Antilles, where the Orinoco River delta lies at the southern terminus of this trench, the Aleutians, where sediment from Alaskan glaciers and rivers flows longitudinally westwards along the trench, and in the Andaman–Nicobar region, where the trench is fed by the Ganges–Brahmaputra river system. The sediment volume in the trench controls whether or not the forearc is associated with a significant accretionary prism (Clift & Vannucci 2004; Scholl & von Huene 2007). Because most IOAS trenches are starved of sediment, accretionary prisms are not generally found in the inner trench.
Fig. 3. (Continued) intraoceanic arc (modified after Stern 2003). The asthenosphere is shown extending up to the base of the crust; delamination or negative diapirism is shown, with blocks of the lower crust sinking into and being abraded by convecting mantle. Regions where degassing of CO₂ and H₂O is expected are also shown.
wall of IOASs, with the exceptions noted above. Instead, igneous basement, generally basalt, boninite, diabase, gabbro and serpentinized peridotite, is exposed in the inner trench wall (Reagan et al. 2010). These rocks make up an \textit{in situ} ophiolite, produced when subduction began, as discussed below. Such exposures are important sources of information about the nature of forearc crust, along with drilling and geophysical studies (Clift & Vannucchi 2004; Scholl & von Huene 2007).

IOAS inner trench walls are highly fractured by continuing deformation and earthquakes associated with plate convergence. Subduction of seamounts and other bathymetric highs further fracture the inner trench wall. The cumulative effect is that much of the inner trench wall slope is defined by the angle of repose for fractured igneous rocks, especially at its base. The base of the inner trench wall consists of a talus prism of this material. This loose talus is carried into the subduction zone, resulting in significant ‘tectonic erosion’ of the outer forearc.

\textbf{Forearc}

The forearc is the lithosphere that lies between the trench and magmatic arc, generally 100–200 km wide. Its most important characteristic is a lack of recent igneous activity and remarkably low heat flow, even though forearc crust preserves strong evidence that igneous activity of unusual intensity occurred when subduction began. Morphologically, the forearc slopes gently towards the trench. It can be subdivided into a more stable inner forearc and a more deformed outer forearc. Tectonically stable forearcs, such as the Izu forearc, are also commonly deeply incised by submarine canyons. In contrast, actively deforming forearcs, such as the Mariana forearc, generally lack a well-developed canyon system.

Geophysical soundings of a typical IOAS forearc such as that of Izu–Bonin–Marianas reveal a pre dominantly mafic crust that tapers towards the trench, such that mantle peridotite is often exposed in the inner trench wall (Fisher & Engel 1969; Bloomer & Hawkins 1983; Pearce et al. 2000). \(P\)-wave velocities increase vertically downward from \(V_p\) consistent with fractured basaltic and boninitic lavas to \(V_p\) consistent with diabase and gabbro (Figs 4 & 5). This \(P\)-wave velocity structure is similar to that of oceanic crust or ophiolites, consistent with sampling of the IOAS inner trench walls. The significance of the ophiolitic nature of IOAS forearc crust is explored further in the final section ‘IOAS forearc structure preserves its early history’.

The upper mantle beneath IOAS forearc crust is serpentinized, as demonstrated by low seismic velocities and the presence of serpentine mud volcanoes in the outer forearc of at least one IOAS, the Marianas (Stern & Smoot 1998; Oakley et al. 2007). The extent of serpentinization seems to vary with distance from the trench; rocks are more serpentinized towards the trench, and less serpentinized away from it. This is shown by upper mantle \(P\)-wave velocities, which decrease significantly from \(V_p\) c. 8.1 km s\(^{-1}\) beneath the inner forearc to 6.4 km s\(^{-1}\) towards the trench. The greater serpentinization beneath the outer compared with the inner forearc makes the outer forearc weaker and easier to deform. Forearc mantle is composed of strongly depleted harzburgite, characterized by spinels with Cr-number \(\equiv 100Cr/(Cr + Al) > 50\) (Bloomer & Hawkins 1983; Parkinson & Pearce 1998; Okamura et al. 2006).

\begin{figure}[h]
\centering
\includegraphics[width=\textwidth]{fig4.png}
\caption{Simplified \(P\)-wave velocity structure beneath Izu forearc (roughly east–west line at 30°50' E) modified after Kamimura et al. (2002), with interpreted lithologies. Forearc crustal thickness decreases from c. 11 km near the volcanic arc to 5 km or less near the trench. Also, \(V_p\) is significantly lower in uppermost mantle beneath the outer forearc (6.4–6.8 km s\(^{-1}\) v. 7.3–8.0 km s\(^{-1}\)), probably reflecting greater serpentinization beneath the outer forearc.}
\end{figure}
Sediments are a relatively unimportant part of the forearc, but there are some. The relatively stable inner forearc may host a thin forearc basin, with a few hundred metres to few kilometres of sediment thickness. This paucity of sediment in IOAS forearcs is consistent with the fact that IOAS trenches are empty, and contrasts with the robust accretionary prisms and thick forearc basins characteristic of many Andean forearcs.

Volcanic–magmatic arc

The locus of continuing igneous activity in a mature IOAS defines the magmatic arc. This is recognized in modern IOASs as a linear or arcuate array of volcanoes that parallel the trench (England et al. 2004; Syracuse & Abers 2006). These are the largest and most productive volcanoes in a mature IOAS and are often the only feature that rises above sea level. These volcanoes are underlain by plutons and hypabyssal intrusions exposed by erosion in ancient, fossil IOASs. The trenchward limit of young igneous activity is referred to as the ‘volcanic front’ or ‘magmatic front’, and marks a steep gradient in heat flow, which is low towards the trench and high towards the back-arc region. The style of IOAS volcanism is fundamentally different from that of the other two great classes of oceanic volcanism, mid-ocean ridges and hotspots. Mid-ocean ridge volcanism is entirely volatile-poor tholeiitic basalt, which produces crust of nearly constant thickness (5–7 km) when magma fills the gap between two plates being pulled apart. Hotspot volcanism typically builds a linear chain of tholeiitic and/or alkalic lavas, with a progression of ages that increases in the direction of plate motion; generally only a volcano or two at one end of the hotspot chain are active. IOAS volcanoes, in contrast, remain relatively fixed with respect to the trench, so that the crust thickens at a rate of several hundred metres per million years as a result of the cumulative effects of magmatic addition at essentially the same place. Furthermore, IOAS magmas are relatively rich in silica and in volatiles, especially water, so that these eruptions are more violent and lavas are dominated by fragmental material, except in deep water, where the pressure suppresses violent degassing of magmas. This can sometimes be seen in the deposits of a growing IOAS volcano, as it grows from a base at 2–4 km below sea level to shallower water and then becomes an island. Eruptions from such a volcano are likely to change progressively with time from more effusive flows to increasingly fragmental as eruptions occur at increasingly lower $P$ environments over time.

Hydrothermal activity and ore deposits associated with IOASs are also distinct from those associated with other oceanic igneous settings. Hydrothermal mineralization at mid-ocean ridges is controlled by the heating of seawater by hot rocks, which sets up a hydrothermal circulation that also leaches metals from the fractured basalts as it passes through these; these dissolved metals precipitate when hydrothermal fluids vent on the sea floor. Such circulation and leaching also occurs in association with IOAS submarine volcanoes, but in addition, the volatile-rich nature of IOAS magmas contributes directly to mineralization when these magmas degas (de Ronde et al. 2003; Baker et al. 2005). There are thus fundamental differences expected for mineral deposits that are related to mid-ocean ridges and IOAs.

IOAS arc volcanoes are built on a platform that varies in depth but that is characteristically well below sea level. The depth of this platform depends mostly on crustal thickness, and is shallower above thicker arc crust and deeper above

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Fig. 5. Interpretation of the seismic structure of the crust and upper mantle of a typical intra-oceanic arc, modified after DeBari et al. (1999). This section is interpreted from the seismic refraction study of Suyehiro et al. (1996) for the Izu arc. The presence of a mid-crustal ‘tonalite’ layer should be noted.
thinner arc crust. Correspondingly, whether or not IOAS volcanoes rise above sea level depends not only on the size of the volcano (reflecting age and magma production rate) but also the thickness of the underlying crust. There is no characteristic spacing between volcanoes along the magmatic front (d’Ars et al. 1995). Where the sea floor has been imaged between volcanoes, there is little evidence of young lavas.

IOAS volcanoes erupt mostly basalts but these are distinct because they are characterized by vesicular, porphyritic, and fractionated. This is evidence that one or more magma reservoirs exist within the arc crust, as depicted in Figure 3c. Processing of magmas and crust in the middle crust is probably responsible for the formation of tonalitic middle crust. Thus IOAS crust thickens by addition of lavas on top as well as plutonism within the crust. IOAS crustal growth is discussed further below. Felsic magmas also erupt in IOASs, usually from large, submarine calderas. Such submarine calderas are now well documented in the Izu and Kermadec IOASs (Fiske et al. 2001; Graham et al. 2008). The origin of felsic melts in IOASs is controversial, and partly depends on the nature of lower arc crust, as discussed below.

IOAS magmatic arcs show strong asymmetries in the volume and composition of magmatic products. Melts along the volcanic–magmatic front reflect the highest degree of melting and much larger volcanoes compared with those at a greater distance from the trench.

**Back-arc basins and intra-arc rifts**

Convergent margins can show extension or contraction, or be strain-neutral. These strain regimes are most clearly manifested in back-arc regions. More often than not, IOASs are associated with active back-arc basins (BABs); such arcs include the Marianas (Mariana Trough; Fig. 7), Tonga–Kermadec (Lau–Havre Basin), New Britain (Manus Basin), Vanuatu (North Fiji Basin), Andaman–Nicobar (Andaman Sea), and South Scotia (East Scotia Basin), so this is shown as part of a typical IOAS in Figure 3a. Other IOASs are associated with extinct BABs, including the Lesser Antilles, Western Aleutians, and Izu–Bonin arcs. Active BABs form by sea-floor spreading, which can be fast (>10 cm a⁻¹) or slow (1–2 cm a⁻¹; Stern 2002; Martinez & Taylor 2003). Spreading results from extensional stresses that split the arc, mostly as a result of to trench rollback. Rifting to initiate a BAB can begin in the inner forearc, along the arc, or immediately behind the arc, as summarized in Figure 6. In any case, arc igneous activity may be extinguished temporarily as mantle-derived magmas are captured by the extension axis. The rifted part of the arc is progressively separated from the volcanic–magmatic front, forming a remnant arc that subsides as spreading continues (Karig 1972). BAB spreading produces sea floor with a crustal structure that is largely indistinguishable from that produced by mid-ocean ridges. Inter-arc rifts create significant basins within IOASs but no sea-floor spreading. These may or may not evolve into BABs; typical examples of inter-arc rifts are found in the Izu and Ryukyu arcs.

**IOAS sediments**

As mentioned before, IOASs are characterized by slow sedimentation rates and relatively thin sediments. Unless an IOAS lies near a continent, its trench will be empty and it will not have an accretionary prism. Significant sediment accumulations in IOASs occur only near the volcanic front, both on the back-arc side and forearc side. Sedimentation rates here will also reflect prevailing wind directions, which control the direction of volcanic ash dispersal and which flanks of subaerial volcanoes will be preferentially eroded by waves. Larger volcanoes are increasingly affected by flank collapse, which episodically sends tremendous volumes of sediment downslope.

Two sites of sedimentary basins in IOASs are worthy of further discussion: the forearc basin and the BAB immediately adjacent to the volcanic arc. Figure 8 shows typical examples of both, using a c. 350 km long multi-channel seismic reflection profile across the Mariana arc–trench system. Most IOASs have an elongated basin parallel to the volcanic arc on the inner forearc. Such forearc basins are 50–80 km wide (Chapp et al. 2008). Sediments in the Mariana forearc basin have a maximum thickness of 1.5 s (two-way travel time; Fig. 8). Deep Sea Drilling Project (DSDP) Site 458 drilled into the distal edge of this forearc basin, penetrating c. 250 m of Oligocene–Pleistocene sediments and into Eocene basement (Shipboard Scientific Party 1982b). Physical property measurements indicate $V_p < 2.0$ km s⁻¹ for these sediments, implying a maximum thickness of c. 1.5 km for this basin. These sediments have accumulated over the past 35 Ma, implying a maximum sedimentation rate of c. 43 m Ma⁻¹.

The other important sedimentary basin lies at the juncture between thin BAB crust and thicker crust beneath the volcanic–magmatic arc (Fig. 7). This results in a deep basin where arc-derived volcanlastic sediments are deposited. In the case of the Mariana Trough profile shown in Figure 8, the sediments are <0.75 s thick. DSDP Site 455 drilled c. 100 m into a similar sediment pile just to the north of this profile (Shipboard Scientific Party...
and DSDP Site 456 (about 20 km farther west, near the distal edge of the BAB sedimentary basin) penetrated 170 m of sediment and into BAB basaltic crust. Physical property measurements indicate $V_p \approx 2 \text{ km s}^{-1}$ for DSDP 455 sediments, indicating a maximum thickness of $\sim 750 \text{ m}$. These sediments have accumulated over the past 7 Ma (since the Mariana Trough back-arc basin began to open), implying a maximum sedimentation rate of $\sim 100 \text{ m Ma}^{-1}$. The greater sedimentation rate for the back-arc basin adjacent to the arc relative to the forearc basin sedimentation rate reflects the fact that this basin is closer to the active arc than the forearc basin, which is separated from...

Fig. 6. Three ways in which an IOAS can rift to form a back-arc basin, modified after Martinez & Taylor (2006). The initial slab and arc lithospheric geometry is shown at the top. The three sets of panels depict the evolving geometry and interaction with arc melt sources (tied to the subducting slab) and back-arc melt generation (tied to mantle advection driven by separation of the overriding lithosphere). Left panels show BAB formation by rifting behind the active magmatic arc, in which case arc volcanism will not be interrupted and a narrow, thin remnant arc is produced. The middle panels show BAB formation by rifting the active arc, whereby the arc magmatic budget will initially be captured by the extension axis, shutting down the arc and forcing re-establishment of a new magmatic arc. The right panels show BAB formation by forearc rifting, whereby the active magmatic arc will slowly migrate away from above the site of melt generation in the mantle and be extinguished, at the same time as a new magmatic arc forms above the site of mantle melt generation. In this case, a very broad and thick forearc is formed.
Neither basin shows sedimentation rates that are particularly impressive. Sediments in the Mariana forearc and back-arc basins are dominantly volcanogenic and derived from the active volcanic arc. These basins have sea floor that lies c. 3–4 km below sea level, mostly above the carbonate compensation depth so that carbonate sediments could be preserved, but biological productivity in this part of the ocean is low and biogenic sediments form a minor part of the fill of either basin. Forearc and back-arc basin sedimentation where biological productivity is higher (at high latitudes and equatorial regions) may have considerably larger proportions of biogenic components (Marsaglia 1995).

Fig. 7. Simplified tectonic map of the Mariana IOAS, showing the distribution of active arc volcanoes, back-arc basin spreading ridge (modified after Baker et al. 2008), frontal arc uplifted islands, remnant arc, major sedimentary basins, and serpentinite mud volcanoes (P. F. Fryer, pers. comm.). Location of boundaries between subducting Pacific plate (PA) and overriding Mariana plate (MA) and Philippine Sea plate (PSP) are also shown. Dashed line shows location of multichannel seismic profile shown in Figure 8; collinear fine continuous line shows position of lithospheric sounding profile in Figure 13. Box shows location of HMR-1 sonar image shown in Figure 9.
Volcanogenic sediments are transported from the active arc to forearc and back-arc basins probably by turbidity currents and by ash settling out of the water column. Many IOAS volcanoes are surrounded by submarine giant dunes, with wavelengths of up to 2 km, which define proximal sediment aprons that extend up to 60 km. The origin of these is poorly understood (Embley et al. 2006). Much other sediment must be transported by turbidites. Sonar backscatter images reveal well-developed, sinuous channels up to 100 km long, flowing downslope from the base of Mariana arc volcanoes towards the BAB to the west (Fig. 9). These may have evolved to transport turbidites produced by collapse of oversteepened arc volcano slopes, allowing mobilized sediments to move to depocentres in the eastern part of the Mariana Trough BAB. The relatively shallow depth of the BAB spreading ridge effectively partitions the two sides of any BAB into a sediment-starved portion that is distal to the active volcanic arc and an arc-proximal portion that is sediment-filled.

**IOAS melts, lavas, and plutons**

IOAS lavas are characteristically subalkaline and dominated by basalts and their fractionates. High-Mg andesites and rhyodacitic magmatic products also occur (Kelemen et al. 2003), with a minor amount of shoshonitic lavas in some cross-chains and along tectonically complex magmatic fronts. Boninites sometimes are formed when subduction zones begin. Tholeiitic basalts dominate BAB lavas, but the terms ‘tholeiitic’, ‘calc-alkaline’ and ‘high-alumina basalt’ all are used to describe the dominant IOAS volcanic front lavas. Arculus (2003) criticized especially the way that the term ‘calc-alkaline’ is commonly used, and proposed a new classification scheme for convergent margin igneous rocks. This scheme recognizes the importance of oxygen fugacity in controlling FeO/MgO v. silica variations in arc lavas as a complement to already established potassium–silica variations (high-Fe, medium-Fe, and low-Fe along with high-K, medium-K, and low-K suites).

The composition of these rocks reflect the unusual nature of subduction-related magmatic processes that make IOAS magmas. These begin with progressive metamorphism and perhaps melting of water-saturated subducted sediments, altered oceanic crust, and serpentinites as these are squeezed and heated in the subduction zone. Increasing pressure and heat closes fractures and pores, then dehydrates hydrous minerals. The result is a continuous release of water during progressive metamorphism to denser and drier minerals. IOAS lavas are characteristically subalkaline and dominated by basalts and their fractionates. High-Mg andesites and rhyodacitic magmatic products also occur (Kelemen et al. 2003), with a minor amount of shoshonitic lavas in some cross-chains and along tectonically complex magmatic fronts. Boninites sometimes are formed when subduction zones begin. Tholeiitic basalts dominate BAB lavas, but the terms ‘tholeiitic’, ‘calc-alkaline’ and ‘high-alumina basalt’ all are used to describe the dominant IOAS volcanic front lavas. Arculus (2003) criticized especially the way that the term ‘calc-alkaline’ is commonly used, and proposed a new classification scheme for convergent margin igneous rocks. This scheme recognizes the importance of oxygen fugacity in controlling FeO/MgO v. silica variations in arc lavas as a complement to already established potassium–silica variations (high-Fe, medium-Fe, and low-Fe along with high-K, medium-K, and low-K suites).

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the temperature of peridotite and metasomatize the mantle source with fluid-mobile trace elements (K, Rb, U, Pb, etc.), giving the hybridized mantle source the distinctive trace element signature of arc melts, as discussed below. Melting as a result of decompression may also be important (Plank & Langmuir 1988; Conder et al. 2002).

The trenchward limit of asthenospheric flow in the core of the mantle wedge determines the location of the volcanic–magmatic front (Cagnioncle et al. 2007). Because fluid flux from the subducted slab decreases with increasing depth, the interaction of asthenosphere with the greatest fluid flux maximizes melt generation; because fluid flux diminishes with greater depth to the subducted slab and distance from the trench, melt generation also decreases rapidly in this direction. The characteristic depth to the subduction zone of c. 110 km below the volcanic–magmatic front probably reflects the minimum thickness of overriding crust and mantle that is required to allow asthenosphere to circulate.

Strong asymmetries are seen in IOAS magmatic products. Igneous activity in the forearc occurs only early in the life of an IOAS, as discussed below. Lavas erupted along the magmatic–volcanic front are dominated by tholeiitic and high-Al basalts and calc-alkaline andesites (and their differentiates;
Rear-arc cross-chains are dominated by more enriched calc-alkaline and shoshonitic lavas. Back-arc basin basalts are tholeiitic, similar in most respects to MORB. These magmatic products are discussed in greater detail below.

Relative to other mantle-derived melts [MORB and ocean island basalt (OIB)], arc melts are enriched in silica, water, and large ion lithophile elements (LILE; e.g. K, U, Sr, and Pb). Figure 11 summarizes the distinctive trace element patterns of primitive (i.e. unfractionated) basalts from IOAS volcanic fronts. The trace element patterns observed for these lavas (and the derivatives of these melts, including fractionates and anatectic melts) show distinctive and similar trace element characteristics that testify to the processes shown in Figure 10, including high degrees of melting of hydrous metasomatized mantle in the spinel peridotite facies. First, the mantle melts more beneath a typical IOAS volcanic front than does the mantle associated with mid-ocean ridges or oceanic hotspots. This is shown by low Na₂O contents of primitive arc basalts (which vary inversely with extent of melting) and by low contents of incompatible trace elements that are not fluid-mobile [i.e. Ti, Zr, Nb, and heavy rare earth elements (HREE)]. For example, the dashed horizontal line approximates the abundances of fluid-immobile elements Zr to the HREE, which are c. 60% of those found in MORB. This implies that the mantle melts c. 70% more beneath IOAS volcanic fronts than it does.

Kelemen et al. (2003). Rear-arc cross-chains are dominated by more enriched calc-alkaline and shoshonitic lavas. Back-arc basin basalts are tholeiitic, similar in most respects to MORB. These magmatic products are discussed in greater detail below.

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beneath mid-ocean ridges. Second, the greater melting beneath arcs reflects the important role that water plays in mantle melting and the much greater abundance of water in IOAS magmas relative to MORB or even hotspot basalts (Stern 2002). IOAS volcanic front lavas are typically coarsely porphyritic, consisting of phenocrysts of An-rich plagioclase and Fe-rich olivine (Fig. 12). This is best explained by high magmatic water contents suppressing the crystallization of plagioclase and rotating its solvus. IOAS volcanic front lavas are typically much more porphyritic than MORB or OIB, consistent with an interpretation that loss of magmatic water during ascent drove rapid crystallization. Third, elevated abundances of fluid-mobile elements such as LILE, shown as positive concentration spikes in Rb, Ba, U, Pb, and Sr (Fig. 11), are best explained as resulting from metasomatism of the mantle source region by hydrous fluids or melts. Finally, the flat HREE patterns shown by IOAS basalts are inconsistent with residual garnet in the mantle source region, instead suggesting melt generation in the spinel peridotite stability field. Spinel peridotite is stable at pressures corresponding to depths of 15–80 km, consistent with the expected position of the asthenosphere in the mantle wedge.

Primitive IOAS magma rises into the arc crust, where it may be temporarily stored in magma reservoirs, differentiate and mix with other magmas. The tremendous heat associated with the magmatic front also can lead to crustal anatexis and delamination to generate more felsic arc crust. The twin processes of magmatic fractionation and anatexis, responsible for producing IOAS felsic melts, occur in the crust.

To a certain extent, IOAS magmas reflect the arc’s evolutionary stage as well as depth to the subducting slab. Early in the history of an IOAS, lavas are dominated by high-degree melts such as boninites and tholeiites (and their differentiates). It is not yet clear whether melts of the magmatic front evolve systematically as the arc ages. The Izu–Bonin–Mariana IOAS shows no systematic compositional evolution once the magmatic front stabilized c. 40 Ma ago (Straub 2003), whereas the Greater Antilles IOAS seems to have evolved from early tholeiites to later calc-alkaline and then shoshonitic lavas (Jolly et al. 2001). Systematic changes in melt composition may reflect changing upper plate stresses (lithospheric extension v. compression), or may be related to evolving crustal thickness (thicker crust encourages melt stagnation, enhancing fractionation and anatexis), or thermal conditioning (mature arcs with thick crust are

![Composition ranges for olivine-plagioclase assemblages for intra-oceanic arc and back-arc basin basalts](image-url)
more likely to experience anatexis around magmatic conduits and reservoirs). Systematic changes in the age of subducting lithosphere could also change IOAS active arc melt compositions, because subduction of very young (hot) oceanic crust (<20 Ma old) can melt to produce adakites (Defant & Drummond 1990). Because IOASs are concentrated where old, dense oceanic lithosphere is subducted, this mode of magmagenesis is not a significant aspect of modern IOASs.

IOAS igneous activity in mature systems is concentrated at the volcanic–magmatic front but can also occur farther away from the trench, along rear-arc or cross-chain volcanoes and in back-arc basins. These two magmatic regimes are very different. Rear-arc volcanoes are common in some IOASs such as the Izu, Bismarck, or Kurile arcs but not in others such as Tonga–Kermadec. Where cross-chains occur they comprise a few to several smaller volcanoes that extend away from a large volcanic front edifice at high angles. Rear-arc volcanoes erupt lavas that are more primitive (Mg-number >60) than volcanic front lavas. These rear-arc lavas are also less porphyritic and do not show the plagioclase–olivine compositional relationships characteristic of IOAS volcanic front lavas. These characteristics, along with the generally more enriched nature of cross-chain lavas, reflect diminished extent of melting as a result of lower fluid flux from the slab with greater distance from the trench. This is the basis for the K–h relationship, which describes the relative enrichment in K and other incompatible elements in arc lavas as a function of depth to the underlying subduction zone (Dickinson 1975; Kimura & Stern 2008).

The final locus of IOAS igneous activity occurs in BABs. Back-arc basin magmas are dominated by tholeiitic basalts that range from MORB-like to arc-like (Pearce & Stern 2006). BAB basalt (BABB) magmas are typically more hydrous than MORB (1–2% v. <0.5% water; Stern 2002). Compositions of BABBs result from four factors: (1) the composition of asthenosphere flowing into the region of melting beneath the BAB spreading ridge; (2) the composition and relative impact of a fluid- or melt-dominated subduction component; (3) how asthenosphere and subduction components interact; (4) the melting of water-rich mantle and the assimilation–crystallization history of the resulting hydrous magma. Trace element and water contents of BAB glasses indicate that decompression melting (similar to that beneath mid-ocean ridges) beneath back-arc basins is augmented by flux melting as a result of the abundance of hydrous fluids.

BAB spreading is generally asymmetric, with spreading axes often lying closer to the active volcanic arc than the remnant arc. BABs have relatively short lifespans compared with volcanic arcs, up to a few tens of million years. This is because as a BAB matures and widens the neovolcanic zone increases in distance from the trench, so that the water flux into the melting region decreases with time. Correspondingly, magma production diminishes over a BAB’s life. Such a progression is observed for the 30–15 Ma old Parece Vela Basin, which progressed from magmatic spreading to amagmatic rifting before spreading stopped completely (Ohara 2006). These controls on melting and extension lead to the cessation of BAB spreading after a few tens of million years.

Because BAB basalts and gabbros have compositions and crustal structures that are similar to those of typical oceanic crust, but with trace element enrichments only found in convergent margins, they are often thought to be where ophiolites with ‘supra-subduction zone’ (SSZ) compositions form (Dilek 2003; Pearce 2003). Some ophiolites may originate in back-arc basins but most probably form in forearcs, where they are in a tectonic position that is much more likely to be emplaced on top of buoyant crust (obducted) if and when this arrives at a subduction zone (Stern 2004). The presence of boninites, which are common in forearcs but rare in BABs, may help resolve such uncertainty, but boninites are found only in some forearcs. Correspondingly, the absence of boninites in an SSZ ophiolite should not be taken as necessarily indicating that the ophiolite formed in a BAB instead of a forearc. Another way to distinguish BAB from forearc crust is that BAB crust forms at intervals after the magmatic arc, whereas forearc crust forms before the magmatic arc.

The nature of IOAS lower crust and upper mantle

The ‘classic’ Moho is strictly defined on the basis of seismic velocities to separate crust ($V_p \approx 6–7 \text{ km s}^{-1}$) from upper mantle ($V_p > 8 \text{ km s}^{-1}$). This generalization does not hold for IOASs, where such Moho is rare beneath the forearc and magmatic front, although the sub-BAB Moho is clear. The variable expression of Moho beneath IOAS magmatic front, forearc, and back-arc basins reflects different compositions and alteration histories. Simply put, forearc peridotites are dominated by harzburgite (Parkinson & Pearce 1998; Pearce et al. 2000) and are most serpentinized; peridotites associated with the active magmatic arc contain abundant pyroxenite and pyroxene-rich lithologies in addition to lherzolite and harzburgite; and back-arc basin peridotites are mostly lherzolites, similar
in most regards to those associated with slow-
spreading mid-ocean ridges.

Because the BAB Moho is the best defined, we
discuss the composition of the upper mantle below
this first. Until recently we had no samples of
BAB peridotite to study but we now have had a
chance to study peridotites from the northern
Mariana Trough (Ohara et al. 2002) and the
Parece Vela Basin (Ohara 2006). The Parece Vela
Basin formed in Miocene time by fast to interme-
diate rates of sea-floor spreading and includes residual
herzolite and harzburgite, along with plagioclase-
bearing harzburgite and dunite. These peridotites
are similar to those of the Romanche Fracture
Zone in the Mid-Atlantic Ridge and Indian and
Arctic Ocean slow-spreading ridges. In contrast,
the Mariana Trough is a typical slow-spreading
ridge and it locally yields residual harzburgite that
is also similar to that recovered from slow-spreading
mid-ocean ridges. These two examples of BAB
peridotites reveal different melt–peridotite inter-
actions. The Parece Vela peridotites were affected
by diffuse porous melt flow and pervasive melt–
mantle interaction. Abundant dunites represent
crystallisation of melt conduits, where melt and pyroxene reacted to
form olivine. In contrast, the Mariana Trough peri-
dotites are characterized by channelled melt–fluid
flow and limited melt–mantle interaction.

Forearc peridotites (recovered from inner trench
walls) are mostly harzburgites that are more
depauperated than abyssal peridotites from mid-ocean
ridges. Forearc peridotites generally are the most
depauperated peridotites that form on Earth today
(Bonatti & Michael 1989). This is shown by their
low Al2O3 and elevated SiO2, reflecting the fact
that these are more orthopyroxene-rich than resi-
dues of fertile mantle peridotite (Herzburg 2004).
Herzburg (2004) suggested on this basis that they
were produced by melt–rock reaction. The extreme
depauperation of forearc peridotites is also shown by the
composition of their spinels. Spinel is sensitive to
depauperation, which is reflected in Cr/(Cr + Al).
Spinel is also very resistant to alteration and gener-
ally preserves petrogenetic information in spite of
significant serpentinization (Dick & Bullen 1984).
Forearc peridotites have high-Cr-number (Cr/
(Cr + Al) > 0.4) spinel. These forearc peridotites are variably serpentinized, being more altered near
the trench, and less altered towards the magmatic
front. Serpentinization makes seismic recognition
of the Moho difficult.

Multiple perspectives are needed to understand
the mantle beneath IOAS magmatic arcs. One per-
spective comes from rare mantle xenoliths brought
up in arc lavas. Arai & Ishimaru (2008) summarized
these for West Pacific arc lavas. These are mostly
from arcs built on continental crust but still
provide an important glimpse of the mantle
beneath the magmatic front. Almost all the perido-
tite xenoliths are spinel-bearing varieties without
garnet or plagioclase. Arai & Ishimaru (2008)
emphasized that mantle metasomatism is most perv-
asive and long-lived beneath the volcanic front,
where magma transit is very long-lived. Metasoma-
tism here reflects the importance of silica-rich,
hydrous melts, and is likely to be manifested in the
formation of secondary orthopyroxene at the
expense of olivine. A large range of fertility and
depauperation is seen in arc xenoliths. Some are more
depauperated than the most fertile abyssal iherzolites,
whereas others are more depauperated than the most
depauperated abyssal peridotite. This lithological vari-
ation is probably due to the complex tectonic
history of each as well as to across-arc variation in
magma production conditions.

A second useful insight into the nature of the
lower crust and upper mantle beneath the magmatic
front comes from active source geophysical soun-
dings of IOAs; investigations of the central Aleutian
IOAS are particularly useful. Fliedner & Klemperer
(2000) interpreted these results to indicate c. 30 km
thick crust underlain by heterogeneous uppermost mantle
(Vp = 7.6–8.2 km s−1). Shillington et al. (2004)
reinterpreted the same dataset to indicate
that the crust was 20% thicker than concluded
from the earlier study. They interpreted the
regions with velocities of 7.3–7.7 km s−1 as lower
crust made up of ultramafic–mafic cumulates and/garnet granulites. They also found upper mantle velocities of 7.8–8.1 km s−1 at greater depth, sig-
nificantly higher than that inferred from the Fliedner
& Klemperer study. A similar situation is seen for the
Izu–Bonin–Mariana IOAS. High-resolution
seismic profiling (Kodaira et al. 2007; Takahashi
et al. 2007) reveals an unusually low-velocity
(7.2–7.6 km s−1) material in the lower crust–
upper mantle transition zone. This is interpreted at
present as an uppermost mantle layer between the
Moho above and reflectors in the upper mantle
below. These velocities are best explained by a
broad crust–mantle transition zone that is rich in
pyroxene and/or amphibole.

P-wave velocity structure of the Mariana IOAS
and complementary interpretation is shown in
Figure 13 (Takahashi et al. 2008). Takahashi et al.
(2008) found that the P-wave mantle velocity
beneath the magmatic front is 7.7 km s−1, markedly
less than the normal upper mantle Vp of 8.0 km s−1.
Based on its continuity, Takahashi et al. (2008)
identified the base of the 6.7–7.3 km s−1 layer as
Moho. It should be noted that the sub-Moho reflec-
tors lie beneath the Mariana arc magmatic front, the
West Mariana Ridge (remnant arc), and the Mariana
Trough spreading axis, in a part of the mantle that is
characterized by low Vp (<8 km s−1). These reflec-
tions might indicate rising magma bodies in the
upper mantle, or ghosts of dense lower crust that has sunk back into the mantle. The slow mantle region does not extend down to the upper mantle reflector. The 8.0 km s\(^{-1}\) velocity contour lies between the Moho and the deep reflector, which Takahashi et al. (2008) inferred has a velocity contrast of about 0.7 km s\(^{-1}\). These observations are consistent with an interpretation that mantle peridotite dominates beneath the arc at depths >40 km, but the shallower mantle contains a significant proportion of pyroxenite and/or amphibolite.

The velocity structure of the Mariana IOAS is similar to that of other IOASs, at least those that have been studied in similar detail. Figure 14 compiles the P-wave velocity structure for 10 sites of three IOASs. These velocity profiles are beneath the active arc, except for one profile beneath the Mariana forearc. Comparison of the velocity structure beneath the Mariana active arc–forearc pair suggests that the forearc structure is c. 0.25–0.5 km s\(^{-1}\) slower than the active arc, but it is not yet known whether this is a systematic difference between forearc and active arc crustal structure. Most of the profiles show increases to \(V_p\) c. 7.8 km s\(^{-1}\) or more, interpreted to correspond to upper mantle and thus marking the Moho, at depths between 25 and 35 km. This is where IOASs are thickest, but is nevertheless significantly thinner than continental crust, which is typically c. 40 km thick. IOAS crust also differs from continental crust in being significantly faster (by 0.5–1.0 km s\(^{-1}\)) than typical continental crust. This indicates that IOAS crust is more mafic than continental crust.

A complementary perspective comes from fossil arcs in orogens, where deep crust and upper mantle
which are significantly faster (by 0.5–1.0 km s$^{-1}$) than typical continental crust (shown with 1 standard deviation, from Christensen & Mooney 1995). All profiles are beneath active arcs except Mariana forearc, where upper crustal P-wave velocities are 240–360 m s$^{-1}$ slower than beneath the active Mariana arc at equivalent depths.

are often exposed. The best studied examples are Mesozoic IOASs of Talkeetna (Alaska) and Kohistan (Pakistan). Reconstruction of the fossil (Jurassic) Talkeetna arc by Hacker et al. (2008) indicated c. 35 km crustal thickness. Hacker et al. (2008) reconstructed the following major layers, from top to bottom (Fig. 15): (1) a c. 7 km thick volcanic section (Clift et al. 2005); (2) intermediate plutonic rocks (tonalites and quartz diorites) at 5–12 km depth; (3) mafic metamorphosed gabbroic rocks from 12 to 35 km depth. These metagabbroic rocks include: (1) hornblende gabbro-norites down to c. 25 km; (2) garnet diorites, first appearing at c. 25 km depth; (3) garnet gabbro, spinel-rich pyroxenite and orthogneiss, and hornblende gabbro-norite at the base of the arc at c. 30–35 km depth (Fig. 15). The upper mantle beneath Talkeetna consists of abundant pyroxenites, consistent with inferences from geophysical studies of the Aleutians and Izu–Bonin–Mariana IOASs. The abundance of pyroxenite in the upper mantle beneath the active magmatic arc contrasts markedly with the more depleted, harzburgitic mantle beneath the forearc. The Border Ranges ultramafic complex fragments in the McHugh mélange complex were suggested by Kusky et al. (2007) to be the ophiolitic forearc to the Talkeetna IOAS.

The upper mantle beneath the Cretaceous Kohistan palaeo-arc also contains abundant dunite, wehrlite, and Cr-rich pyroxenite in the inferred Moho transition zone [Jijal complex; (Garrido et al. 2007, fig. 8) at 162, 260, 307, 365, and 435 km. M, depth range of Moho. A noteworthy feature is the broad similarity of the 10 profiles, all of which are significantly faster (by 0.5–1.0 km s$^{-1}$) than typical continental crust (shown with 1 standard deviation, from Christiensen & Mooney 1995). All profiles are beneath active arcs except Mariana forearc, where upper crustal P-wave velocities are 240–360 m s$^{-1}$ slower than beneath the active Mariana arc at equivalent depths.

Fig. 14. Comparison of average P-wave velocities at 10 sites from three intra-oceanic arcs (Mariana, Izu–Bonin, and Aleutian), modified after Calvert et al. (2008). Profiles for Mariana are from Calvert et al. (2008); for Aleutian 1 after Holbrook et al. (1999); Aleutian 2 after Fliedner & Klemperer (2000); Aleutian 3 after Shillington et al. (2004); and five Izu–Bonin arc profiles are from Kodaira et al. (2007, fig. 8) at 162, 260, 307, 365, and 435 km. M, depth range of Moho. A noteworthy feature is the broad similarity of the 10 profiles, all of which are significantly faster (by 0.5–1.0 km s$^{-1}$) than typical continental crust (shown with 1 standard deviation, from Christiensen & Mooney 1995). All profiles are beneath active arcs except Mariana forearc, where upper crustal P-wave velocities are 240–360 m s$^{-1}$ slower than beneath the active Mariana arc at equivalent depths.
lower than the geotherm (Fig. 16d) and remelting of large parts of the lower crust seems inevitable. Fractional crystallization coupled with anatexis in the lower arc crust seems likely, and dense residues and fractionates are likely to sink back into the upper mantle. In support of this idea, Brophy (2008) concluded from consideration of REE modelling that felsic magmas in IOASs could be produced either by fractional crystallization (with or without hornblende) or by amphibolite melting, and that fractional crystallization seemed to be the dominant mechanism.

The Talkeetna fossil arc again provides important insights into the fate of lower crust beneath IOAS magmatic fronts. Field studies show that the great volumes of mafic and ultramafic cumulates (gabbronorite and pyroxenite) expected to have been produced by fractionation of primitive, mantle-derived melts are missing from the lower crust. Behn & Kelemen (2006) argued that lower crustal \( V_p > 7.4 \text{ km s}^{-1} \) in modern arcs indicates material that is gravitationally unstable relative to underlying asthenosphere; such material is likely to sink into the mantle. They also noted that the lower crust beneath modern arc magmatic fronts has \( V_p < 7.4 \text{ km s}^{-1} \), indicating that pyroxene-rich lower crust is generally not present. These observations support the conclusion that large volumes of lower arc crust founder rapidly on geological time scales, or that a significant proportion of high-\( V_p \) cumulates form beneath the Moho (Behn & Kelemen 2006).

Based on petrological modelling and \( V_p \) estimates for sub-Izu–Bonin–Mariana crust and mantle, Tatsumi et al. (2008) inferred that the uppermost mantle low-\( V_p \) layer was composed of mafic,
pyroxene-rich material that had sunk back into the mantle, not mantle peridotite. It appears that igneous activity beneath IOAS active magmatic arc evolves from mafic to intermediate as a result of continuous loss of mafic and ultramafic lower crustal components to the subarc mantle across a chemically transparent Moho (Tatsumi et al. 2008). This process will slow crustal thickening at the same time that increasingly felsic crust comes to make up the evolving arc crust, leading to a crustal thickening rate of \( c.500 \text{ m Ma}^{-1} \). When IOAS crust thickens sufficiently to stabilize garnet in the lower crust (greater than \( c.30 \text{ km thick} \), magma compositions will become more adakitic at the same time that delamination and crustal processing are likely to be enhanced [as a result of the greater density of garnet (3.6 g cm\(^{-3}\)) relative to pyroxene (3.2 g cm\(^{-3}\))]. This model is summarized in Figure 17.

**IOAS forearc structure preserves early history**

A final important point about IOASs concerns how they form, how the forearc is built and abandoned as a site of igneous activity, and how the active magmatic arc finally comes to the position that it subsequently occupies for as long as the IOAS exists. IOASs begin their evolution when subduction...
begins. Subduction zones can form either as a result of lithospheric collapse (spontaneous nucleation of subduction zone; SNSZ) or by forced convergence (induced nucleation of subduction zone; INSZ; Stern 2004). For SNSZ, IOASs start as broad zones of sea-floor spreading associated with subsidence of the adjacent lithosphere, whereas INSZ IOASs may be built on trapped crust, without construction of new arc substrate by large-scale sea-floor spreading. SNSZ results from gravitational instability of oceanic lithosphere adjacent to lithospheric zones of weakness such as fracture

![Fig. 17. Formation and refinement of juvenile arc crust. (a) Model of arc crust evolution from Tatsumi et al. (2008). (1) Incipient arc magmatism replaces the pre-existing oceanic crust to create (2) the initial mafic arc crust. (3) Continuing arc magmatism causes anatexis and differentiation of the arc crust, along with transformation of mafic crustal component into the mantle through the Moho, finally creating mature arc crust with an intermediate composition similar to the average continental crust. Generalized seismic velocity structure in the IBM arc after Suyehiro et al. (1996), Kodaira et al. (2007) and Takahashi et al. (2007). (b) Schematic illustration of progressive burial of early formed plutonic and volcanic rocks within a growing arc (modified after Kelemen et al. 2003). Pyroxenites near the base of the crust are denser than the underlying mantle, and temperatures are high near the Moho, so that these ultramafic cumulates may delaminate whenever their thickness exceeds some critical value (e.g. Jull & Kelemen 2001). Increasing pressure forms abundant garnet in Al-rich gabbroic rocks near the base of the section, and these too may delaminate. Intermediate to felsic plutonic rocks, and even volcanic rocks, may be buried to lower crustal depths, where they are partially melted. (c) Four-stage schematic evolution of the Kohistan arc during the period 117–90 Ma (after Dhiume et al. 2007). Four evolutionary stages span the history of the arc, from spontaneous initiation of subduction associated with extensive boninitic magmatism (stage 1) to arc building by tholeiitic basalts (stages 2 and 3) and a fourth stage of intracrustal differentiation. Stage 4 culminated in the intrusion of granitic rocks akin to continental crust in the upper level of the metaplutonic complex.](image-url)
Fig. 18. Generation of forearc during subduction initiation and retreat of magmatic arc to the approximate position that it will subsequently occupy. (a–e) Subduction infancy model of Stern & Bloomer (1992), modified to show the third dimension. Left panels are sections perpendicular to the plate boundary (parallel to spreading ridge) and right panels are map views. (a, b) Initial configuration. Two lithospheres of differing density are juxtaposed across a transform fault or fracture zone. (c, d) Old, dense lithosphere sinks asymmetrically, with maximum subsidence nearest fault. Asthenosphere migrates over the sinking lithosphere and propagates in directions that are both orthogonal to the original trend of the transform or fracture zone as well as in both directions parallel to it. Strong extension in the region above the sinking lithosphere is accommodated by sea-floor spreading, forming infant arc crust of the proto-forearc.
zones (Fig. 18a, b). Because modern IOASs are concentrated in regions where the oceanic lithosphere is ancient, dense, and prone to collapse, it is likely that most IOASs form by SNSZ.

Lithospheric collapse to initiate SNSZ allows lighter asthenosphere to exchange places with denser, sinking lithosphere (Fig. 18c, d). This results in sea-floor spreading above the sinking lithosphere, and the combined result of decompression of rising asthenosphere and fluids released from the sinking slab is unusually high degrees of mantle melting. This produces very depleted tholeiites (low TiO$_2$ and K$_2$O) and subordinate boninites that make up the forearc crust. Felsic magmas are also produced during this ‘infant arc’ stage. Oceanic crust made up of these lavas and intrusions is underlain by genetically related ultradepleted harzburgitic mantle residues, and these rocks form what becomes the forearc. It is important to note that sea-floor spreading to form the forearc lasts only a few million years, while the subjacent lithosphere continues to sink and asthenospheric mantle flows above the sinking slab. This is the origin of most boninites and ophiolites. Stern & Bloomer (1992) argued that the Izu–Bonin–Mariana intra-oceanic arc system in the Western Pacific formed by collapse of old Pacific lithosphere east of a major transform, with much younger sea floor and hotter shallow mantle to the west. Stern & Bloomer (1992) estimated that during the first 10 Ma in the life of the Izu–Bonin–Mariana arc, melt generation occurred at a rate of c. 120–180 km$^3$ km$^{-1}$ Ma$^{-1}$. This may be a significant underestimate, because recent geophysical studies of the Izu–Bonin–Mariana and Aleutian arcs infer a mean crustal growth rate of c. 100 km$^3$ km$^{-1}$ Ma$^{-1}$ or more (Dimalanta et al. 2002; Jicha et al. 2006), orders of magnitude larger than mature arc eruption rates of <10 km$^3$ km$^{-1}$ Ma$^{-1}$ (Kimura & Yoshida 2006; Carr et al. 2007).

Once down-dip motion of the slab (true subduction) begins, the asthenosphere is no longer able to flow into this region and the sub-forearc mantle rapidly cools. Slab-derived fluids migrating into the sub-forearc mantle no longer cause melting, only serpentinization. Igneous activity migrates away from the trench and eventually forearc igneous activity stops (Fig. 18e, f). A typical IOAS begins to mature when true subduction (i.e. down-dip motion of the sinking slab) begins and the subducted slab reaches c. 130 km depth, focusing magmatism to begin building the magmatic arc and allowing the forearc lithosphere to cool and hydrate. The zone of infant arc crust produced by sea-floor spreading above the subsiding lithosphere becomes forearc as the IOAS matures and the locus of magmatism retreats from the trench. This makes up the forearc, lithosphere that is emplaced as an ophiolite when buoyant lithosphere is subducted beneath it (Pearce 2003). Thus the forearc preserves the early history of the arc.

**Intra-oceanic arc systems and the growth of the continental crust**

In contrast to continuing controversy about how the modern inventory of continental crust came to be and what it was in the past (Dewey & Windley 1981; Armstrong 1991; Windley 1993), a broad geoscientific consensus exists that continental crust today is mostly generated above subduction zones, with secondary sites associated with rifts, hotspot volcanism, and volcanic rifted margins. Subduction zones are also the most important sites of crust removal, by sediment subduction, subduction erosion, and deep subduction of continental crust. A comprehensive overview of how continental crust is created and destroyed by plate-tectonic processes has been given by Stern & Scholl (2010).

Continental crust mostly forms today and over the lifetime that plate tectonics has operated by magmatic additions from water-induced melting of mantle above subduction zones (Coats 1962; Dimalanta et al. 2002; Davidson & Arculus 2005). This produces juvenile IOAS arc crust, and magmatically adds to pre-existing continental crust at Andean-type arcs. Scholl & von Huene (2009) estimated that over the last tens to hundreds of million years, c. 2.7 km$^3$ of juvenile continental crust has been generated by convergent margin igneous activity each year. This crust is mostly extracted from the mantle as mafic (basaltic or boninitic) magma, which must be further refined by anatectic remelting of the crust and delamination (foundering) to yield andesitic material approximating the

Fig. 18. (Continued) (e, f) Beginning of down-dip component motion in sinking lithosphere marks the beginning of true subduction. Strong extension above the sunken lithosphere ends, which also stops the advection of asthenosphere into this region, allowing it to cool and become forearc lithosphere. The locus of igneous activity retreats to the region where asthenospheric advection continues, forming a magmatic arc. (g) A detailed view of how magmatism in the arc evolves with time; 1, infant arc crust of the forearc ophiolite forms by sea-floor spreading during first c. 5 Ma of subduction zone evolution, as shown in (c) and (d); 2, retreat of magmatic activity away from the trench (Tr) during the second c. 5 Ma; 3, focusing of magmatic activity at the position of the active magmatic arc, resulting in crustal thickening and delamination.
bulk composition of continental crust (Rudnick & Gao 2003; Tatsumi 2005; Tatsumi et al. 2008). Although juvenile continental crust is mafic when first formed, it inherits from subduction zone processes the distinctive geochemical characteristics of subduction-related magmas: enrichment in silica and LILE (e.g. K, U, Sr and Pb) and depletion of high field strength incompatible elements (e.g. Nb, Ta and Ti). The similarity of trace element variations in arc rocks to that of bulk continental crust strengthens the inference that subduction-related magmatism is an important way to make continental crust today (Kelemen 1995; Rudnick & Gao 2003).

The distinctive chemical characteristics of IOAS magmas (Fig. 11) are likely to persist through later processing by anatexis within the crust and foundering of its base, so that juvenile mafic crust slowly matures into continental crust with andesitic bulk composition. The formation of continental crust at IOAs requires two stages: mantle melting to generatebasalt and mafic crust and further refinement of this to yield granitic crust and refractory material, much of which founders back into the mantle (Hawkesworth & Kemp 2006). This is accomplished by distilling felsic magmas from mantle-derived juvenile crust via anatexis of amphibolites and fractional crystallization of mafic magmas. Much basaltic magma crystallizes as pyroxene-rich material near the Moho. Magmatic maturation of IOAs is also likely to occur in tandem with terrane accretion, whereby various juvenile crustal fragments coalesce to build progressively larger tracts of increasingly differentiated crust.

Conclusions

Intra-oceanic arc systems are enormous tracts of proto-continental crust, hundreds to thousands of kilometres long, several hundred kilometres wide, and up to 35 km thick. They tend to form where very old (>100 Ma) oceanic lithosphere is subducted, typically showing strong evidence for extension (normal faulting, intra-arc rifts, back-arc basins). Morphological asymmetries (trench, fore-arc, volcanic front, rear-arc volcanoes, and back-arc basin) reflect asymmetrical processes controlled by the dipping subduction zone. Crustal structure reflects this asymmetry, with IOAS crust thickening from the trench towards the magmatic front, and thinning again beneath back-arc basins. Distinctive compositions of igneous rocks and associated mantle also reflect the subduction-imposed asymmetries that reflect in part the evolution of the IOAS. Igneous rock compositions reflect decreasing water flux from the slab with distance from the trench, resulting in lower-degree melts and more enriched igneous rocks with this distance. This is the basis of the K–h relationship. In addition, ultrahigh-degree melting occurs early in the history of the IOAS to yield ultra-depleted tholeiites and some boninites to form what will become forearc crust. Magmatic products of a mature subduction zone manifest combination of a continuous flux of hydrous basalt from the mantle wedge, crustal magmatic process of fractionalation and anatexis, and loss of dense pyroxene-rich cumulates from the base of the crust back to the mantle. The result of these processes is to produce crust that becomes increasingly felsic as it ages and slowly thickens. Mantle peridotites reflect these subduction-related asymmetries in melting and evolution: ultra-depleted harzburgites that are strongly serpentinitized beneath the forearc, pyroxene-rich ultramafic rocks beneath the magmatic front, abyssal peridotite-like lherzolites and moderately depleted harzburgites beneath the back-arc basin. IOAs cannot be understood without understanding the remarkably different processes and magma production rates that accompany their early development. The characteristic dimensions and asymmetries of IOAS systems should be at least partly preserved in orogenic belts and discernible to careful observers.

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