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Differential preservation in the geologic record of intraoceanic arc sedimentary and tectonic processes

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ABSTRACT

Records of ancient intraoceanic arc activity, now preserved in continental suture zones, are commonly used to reconstruct paleogeography and plate motion, and to understand how continental crust is formed, recycled, and maintained through time. However, interpreting tectonic and sedimentary records from ancient terranes after arc–continent collision is complicated by preferential preservation of evidence for some arc processes and loss of evidence for others. In this synthesis we examine what is lost, and what is preserved, in the translation from modern processes to the ancient record of intraoceanic arcs.

Composition of accreted arc terranes differs as a function of arc–continent collision geometry. 'Forward-facing' collision can accrete an oceanic arc on to either a passive or an active continental margin, with the arc facing the continent and colliding trench- and forearc-side first. In a 'backward-facing' collision, involving two subduction zones with similar polarity, the arc collides backarc-first with an active continental margin. The preservation of evidence for contemporary sedimentary and tectonic arc processes in the geologic record depends greatly on how well the various parts of the arc survive collision and orogeny in each case.

Preservation of arc terranes likely is biased towards those that were in a state of tectonic accretion for tens of millions of years before collision, rather than tectonic erosion. The prevalence of tectonic erosion in modern intraoceanic arcs implies that valuable records of arc processes are commonly destroyed even before the arc collides with a continent. Arc systems are most likely to undergo tectonic accretion shortly before forward-facing collision with a continent, and thus most forearc and accretionary-prism material in ancient arc terranes likely is temporally biased toward the final stages of arc activity, when sediment flux to the trench was greatest and tectonic accretion prevailed. Collision geometry and tectonic erosion vs. accretion are important controls on the ultimate survival of material from the trench, forearc, arc massif, intra-arc basins, and backarc basins, and thus on how well an ancient arc terrane preserves evidence for tectonic processes such as subduction of aseismic ridges and seamounts, oblique plate convergence, and arc rifting. Forward-facing collision involves substantial recycling, melting, and fractionation of continent-derived material during and after collision, and so produces melts rich in silica and incompatible trace elements. As a result, forward-facing collision can drive the composition of accreted arc crust toward that of average continental crust.

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1. Introduction

Geologic records from intraoceanic arcs are commonly used to interpret the tectonic history of convergent margins. Ancient arc terranes preserved in suture zones within continents reveal patterns of plate motion and supercontinent assembly that are significant controls for paleogeographic reconstructions (Cawood and Buchan, 2007; Murphy et al., 2009; van der Meer et al., 2012), and contain rock assemblages that are economically valuable (Cooke et al., 2007; Glen et al., 2011; Herrington et al., 2011; Wainwright et al., 2011). Moreover, intraoceanic arcs have been producing and refining Earth's crust since Archean time (Kimura et al., 1993; Polat and Kerrich, 2002), and are key to understanding the origin and evolution of the continental crust (Rudnick and Fountain, 1995; Suyehiro et al., 1996; Holbrook et al., 1999; Draut et al., 2002; Kelemen et al., 2003a; Davidson and Arculus, 2006; Hawkesworth and Kemp, 2006; Clift et al., 2009; Draut et al., 2009; Stern and Scholl, 2010). Intraoceanic arcs, which comprise 35–40% of active subduction-zone length globally, are most common today in the western Pacific Ocean (Fig. 1). Ancient, accreted oceanic arc terranes occur within orogenic belts on every continent.

Although much of our understanding of Earth's history relies on the geologic record of ancient subduction zones (Scholl and von Huene, 2010; Dilek and Furnes, 2011; Korsch et al., 2011; Murphy et al., 2011), it is likely that the ancient arc terranes now preserved in continental suture zones lack, or represent disproportionately, evidence for sedimentary and tectonic processes that characterized the intraoceanic arc while it was active prior to collision. Such a disparity likely arises because the different tectonic regions of an arc have variable preservation potential following collision with a continent (Scholl and von Huene, 2010; Draut and Clift, 2012). As a result, some tectonic events and processes that commonly affect the development of intraoceanic arcs over millions to tens of millions of years, and that are notable and prominent in the modern oceans, may leave little or no trace in the rock record after the arc accretes onto a continental margin.

In this paper we review and summarize major tectonic, geomorphic, and sedimentary characteristics of active intraoceanic arcs, and discuss the means by which they reflect and record significant tectonic processes that commonly characterize arc development, such as subduction of high bathymetric features (i.e., seamounts, aseismic ridges, and fracture zones), oblique plate convergence, and arc rifting. Different regions of an oceanic arc have variable preservation potential after collision with a continental margin, controlled in part by collision geometry. In light of these complications, we discuss whether and how completely the record of arc sedimentary and tectonic history is preserved through geologic time. We thereby evaluate the benefits and limitations of interpreting ancient arc terranes in the geologic record.

2. Intraoceanic arc morphology and sedimentary processes

Seafloor and sub-bottom mapping of modern, active arc regions, combined with dredging and drill-core sampling, documents their

Fig. 1. Map showing active global subduction zones and the ancient arc systems referenced in the text. Accretionary plate margins are shown with solid triangular marks along the plate boundary, and tectonically erosive margins are indicated by empty triangles.

morphology and sedimentary environments. From pioneering sedimentological and geophysical work by Karig (1971), Grow (1973), Marlow et al. (1973), Dickinson (1974), and Scholl et al. (1983), among others, to measurements of active arc deformation by more recent geodetic techniques (Hu et al., 2001; Freymueller et al., 2008), tectonic and sedimentary processes around subduction zones have been characterized sufficiently to inform not only direct investigations of convergent-margin tectonics, but also related studies of climate–tectonic coupling, which may be important processes in collision zones where an orogen undergoes substantial subaerial erosion (Dadson et al., 2003; Kimura et al., 2008).

Intraoceanic subduction zones undergoing tectonic erosion exhibit important differences from those in a state of tectonic accretion (Fig. 2). Approximately one quarter of modern oceanic arcs experience long-term accretion from the downgoing plate (Fig. 2A), whereas three quarters are of the erosional type (Fig. 2B) (von Huene and Scholl, 1991; Stern, 2010). Almost all forearcs experience phases of crustal addition or loss, but for the purpose of this paper we define a tectonically erosive margin to be one on which there is a trenchward migration of rock in the forearc as a result of net crustal loss from the front or base of the forearc wedge on time scales of 10^6–10^7 yr. Tectonic erosion progressively removes material from the upper plate (trench slope, outer forearc, and beneath the forearc), causing the arc volcanic front to migrate landward over time, away from the present location of the trench. While tectonically eroding margins may undergo brief periods of accretion, and vice versa, or even have areas of the forearc that show the opposing tectonic style for some time (Wagreich, 1993), the key difference between eroding and accreting margins lies in whether the margin is losing or gaining net crustal volume over time intervals >5 m.y. In accretionary subduction zones the width of sedimentary rock frontally accreting at the trench can be >50% of the forearc width, compared to <25% at tectonically eroding margins (Scholl and von Huene, 2010).

The tectonic-erosion model thus proposes a fairly constant forearc width, in which individual packages of rock migrate closer to the trench as ongoing tectonic erosion progressively strips material from the toe and base of the forearc wedge. This occurs only if sediment thickness on the lower plate (and so in the trench) is not great enough to form a frontal accretionary prism to serve as a buffer between the two plates. On accretionary subduction margins the sediment thickness is sufficient (>1 km) to form a buffering accretionary prism over timescales of >5 m.y. owing to progressive offscraping of sediment from the subducting plate against the forearc backstop, usually formed from arc rocks at the start of an accretionary phase (Clift and Vannucchi, 2004). It is unclear why a 1-km trench sediment thickness seems to be critical, although it is noteworthy that seismically imaged subduction channels containing sediment on modern active margins are approximately 900–1000 m thick (Scholl and von Huene, 2007; Collot et al., 2011) (Fig. 3). Preserved subduction channels may be thicker, up to 3 km in the case of the McHugh Complex in southern Alaska (Amato and Pavlis, 2010; Clift et al., 2012), or thinner, e.g., 500 m thick in the Apennines (Vannucchi et al., 2008). However, deformation during and after collision may alter the preserved channel width significantly. The width of the subduction channel may limit the amount of trench sediment that can be subducted readily, forcing the excess to be accreted frontally or beneath the forearc wedge. Thus, whether an active margin is accretionary or erosional depends on whether the sediment supply rate to the trench and convergence rate allow a thick sediment pile to

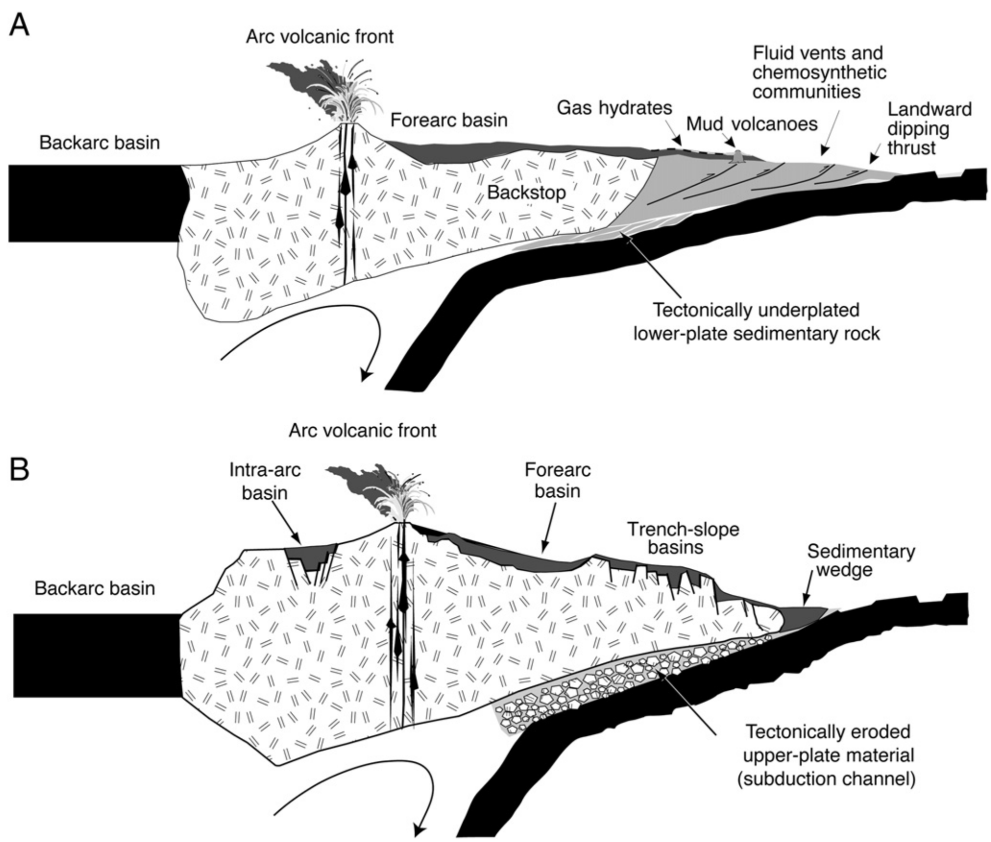

Fig. 2. Schematic cartoons showing the two basic types of active margin: tectonically accreting and tectonically eroding, modified after Clift and Vannucchi (2004). (A) Accretionary margins are characterized by forearc regions comprising thrusted and penetratively deformed trench and oceanic sediments that often develop mud diapirism and volcanism due to sediment over-pressure. Gas-hydrate zones also are commonly associated with structures in the accretionary wedge. (B) Erosive plate margins, such as Tonga, are marked by steep trench slopes composed of volcanic, plutonic and mantle rocks. Sedimentary rocks typically are limited to the forearc basin, where they may be faulted but are not strongly sheared as in an accretionary wedge. In the Mariana arc serpentinite mud volcanism occurs. Reprinted with permission from the American Geophysical Union.

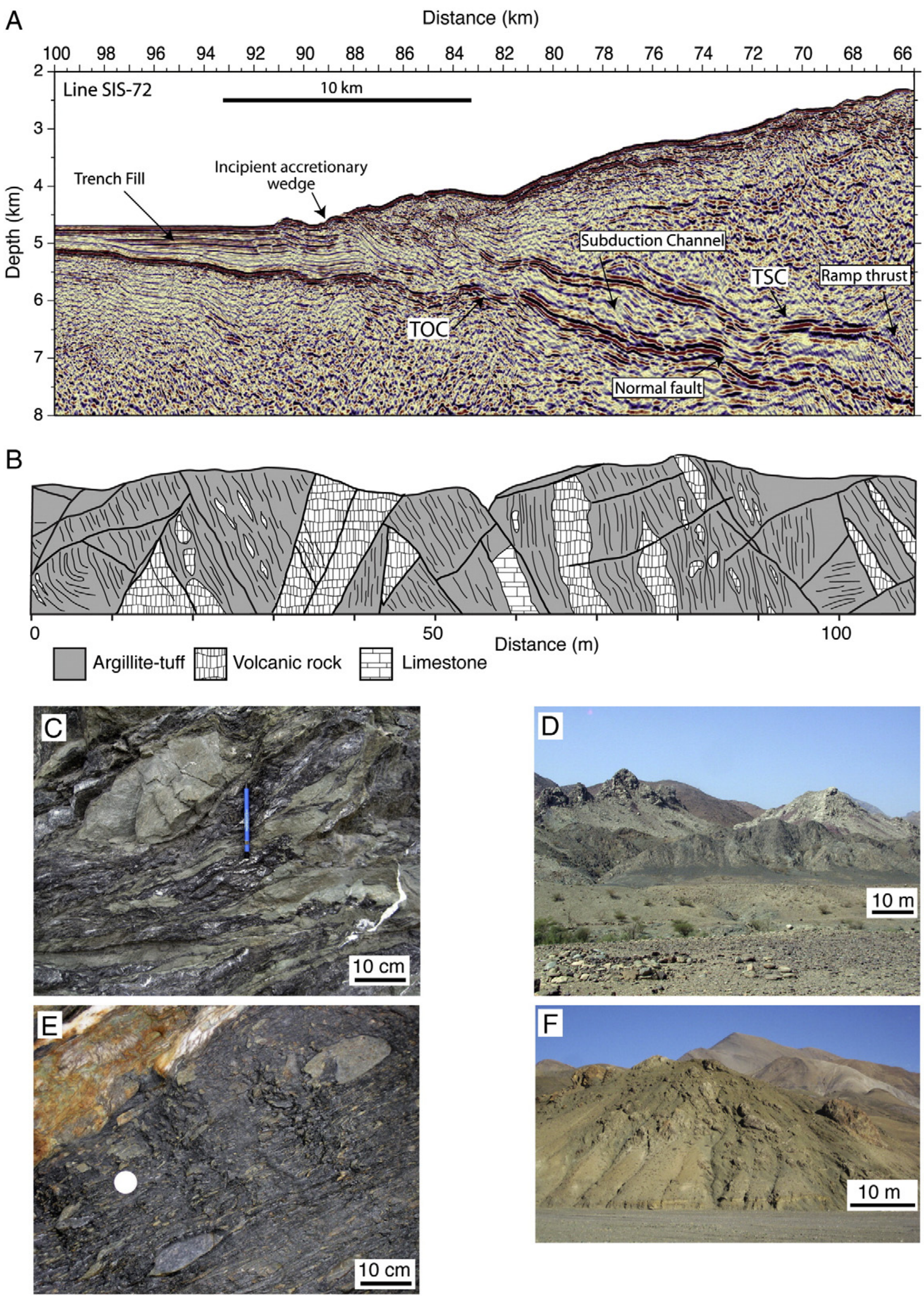

Fig. 3. Accreted sediment at modern and ancient subduction zones. (A) Pre-stack depth-migrated (PSDM) calibrated multi-channel seismic (MCS) line SIS-72 from the Ecuador convergent margin showing a well developed subduction channel ~1 km thick. TOC = Top of Oceanic Crust reflector; TSC = Top of Subduction Channel reflector (from Collot et al., 2011, used with permission). (B) Section of the mesomélange from the McHugh Complex, southern Alaska, a preserved Cretaceous subduction channel sequence (Clift et al., 2012), shown without vertical exaggeration. (C) Sheared argillite tuff from the mesomélange, McHugh Complex, southern Alaska (Clift et al., 2012). (D) Coherent, hard-weathering limestone blocks in ophiolitic Haybi Mélange, Oman. The green and red weathered, poorly exposed material is a sheared shaley matrix that dominates the unit. (E) Tienhsiang Formation mélange, Taiwan (Hsu, 1988), and (F) Mélange dominated by basaltic blocks (coherent masses) in Yarlung Tsangpo Suture Zone, near Gyantse, Tibet. The bulk of this mélange comprises a sheared, volcaniclastic sand and shales supporting the blocks of basalts that range up to 15 m across and show no sorting, but have a rough alignment to the shale fabric.

accumulate (von Huene and Scholl, 1991; Clift and Vannucchi, 2004). These factors affect the geomorphic and sedimentary development of the arc both during its intraoceanic activity and after collision with a continent.

In this section we summarize briefly the morphology and major sedimentary processes that characterize intraoceanic arcs prior to collision with continents, considering four regions: (1) trenches and trench-slope basins; (2) forearcs and forearc basins; (3) arc massifs and intra-arc basins; and (4) backarc basins. Detailed discussions of these have been well covered by, e.g., Clift (1995), Dickinson (1995), Marsaglia (1995), and Underwood et al. (1995), and in a comprehensive review by Stern (2010). We largely restrict this discussion to intraoceanic arcs, i.e., those formed by subduction within oceanic lithosphere and not founded on continental crust, most examples of which are also known as 'island arcs'. Continental arcs and their sedimentary record are beyond the scope of this review, though we reference them for some informative comparisons with oceanic arcs; see, e.g., Wilson (1991), Jordan (1995), Fildani and Hessler (2005), Kay and Ramos (2006), Lamarche et al. (2008), Trop (2008), and Nester and Jordan (2012) for more detail on continental arc processes and their records. Focusing on the sedimentary and tectonic record, we also omit extensive discussion of petrologic and geochemical data except as they directly inform these topics.

2.1. Trenches and trench-slope basins

Trenches and trench-slope basins form along the boundary between the overriding and underthrusting plates, with morphology controlled by flexure of the subducting slab and, at tectonically eroding margins, extension of the upper plate (von Huene and Scholl, 1991; Underwood and Moore, 1995). If sufficient sediment is present, the trench contains an accretionary prism (Fig. 2A).

In both erosional and accretionary subduction zones, structurally controlled basins on the trench slope of the deforming upper plate form relatively small, isolated sedimentary depocenters (Fig. 4). These perched trench-slope basins may be quite deep and narrow; in the Tonga trench some basins measure as large as 10×50 km and have complex stratigraphy, reflecting their dynamic structural setting (Clift et al., 1998). Because many erosive plate margins are retreating at long-term rates as great as ~3 km/m.y. (Clift and Vannucchi, 2004), basins in the trench slope and outer forearc have a limited life span even before the trench collides with a continental margin. In some arcs a broad, elevated forearc platform separates the arc front from the trench, limiting arc-derived sediment from reaching the trench and trench-slope basins. However, sediment can cross the forearc platform if delivered through submarine canyons (e.g., Tonga, Hellenic, and Lesser Antilles arcs) (Scholl and Vallier, 1985; Marsaglia and Ingersoll,

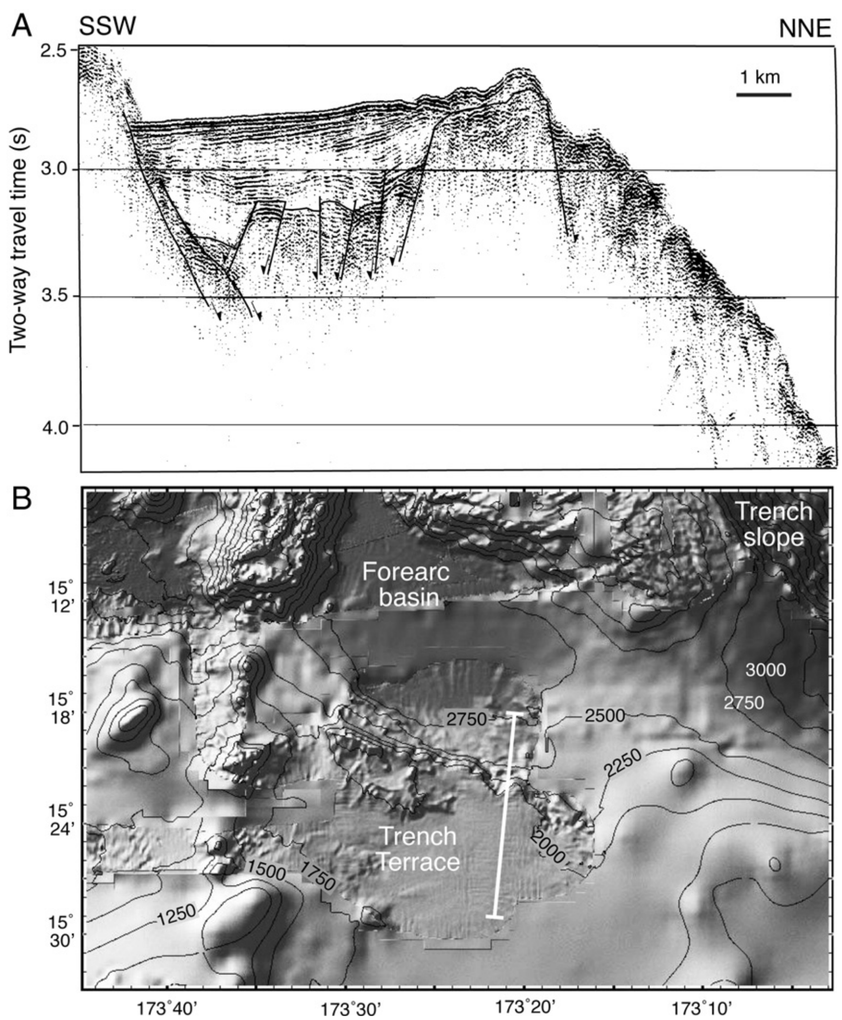

Fig. 4. (A) Seismic profile through a trench slope basin from the northern Tonga arc (Clift et al., 1998). Note the moderate thickness and width of the basin. Basins of this type would have low preservation potential after collision with a continental margin. (B) Shaded bathymetric map of the region across the sub-basin shown in part A. Map made by GeoMapApp™ with data from Wright et al. (2000). See Fig. 1 for approximate location.

1992; Wright et al., 2000; Draut and Clift, 2006), or if unconfined turbidity currents flow across the forearc, over the outer forearc platform, and down the trench-slope break, as occurs in the Aleutian arc (Underwood, 1986). Sediment transport also occurs along the axis of the trench, because the trench axis slopes downward toward older, underthrusting crust (e.g., Sumatra, Alaska–Aleutian margin, and central and southern Chile trenches). Bathymetric relief in the trench and the regional slope of the lower plate can control the sediment-transport distance and thickness of deposits; the resulting spatial variation in trench-sediment thickness affects the composition of arc volcanism (Scholl et al., 1982; Plank and Langmuir, 1993; Kelemen et al., 2003b).

Sedimentary deposits on trench slopes of tectonically erosive margins (e.g., most western Pacific arc settings) subside rapidly in response to basal erosion and thinning of the forearc crust, creating abundant accommodation space near the trench, and thus the trench and trench-slope generally are areas of slow, deep-water sedimentation (Kaiho, 1992; Clift and MacLeod, 1999). Trench-slope basins and their sedimentary record are progressively destroyed by subduction erosion in such cases (Collot et al., 2008), exposing old lavas, lower crust, and mantle rocks at the trench (Bloomer and Hawkins, 1983; Bloomer and Fisher, 1987). The forearc is nonetheless maintained in a nearly constant state—as one set of basins approaches the trench and is destroyed, new basins form on the upper trench slope by extension driven by ongoing basal subduction erosion. In contrast, in accretionary margins such as the northern Luzon arc, Nankai, or the Lesser Antilles, accommodation space may be limited; sediment can be deposited in perched basins overlying the accretionary wedge, or can move downslope via mass-movement slides to be deposited on the trench floor (Beaudry and Moore, 1985; Underwood, 2003; Bangs et al., 2004). At accretionary margins, uplift can culminate in subaerial emergence and recycling of sediment from the accretionary prism that can mix with arc volcaniclastic material, forming shallow-water facies. However, advanced uplift at the trench slope would not be expected in an intraoceanic arc unless collision with a continent were imminent (Abbott et al., 1994; Warren and Cloos, 2007) or unless the arc lies along strike from a sediment-productive continental margin (e.g., eastern Aleutians or Lesser Antilles), because the sediment supply needed to form a large accretionary prism is most readily available near continental margins. Because oceanic arcs are largely submarine, they undergo little erosion (and produce little sediment) from subaerial weathering, and generate only modest volumes of volcaniclastic sediment by explosive eruption, though submarine volcaniclastic deposits can still be thick near volcanic centers (Gill et al., 1990) especially along basin edges during arc rifting (Sigurdsson et al., 1980; Taylor et al., 1991).

2.2. Forearcs and forearc basins

The forearc region, on the upper plate between the arc and the outer-arc high (trench-slope break; (Dickinson, 1995), is commonly the site of basin formation where sediment can accumulate to thicknesses of several kilometers. Structural relief and subsidence of forearc basins can be a function of crustal thinning by basal subduction erosion (Oncken, 1998; Clift et al., 2003; Wells et al., 2003) and relative uplift of the outer-arc high, in turn controlled by convergence rate and angle (Jarrard, 1986), accretionary-complex formation (Harbert et al., 1986; Dickinson, 1995), and inter-plate coupling (Ryan et al., 2012). Although numerical modeling indicates that forearc basins do not require focused subduction erosion (Fuller et al., 2006), both subduction erosion and underplating beneath the forearc contribute to vertical tectonic motions and thus to forearc-basin and outer-arc-high topography (Sample and Moore, 1987; Moore et al., 1991). Consequently, subsidence and relative elevation of forearc basins are controlled mostly by trench tectonics, and thus are sensitive to subduction of large ridges or seamounts, though they can also be linked to arc rifting during formation of backarc basins (Section 3) (Austin et al., 1989; Clift et al., 1994). Forearc-basin development can postdate subduction initiation by several million years (Scholl et al., 1983; Harbert et al., 1986; Scholl et al., 1987). As mentioned above, some arcs (e.g., Tonga, the Marianas, and parts of the Aleutians) have an elevated platform, rather than a basin, in the forearc region (Austin et al., 1989; Tappin and Ballance, 1994; Draut and Clift, 2006; Dickinson and Burley, 2007; Ryan et al., 2012).

Sediment in oceanic forearc basins can include lava flows and proximal mass-flow deposits from the steep slopes of volcanic centers. Closer to the trench, the stratigraphy is dominated by distal volcaniclastic turbidites sourced from the arc, and even debris reworked from the trenchward side of the basin (the outer-arc high) (Hussong and Uyeda, 1982; Ballance et al., 1994, 2004). Mass-transport deposits in forearc basins can occupy hundreds of cubic kilometers in volume, and substantially rework underlying turbidite sequences by basal erosion (Lamarche et al., 2008; Ryan et al., 2012). Pelagic sedimentation tends to be significant only at the outer forearc, which is the only area likely to record distal tephra fallout without disruption by mass wasting (Dickinson, 1974; Larue et al., 1991; Marsaglia and Ingersoll, 1992; Underwood et al., 1995; Ballance et al., 2004; Draut and Clift, 2006). Outer-forearc sedimentary assemblages thus can include diatomaceous silt and clay interbedded with airfall ash and pumice, as well as laminated distal volcaniclastic turbidites (Scholl and Creager, 1973; Stewart, 1978). In some outer forearc areas, sedimentation rates can be so low that Mn crust develops on the volcanic basement (Cronan et al., 1984). Submarine canyons can transfer sediment off the arc massif and into forearc basins (Kopp et al., 2006), just as canyons feed trenches elsewhere. Owing to the diversity of their sediment sources and the fact that they usually are not strongly deformed by arc rifting, forearc basins may contain a sedimentary record of much longer duration than the time the arc front has been active in any one place.

Tectonically erosive and accreting margins differ in the accommodation space available for sediment storage, with greater water depths maintained in forearc basins of erosive margins owing to ongoing basement subsidence. Because tectonic erosion of the forearc causes gradual migration of the arc magmatic front away from the trench, if the slab dip remains relatively constant, the locus of forearc sedimentation would be expected to migrate landward with time (Tagudin and Scholl, 1994; Scholl and von Huene, 2010). Although this can be true locally and over intervals of a few million years, seismic stratigraphy in the Tonga forearc indicates that forearc depocenter locations are often controlled largely by extensional structures rather than by proximity to sediment sources (Austin et al., 1989; Tappin, 1993).

2.3. Arc massifs and intra-arc basins

The arc volcanic front is a major source of sediment in intraoceanic subduction zones, from primary volcanic products—lava, airfall tephra, and ash—and as mass wasting and volcanic collapse episodically generate large volumes of material (Coombs et al., 2007; Silver et al., 2009; Watts et al., 2012). Much of the arc sediment, in particular the proximal mass-transport deposits, accumulates in basins that form among volcanic edifices on the arc platform as part of the arc massif (Smith and Landis, 1995). These intra-arc basins, with smaller area and often shallower water depths than forearc or backarc basins, can be bounded by volcanic centers or by faults, and may form as part of a rifting event whereby arc-normal extension (such as incipient backarc-basin spreading) generates new basins within the arc massif (Busby, 2004; Busby et al., 2006). Other intra-arc basins develop as transpressional or transtensional features caused by strike-slip faulting (Sarewitz and Lewis, 1991) or large-scale block rotation within the arc as plate motions change through time (Geist et al., 1988). In the latter cases, the extension that forms the intra-arc basins is oriented obliquely or subparallel to the arc massif, in contrast to the arc-normal extension described by Busby (2004).

Intra-arc sedimentation consists of proximal volcanic and volcaniclastic material in large debris aprons derived from the arc massif, fining away from eruptive centers into turbidite and deep-water drift facies. Bottom currents can rework the volcaniclastic sediment into fields of sand waves that show the direction of sediment distribution away from the sediment-producing volcanic centers (Draut and Clift, 2006; Hoffmann et al., 2008). Intra-arc volcanic and volcaniclastic deposits may be intercalated with reef carbonates, reflecting changes in the intensity of eruption history and in relative sea level (local uplift and subsidence) that lead to episodic reef formation and destruction during arc activity and basin sedimentation (Austin et al., 1989; Busby et al., 2006; Dorobek, 2008; Hoffmann et al., 2009). On the forearc side of the arc massif, volcanic and volcaniclastic material interbeds with and grades into forearc-basin fill, such that distinguishing between intra-arc and proximal forearc deposits in the geologic record could be problematic (Dickinson, 1995).

2.4. Backarc basins

Basins commonly occur behind intraoceanic arcs (Fig. 2), originating either from rifting and extension after arc development, as has occurred twice in the Izu–Bonin–Mariana (IBM) arc system and twice in the Tonga–Kermadec arc (Hawkins, 1974; Taylor, 1992; Bevis et al., 1995; Clift, 1995; Hawkins, 1995), or as older ocean floor that pre-dates the arc and subduction zone (Karig, 1971; Taylor and Karner, 1983). In backarc basins that form by arc rifting, each rifting event splits the arc so that fragments of the older arc are carried trenchward, leaving remnant arcs in the backarc region (e.g., in the IBM arc, the Palau–Kyushu Ridge and West Mariana Ridge). As a result, an arc that has undergone arc-normal or longitudinal extension to form a backarc basin will contain thinner crust than does an un-rifted arc (Calvert, 2011). Rearrangement of the arc volcanic front and backarc geometry in this manner can lead to an intraoceanic arc having a sedimentary record (preserved in the forearc) that spans a much longer time interval than the activity of the volcanic arc in its present location (Clift, 1995).

Sedimentation in retro-arc basins is dominated by volcanic and volcaniclastic products of the arc, including pyroclastic-flow deposits and lapilli tuffs and breccias, with facies indicating water depths that increase with distance from the arc (Bednarz and Schmincke, 1994; Busby, 2004). Klein (1985) inferred submarine-fan turbidites to be the dominant depositional facies for backarc volcaniclastic material, with less sediment volume contained in pyroclastic deposits, debris flows, and silty basinal turbidites. Pelagic and hemipelagic clays, biogenic material, and resedimented carbonates also contribute substantially to backarc-basin sediment flux (Klein, 1985; Marsaglia, 1995). Irregular, faulted topography in the backarc region can limit the transport distances of mass-wasting deposits or turbidity currents, leading to sediment being trapped near the volcanic centers, as occurs in the Mariana arc (Draut and Clift, 2006). Where backarc basins have formed by longitudinal arc rifting, silicic volcanic products can be abundant in the backarc stratigraphy, especially during the earliest phases of extension (e.g., Izu–Bonin arc) (Packer and Ingersoll, 1986; Nishimura et al., 1992; Marsaglia and Devaney, 1995; Iizasa et al., 1999; Fiske et al., 2001; Critelli et al., 2002).

3. Intraoceanic arc response to significant tectonic processes

In order to understand how accreted arc terranes within continental suture zones record evidence for major tectonic events during the pre-collisional activity of the arc, it is necessary first to document the geomorphic and sedimentary responses of modern arcs to those events and processes. We consider the effects of subduction of bathymetric highs, convergence angle and rate, and arc rifting on modern, active arc environments. By summarizing how each of these processes affects the arc prior to arc-continent collision, we then can evaluate whether and how completely such records survive collision and orogenesis.

Analyzing the sedimentary record in an active arc presents some disadvantages over examining ancient, accreted arcs. It is obviously not possible to access modern, submarine sedimentary deposits in as much detail as would be achieved by mapping and sampling subaerially exposed outcrops of an accreted arc terrane. In active, submarine arcs outcrop-scale features such as bedding style and sedimentary structures indicating paleocurrent direction commonly are not observable, but instead must be inferred, where possible, from larger-scale bathymetry, seismic stratigraphy and rarely from smaller-scale drill cores (Draut and Clift, 2006; Hoffmann et al., 2008). However, the difficulty of direct access is offset by the advantages of utilizing seafloor imagery, earthquake seismology, and geodetic strain measurements to characterize tectonic and sedimentary processes at modern arcs with accuracy and resolution that would be impossible in accreted terranes where arc activity has long since ceased.

3.1. Subduction of bathymetric highs

It is common for modern intraoceanic arcs to accommodate subduction of a bathymetrically high feature on the downgoing plate such as a seamount, aseismic ridge, or large fracture zone. Examples include subduction of the Louisville Ridge beneath the Tonga arc (Dupont and Herzer, 1985; Ballance et al., 1989), subduction of the D'Entrecasteaux Ridge beneath the New Hebrides arc (Fisher, 1986), and subduction of the Magellan seamounts and the Ogasawara Plateau beneath the IBM arc and Japan (Fryer and Smoot, 1985; Lallemand and Le Pichon, 1987; Dominguez et al., 1998). Other prominent modern examples that affect continental rather than oceanic arcs include fracture zones, the Cocos Ridge, and seamounts subducting beneath central America (Gardner et al., 1992; Moore and Sender, 1995; Gardner et al., 2001; Vannucchi et al., 2006; Morell et al., 2008), the western end of the Aleutian arc underthrusting Kamchatka (Geist and Scholl, 1994; Scholl, 2007), and aseismic ridges subducting beneath Peru and Chile (Laursen et al., 2002; Hampel et al., 2004). In each of these examples the trench absorbs a lengthy structure on the downgoing plate measuring 1.5 to 4.0 km high and 100 to 200 km wide, large enough to affect the arc morphology but not great enough to block the subduction zone and cause wholesale collision, as would a larger plateau (e.g., Ontong-Java Plateau) or a continental margin (Mann and Taira, 2004; Taylor et al., 2005; Kopp et al., 2006).

The effects of subducting bathymetric highs vary depending on the size and depth of subducting features, the strength of the upper plate, and the amount of sediment present (Trehu et al., 2012). Responses of the forearc region to subduction of bathymetric highs are generally similar for oceanic and continental arcs (McCann and Habermann, 1989; Rosenbaum and Mo, 2011), as informed not only by observations in the modern oceans but also by physical experiments Lallemand et al. (1992) and numerical modeling (Geist et al., 1993). Notably, Geist et al. (1993) suggested that the response to ridge collision depended on convergence velocity. Rapid collisions, such as that of the Louisville seamounts with the Tonga arc, caused arc-parallel tension in the wake of ridge subduction, whereas slower collisions resulted in compressional deformation directly arcward of the collision zone and transverse strike-slip faulting next to the zone of compression. The angle of convergence and collision between the bathymetric high and the arc also play an important role in deformational geometry, with orthogonal collisions (such as the Cocos Ridge at Costa Rica, or the western Aleutian Ridge colliding with Kamchatka) producing a localized coastal thrust belt. GPS data show that the Cocos Ridge acts as an indentor into the Costa Rica margin, forcing forearc blocks away from the collision zone and opening narrow pull-apart basins in the forearc wedge (LaFemina et al., 2009). Because the orientation of subducting bathymetric features commonly is not parallel to the direction of plate motion, collisions between the arc and bathymetric feature are commonly oblique, with the collision zone migrating along the margin over time even when the overall plate convergence direction is essentially orthogonal. Examples of oblique collision include where the Louisville Ridge moves southward along the Tonga trench, and where oblique collision with the Nazca Ridge moves southeastward along the Peruvian margin (Hampel, 2002). Oblique subduction of obstacles on the underthrusting plate can compress the forearc wedge at the leading edge of the indenting object, transporting accretionary-prism material along strike (McCann and Habermann, 1989).

Forearc regions can respond to subduction of bathymetric highs by shortening, uplifting, and then subsiding as the obstacle passes obliquely under a given area. This creates a thrust belt that subsequently becomes extensional and may include differential uplift and rotation of fault-bounded blocks. At the Tonga arc, subduction of the Louisville Ridge apparently uplifted not only the forearc, but also parts of the arc massif and even the backarc basin (Lallemand et al., 1992), although the amount of uplift (estimated to be <300 m) was insufficient to be detected in sediments of ODP Site 840 located in the Tonga forearc close to the modern arc volcanic front (Clift et al., 1994). This indicates that uplift is more pronounced on the trenchward side of the forearc basin, although the trench-slope region opposite the Louisville Ridge (5 km water depth around ODP Site 841) (Clift et al., 1994) did not shoal sufficiently for uplift to be recorded in either the sedimentary facies or the preservation state of microfossils.

Other forearc responses to subducting bathymetric highs include forming indentations or reentrants in the upper plate, such as the 80-km retreat of the Tonga forearc opposite the Louisville Ridge collision. Clift and Macleod (1999) inferred a paleo-ridge collision in the Tonga trench at 16 Ma based on arcward backtilting of the forearc sediments at ODP Site 841, a major shoaling in paleo-water depth, and subsequent mass wasting. Enhanced tectonic erosion also results; subduction of the Louisville Ridge increased tectonic-erosion rates by a factor of ~50 (Ballance et al., 1989), and erosion rates along the Peru margin increased 10-fold in the wake of Nazca Ridge collision (Clift et al., 2003). Mass wasting of the Costa Rica forearc following seamount collisions also is well documented (Ranero and von Huene, 2000). Less well understood, however, is how far into the overriding plate the effect of collision extends; most studies argue that ridge subduction drives uplift and deformation of the overriding plate only <200 km inland from the trench (Vannucchi et al., 2006; Clift and Hartley, 2007). For the effects of aseismic ridge subduction to extend very far inland may require flat-slab subduction (Gutscher et al., 2000; Ridgway et al., 2012). Flattening of subduction angle associated with the underthrusting of a large, buoyant bathymetric element on the lower plate has also been linked to cessation of arc volcanism (Rosenbaum and Mo, 2011) and enough coastal uplift to inhibit foreland-basin formation (Nur and Ben-Avraham, 1983; Gutscher et al., 2000). Uplift of the overriding plate can be great enough, at least at continental active margins, to increase sediment flux substantially to the trench and to enlarge the accretionary prism, as well as increasing tectonic erosion of the overriding plate (von Huene et al., 1996; Clift et al., 2003). Seamount subduction also may cause along-strike variations in subduction-channel thickness (Trehu et al., 2012), contributing to variations in stress concentration, fault segmentation, seismicity, and inter-plate coupling (Moyer et al., 2011; Singh et al., 2011; Trehu et al., 2012).

Uplift and unconformity development across the forearc might be expected as a bathymetric high subducts (Hsu, 1992), but seismic-stratigraphic work in Tonga instead attributed many forearc unconformities to rifting events rather than collisions when the unconformities extended spatially across the entire forearc (Austin et al., 1989). It is clear that ridge and seamount collisions can have a major and potentially complicated impact on forearc sedimentation, but that attributing changes in forearc sedimentation unequivocally to a ridge or seamount collision can be challenging, especially in ancient systems where evidence of the underthrusting bathymetric feature itself has been lost.

3.2. Accommodation of oblique convergence

Almost all active margins are affected by some obliquity in their convergence direction (Fitch, 1972; Jarrard, 1986). Even in the Tonga arc, Kamchatka, and the Costa Rica sector of Central America, where convergence is now orthogonal, this situation evolves as plate motions change through time. Forearcs affected by the stresses resulting from oblique convergence, and by associated variation in the degree of coupling across the plate boundary (Cross and Freymueller, 2007), can accommodate these stresses by means of strike-slip faulting (Jarrard, 1986; McCaffrey, 1992). The forearc above an oblique convergence zone commonly is carried along by the subducting plate as a block or sliver that behaves semi-independently from the rest of the overriding plate. In Sumatra, modern slip partitioning has long been recognized and confirmed by GPS surveys (McCaffrey et al., 2000). There, the margin-parallel Sumatra and Mentawai fault zones, along the forearc and the trenchward side of the arc massif respectively, play a key role in dextral slip partitioning (Fig. 5). Pull-apart basins can open along the length of the fault (Nakano et al., 2010) and also occur at the southeastern termination of the Sumatra fault zone, forming the Sunda Strait pull-apart basin. Seismic surveys of the Sumatran forearc also show that deformation influences the perched forearc basin between the arc edifice and the uplifted accretionary prism (Kopp and Kukowski, 2003). Fig. 5A shows an example of a transpressional fault cutting through the center of the forearc basin, driving localized basin inversion that substantially affects the basin stratigraphy (Schlüter et al., 2002). Shear-sense indicators and metamorphic lineations suggest that a similar oblique subduction fabric is preserved within Cretaceous accretionary-complex exposures of California (Wakabayashi, 1992).

As well as localizing strain along major trench-parallel faults, some forearcs accommodate oblique convergence through a series of smaller, rigid fault blocks that rotate relative to one another. Seismologic and geodetic measurements from the Aleutian forearc indicate a series of blocks rotating clockwise above the obliquely subducting Pacific plate (Geist et al., 1988; Ruppert et al., 2012). Locally, oblique convergence in the central Aleutian arc also causes dextral strike-slip offset along the Hawley Ridge shear zone, which disrupts stratigraphy and structure of the forearc basin and controls lateral and vertical displacement of the outer-arc high (Ryan and Scholl, 1989). Effects of oblique convergence on the upper plate are not restricted only to the forearc region; in the central Aleutian arc, sinistral strike-slip seismicity within the volcanic arc massif itself and even behind it reflects Riedel shearing caused by slip partitioning associated with oblique convergence (Ruppert et al., 2012), showing that the plate boundary behaves as a broad zone encompassing the entire forearc–arc region.

3.3. Arc rifting

Arc extension and rifting is a common process in oceanic systems and has affected most western Pacific arcs more than once in the life cycle of each. Because the arc magmatic front tends to have the weakest lithosphere, extension caused by slab rollback often is localized there, though rifting can also occur behind or in front of the arc axis (backarc rifting or forearc rifting; (Taylor and Karner, 1983; Marsaglia and Devaney, 1995). In initial phases arc rifting is manifest as intra-arc grabens, such as the Sumisu Rift of the Izu–Bonin Arc (Taylor et al., 1991). These basins are structurally segmented and sedimentation is dominated by pumice and proximal volcaniclastic breccias, mostly of a basaltic composition at least initially (Gill et al., 1990; Taylor et al., 1990); as rifting progresses, the volcanic and volcaniclastic products can be intermediate to felsic (Marsaglia and Devaney, 1995). At depth, diking and intrusion are presumed to be extensive, increasing crustal volume (Kodaira et al., 2008; Takahashi et al., 2011) and major igneous complexes in accreted arcs may have formed during this stage of development (Jagoutz et al., 2007). Extension along the magmatic front drives subsidence of the volcanic

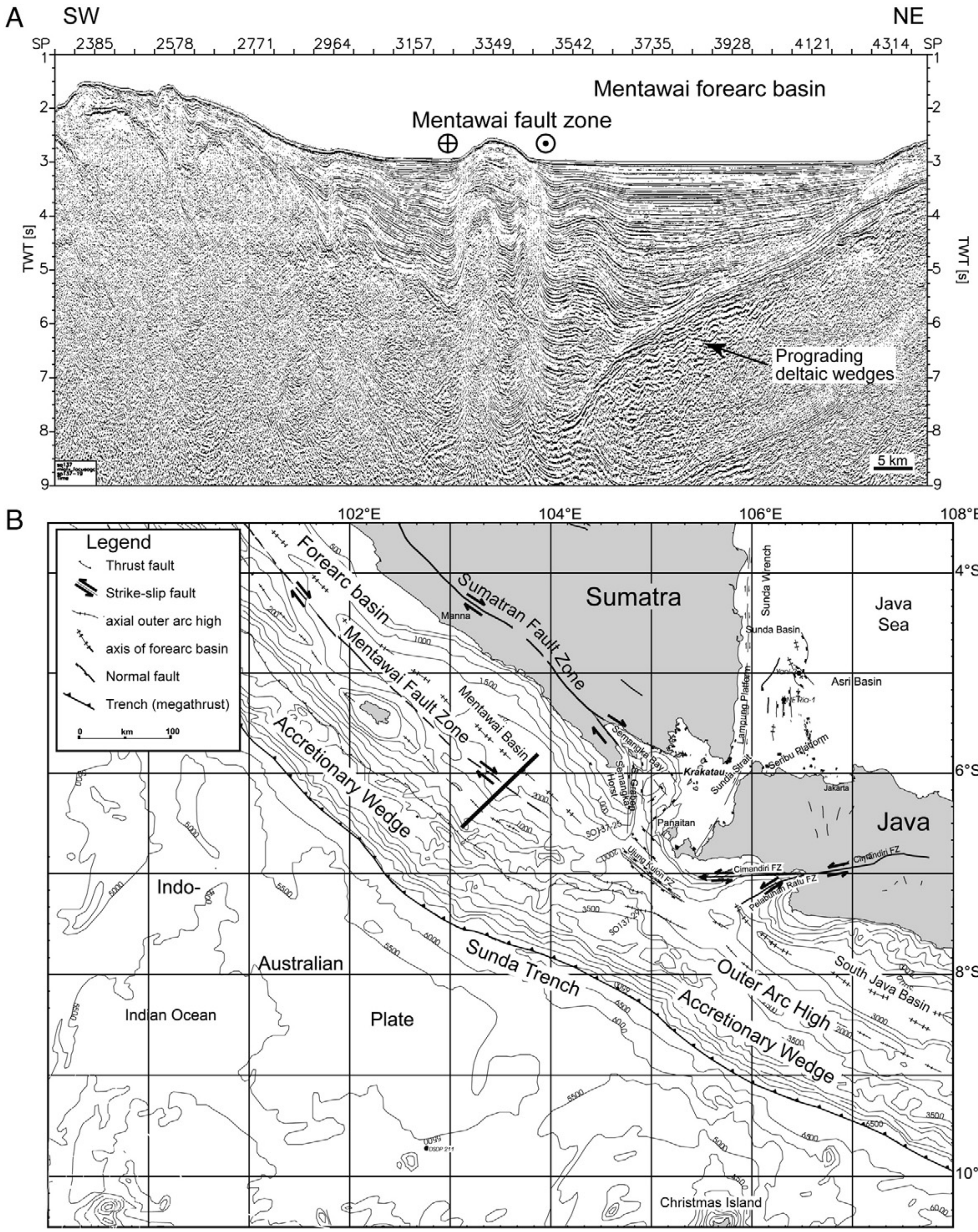

Fig. 5. (A) Seismic-reflection profile SO137-19 off Sumatra (Mentawai Basin, Paleogene deltaic sequences, Neogene basin infill; (Schlüter et al., 2002). Basin is locally inverted by margin-parallel strike-slip faults linked to oblique convergence, but the basin as a whole is expected to have high preservation potential. (B) Bathymetric map showing the region of the central Sunda Arc between Java and Sumatra. Solid line shows the location of Profile SO137-19. Modified from Schlüter et al. (2002).

arc and the drowning of subaerial volcanic centers, in turn preventing airfall tephra sedimentation. Rifting of the Lau Basin, for example, resulted in a hiatus in volcanic sedimentation as the remnant Lau Ridge separated from the forearc platform of the Tonga arc (Clift, 1995). Because rifting is commonly asymmetric, sedimentation rates can vary substantially between the inner and outer rift flanks, as documented during drilling into the Sumisu Rift in the IBM arc (Klaus et al., 1992; Marsaglia, 1995). In the Tonga arc the largest part of the arc sediment source was left on the backarc side of the newly opening basin, whereas the forearc basin received only pelagic sedimentation for 3–4 m.y. until a new magmatic front formed on the trenchward side of the basin (Parson and Hawkins, 1994; Clift, 1995).

Regional unconformities in forearc basins have been linked to arc rifting events (Austin et al., 1989; Tappin et al., 1994). These disconformities imaged on seismic lines must form by tilting of the forearc, but apparently not to such a degree as to cause uplift and subaerial exposure, because sediments on the forearc show continuous water depths of hundreds of meters (Hussong and Uyeda, 1982; Clift et al., 1994) and modern arc extensional regions (such as the northern Marianas or Kermadec arc) do not show water depths greatly different than other parts of the arc system. Following initial rifting, subsidence rates accelerate but then slow again as backarc extension focuses along seafloor spreading centers. Although extension of the forearc is spatially widespread during rifting, the degree of extension can be modest, such that strain is accommodated mostly in a central intra-arc rift zone. The central Tonga forearc includes major along-strike segmentation of depocenters linked to trench-perpendicular faults accommodating the motion of rigid blocks 50–100 km across, inferred to result from structural fragmentation of the arc during Eocene arc rifting (Tappin et al., 1994). The oldest backarc crust contains small grabens separated by horsts and short-lived volcanic seamounts that shed proximal volcaniclastic debris and thick turbidites into adjacent basins. Drilling of these basins, such as the Lau Basin in the Tonga backarc, indicates that they are short-lived (3–5 m.y.) and rapidly revert to slow, pelagic sedimentation as the locus of extension migrates. A zone of extended arc crust ~90 km across now dominates the western side of the Lau Basin (Taylor et al., 1996).

The culmination of rifting is the establishment of a seafloor spreading axis, usually as a result of propagation in a V-shaped basin. Seafloor spreading typically parallels the establishment of a new magmatic arc on the trenchward side of the basin (Parson and Hawkins, 1994). Drilling on the spread crust reveals a sediment cover similar to those on mid-ocean ridges, i.e., basal hyaloclastite breccias, hydrothermal deposits, and pelagic sediment with rare airfall tephra deposits (Parson et al., 1994; Marsaglia and Devaney, 1995). Because of the rough, tectonically induced topography at new spreading centers, turbidites

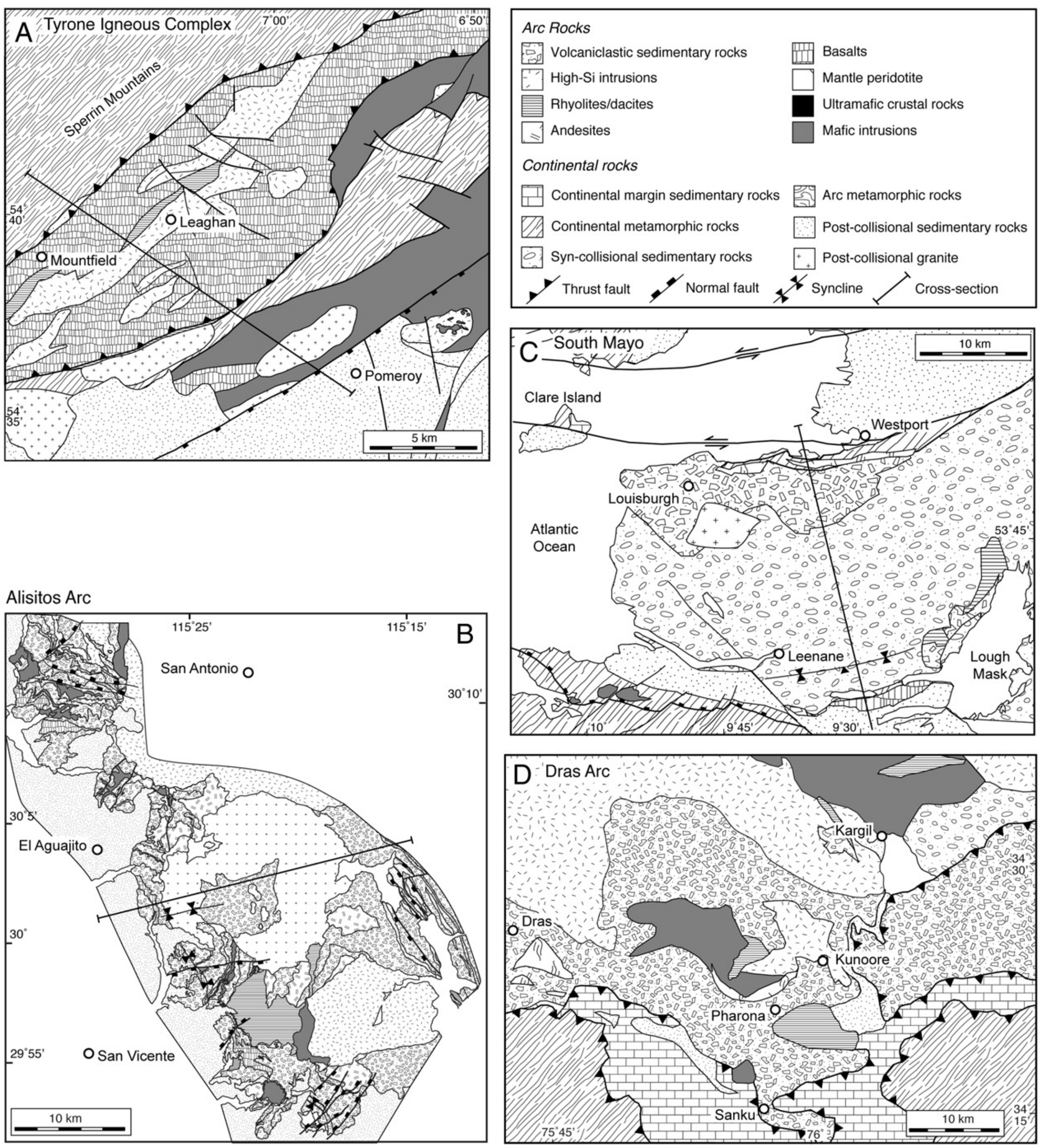

Fig. 6. Geologic maps of accreted oceanic arc terranes showing the main lithologies and crustal units preserved. (A) Tyrone Igneous Complex, Ireland (Hartley, 1933; Cooper et al., 2011), (B) Alisitos arc, Baja California (Busby et al., 2006), (C) South Mayo, Ireland (Graham et al., 1989), (D) Dras arc, India (Reuber, 1989), (E) Talkeetna arc, Alaska (Rioux et al., 2010), (F) Kohistan arc, western Himalaya (Burg et al., 2006), (G) Taiwan Coastal Ranges (Barrier and Angelier, 1986), (H) Kamchatka–Olyutorsky arc (Hourigan et al., 2009), (I) Magnitogorsk Arc, Urals, Russia (Brown et al., 2006b), and (J) Macquarie arc, SE Australia (Glen et al., 2007b).

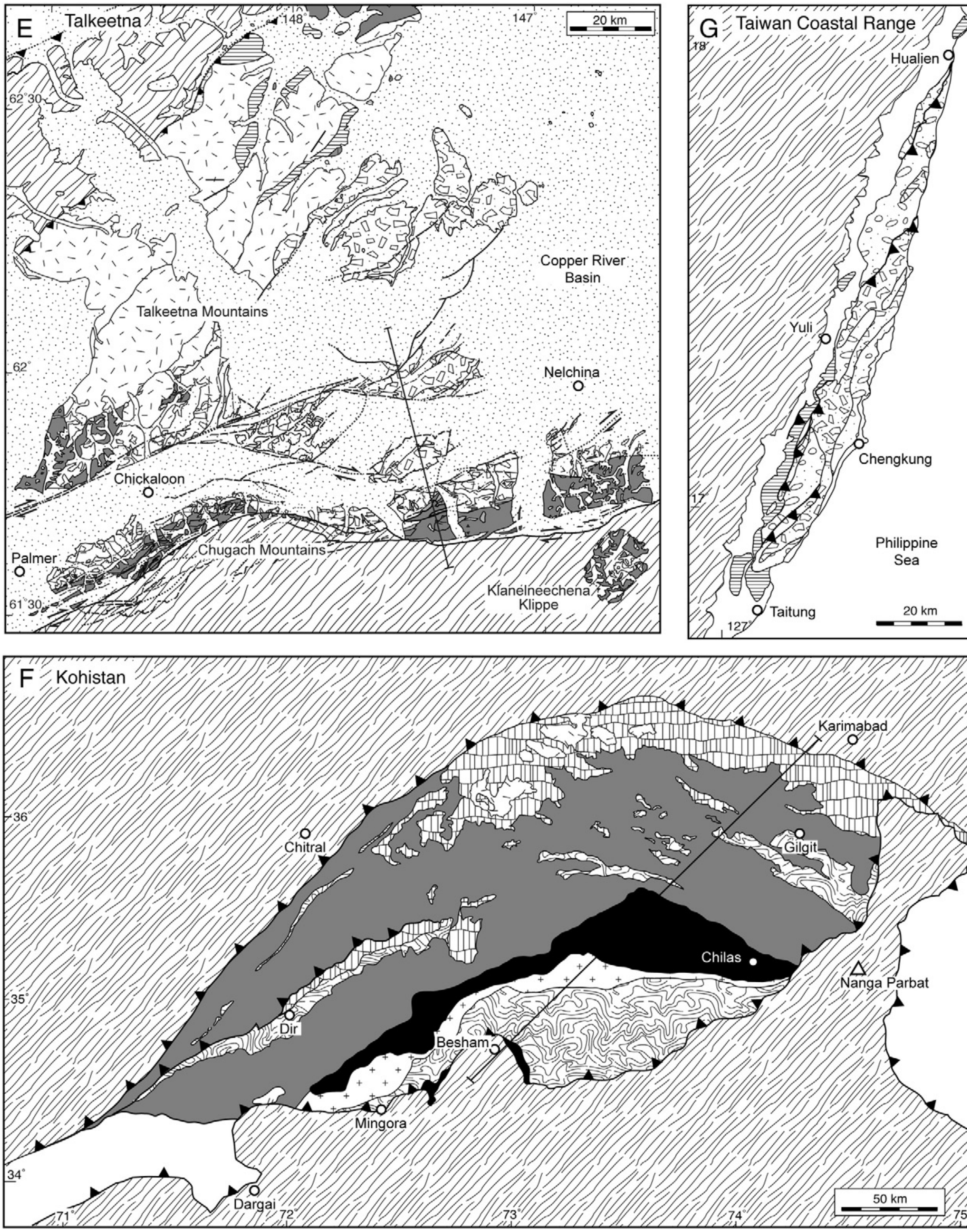

Fig. 6 (*continued*)

from the remnant or new arc generally do not occur in the basin center, but pond close to the basin edges (Clift, 1995; Draut and Clift, 2006). Steep, tectonically controlled backarc topography also controls the distribution of pyroclastic and mass-wasted debris flows (Sigurdsson et al., 1980). Pelagic sedimentation is partly controlled by the water depth. New backarc basins are found at a range of water depths that appear linked to the dip of the subducting plate (gentle dips corresponding to shallower basins) (Park et al., 1990), but older basins, such as the Philippine Sea, often are deeper and so contain less carbonate material, having subsided below the carbonate compensation depth (Klein, 1985; Higuchi et al., 2007). Sediment cover in backarc basins is generally thin (<100 m) in distal, oceanic settings, but thicker close to the arc. In examples where an intraoceanic arc develops near a continental margin, sediment supply can be much greater and the backarc crust has a clastic turbidite cover dominated by terrigenous continental rather than volcaniclastic sediments (Packer and Ingersoll, 1986), e.g., Shikoku

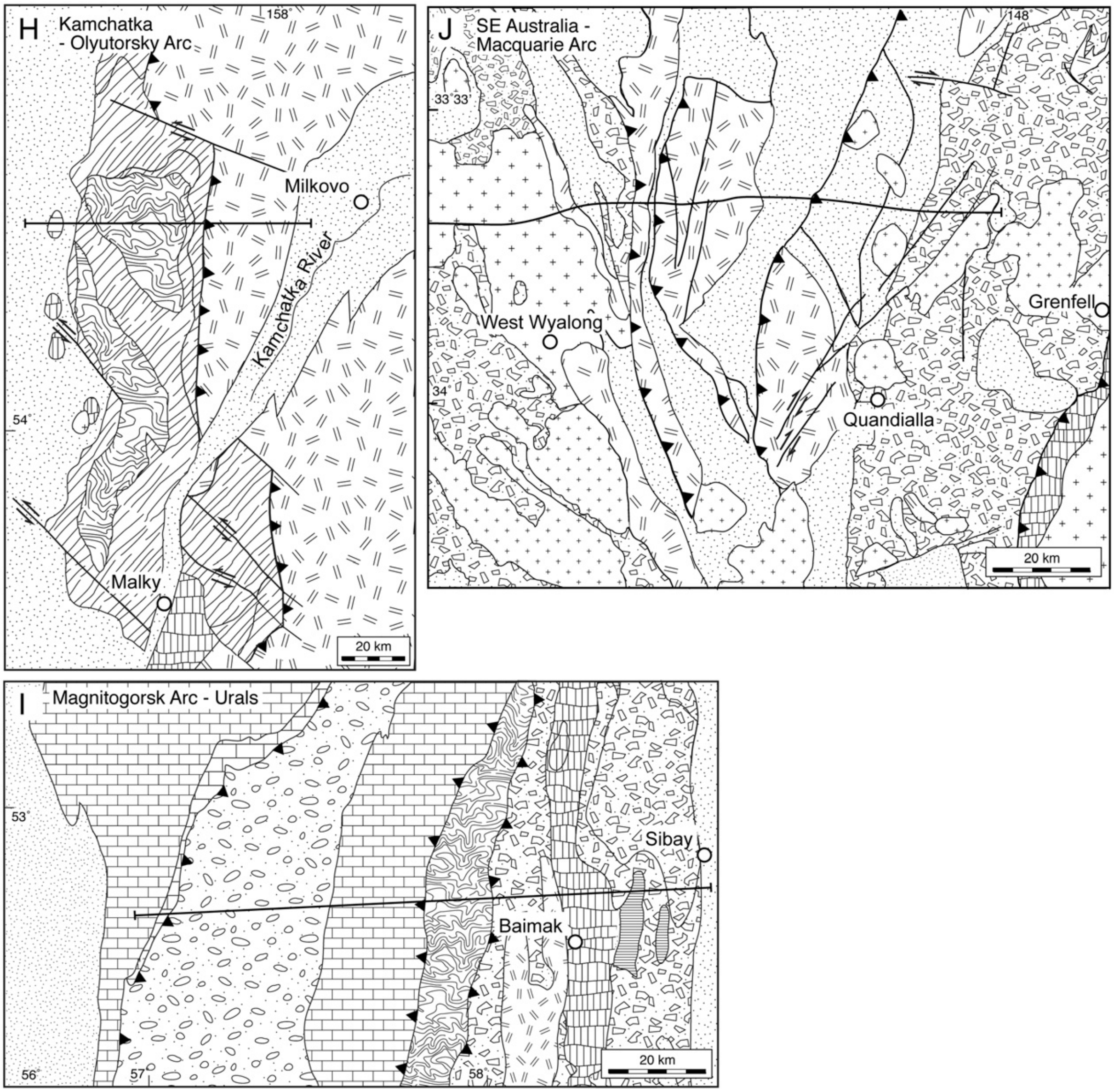

Fig. 6 (*continued*)

Basin, Sea of Japan, and Okhotsk Sea, the fringing Alisitos arc terrane of Baja California (Centeno-Garcia et al., 2011), or the Ordovician Macquarie arc of Australia (Glen et al., 2007a). This is true particularly in high latitudes during continental glaciation, even in retro-arc basins where arc rifting did not affect the sedimentary fill—e.g., the Bering Sea and Kamchatka Basin, which represent oceanic crust that was trapped during plate reorganization (Scholl and Creager, 1973).

4. Modification of arc terranes during arc–continent collision

If an intraoceanic subduction zone remains active long enough, eventually it will collide with a continental margin. Arc–continent collision then forms an orogenic belt that may contain one or more accreted arc terranes, of which there are numerous examples now exposed on land (Figs. 6 and 7). Below, we summarize arc–continent collision processes and the resulting fate of trenches and trench-slope basins, forearcs, arc massifs and intra-arc basins, and backarc basins. Structural and stratigraphic alteration, loss (by partial subduction or tectonic dismemberment), metamorphism, and differential preservation of these regions and deeper crust and upper mantle as a result of collision determine how faithfully accreted terranes preserve sedimentary, structural, and magmatic records of the original intraoceanic arc, and so control what evidence of the subduction zone remains after its transition from activity into the geologic record. The final composition of accreted arc terranes is of particular importance in the post-Archean assembly of continents (Whitmeyer and Karlstrom, 2007), as arc–continent collision is thought to be a key process controlling the formation of continental crust (Pearcy et al., 1990; Rudnick, 1995; Holbrook et al., 1999).

4.1. Collision processes

The likelihood that the geologic record will preserve evidence for contemporary tectonic processes of an active arc depends greatly on how well the various parts of the arc survive collision and orogeny. ‘Forward-facing’ collision, in which the arc faces the continent and collides trench- and forearc-first, involves a substantially different geometry than does ‘backward-facing’ collision, a situation involving two subduction zones with similar polarity and in which the arc backs into the continent (Fig. 8). Forward-facing collision can accrete an oceanic arc onto either a passive or an active continental margin,

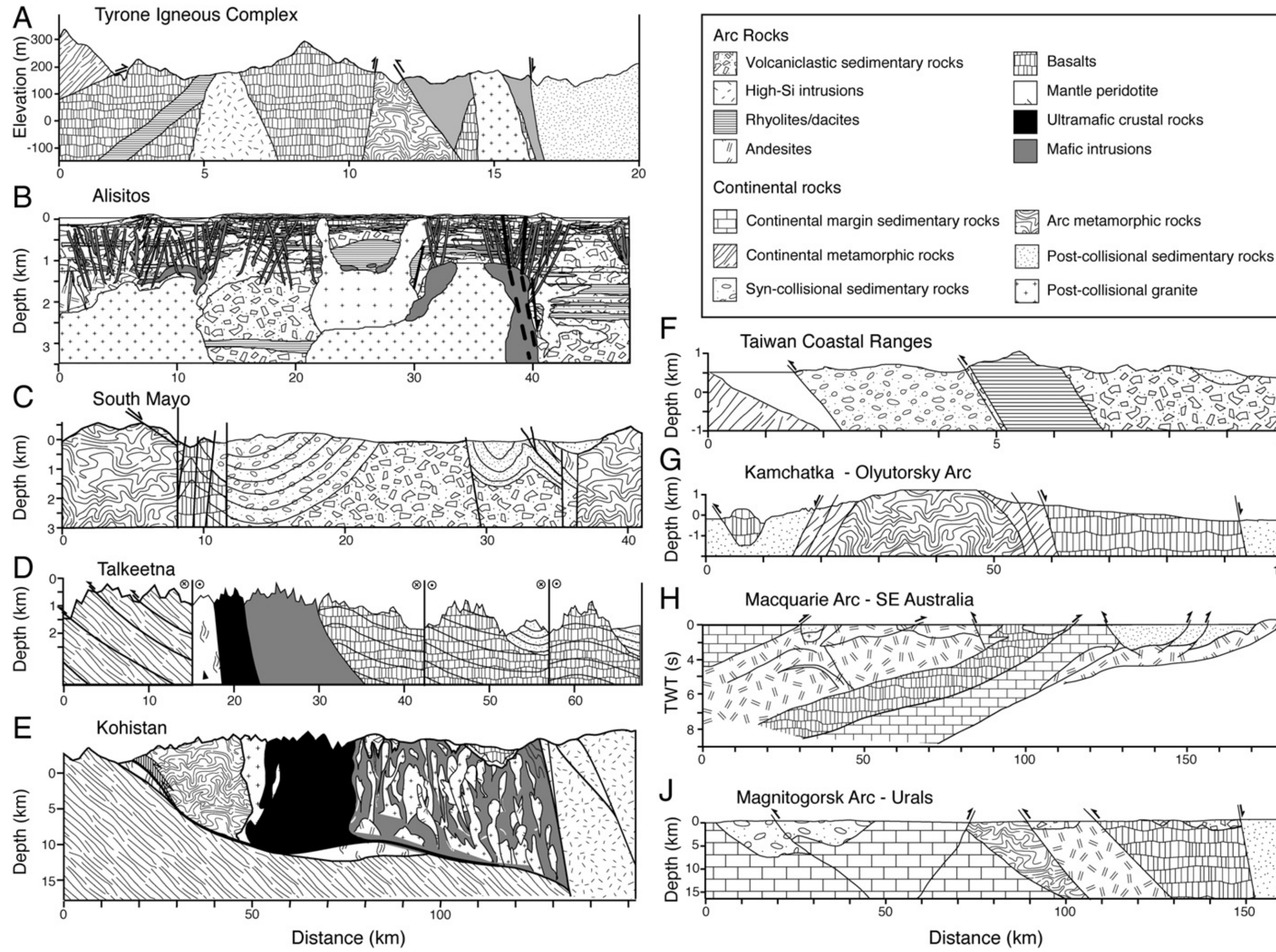

Fig. 7. Cross sections through a series of accreted oceanic arcs. (A) Tyrone Igneous Complex, Ireland (Chew et al., 2008), (B) Alisitos Arc, Baja California (Busby et al., 2006), (C) South Mayo, Ireland (Graham et al., 1989), (D) Talkeetna Arc (Clift et al., 2005a), (E) Kohistan, western Himalaya (Burg et al., 2006), (F) Taiwan Coastal Ranges (Barrier and Angelier, 1986), (G) Kamchatka–Olyutorsky Arc (Hourigan et al., 2009), (H) Macquarie Arc, SE Australia (Glen et al., 2007b), and (J) Magnitogorsk Arc, Urals, Russia (Brown et al., 2006b).

whereas backward-facing collision must join the oceanic arc to an active continental margin. As discussed below, the post-collisional composition of an arc terrane differs substantially as collision occurs in each case.

Examples of ancient intraoceanic arcs thought to have accreted by forward-facing collision with a passive margin (Fig. 8A) occur in Devonian terranes of the Urals (Brown et al., 2006a, 2011a; Puchkov, 2009), in an Eocene terrane on New Caledonia (Aitchison et al., 1995; Spandler et al., 2005), in the Precambrian Anti-Atlas of Morocco (Thomas et al., 2002), within the extensive Ordovician Appalachian–Caledonide suture of North America and the British Isles (Dewey and Ryan, 1990; van Staal et al., 1998, 2007), and in the Neotethyan suture (Robertson, 2002; Dilek and Flower, 2003). Forward-facing collision of an oceanic arc with an active continental margin (Fig. 8B) may have occurred during the Jurassic Nevadan Orogeny, emplacing arc-derived ophiolites along the west coast of North America (Garcia, 1982; Ingersoll and Schweickert, 1986; Godfrey and Klemperer, 1998), though this tectonic model is not universally accepted. Ancient arcs that collided by backing into continents (Fig. 8C) include the Jurassic Talkeetna–Bonanza arc of Alaska and British Columbia (Burns, 1985; Plafker et al., 1989; Clift et al., 2005b), the Cretaceous Alisitos fringing arc of Baja California (Busby, 2004; Busby et al., 2006; Centeno-Garcia et al., 2011), and possibly the Proterozoic Nahanni arc terrane of the Wopmay orogen, Canada (Cook, 2011). The Cretaceous Dras–Kohistan arc of the western Himalaya is thought to have undergone first a forward-facing collision with India and then backward-facing collision with the Asian (Karakoram) margin (Burg, 2011), although several alternatives have been proposed (Khan et al., 1997). Arc–continent collision in Eocene sutures of Kamchatka has been variously interpreted as forward-facing (Konstantinovskaia, 2001; Konstantinovskaya, 2011) and backward-facing (Geist and Scholl, 1994), whereas Hourigan et al. (2009) concluded that field evidence does not allow the polarity of that arc to be resolved clearly. Collision geometry remains similarly unresolved for a Permian accreted arc in the Altaids of Mongolia (Yang et al., 2012); (Heumann et al., 2012).

Two prominent examples of forward-facing collision between oceanic arcs and passive margins occur in the modern ocean: the high-angle, nearly orthogonal collision of the Luzon arc with the passive margin of Eurasia at Taiwan (Suppe, 1984; Teng, 1990; Huang et al., 2006; Byrne et al., 2011), and collision of the Banda arc with the northern passive margin of Australia (Abbott et al., 1994; Snyder et al., 1996; Lüschen et al., 2011). Ongoing collision of the intraoceanic Izu–Bonin arc with the microcontinental Honshu arc involves a high-angle backward-facing collision at a trench–trench–trench triple junction (Ogawa et al., 1985; Marsaglia, 2012). Also noteworthy is the ongoing collision of the Halmahera and Sangihe arcs in the Molucca Sea (Lallemand et al., 1998; Hall and Smyth, 2008). The latter involves collision of two active margins (similar to Fig. 8B), but because neither arc there is accreting onto a continent evidence of this collision is unlikely to survive in the long-term geologic record. Orthogonal collision of the intraoceanic Aleutian arc with Kamchatka is a unique active arc–

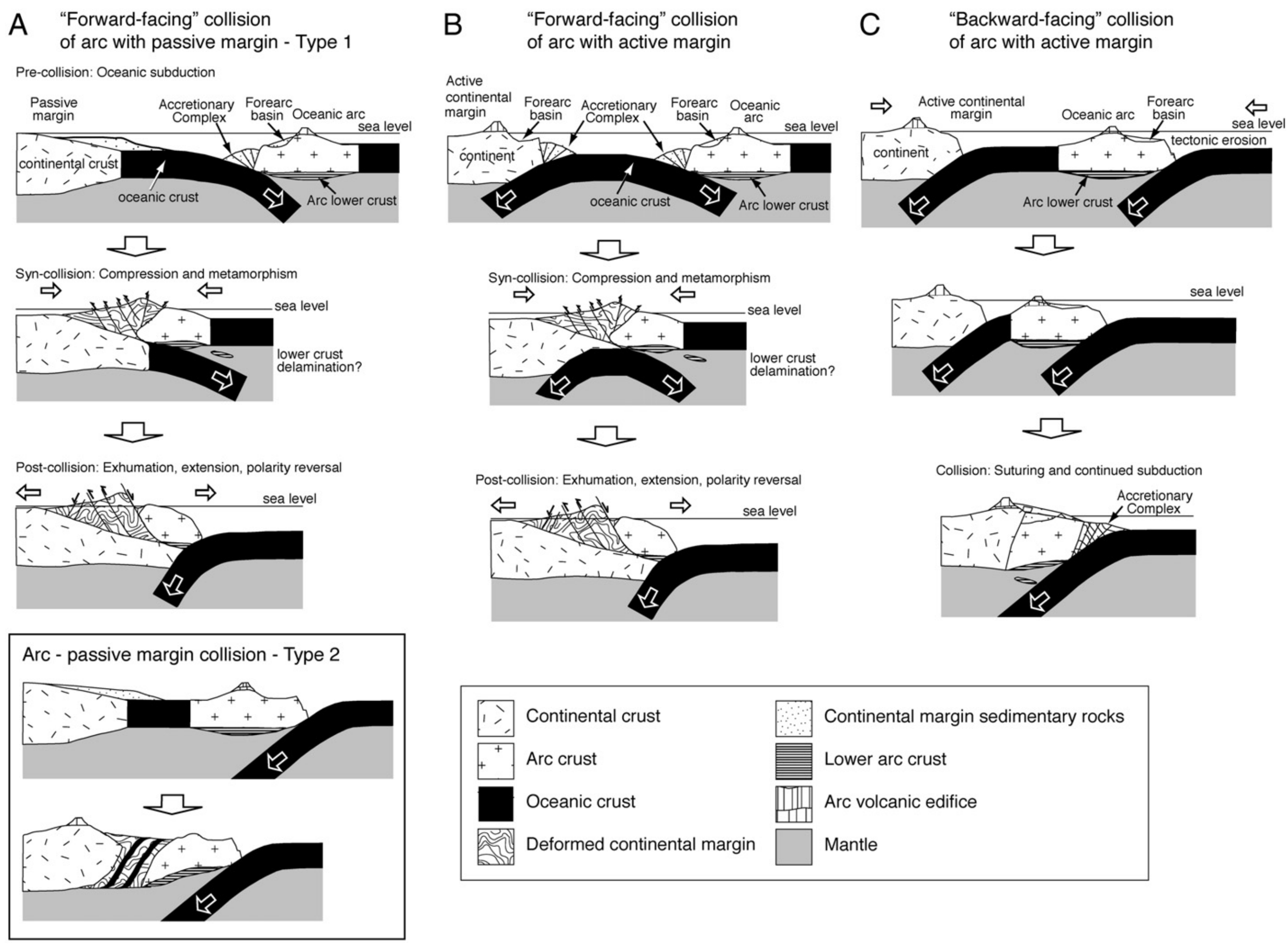

Fig. 8. Tectonic diagrams showing different geometries of arc–continent collision. (A) "Forward-facing" collision (in which collision occurs on the trench and forearc side) of an oceanic arc with a continental passive margin, resulting in a collisional orogen, subduction polarity reversal, and orogenic collapse (e.g., the modern high-angle collision at Taiwan, or Ordovician collision preserved in County Tyrone and South Mayo, Ireland). Figure is modified from Clift et al. (2008). The lower panel (labeled type II) shows the special example of arc–passive-margin collision where the oceanic arc formed close to the continental margin (e.g., Ordovician Macquarie arc, SE Australia). (B) Forward-facing collision of an oceanic arc with a continental active margin. Examples include the Jurassic Nevadan orogeny of western North America (Godfrey and Klemperer, 1998), or the modern orthogonal collision of the Aleutian arc at Kamchatka (although the oblique convergence at the western Aleutians has effectively turned it into a transform margin). (C) "Backward-facing" collision between a continental active margin and the back-arc region of an oceanic arc, resulting in continued convergence (e.g., Kohistan and Talkeetna arcs, or modern high-angle collision between the Izu–Bonin and Honshu arcs), modified after Clift et al. (2005b).

continent collision in that Pacific Plate–Aleutian convergence there is so oblique that the western Aleutians comprise a transform plate boundary rather than active subduction (Geist and Scholl, 1994; Gaedicke et al., 2000; Scholl, 2007). If the western Aleutian arc were approaching Kamchatka at a more acute angle, the collision-zone geometry would resemble the Molucca Sea case and Fig. 8B.

As plate convergence brings an intraoceanic arc into proximity with a continental margin, in the case of forward-facing collision, increased sediment flux to the trench causes important changes even before collision begins. Oceanic arcs remote from continental margins tend to have low rates of sediment supply to the trench. Being mostly submarine, intraoceanic arcs also do not undergo rapid weathering and subaerial erosion, at least not until they begin to uplift while colliding with a continental margin, and so sediment production around arcs generally is low. An exception is the Aleutian arc, which not only receives sediment supply along the trench from the glaciated orogen of western North America, but also has an arc massif that has been subaerially eroded to a great degree since Eocene time (Scholl et al., 1983). For most arcs, even if the open-ocean phase of activity was one of tectonic erosion with little sediment entering the trench, as the arc nears a continental margin in a forward-facing configuration (Fig. 8A,B) the flux of continent-derived sediment to the trench must increase. Increased subduction of continental sediment just before collision can build or enlarge an accretionary prism, though accreted oceanic arcs are more likely to contain metasedimentary rocks in preserved subduction channels than they are to have major accretionary complexes (discussed below). In contrast, arcs that collide with continents in a backward-facing configuration (Fig. 8C) may receive no comparable increase of sediment flux to the trench; pre- and syn-collisional accretionary metasedimentary rocks are minimal or absent in accreted terranes of backward-facing arc collisions (Clift et al., 2005b). If subduction continues at the active continental margin after backward-facing arc collision, however, an accretionary complex can form after collision, built using sediment supply from the newly formed orogen. The Cretaceous Chugach Complex of southern Alaska is one example, comprising accretionary material that formed after the Talkeetna arc accreted (Clift et al., 2012).

Perhaps the most substantial difference between forward- and backward-facing arc-continent collision involves the composition of magmatism during and after collision, with important implications for the formation and maintenance of continental crust in post-Archean time. Although some geochemical similarities indicate that

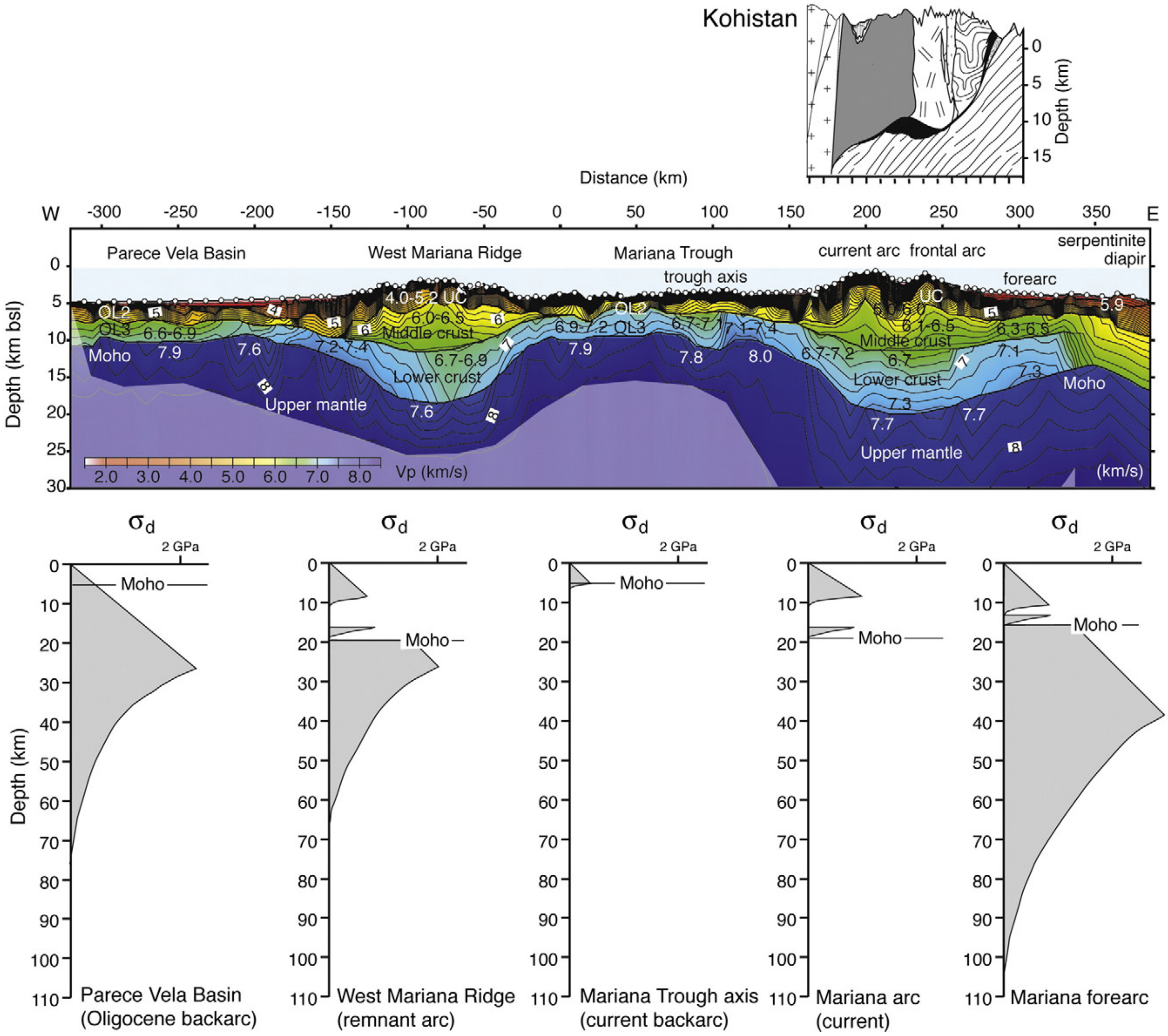

Fig. 9. (A) Seismic-velocity model through the Mariana arc and backarc region, based on refraction data, with p-wave velocities shown in km/s. The scale of the arc complex is shown compared to the largest preserved arc unit from Kohistan (Fig. 7E). Seismic section is from Takahashi et al. (2007). UC = upper crust; OL = oceanic lithosphere. (B) Schematic diagrams showing yield-strength envelopes for arc lithosphere in forearc, backarc, and arc-massif regions. Basins located close to the arc volcanic front or in young backarc rifts tend to be weak (Watts et al., 1982; Watts, 2001), whereas forearc basins and older backarc basins have stronger underlying mantle because they are thermally mature, i.e., cooler (Lin and Watts, 2002; Crawford et al., 2003). Forearc basins are thus presumably less likely than are most backarc basins to be deformed and destroyed during collision and orogenesis. See also modeling by Boutelier et al. (2003).

oceanic subduction-zone magmatism can produce continental crust, most modern arc crust is thought to be too mafic and too depleted in light rare earth elements (LREE) to be an obvious precursor to continental crust, a problem that may be reconciled by geochemical changes during arc-continent collision (Pearcy et al., 1990; Holbrook et al., 1999; Draut et al., 2002). Forward-facing arc–continent collision includes subduction of continental sediment and probably outer continental crust just before collision—materials rich in silica and incompatible trace elements. The continent-derived materials melt, mix with arc-derived magmas, and undergo enough crystal fractionation that syn- and post-collisional mid- and upper-crustal rocks can be higher in silica and more LREE-enriched than is average continental crust, driving the bulk content of accreting arc crust toward that of typical continental crust (Draut and Clift, 2001; Draut et al., 2002, 2004). This geochemical evolution occurred in multiple areas of the Appalachian–Caledonide suture associated with forward-facing arc-continent collision (Draut et al., 2009). In contrast, during backward-facing collision arc geochemistry can remain largely unchanged from its oceanic composition, as in the Talkeetna and Kohistan arc accretions (Clift et al., 2000, 2005a). Production of highly enriched melts from forward-facing, but not backward-facing, collision implies that if arc–continent collision is important to the genesis of andesitic, LREE-enriched continental crust, this process most likely depends heavily upon forward-facing collisions.

In describing the stages and styles of arc–continent collision (Brown et al., 2011b), previous papers have distinguished between 'soft' and 'hard' collision, but with different authors assigning different meaning to those terms. Draut and Clift (2001) defined 'soft' collision as initial subduction of continental sediment and outermost continental crust before orogeny begins, and 'hard' collision as orogeny involving subsequent regional deformation and metamorphism. Van Staal et al. (1998) and Zagorevski and van Staal (2011) used 'soft' and 'hard' collision to specify not a sequence of events, but rather behavior of the upper plate during a collision event—a hard collision being one in which the upper plate remains essentially intact but undergoes tectonic thickening associated with deformation and metamorphism, whereas soft collision would involve no significant

Table 1
Proportions of different arc lithologies found in some of the arcs profiled in this synthesis.

Name	Location	Age	Collision mechanism	Volcaniclastic sedimentary rock	Basalt	Andesites	Rhyolites/dacites	High-Si intrusion	Mafic intrusions	Ultramafic crust	Mantle peridotite
Tyrone Igneous Complex	Northern Ireland, U.K.	Ordovician	Arc-passive margin, Type 1	0	45	0	5	12	38	0	0
Lough Nafooey arc	County Mayo, Ireland	Ordovician	Arc-passive margin, Type 1	58	5	15	15	0	5	0	2
Alisitos arc	Baja California, Mexico	Early Cretaceous	Arc-passive margin, Type 2	30	5	10	10	40	5	0	0
Dras arc	Kashmir, India	Early Cretaceous	Arc-Arc, Backward-facing	42	20	0	8	5	20	0	5
Kohistan arc	Pakistan	Early Cretaceous	Arc-Arc, Backward-facing	0	20	0	0	22	35	15	8
Talkeetna arc	Southern Alaska	Jurassic	Arc-Arc, Backward-facing	30	8	12	7	30	10	2	1
Taiwan Coast Ranges	Taiwan	Miocene-Pliocene	Arc-passive margin, Type 1	65	0	15	20	0	0	0	0
Olyutorsky arc	Kamchatka, Russia	Cretaceous–Eocene	Unclear	20	5	60	15	0	0	0	0
Magnitogorsk arc	Ural Mountains, Russia	Devonian	Arc-passive margin, Type 1	35	30	20	15	0	0	0	0
Macquarie arc	Australia	Ordovician	Arc-passive margin, Type 2	45	5	30	0	20	0	0	0
Average				33	14	16	10	13	11	2	2

thickening or metamorphism of the upper plate, and some subduction of the forearc region. 'Soft' collision also has been used to describe the Kohistan–India arc–continent collision that occurred before final closure of the Tethys Ocean (Burg, 2011). Zagorevski and van Staal (2011) noted that the surface manifestation of these soft and hard collision styles depends considerably on the erosion level of the orogen. Interpreting how substantially collision altered either plate, and with what timing, is challenging not only because of surficial exposure but also because arc crust is inherently variable and complex (Calvert, 2011; DeBari and Greene, 2011), because collision timing and geometry are complicated by promontories and embayments in most margins (Brown et al., 2011b), and because compressive and extensional forces can cause great spatial variation in the collision-zone morphology (Whitmore et al., 1997). Collision timing is almost always diachronous along strike because of oblique convergence directions and non-linear continental margins.

Field investigations of accreted arc terranes have shown clearly that preserved arc sequences are not as complete (vertically or laterally) as those of active arcs in the modern oceans. Only one ancient arc comes close to preserving a complete arc volume, namely the Dras–Kohistan of the western Himalaya (Fig. 7E). In terms of its mass the Kohistan block is comparable to the mass of the arc edifice in the Marianas (Fig. 9) (Takahashi et al., 2007), but even in this respect the Kohistan arc is missing much of the forearc and any remnants of the backarc basin. Surface exposures of most other preserved, accreted arc complexes tend to be dominated by volcanic rocks of various compositions and their associated volcaniclastic sedimentary aprons (Aitchison et al., 1995; Dewey and Mange, 1999; Critelli et al., 2002; Clift et al., 2005a; Trop et al., 2005; Brown et al., 2006b; Zagorevski et al., 2009), whereas lower and mid-crustal rocks are more poorly represented in surface exposures and are volumetrically minor, at least in outcrop (Fig. 6, Table 1). With some notable exceptions (DeBari and Coleman, 1989; Leake, 1989; Draper et al., 1996; Burg et al., 2006; Greene et al., 2006; DeBari and Greene, 2011), knowledge of the deeper crust in accreted arc terranes tends to be poor or non-existent (Fig. 7). Precambrian ophiolites may represent an exception, with examples such as the Amalaoulaou Complex within the Pan-African belt of Mali exposing extensive lower and middle crustal intra-oceanic arc rock (Berger et al., 2011). The geochemistry there points to melting from a depleted mantle source with crystallization at depths of 25–30 km. Other Pan-African ophiolites also show lower crust and mantle sections, but additionally preserve higher level units, including pillow basalts, to form a classic Penrose-type assemblage (Stern et al., 2004). Geochemically a range of compositions has been identified spanning MORB, backarc-basin basalt, arc tholeiite, and boninite, interpreted to indicate their origin within a forearc setting (Stern et al., 2004). However, in Phanerozoic accreted arcs lower crust and mantle exposures are generally uncommon. An example comes from the Coastal Ranges of Taiwan, where the intraoceanic Luzon arc is colliding with the Eurasian continental margin. There, outcrop exposures include a relatively small volume of geochemically evolved volcanic and volcaniclastic rocks (Fig. 6G). However, mass balancing across Taiwan suggests that much of the root of the island comprises the colliding Luzon arc and that this is simply buried under the thrust wedge of deformed passive-margin material (Fig. 10) (Clift et al., 2009). In many accreted arc sections so little is known of the deep crustal structure that a mass balance is impossible and the possible presence of lower arc crust at depth remains untested.

Petrologic calculations, geodynamic modeling, and seismic structure indicate that even in accreted arcs with exposures thick enough to include mid–lower crust and upper mantle, some additional amount of lower crust likely foundered and sank deeper into the mantle, lost by delamination or convective instability of dense cumulates (Klemperer et al., 1991; Kay and Kay, 1993; Jull and Kelemen, 2001; Behn and Kelemen, 2006; Greene et al., 2006; DeBari and Greene, 2011). Seismic tomography reveals such processes beneath the active collision zone of northern Taiwan, where lower crust and upper mantle apparently delaminate and subduct (Wu et al., 2007); ongoing lithospheric delamination also has been inferred for the collision zone along the island of New Guinea (Cloos et al., 2005). Loss of some dense, lower-crustal matter can occur not only by density instability at depth, but also via exposure and erosion. Heavy-mineral studies from the McHugh Complex, an uplifted Cretaceous accretionary prism in south-central Alaska, indicate that shortly after final accretion of the Talkeetna arc to North America, high-Mg diopsides and garnets, typical components of lower crustal rocks, were delivered to the trench. This requires that the lower parts of the Talkeetna arc crust were exposed and eroded at that time. Deformation associated with backward-facing collision of the Talkeetna arc evidently had uplifted deep regions of the arc section, allowing lower-crustal material to be recycled rather than preserved intact in the newly amalgamated crust (Fig. 11) (Clift et al., 2012). Loss of dense lower arc crust after collision is thought to help drive the composition of continental crust toward its modern bulk andesitic content

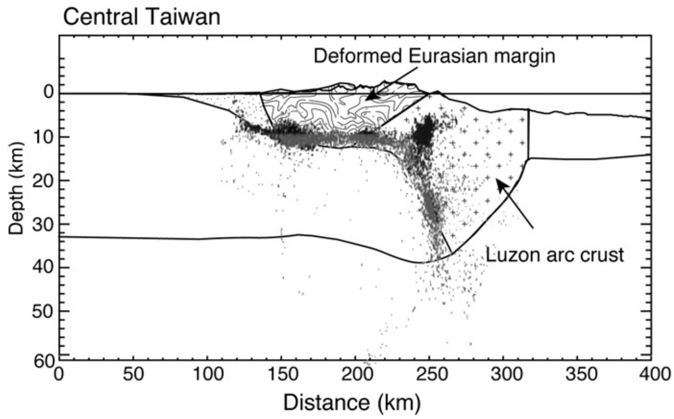

Fig. 10. Cross section through the Taiwan collision zone, at the collisional maximum in central Taiwan. Crustal structure under Taiwan is inferred from the seismic evidence of Carena et al. (2002) shown as black and gray dots projected on to the section from a number of major faults in the central Taiwan region. This section shows that much of the deep crust under Taiwan is arc-derived and sits above the main detachment surface.

(Rudnick and Fountain, 1995; Holbrook et al., 1999; Behn and Kelemen, 2006). Hacker et al. (2011) proposed that an inverse process, relamination, also could be important, wherein felsic and intermediate material derived from subducted arcs, sediment, or continents form silica-rich melts that rise buoyantly and attach onto the lower crust; the extent to which this occurs as arcs accrete onto continents (thus becoming part of continental crust) is presently unclear.

4.2. Post-collisional fate of arc regions

Complete lateral sections of intraoceanic arcs (from backarc to forearc) seem never to survive collision with a continent intact, but the fates of trenches and trench-slope basins, forearc regions, arc massifs and intra-arc basins, and backarc basins differ substantially depending on whether collision is forward- or backward-facing, and on whether the oceanic arc was tectonically eroding or accreting for a long time before collision.

It is unlikely that trench-slope basins can be preserved intact or remain identifiable after arc–continent collision, although sediment that was underplated or formed part of a frontal accretionary prism may be. In a forward-facing collision, the trench region bears the brunt of earliest collision, and can be rapidly uplifted and eroded. In a backward-facing collision, the trench is sheltered from the main collision zone (Fig. 8). Nonetheless, if subduction continues at the trench of an intraoceanic arc subduction erosion will progressively destroy the trench slope. Because trenches and their sedimentary records are destroyed continually by tectonic erosion throughout most of the life of an oceanic arc, long-term preservation of any trench sediment is biased toward margins that were tectonically accreting for a substantial length of time before collision.

If accretionary-prism material is preserved in an orogen, it may be difficult to distinguish from volumetrically less important trench-slope-basin deposits (Dickinson, 1982), although the accretionary prism may be more deformed. Because of their greater sediment supply to the trench shortly before collision, arc terranes that accreted by forward-facing collision are more likely than those accreted by backward-facing collisions to include trench or subduction-channel metasedimentary rocks, even though the trench is commonly the first region to be deformed by forward-facing collision (Abbott et al., 1994; Snyder et al., 1996). The most substantial difference expected between forward-facing collision of an oceanic arc with a passive margin compared to with an active margin (Fig. 8A vs. 8B) would be the greater amount of accretionary-complex metasediment involved in collision with an active margin, as two trenches contribute material (Fig. 8B). This is consistent with the great thickness of accretionary-complex exposures along the west coast of North America that likely formed in part by this collision geometry (Ingersoll and Schweickert, 1986), although they were also subsequently overprinted and added to by post-collisional Cretaceous (Franciscan) subduction accretion (Wakabayashi, 1992; Tagami and Dumitru, 1996).

Several arc terranes that accreted by forward-facing collision include deformed accretionary metasedimentary rocks, such as the Clew Bay complex of the Ordovician Lough Nafooey arc (South Mayo Trough) in western Ireland (Dewey and Ryan, 1990; Ryan and Dewey, 2011) and the Southern Urals accretionary complex of the Devonian Magnitogorsk arc terrane (Alvarez-Marron et al., 2000). Large accretionary complexes in arc terranes probably formed in a manner comparable to modern uplift of the 6–7-km-thick accretionary wedge in the Timor Trough and Erap Complex of Papua New Guinea against the Australian continental margin (Abbott et al., 1994; Snyder et al., 1996; Roosmawati and Harris, 2009); underplated metasedimentary rocks also have been uplifted and exposed by that collision (Warren and Cloos, 2007). However, the suture between the Dras–Kohistan arc and the passive margin of India, a forward-facing collision (Burg, 2011), shows no accretionary prism between the forearc (Indus Group) and the telescoped Indian passive margin slope (Lamayuru Complex) (Garzanti et al., 1987; Robertson and Degnan, 1993). Those two units are separated only by an ophiolitic mélange several hundred meters wide. Likewise, accretionary metasedimentary rocks are largely missing from Ordovician arc exposures resulting from forward-facing collision in the northern Appalachians (Zagorevski and van Staal, 2011). Accretionary prisms that do survive in forward-facing arc–continent collision zones can be greatly structurally altered, being sliced into imbricated thrust sheets and overthrust by part of the arc itself, as in collision between the Bismarck arc and Australia at Papua New Guinea (Abbott et al., 1994).

One feature of trench regions that can be well preserved is the subduction channel, a zone generally 100 to 1000 m thick between the subducting and overriding plates containing sheared sediment (Shreve and Cloos, 1986; Vannucchi et al., 2012) and commonly mapped as ophiolitic mélange (Robertson, 2000; Vannucchi et al., 2008; Bachmann et al., 2009). Whereas subduction-channel sediment may be underplated below the forearc in accretionary margins, it is presumed to subduct into the upper mantle at tectonically erosive margins. Subduction channels are naturally dynamic features (Collot et al., 2011), but can be preserved because subduction accretion developed just before forward-facing arc–continent collision. Introducing large volumes of sediment into a trench formerly dominated by tectonic erosion results in accretionary-prism development by both underplating and frontal accretion. This effectively removes pre-existing subduction-channel material from the dynamic plate boundary and preserves it as the structurally uppermost and oldest part of the new accretionary prism. A good example is the mesomélange of the McHugh complex, Alaska, preserved because collision caused tectonic accretion as more sediment was delivered to the trench (Amato and Pavlis, 2010; Clift et al., 2012). Fig. 3B shows a typical outcrop of sheared mudstones enclosing blocks of basalts, radiolarian cherts, and limestone olistoliths. Ophiolitic mélange zones in other mountain belts worldwide are recognized as remnants of subduction channels, distinguished by lithology and modest structural thickness—tens to hundreds of meters, in contrast to the 40- to >100-km width of accretionary wedges in the Nankai Trough, Japan, or offshore continental Alaska (Brocher et al., 1994; Gulick et al., 2004; Underwood and Moore, 2012). The example shown in Fig. 3B is believed to be a subduction channel because of its strongly sheared character and its location against the Border Ranges Fault, interpreted as a paleo-subduction-zone thrust (Clift et al., 2005b). Other examples occur in the sub-ophiolitic Haybi Mélange of Oman (Robertson, 1987), the Maksutovo Complex in the Urals (Brown et al., 2011a), the Tienhsiang Formation mélange of Taiwan (Hsu, 1988), the South

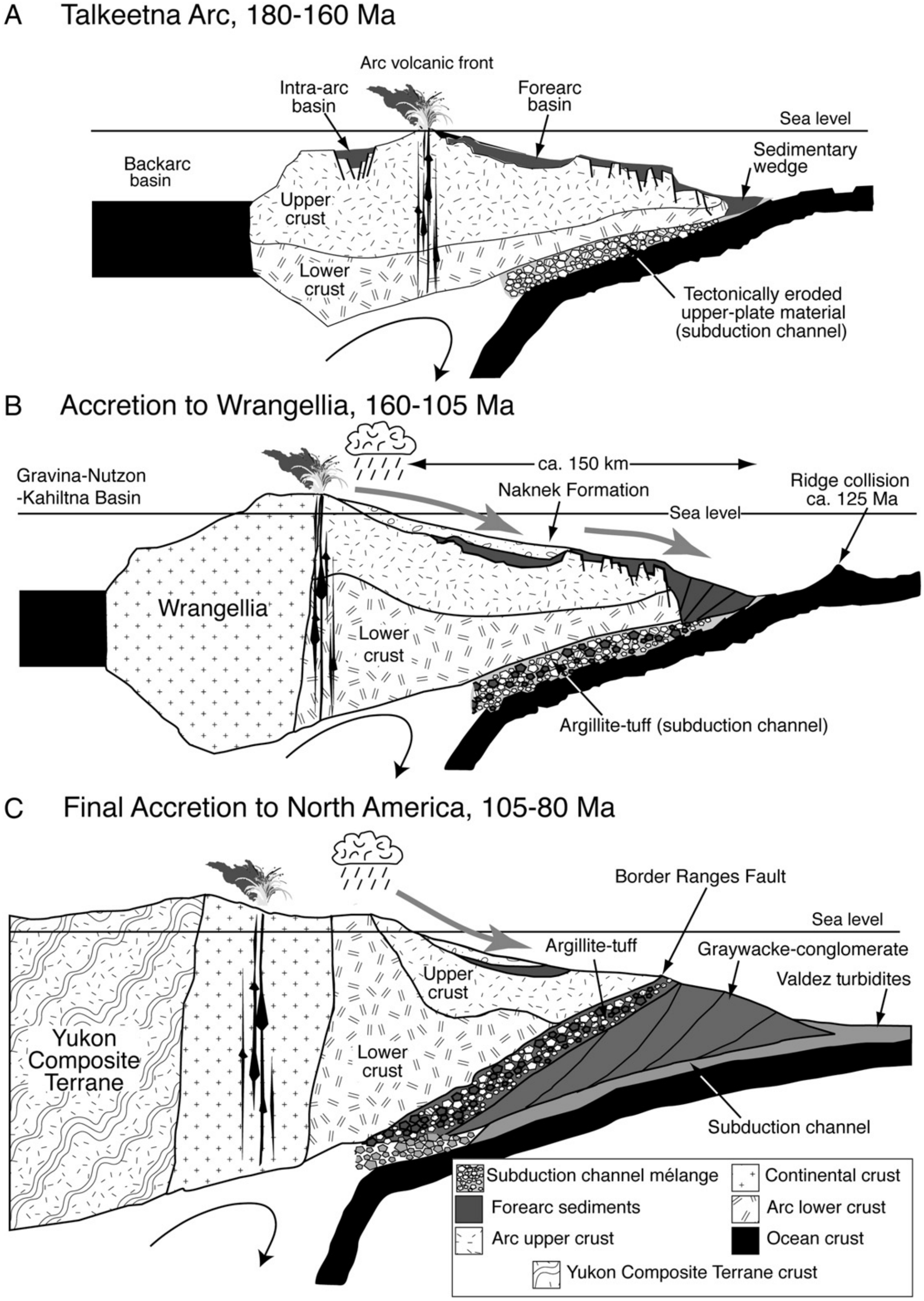

Fig. 11. Example of arc–continent collision resulting in accretion of a large crustal section to the continental margin. Model is from the Jurassic Talkeetna arc terrane of Alaska (Clift et al., 2012). Note that accretion results in a change from tectonic erosion to tectonic accretion, preservation of the original subduction channel, and progressive deformation and exhumation of the oceanic arc lower crust.

Penninic mélange of the European Alps (Bachmann et al., 2009), and a basaltic mélange in the Indus–Yarlung suture zone in central Tibet (Aitchison et al., 2000) (Fig. 3).

The forearc region of an arc undergoing backward-facing collision, being on the opposite side of the arc from that which encounters the continent, would change little tectonically until some future collision affects the second, outboard subduction zone (Fig. 8C), although the forearc-basin sediment supply will change. Post-collisional sediment supply to the forearc will be dominated by erosion of the colliding, exhumed arc, and greater sediment supply could cause a shift from tectonic erosion to tectonic accretion. This change in sediment provenance is evident from stratigraphic studies of forearc basins to the Dras–Kohistan and Talkeetna arcs (Clift et al., 2000; Trop et al., 2005).

Forearc basins involved in forward-facing arc-continent collisions, however, can respond to collision in one of four ways: (1) intact survival and continued sedimentation, (2) burial beneath the accretionary prism, (3) being uplifted, inverted, and eroded, or (4) being destroyed by subduction. Forearc basins in tectonically erosive oceanic arcs usually are floored by mechanically strong lithosphere (Fig. 9) and may well survive forward-facing collision with a passive margin, sometimes continuing to subside and accumulate sediment throughout collision and orogeny. For a forearc basin to survive forward-facing collision, topography likely is isostatically suppressed by eclogite formation in the underlying slab (Cloos, 1993; Dewey et al., 1993; Ryan, 2008). Along with potentially successful accretion of forearc basin sediments, arc-continent collision can also involve preservation of intraoceanic forearc crust obducted as ophiolites (Stern and Bloomer, 1992; Robertson, 2002; Dilek and Furnes, 2011).

Forearc basins that survive collision contain material from the deformed, metamorphosed passive margin and exhumed accreted arc, and include turbidite, mass-wasted, and fan facies. The South Mayo Trough, the 9-km-thick forearc basin to the Lough Nafooey arc of Ireland, which collided with the Laurentian margin in Ordovician time, is a well-studied example containing pre-, syn- and post-collisional sedimentary and volcanic deposits with no major unconformities (Dewey and Ryan, 1990; Mange et al., 2010). Sedimentation rates there were rapid during and shortly after collision (~1 km/m.y.; (Dewey and Mange, 1999). The 5-km-thick Aktau Formation, pre- and syn-collisional forearc-basin deposits of a Devonian arc terrane in the Urals (Brown et al., 2011a), similarly recorded rapid sedimentation during arc-continent collision and was preserved relatively intact, as were boninitic forearc materials (Baimak-Buriabi Formation) (Brown et al., 2011a). Thus, although not all forearc regions form a true basin morphology while the oceanic arc is active (some instead have platforms and are zones of sediment bypass; Section 2.2), those that form deep basins and accumulate long-term sedimentary records also seem the most likely to survive collision intact (see also Oncken (1998) for an example of a forearc basin preserved within the European continent). Dickinson (1995), having tabulated forearc-basin fill in oceanic and continental arcs, noted that preserved ancient forearc basins, on average, contain thicker sedimentary fill than do modern ones, reflecting a tendency for mature basins to be preserved after their accommodation space has filled completely and creating a bias toward description of thick forearc sedimentary packages in the rock record.

Although the Luzon–Eurasia collision is tectonically similar to Paleozoic collisions in western Ireland and the Urals, the Luzon forearc basin does not continue to accumulate sediment during this collision, but instead is buried by the accretionary complex south of Taiwan (Lundberg et al., 1997). There, a large accretionary wedge has grown from imbrication of passive-margin sediment and Taiwan-derived sediment entering the Manila Trench; the Central Range is the onshore continuation of uplifted accretionary material. Although the dominant thrust direction is toward the trench, backthrusts also push accretionary material over the old North Luzon Trough forearc basin, resulting in its burial and cessation of sedimentation (Lundberg et al., 1997; Hirtzel et al., 2009). Basin remnants are exposed onshore, but their width is much reduced and the sediment is strongly deformed. The preservation potential of sediment in the forearc basin south of Taiwan therefore is probably good because the overthrusting accretionary prism shields it from erosion.

Forearc crust preserved and exposed after arc–continent collision may be mapped as ophiolites. It has been recognized for some time that many ophiolites do not represent mid-ocean ridge crust as previously believed but, based on their geochemistry, are pieces of forearc crust (Pearce and Cann, 1973; Stern and Bloomer, 1992). Not all ophiolites are clearly linked to an arc, such as the iconic Troodos and Oman ophiolites (Dilek et al., 1990; Searle and Cox, 1990), and even when they are juxtaposed against a preserved arc complex it is not always clear whether the ophiolite represents a separate subduction system or the forearc to the accreted arc, e.g., Spontang Ophiolite of the western Himalayas and the Dras–Kohistan Arc (Corfield et al., 2001; Pedersen et al., 2001). Nonetheless, it is clear from the worldwide abundance of forearc ophiolites (Dilek and Furnes, 2011) that forearc crust has a relatively high preservation potential.

In either of the previous two scenarios, whether a forearc basin survives forward-facing collision and continues to accumulate sediment, or is buried under an accretionary prism, the basin configuration can survive to be preserved relatively intact in the geologic record. In contrast, the third possible response of forearc regions to forward-facing collision is to be uplifted, inverted, deformed, and eroded, destroying the original basin configuration. Examples have been described from the Eocene–Miocene Sarawaget Formation of the Finisterre Range, Papua New Guinea (Abbott et al., 1994) and from rapidly uplifted forearc nappes in the Lolotoi Complex of East Timor (Standley and Harris, 2009). Interpreting original structure and arc sedimentary processes in forearc strata of ancient, entirely accreted terranes could prove nearly impossible if the forearc has had a history comparable to that of East Timor—rapid uplift of ~3000 m, multiple deformation episodes, and erosion that has turned the already-complex metasedimentary units into isolated klippen (Standley and Harris, 2009). Even if uplifted forearc basins are not severely altered during collision and orogeny, sedimentation most likely ceases owing to loss of accommodation space, as has happened at the Miocene South Savu Basin, Indonesia. Australian continental crust underthrusts that basin, inverting and uplifting the outer forearc and causing arcward migration of the forearc depocenter (van der Werff, 1995; Lüschen et al., 2011).

As a final alternative, forearc regions can be subducted and destroyed during forward-facing collision, with the forearc overthrust by the arc massif (Cloos, 1993; Boutelier et al., 2003; Hall and Smyth, 2008; Boutelier and Chemenda, 2011). The younger and thinner the arc crust, the more likely it is to be subducted; therefore, basins of young, thin arcs are less likely to be preserved in orogenic belts than basins of older, thicker arcs (Cloos, 1993). Forearc and arc crust appear to have subducted where the northern Izu–Bonin arc collides with Honshu (Otsuki, 1990), where the Sangihe arc overthrusts the Halmahera arc in the Molucca Sea (Hall and Smyth, 2008), and where the Aleutian arc is underthrust below Kamchatka (Scholl, 2007). Partial forearc subduction also has been postulated for the late Cretaceous Kohistan–India collision (Burg, 2011). However, whether the forearc-arc crust is deeply subducted is unclear, as it may instead be underplated beneath tectonized overriding crust or partially accreted (Ogawa et al., 1985; Tani et al., 2007). Mass-balancing calculations for the continental crust suggest that major arc crustal subduction is not common (Clift et al., 2009). Whether ultimately subducted or buried at shallow depths to be later uplifted by orogeny, forearc crust and basin material can be substantially deformed and faulted in collision zones, as in the Kuril–Hokkaido Arc since the Late Miocene (Kusunoki and Kimura, 1998) and in the collision of the Aleutian arc with Kamchatka (Gaedicke et al., 2000).

The apparent lack of forearc crust in an accreted arc terrane need not mean that the forearc was subducted during collision, however. Because most oceanic arcs are in a state of tectonic erosion, the early magmatic front is progressively displaced toward the trench as material is removed. Disruption by tectonic erosion and strike-slip faulting complicate interpretation of accreted forearc and trench material (Grove et al., 2008) and rocks identified as having formed at the magmatic front actually may have been in the forearc at the time of collision. The accreted Talkeetna arc, Alaska (Figs. 6E and 7D), forms a good example because arc plutons are directly juxtaposed against the younger, post-collisional Chugach–McHugh accretionary complex and, by definition, form the structural backstop to the accretionary prism in the forearc. This is possible because by the time the accretionary prism formed, tectonic erosion had removed the original forearc and displaced the magmatic front landward. Thus, volcanic overprinting by a migrating arc axis may make distinguishing original arc from forearc rocks very difficult in accreted terranes. The sedimentary cover may be more diagnostic, being finer grained and more hemipelagic in the forearc farther from the arc volcanoes. Of the examples considered here, the South Mayo Trough of western Ireland best preserves a basin that is clearly in part pre-collisional forearc, bracketed as it is between a volcanic complex and a peridotite and accretionary-mélange subduction channel assemblage (Dewey and Ryan, 1990).

The effects of collision on trench and forearc regions bear significantly on how well the geologic record ultimately represents original arc structure and sedimentary deposits, especially those pertaining to subduction of bathymetric high regions or accommodating oblique convergence. To record subduction of aseismic ridges or seamounts in the long term requires enough post-collisional preservation of the trench and forearc to show evidence for shortening and uplift (and possibly geographic re-entrants) that predate the arc–continent collision, accompanied by increased mass wasting, increased sediment flux to the trench, and local cessation of volcanism. Only one of four possible outcomes of the forearc in forward-facing arc–continent collision (intact survival and continued sedimentation; burial beneath the accretionary prism; being uplifted, inverted, and eroded; or being destroyed by subduction) preserves forearc structure and sediment sufficiently to identify such processes—that of continued subsidence and sedimentation. Even in a well preserved forearc basin, the cause of a change or hiatus in sedimentation may never be entirely clear—for example, local cessation of volcanism, uplift, and forearc-basin unconformities can result either from subduction of a bathymetric high region or from arc rifting (Section 3).

Despite these complications, some studies have inferred subduction of a bathymetric feature from examining ancient terranes. Recent work in the McHugh accretionary prism, along a Cretaceous accretionary margin that just postdates collision of the Talkeetna arc with Alaska, shows a sharp tectonic contact between two units within the forearc prism. Amato and Pavlis (2010) inferred from this that an interval of tectonic accretion likely was truncated by subsequent tectonic erosion triggered by seamount collision. Interestingly, the timing of this truncation at the trench correlates with an Early Cretaceous unconformity in the adjacent Matanuska forearc basin, ~70 km inboard from the trench (Trop and Ridgway, 2007). The coincidence of trench erosion and short-lived forearc uplift and inversion there suggests that a collisional event generated uplift across a wide region of the forearc. In other arc terranes, although little sedimentary detail might survive in the forearc or trench regions, small fragments of seamounts with which the arc collided may be present in the eventual suture. Examples include the Summerford Seamount of the northern Appalachians (van Staal et al., 1998) and Paleogene seamounts exposed on the Azuero Peninsula, Panama (Buchs et al., 2011). Ridge subduction has been inferred from changes in forearc sedimentary facies indicating uplift followed by subsidence in Cretaceous strata of the European Alps (Wagreich, 1993), from extension, metamorphism, and forearc magmatism in the Eocene Chugach complex of Alaska (Sisson and Pavlis, 1993; Scharman et al., 2012), and possibly from changes in detrital sedimentary modes (increased lithic fragments, indicating uplift) in Eocene–Miocene exposures of the Kamchatka forearc (Marsaglia et al., 1999).

To preserve evidence that an intraoceanic arc accommodated oblique convergence, which would prove valuable in reconstructing paleogeography and plate motions, requires evidence for pre-collisional along-strike shear in the outer-arc high (such as the Hawley Ridge shear zone of the central Aleutians), and/or in the forearc basin and trenchward side of the arc massif (as in the Mentawai and Sumatra fault zones), as discussed in Section 3.2. In addition to requiring nearly intact preservation of the forearc, again, this also would be most likely at margins that were accretionary for a long time before collision, or at least that did not destroy the outer-arc high and forearc by rapid tectonic erosion. If both those requirements are met, the size and depth of a feature such as the localized basin inversion along the Mentawai fault-zone (Fig. 5) suggest that its long-term preservation potential could be high. However, evidence for pre-collisional shearing caused by oblique convergence could be difficult to distinguish from strike-slip faulting that disrupts the arc terrane during or after collision, and structural or sedimentary records from accreted terranes have only rarely been used to infer convergence angle during arc activity (Harper et al., 1990). If sufficient along-strike exposure exists, oblique convergence and collision still may be inferred from paleomagnetic data or diachronous orogeny (van Staal et al., 1998, 2007). Because oblique convergence also structurally affects areas within and behind the arc massif, even when the forearc and outer-arc high are lost to tectonic erosion or do not survive collision, it may be possible to find evidence for oblique convergence in intra-arc basins or near the backarc region. However, the arc massif can be greatly altered structurally by collisional orogeny, and backarc regions have probably the lowest preservation potential of any part of an arc because of their propensity to subduct.

Arc massifs and intra-arc basins have higher starting topography and less-dense, more-buoyant crust than other regions of an oceanic arc (Fig. 2) (Smith and Landis, 1995). Therefore, although in some cases arc-massif crust may be subducted or at least overthrust (as in the modern Halmahera–Sangihe and Izu–Honshu collisions discussed above), the arc massif and associated basins most likely will form part of the elevated orogen after collision with a continent. The spatial scale of intra-arc basins is generally much smaller than that of forearc basins; that, along with their higher basal elevation, suggests that intra-arc basins are much less likely than are forearc basins to remain submerged and continue accumulating sediment during collision. As collision begins, intra-arc basins should fill quickly with orogen-derived clastic material, then run out of accommodation space and be inverted as uplift and deformation proceed. Burg (2011) discussed inversion of intra-arc basins in Eocene time as closure of the Tethys Ocean sandwiched the Dras–Kohistan arc between India and Asia. The Dras sequences of this arc, exposed in India, may preserve some proximal deposits of intra-arc basins (Clift et al., 2000), as does the Talkeetna arc terrane of Alaska (Clift et al., 2005a). Syn-collisional intra-arc basins also may form and accumulate sediment over a brief life span, as Huang et al. (1995, 2006) described from the Coastal Ranges of Taiwan (Fig. 6G). There, Pliocene–Pleistocene basins 1.5–10 km wide and 40 km long apparently formed during arc–continent collision by transtensional faulting and rapid subsidence. Those basins accumulated flysch and carbonate deposits, then were inverted and incorporated into the subaerial orogenic belt within only 0.8–3.1 m.y. of the start of collision. Although short-lived, such basins, if preserved in orogens, can provide a record of tectonic and geochemical changes at the active margin during collision. However, the likelihood that intra-arc basins will be uplifted and eroded during collision implies that preservation of their original structure is unlikely, given how significantly collision can deform the upper part of an arc massif (Clift et al., 2005a; Hourigan et al., 2009). If intra-arc basins are preserved in the rock record, their strata could be difficult to distinguish from proximal forearc-basin fill (Dickinson, 1995).

Syn-collisional deformation, uplift, and erosion could readily destroy the original configuration of arc-summit basins and any characteristic structures within them, such as evidence for Riedel shearing. This implies that some of the best evidence for structural accommodation of oblique convergence in an oceanic arc (Geist et al., 1988; Ruppert et al., 2012) (Section 3.2) would be lost or be very difficult to interpret after orogenic deformation overprints the arc massif.

Backarc basins are one of the least well preserved parts of oceanic arc systems after collision. Because arc-rifting events most substantially affect the backarc (Section 3.3), the difficulty of preserving backarc regions after continental collision implies that most evidence for intraoceanic arc rifting does not remain long in the geologic record. Forward-facing collision between an arc and a passive margin commonly is followed by formation of a new subduction zone outboard of the accreted arc, i.e., the former backarc basin becomes the locus of a new trench and begins to be subducted. Examples of such a progression have been identified multiple times in the geologic record and at modern-day Taiwan (Dewey and Ryan, 1990; Teng, 1990; Konstantinovskaia, 2001; van Staal et al., 2007; Dickinson, 2008). In backward-facing collision, as oceanic arcs collide with active margins via closure of a backarc basin, most or all of the backarc region is necessarily subducted before collision even begins. Examples such as the Macquarie arc in southeastern Australia show that while the backarc sediment fill may be offscraped and preserved, the backarc oceanic crust is subducted and lost (Glen et al., 2007b). Although rifted backarc lithosphere is thermally buoyant when young, it is also thin, contains dense mafic-ultramafic minerals, and is weaker than the colder forearc lithosphere (Fig. 9) (Watts et al., 1982; Watts, 2001). Rifted backarc crust thus is likely prone to subduction after it surpasses ~10 m.y. in age (when it has cooled enough to be denser than the asthenosphere; (Cloos, 1993), making it unlikely to survive final closure of the backarc during collision. Even in the early stages of collision before any subduction of backarc crust, the backarc region can be deformed substantially as it is shortened and overthrust by part of the arc. This is evident today along the Flores thrust, behind the incipient arc–continent collision of the Sunda arc with Australia, and farther east in the more advanced stage of this collision along the Wetar thrust behind the Banda arc (Silver et al., 1983; Harris et al., 2009).

If original backarc-rifted crust were to survive collision and become incorporated into an orogen relatively intact, one might expect to find repetition of arc rocks with similar ages (Calvert, 2011); we are not aware of any such well preserved examples in the geologic record. Even without age repetition, geochemical signatures imply a backarc origin for some ophiolite crust in accreted terranes (such as the Ordovician Solund–Stavfjord ophiolite complex of Norway; (Furnes et al., 2012), with geochemistry indicating the clearest distinction between rifted backarc crust and the arc massif (Hawkins and Melchior, 1985; Pearce and Stern, 2006; Dilek and Furnes, 2011). In the Talkeetna arc, for instance, basalts with backarc geochemical signature occur landward of the relict arc volcanic front, confirming the north-dipping polarity of subduction (Clift et al., 2005a), and rare-earth-element chemistry also suggests an arc-rifting event in the Ordovician arc terrane of Ireland (Cooper et al., 2011). Geochemical similarity to Mariana backarc magmatism identifies the Cretaceous Chilas Complex within the Kohistan arc, comprising a large volume of massive gabbronorites and mafic dikes, as having formed by decompression melting in an oceanic back-arc setting (Khan et al., 1997; Burg et al., 2006); the Chilas Complex is perhaps the clearest example of ancient intra-oceanic-arc rifting preserved in an accreted terrane.

Where preserved in orogens, backarc remnants occur not only as plutons and ophiolite crustal sections but also as deformed, offscraped fragments within the accretionary prism of either the new suture zone (if the arc accreted by backward-facing collision) or the accretionary prism of the new subduction zone (if subduction resumed with opposite polarity after forward-facing collision). Deformed backarc remnants can comprise eroded lavas and plutons, such as in the Chilas and Talkeetna examples, with material from them locally deposited as turbidites and conglomerates in slope fan complexes (e.g., Ashigara Basin of Japan (Soh et al., 1998); and with slices of volcanic basement intercalated with thrusted sedimentary cover. For accretion onto a continent of an open-ocean backarc basin like the Lau Basin, this presumably would not preserve much metasedimentary material because the sediment cover is only 100–250 m thick in the basin center, thickening in volcaniclastic sediment aprons close to the basin edge. Only in cases where the backarc contains large sediment volumes by virtue of proximity to a continent would significant backarc sediment be preserved in the geologic record. A good example of this is the Ordovician Macquarie arc in southeastern Australia, which formed in a location marginal to the continent onto which it later accreted (Fig. 8A, type 2). Siliciclastic sedimentary flux from the Australian margin was significant enough that when the basin closed, backarc volcanic rocks were imbricated into a large complex of metamorphosed continental sedimentary rocks penetrated by post-collisional plutons sourced from the new arc magmatic front (Glen, 1992; Glen et al., 2007b; Fergusson, 2009) (Fig. 6J). The Alisitos accreted arc terrane of Baja California also preserves extensive volcanic and volcaniclastic backarc material inferred to represent multiple episodes of arc rifting near a continental margin (Kimbrough, 1984; Busby, 2004). Other examples of oceanic backarc crust in accreted arc terranes include Eocene–Miocene volcanic remnants in northern Australia, caught in the ongoing arc–continent collision there (Glen and Meffre, 2009), Miocene–Pleistocene volcanic exposures on the Izu Peninsula, Japan (Tani et al., 2011), Ordovician volcanic rocks of the Lloyds River ophiolite complex in the Appalachians of Newfoundland (Zagorevski et al., 2006, 2009), and possibly the Coast Range ophiolite of western North America (Godfrey and Klemperer, 1998), though a backarc origin for the latter has been controversial (Stern and Bloomer, 1992).

Because intraoceanic arcs undergoing backarc rifting (such as the modern IBM and Fiji–Tonga–Kermadec arcs) can produce intermediate and felsic magmatism (Gill et al., 1990; Clift, 1995; Marsaglia and Devaney, 1995; Suyehiro et al., 1996; Todd et al., 2012), it is possible that pulses of felsic volcanic and volcaniclastic sedimentation in accreted arc-terrane stratigraphy could reflect arc rifting that occurred before collision, as Critelli et al. (2002) inferred for the remnant backarc basin of the Jurassic Gran Cañon Formation offshore Baja California (Busby-Spera, 1988). However, intraoceanic arcs can also produce highly evolved volcanism without arc rifting. Although rare, this is demonstrable in the western Aleutians, where oceanic-arc lavas can have SiO_2 contents of >70 wt % (Kelemen et al., 2003b); therefore, the presence of felsic volcanism in ancient, accreted intraoceanic arcs does not unequivocally identify an episode of arc rifting.

5. Summary

Whereas the tectonic and sedimentary processes at modern subduction zones can be studied using seafloor bathymetry, passive- and active-source seismology, geodetic measurements, and seismic imagery, the history of ancient convergent margins must be inferred largely from stratigraphic and structural records in mountain belts within continents. The translation of sedimentary and tectonic processes into the geologic record after arc-continent collision involves preferential preservation of evidence of some processes and the nearly unavoidable loss of evidence for others. Collision geometry and the final terrane composition differ substantially depending on whether an intraoceanic arc undergoes forward-facing collision (in which the trench and forearc collide first) with either a passive or an active continental margin, or backward-facing collision (backarc colliding first) with an active continental margin.

In general, preservation of arc terranes will be biased toward those that were in a state of tectonic accretion for a long time before collision rather than tectonic erosion, which reduces the loss of forearc, outer-arc-high, and accretionary-prism material. Preservation of the forearc and accretionary (trench) material after collision can be valuable in reconstructing such processes as subduction of high bathymetric features and accommodation of oblique convergence. However, accretionary arc systems are in the minority among modern, active intraoceanic subduction zones, implying that valuable records of arc activity commonly are destroyed by tectonic erosion even before the arc collides with a continent. Arc systems that undergo tectonic accretion are most likely to do so only in the late stages of their activity, as they approach collision with a continent, and thus most accretionary-prism material in ancient arc terranes (whether underplated subduction channel or frontally accreted sediment) likely is biased toward the latest phase of arc activity before collision, when sediment flux to the trench was greatest.

Much has been learned from studying forearc-basin deposits of ancient arc terranes, which sometimes preserve pre-, syn-, and post-collisional stratigraphy, as well as ophiolite crustal sections. However, intact survival and continued sedimentation is only one of four possible outcomes for forearc basins. The alternatives—burial beneath the accretionary prism, major deformation by uplift and inversion, and partial subduction—leave little intact forearc material in the accreted terrane. Where a paleo-subduction zone successfully preserves comprehensive, undeformed forearc stratigraphy and a thick accretionary prism, it would be possible to infer such processes as subduction of bathymetric highs and oblique convergence.

Arc massifs and intra-arc basins, having begun with higher topography than most of the arc system, commonly survive collision and form much of the resulting orogenic belt. However, these regions can be so strongly deformed by uplift, collision, and basin inversion as to obscure much of their original structure, limiting their utility for deciphering arc response to oblique convergence and arc rifting. Backarc basins of oceanic arcs have relatively low preservation potential after arc-continent collision, especially in the case of rifted backarc crust that is thin, weak, and so can be subducted readily in a backward-facing collision or shortly after forward-facing collision as new subduction begins outboard of the accreted arc. The loss of backarc-basin crust during arc–continent collision removes much of the evidence for arc rifting events. Some traces of backarc crust do remain in ancient arc terranes, identifiable by distinctive geochemical signatures, although the clearest example representing arc rifting (the Chilas Complex, Kohistan arc) occurs where voluminous mafic plutons were preserved adjacent to an arc massif rather than from a more distal rifting center. In cases where an oceanic arc develops near a continental margin, siliciclastic sediment supply to the backarc basin can be voluminous enough also to remain in the accreted terrane after collision.

The difficulty of preserving intact records of intraoceanic arcs is not surprising, given that they represent inherently destructive plate boundaries. However, understanding what remains after collision with a continent, and the associated biases and differential preservation in the rock record, is a valuable part of understanding constructive processes as well, because arc volcanism, terrane composition, and crustal preservation and recycling during collision are critical factors in post-Archean formation and evolution of the continental crust. We propose that if arc–continent collision is an important means to add andesitic, LREE-enriched crust onto continents, this process would depend largely on forward-facing arc–continent collisions.

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