Chapter 17
Basins in arc-continent collisions

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ABSTRACT

Arc-continent collisions occur commonly in the plate-tectonic cycle and result in rapidly formed and rapidly collapsing orogens, often spanning just 5–15 My. Growth of continental masses through arc-continent collision is widely thought to be a major process governing the structural and geochemical evolution of the continental crust over geologic time. Collisions of intra-oceanic arcs with passive continental margins (a situation in which the arc, on the upper plate, faces the continent) involve a substantially different geometry than collisions of intra-oceanic arcs with active continental margins (a situation requiring more than one convergence zone and in which the arc, on the lower plate, backs into the continent), with variable preservation potential for basins in each case. Substantial differences also occur between trench and forearc evolution in tectonically erosive versus tectonically accreting margins, both before and after collision.

We examine the evolution of trenches, trench-slope basins, forearc basins, intra-arc basins, and backarc basins during arc-continent collision. The preservation potential of trench-slope basins is low; in collision they are rapidly uplifted and eroded, and at erosive margins they are progressively destroyed by subduction erosion. Post-collisional preservation of trench sediment and trench-slope basins is biased toward margins that were tectonically accreting for a substantial length of time before collision. Forearc basins in erosive margins are usually floored by strong lithosphere and may survive collision with a passive margin, sometimes continuing sedimentation throughout collision and orogeny. The low flexural rigidity of intra-arc basins makes them deep and, if preserved, potentially long records of arc and collisional tectonism. Backarc basins, in contrast, are typically subducted and their sediment either lost or preserved only as fragments in mélangé sequences. A substantial proportion of the sediment derived from collisional orogenesis ends up in the foreland basin that forms as a result of collision, and may be preserved largely undeformed. Compared to continent-continent collisional foreland basins, arc-continent collisional foreland basins are short-lived and may undergo partial inversion after collision as a new, active continental margin forms outboard of the collision zone and the orogen whose load forms the basin collapses in extension.

Keywords: arc continent collision; forearc basin; backarc basin; trench slope basin; tectonic erosion

INTRODUCTION

The sedimentary record that accumulates around volcanic arcs forms the basis for interpreting the tectonic and magmatic history of active margins, both as an end in itself and as a means to understand the development of Earth’s continental crust, much of which formed in these settings (Rudnick and Fountain, 1995; Hawkesworth and Kemp, 2006). Although arc basins can contain the most complete sedimentary and geochemical history of plate convergence and continental-crust generation...
of any geologic province, their utility in deciphering the ancient record is hindered by their low preservation potential relative to basins that form away from destructive plate boundaries. Nevertheless, arc basins are preserved in some suture zones (Fig. 17.1). Much can also be learned from the sedimentary record of active arc basins (Fig. 17.1), from which a more complete understanding of facies evolution can be obtained. This in turn allows such units to be recognized more easily in accreted terranes and used to make paleogeographic and tectonic reconstructions.

In the 15 years since the contributions of Busby and Ingersoll (1995), new advances in understanding basin formation and evolution have been permitted by improved and more widely applied marine mapping techniques around active arcs. These methods have generated high-resolution images of seafloor features, structure, and sub-bottom stratigraphy (e.g., Gaedicke et al., 2000; Wright et al., 2000; Laursen et al., 2002; Kopp et al., 2006). Geodynamic modeling capabilities have similarly expanded in recent years, leading to a greater understanding of crustal loading and basin development (Tang and Chemenda, 2000; Londono and Lorenzo, 2004; Waltham et al., 2008). Geophysical imaging of active and ancient orogens has greatly improved knowledge of tectonic processes in arc-continent collision zones (Snyder et al., 1996; Gaedicke et al., 2000; Brown et al., 2006a; Wu et al., 2007). Field studies of modern and ancient arc collisions have also extended our understanding of climate-tectonic coupling (Dadson et al., 2003; Johnson et al., 2008; Kimura et al., 2008).

Here we review briefly the major types of basins that form at intra-oceanic arcs and that evolve during arc-arc and arc-continent collision, and assess accreted arc terranes to evaluate the fate of basins during collisions of arcs with continental margins. Our discussion draws upon specific examples, but is not intended to be an exhaustive review of all accreted arcs. Arc-continent collisions not only commonly destroy pre-existing basins but also create new ones, as we review using as an example the ongoing collision of the Luzon Arc with the Eurasian margin at Taiwan. Other examples of presently active arc-continent collision (Fig. 17.1) occur in Papua New Guinea (Melanesian arc; e.g., Cloos et al., 2005) and Japan (Izu-Bonin arc; e.g., Taira et al., 2001). Understanding the evolution of accommodation space,

![Fig. 17.1. World map showing the active arcs and accreted arc terranes discussed in the text. Subduction zones are shown with open triangles indicating erosive margins and filled triangles indicating accretionary margins.](image-url)
together with constraints on lithospheric strength throughout all phases of arc activity and collision, maximizes what can be learned from analyzing the sedimentary and geochemical records of arc basins.

Ultimately, understanding arc-continent collision processes can provide valuable insight into the formation and evolution of continental crust. The growth of continents by accretion of arc terranes, accompanied by siliceous magmatism and loss of dense, mafic lower crust through delamination (Bird, 1978) or convective instability (Jull and Kelemen, 2001), is thought to be a key process by which the continents attained their structure and geochemical composition (Pearcy et al., 1990; Rudnick and Fountain, 1995; Suyehiro et al., 1996; Holbrook et al., 1999; Busby et al., 2006; Brown, 2009; Draut et al., 2009). In addition, most of the continental material that returns to the upper mantle does so via subduction zones, the dynamics of which are recorded in arc and trench-slope basins.

INTRA-OCEANIC ARC BASINS AND THEIR FATE IN COLLISION ZONES

Intra-oceanic arc systems were discussed in detail in Busby and Ingersoll (1995). We review briefly the major basin types formed along intra-oceanic arcs and discuss what is known from accreted arc terranes about how each basin type may evolve during arc-continent collision. Our discussion is restricted to basins that form around intra-oceanic arcs, and which eventually collide with continents; continental arcs and their basins are beyond the scope of this work and are well reviewed elsewhere (Wilson, 1991; Jordan, 1995; Fildani and Hessler, 2005; Centeno-Garcia et al., 2008; Lamarche et al., 2008). Here we consider “intra-oceanic” arcs to be those that formed as a result of subduction initiation within oceanic lithosphere and do not include pre-existing continental crust within their basement prior to the collision event.

Oceanic subduction zones assume a geometry that is a function of plate convergence rate, slab age and dip, as well as the rheology and buoyancy of the upper plate (Turcotte and Schubert, 1982; Jarrard, 1986; Billen and Gurnis, 2003). Cross-sections of erosional and accretionary intra-oceanic subduction systems are shown in Figure 17.2, with representative sedimentary profiles from various arc basins in Figure 17.3. Approximately three-quarters of modern subduction zones are of the erosional type (Fig. 17.2A), whereas one quarter are of the accretionary type (Fig. 17.2B; von Huene and Scholl, 1991). Among the major arc-basin types, it has long been recognized that geodynamic mechanisms of basin formation differ in important ways. Trenches and trench-slope basins form along the boundary between the two plates in response to flexure of the subducting slab and extension of the upper plate driven by basal tectonic erosion (Fig. 17.2; von Huene and Scholl, 1991; Underwood and Moore, 1995; Underwood et al., 2003). Forearc basins form on the upper plate between the arc and the outer-arc high (trench-slope break; Dickinson, 1995). Intra-arc basins form among volcanic edifices on the arc platform, as part of the arc massif (Smith and Landis, 1995). Backarc basins can originate by rifting of the arc, associated with high heat flow as new crust forms, or can be deep-water features formed by trapped old oceanic crust that pre-dates the adjacent arc (Karig, 1971; Taylor and Karner, 1983; Tamaki and Honza, 1991; Clift, 1995; Marsaglia, 1995). Deep basins can also form very close to the arc in response to flexural loading by the volcanic arc edifice (Waltham et al., 2008), as well as from arc extension. Each basin type is characterized by different sedimentary facies, structural controls, and underlying basement composition, and each responds differently to the processes of arc-continent collision. Basin types differ in their potential for preservation, metamorphism, or destruction as the arc in which they formed collides with a continental margin. Basin sedimentation in arc-continent collisions can be thought of as rapid but short-lived; collision, orogeny, and orogenic collapse span typically 5–15 My. Nonetheless, explosive volcanism in the arc coupled with rapid erosion of the uplifting orogen results in voluminous sediment supply into the basins around the collision zone.

Trenches and trench-slope basins

Outboard of the arc and forearc, sediment accumulates in the trench and usually form an accretionary prism if the thicknesses exceeds ~1 km (Clift and Vannucchi, 2004; Fig. 17.2B). Intra-oceanic arcs that collide with a passive (continental) margin typically form an accretionary prism as collision nears, even if the open-ocean phase of arc activity is typified by subduction erosion (Fig. 17.2A). In either case, however, the preservation potential of trench sedimentary sequences is
limited. Structurally controlled basins on the trench slope of subduction zones (Fig. 17.2A) form smaller depocenters; these can occur in both erosional and accretionary subduction zones. These perched trench-slope basins, which in Tonga are as large as $\sim 10 \times 50 \text{ km}$, have complex stratigraphy reflecting their dynamic structural setting (Fig. 17.3; Clift et al., 1998). They are formed by extension of the brittle, deformed crust in the cold forearc region. Because the flexural rigidity of the trench slope is low owing to strong deformation, these basins may be quite deep and yet narrow in their aspect. In some arcs the trench and trench-slope basins receive sediment from a broad, elevated forearc platform (delivered via submarine canyons), but not necessarily from the arc itself because the platform and forearc basin separate the arc front from the trench (e.g., Tonga, Hellenic, and Lesser Antilles arcs; Scholl and Vallier, 1985; Marsaglia and Ingersoll, 1992; Draut and Clift, 2006). High-resolution bathymetric images have shown, for example, the forearc Tongan Platform and canyons that incise it (Wright et al., 2000). The preservation potential of trench-slope basins is very low; in collision they are rapidly uplifted and eroded, and at

Fig. 17.2. Schematic cross sections through (A) a typical western-Pacific-type intra-oceanic arc undergoing subduction erosion, showing basins between the arc and the outer arc high, backarc basin development, as well as extension associated with subduction-erosion tectonics; and (B) an accretionary intra-oceanic arc system (e.g., Luzon or Lesser Antilles) in which sediment subduction has formed a frontal accretionary prism and is being underplated beneath the forearc. Intra-arc basins, shown here inboard of the volcanic centers, commonly occur along the axis of volcanism. Source: Diagrams modified after Clift and Vannucchi (2004). Reproduced with permission of the American Geophysical Union.
erosive margins even in intra-oceanic subduction
they are progressively destroyed by subduction erosion.

Substantial differences are expected between
trench and forearc evolution (and associated
cementation) in tectonically erosive versus
tectonically accreting margins (Fig. 17.2), both
before and after collision. Along accretionary
margins such as North Luzon, Nankai, Sumatra, or the
Lesser Antilles, accommodation space may be lim-
ited and sediment can be deposited in perched
basins overlying the accretionary wedge (Beaudry
and Moore, 1985; Underwood et al., 2003). Uplift can culminate in emergence and recycling of sediment from the accretionary prism that can mix with arc-derived volcanioclastic material forming shallow-water facies. Emergence is most likely to happen where the forearc wedge is affected by substantial underplating and where thrust and antithetic thrust faults shorten the wedge. In contrast, trench slopes along tectonically erosive margins (e.g., western Pacific; Fig. 17.2A) subside fairly rapidly in response to basal erosion and thinning of the forearc crust, creating abundant accommodation space near the trench, and are areas of slow, deep-water sedimentation. Trenches and their sedimentary record are progressively destroyed as subduction erosion causes migration of the arc toward the trench (Collett et al., 2008). Therefore, the preservation of trench sediment and trench-slope basins following arc-continent collision 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 might be expected to be more deformed.

Trench sediment from tectonically accretionary margins (Fig. 17.2B) is found in a number of ancient arc terranes; accretionary prism metasedimentary rock tends to be more altered and deformed than other parts of accreted arcs, having been metamorphosed even before collision. This is a result of burial below the forearc, where the accreted, underplated material can be deformed and heated (Sample and Moore, 1987; Moore et al., 1991). Frontal accretionary prisms, in contrast to underplated material, tend to be deformed but not substantially metamorphosed. The metasedimentary Westport Complex and South Connemara Group of western Ireland, for example, are interpreted as frontal accretionary mélangé material associated with an Ordovician island arc that collided with the Laurentian continental margin ca. 470 Ma (Dewey and Ryan, 1990; Ryan and Dewey, 2004). A rapid transition from tectonic erosion in an oceanic setting to collision of the arc with a passive margin results in zones of subduction mélangé, often ophiolitic in composition, only hundreds of meters thick interposed between the accreted arc and the continent, for example in the Shyok Suture between Kohistan and Eurasia (Robertson and Collins, 2002), and the Lichí Mélange in Taiwan (Chang et al., 2000). This mélangé represents the contents of the subduction channel at the time of collision, frozen in place as plate motion effectively ceased.

The Magnitogorsk accreted arc terrane in the Ural Mountains, the product of diachronous arc-continent collision in Devonian to early Carboniferous time (summarized by Puchkov, 2009) includes metasedimentary rocks of the Southern Urals accretionary complex. This Devonian accretionary complex provides important constraints on the timing of arc-continent collision (Alvarez-Marron et al., 2000; Brown et al., 2006b). There, greywacke turbidites 5–6 km thick have been deformed, brecciated, and ductilely sheared into a large synform, with metamorphic grade ranging from anchizone to greenschist facies. Brown et al. (2006b) interpreted the end of trench sedimentation in this accretionary prism (Frasnian; 385–375 Ma) as indicating uplift and widespread basin instability caused by the arrival of the full thickness of the Baltica continental crust at the subduction zone.

Deformation of the Southern Urals accretionary complex is thought to resemble processes in modern arc-continent collisional settings of the western Pacific and southeast Asia at New Guinea, where northward subduction of the oceanic part of the Australian plate began ca. 30 Ma (Quarles van Ufford and Cloos, 2005). In Timor, the 6-km-thick accretionary wedge in the Timor Trough (trench to the Banda Arc) is emplaced southward over the Australian continental margin as basement-involved thrusting imbricated the accretionary prism (Snyder et al., 1996). To the east, in a related collision between the Bismarck Arc and Australian margin, turbidites and pelagic deposits of a frontal accretionary prism are exposed as the Erap Complex in the Finisterre Mountains of Papua New Guinea (Abbott et al., 1994), interpreted as sedimentary material scraped off the lower (Australian) plate. The Erap Complex has been sliced into imbricate thrust sheets as crustal thickness has doubled; the accretionary prism itself is being overthrust by the Bismarck Arc along out-of-sequence thrusts as both are emplaced over part of the Australian continental margin (Abbott et al., 1994). Sedimentary material that was formerly underplated beneath this same arc is exposed as the Derewo metamorphic belt of northern Papua New Guinea (Warren and Cloos, 2007). Arc-continent collision in the New Guinea region has involved the preservation of
several syn-collisional basin sequences; one, the 7-km-thick siliciclastic Aure Group, is interpreted as having filled the trench (Aure Trough) of the Oligocene Papuan subduction zone as the early arc-continent-collisional orogen underwent unroofing and erosion; the preserved trench fill was uplifted and deformed during mid-Miocene time (Quarles van Ufford and Cloos, 2005).

It is important to note that accretionary complexes are generally constructed in the final stages of collision between an arc and a passive continental margin, whereas open-ocean subduction is more commonly associated with subduction erosion. This difference largely reflects the relative flux of sediment to the trench in each setting. As an arc that faces a passive continental margin nears collision with that margin (as in Taiwan; Fig. 17.4), accretionary processes will dominate because the sedimentary section entering the trench greatly exceeds 1 km thickness. Because arc-continent collision is commonly oblique, with the active collision area migrating progressively along strike, much of the trench sediment that accretes prior to suturing is actually reworked from the active orogen along strike (i.e., material that accreted onto the margin and was subsequently eroded and redeposited along strike in the trench). In cases where the arc collides by backing into an active continental margin (as opposed to facing a passive continental margin), sedimentation would not increase substantially at the colliding arc’s trench, and so a large accretionary complex is less likely to form (e.g., the Jurassic Talkeetna arc, discussed below; Fig. 17.5).

Forearc basins

Accommodation space in forearc basins is controlled by the geometry of the arc massif, the trench-slope break, and their position relative to sea level (Dickinson, 1995). Uplift and faulting of the outer-arc high can modify basin shape over time (Ryan and Scholl, 1989). It has been proposed that structural relief and subsidence of forearc basins, particularly those of continental arcs, are functions of crustal thinning by basal subduction erosion (Oncken, 1998; Clift et al., 2003; Wells et al., 2003). Although recent numerical modeling indicates that forearc basins can form as a natural consequence of intra-plate coupling at subduction zones and do not require focused subduction erosion (Fuller et al., 2006), both subduction erosion and underplating beneath the forearc may still contribute to vertical tectonic motions and thus to forearc-basin and outer-arc-high topography (Sample and Moore, 1987; Moore et al., 1991). Forearc basins do not require active subsidence to form, but can reflect relative uplift of structural highs that allows sediment to pond between the uplifted region and the arc itself. Forearc basin geometries can vary greatly depending on the mechanical character of the basin lithosphere. Low flexural rigidity (relatively weak lithosphere) would be anticipated for basins overlying tectonized accretionary complexes, whereas basins overlying mature, igneous, oceanic forearc crust would be less likely to deform because of their cold, stiff lithospheric roots (Fig. 17.6); the latter are more likely to survive collisional orogenesis, even if juxtaposed against weaker accretionary complexes.

Sediment in the forearc basin before collision (Fig. 17.3D–F) includes proximal mass-flow deposits from the steep slopes of volcanic centers, distal volcaniclastic turbidites, reworked ophiolitic debris from the trenchward side of the basin (the outer-arc high), with pelagic sedimentation significant only at the outer forearc, 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). Submarine canyons can transfer sediment into forearc basins (Kopp et al., 2006), just as canyons feed trenches elsewhere. Forearc basins may contain a sedimentary record of much longer duration than the time the arc front has been active in any one place, as in the Mariana arc, which has been split twice by rifting while providing sediment to the same forearc basin for 45 My. As discussed above for trenches, tectonically erosive and accreting margins (Fig. 17.2A and 17.2B, respectively) differ in the accommodation space available for sediment storage, with greater water depths maintained in forearc basins of erosive margins (Fig. 17.3E) because of ongoing basement subsidence.

Forearc basins not only can be preserved during arc-continent collision, but also may even continue to subside and accumulate sediment throughout the collision. This partly reflects the strong mechanical character of oceanic forearc lithosphere (Fig. 17.6). Ryan (2008), using the Ordovician Grampian orogen of western Ireland as an example, showed this to be possible if eclogite formation in the subducting outer continental...
Fig. 17.4. Schematic depiction of arc-continent collision at Taiwan, in which the Luzon arc collides with the passive continental margin of Eurasia. Inset map shows locations of the three cross-sections. The origin of the Ilan Basin is interpreted as a result of gravitational collapse of the Taiwan Central Ranges during a SW-migrating collision of the Luzon Arc and mainland China. Source: Clift et al. (2008). Reprinted with permission of the Geological Society of America.
margin sufficiently reduces buoyancy of the lower plate and causes concurrent forearc subsidence. In the Grampian orogen, the South Mayo Trough comprises a 9-km-thick forearc-basin succession that includes, with no major unconformities, pre-, syn-, and post-collisional sedimentary and volcanic deposits (Dewey and Ryan, 1990; Draut and Clift, 2001). Modeling by Ryan (2008) suggests that forearc basins in the hanging wall of a subduction zone are unlikely to survive a collision unless their topography is isostatically suppressed, such as by substantial volumes of eclogite forming in the footwall (Cloos, 1993; Dewey et al., 1993). Otherwise, even if forearc basins are not destroyed during orogeny, basin sedimentation most likely ceases owing to uplift and loss of accommodation space during collision.

The uplifted, tilted Miocene South Savu Basin, Indonesia, is one such basin, where rapid uplift (5 mm/yr) and subaerial exposure are attributed to underthrusting of rigid forearc basement by buoyant continental crust of the Australian margin (van der Werff, 1995). Part of that forearc basin remains seismically active now (North Savu Basin), but it is progressively losing accommodation space as arc-continent collision proceeds.

Alternatively, forearcs and their basins can be subducted and destroyed during arc-continent collision. Physical experiments by Shemenda (1994) and Boutelier et al. (2003) indicated that if collision stress ruptures the upper plate at the arc front, the forearc region could be entirely overthrust by the arc massif and subducted, leaving perhaps only the most trenchward part of the forearc crust and basin to be obducted and incorporated into an orogen. Although such a scenario has been invoked for the intra-oceanic Achaivayam-Valaginskaya arc terrane that collided obliquely with continental crust at Kamchatka in Paleocene to Eocene time (Konstantinovskaia, 2001), the field evidence supporting this model is unclear. Crustal thickness and density analyses by Cloos (1993) showed that some western Pacific arcs have crust thin enough (<17 km for basaltic crust, <15 km for granitic arc crust) for the arc potentially to be subducted in its entirety.

Arc subduction appears to have occurred in several collision zones, for example the northern Izu-Bonin arc subducting under Honshu (Otsuki, 1990) and the Aleutian arc being underthrust below Kamchatka (Scholl, 2007). However, whether the arc crust is actually subducted to great depth is far from clear, as it may instead be underplated below the tectonized crust of the overriding arc. Mass-balancing arguments for the continental crust suggest that major arc crustal subduction cannot be common (Clift et al., 2009a). In the case of Izu, it may be argued that most of the arc crust is accreted and underplated to the edge of the colliding Japanese margin. In the Izu-Honshu collision zone, the lower crust of the Izu arc is exposed in the Tanzawa Mountains (Tani et al., 2007) and the forearc basin is exposed in the Miura Peninsula (Ogawa et al., 1985), indicating that at least some of the arc is being accreted. Arc thrusting is also known in the Molucca Sea, where the Sangihe arc is thrust over the Halmahera arc (Hall and Smyth, 2008). The same process is inferred in Ordovician accreted terranes in Newfoundland, where partial subduction of one arc under another is inferred to have caused rapid subsidence,
loading, and deposition of black shales in a newly formed basin above the subducting arc and backarc (Zagorevski et al., 2008). 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). Whether they are ultimately subducted or preserved in an orogen, forearc crust and basin material can be substantially deformed and offset by strike-slip faulting in collision zones, as in the Kuril-Hokkaido Arc since the Late Miocene (Kusunoki and Kimura, 1998) and in the oblique collision of the Aleutian arc with continental crust at the Mys Peninsula, Kamchatka (Gaedicke et al., 2000).

Another fate for pre-collisional forearc basins can be seen in the active example of Taiwan (Figs. 17.4 and 17.7). In this area the accretionary complex grows as the Luzon arc progressively collides with the passive margin of China (Suppe, 1981; Huang et al., 2000). Imbrication of Chinese passive-margin sediment, as well as Taiwan-derived sediment fed into the Manila Trench, leads to formation of a large accretionary ridge and eventually the Coast Ranges onshore (Fig. 17.7). Although the dominant thrust direction is toward the trench, backthrusts also push accretionary material over the older North Luzon Trough forearc basin, resulting in its eventual 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 preserved is commonly strongly deformed (Fig. 17.7). Although the arc locally overthrusts the basin, most of the arc crust lies at depth under the mountains and is efficiently accreted (obducted) onto the margin of Asia (Clift et al., 2009a). The preservation potential of sediment in the forearc basins south of Taiwan is modest compared with the foreland basin and post-collisional collapse basins, which are nonetheless filled owing to along-strike transport of syn-collisional sediment.

Preserved forearc basins provide valuable data concerning the timing and dynamics of arc-continent collision. The aforementioned South Mayo Trough has yielded a detailed record of volcanic geochemical changes during arc-continent collision, indicating the degree and timing of continental-crust subduction (Draut and Clift, 2001). In the Urals, the Devonian Aktau Formation represents 5-km-thick forearc basin deposits of cherty and volcanioclastic sediment (Brown et al., 2006b). As the passive continental margin arrived at that trench, forearc basin sedimentation increased and arc-derived volcanioclastic turbidites were deposited across what had been the subducting slab. Rapid sedimentation, soft-sediment deformation of poorly consolidated forearc sediment, and emplacement of large olistostromes (~10 km²)
are attributed to seismicity caused by continental crust underthrusting the arc for 3–5 My (Brown and Spadea, 1999).

In southern Alaska, where various arc and microcontinental terranes have been accreted to the active margin of North America since Jurassic time (Plafker and Berg, 1994), sedimentary basins that recorded the collisional history are exposed subaerially, allowing exceptionally detailed facies analysis and reconstruction of sediment-transport pathways (Trop et al., 2002; Fig. 17.5). Syn-collisional forearc basin deposits there include the Upper Jurassic Naknek Formation, a >900-m-thick belt exposed for ~1200 km total length along the margin, that recorded crustal-scale compression and exhumation of plutonic sources during collision of the Talkeetna-Chitina arc either with the North American continent or with smaller terranes during amalgamation of the Wrangellia Composite Terrane before its final accretion to North America (Trop et al., 2005). Naknek Formation facies indicate deposition on a steep basin floor,

Fig. 17.7. Cross sections through the North Luzon Trough (NLT), the forearc basin to the Luzon Arc. Insert map shows the locations of the cross sections. HB = Huatung Basin, MT = Manila Trench, IP = Ilan Plain, LF = Lishan Fault. (A) The telescoped forearc basin onshore in central Taiwan (after Huang et al., 2000). (B) The basin just south of Taiwan, showing a deeper fill and much wider extent, and (C) in deep water remote from the collision zone but showing its western edge already impinged by the Manila Accretionary Prism.
with submarine mass-flow conglomerates and proximal fan-delta units transitioning to deeper-water turbidites and muddy pro-delta deposits (Fig. 17.8). An upper unconformity reflects subaerial uplift and erosion following arc accretion.

In general, the source of sediment to basins involved in arc-continent collision depends greatly on the tectonics of the collision. In an arc/passive-margin collision in which the forearc faces the collision zone (as in Fig. 17.4), topography appears to be built largely from imbricated, deformed, and metamorphosed sedimentary rocks of the passive continental margin. This means that syn- and post-collisional forearc-basin-filling sediment, such as the Paliwan Formation of Taiwan (Dorsey, 1992) or the Rosroe Formation of South Mayo, Ireland, is largely derived by erosion of the deformed continental margin (Fig. 17.8). Sediment provenance from South Mayo (Clift et al., 2009b) shows that the arc itself must have formed only modest topography, as in modern Taiwan. In contrast, when an arc collides by backing into an active continental margin, such as in Honshu-Izu, Kohistan, or Talkeetna (Fig. 17.5), sedimentation is dominated by erosion of the colliding and exhuming arc complex. Evidence from these regions suggests that the forearc basin of the oceanic arc can remain relatively undeformed as it continues to face an open ocean basin (Clift et al., 2000). Sedimentation in such a forearc basin usually continues, albeit with a change in provenance as the composition of the arc volcanism changes, while the backarc region (which faced the collision zone) is strongly deformed and usually subducted (discussed below).

Other forearc-basin sedimentary sequences preserved in arc collision zones include the Sarhro Group of Morocco, where folded greywacke turbidites and volcanic rocks several kilometers thick record arc-continent collision in the Anti-Atlas orogen ca. 660 Ma (Thomas et al., 2002), the Eocene to Miocene Sarawaget Formation of the Finisterre Range, Papua New Guinea (Abbott et al., 1994), and possibly Upper Permian distal turbidites with arc provenance representing a closing ocean basin between the Altaids of southeastern Mongolia and the North China microcontinent (Johnson et al., 2008).

**Intra-arc basins**

Basins can form within the arc edifice, typically covering less area and with shallower water depths than forearc and backarc basins, and can be “perched” between bathymetric highs (Smith and Landis, 1995). The structural origin of intra-arc basins, which can be bounded either by volcanic centers or by faults (Busby, 2004), is complex and diverse among basins. These basins can form by rifting, which may occur multiple times during the activity of an intra-oceanic arc (as in the Ali-sitos accreted arc terrane of Baja California; Busby et al., 2006) and may produce new oceanic crust when opening intra-arc basins within a continental arc (as during the evolution of the complex Guerrero Composite Terrane of Mexico; e.g., Centeno-Garcia et al., 2008). Other intra-arc basins develop as transpressional or transtensional basins 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). An impediment to preservation of intra-arc basins is the fact that close to the arc volcanic front, flexural rigidity is low because of the lack of a major lithospheric root, such that basins may be readily deformed, inverted, and destroyed.

Intra-arc sedimentation consists of proximal volcanic and volcaniclastic material in debris aprons derived from the arc massif that fine away from the eruptive centers into debris aprons and deep-water drift facies (Fig. 17.3C; Draut and Clift, 2006). Proximal Cretaceous intra-arc-basin facies exposed in Baja California include silicic subaqueous pyroclastic flow deposits, tuffs and tuff turbidites, hyaloclastic breccias, and evidence for mixing of basaltic and silicic lavas with wet marine sediment; the volcanic material in those basin deposits are intercalated with rudist reef deposits and marine mudstones, reflecting changes in relative sea level during basin sedimentation (Busby et al., 2006). During arc-continent collision, intra-arc basins can fill with orogen-derived clastic material and, if preserved, can record tectonic and geochemical changes in the active margin during arc collision. Huang et al. (1995) interpreted subaerially exposed Pliocene and Pleistocene intra-arc basins in Taiwan’s Coastal Ranges to have formed because of strike-slip transtension faulting during active arc-continent collision. These syn-collisional basins are of similar size to those of modern intra-arc basins not yet involved in collision (1.5–10.0 km wide, 40 km long), and are predicted to be short-lived as collision progresses between the Luzon Arc and Eurasian margin, resulting in their rapid inversion soon after formation and
Fig. 17.8. Field photographs of sediment deposited in arc collisional settings. (A) Submarine fanglomerates (clasts are metasedimentary rocks) and (B) sandy turbidites from the early syn-collisional Paliwan Formation of Taiwan; these sediments, which contain material eroded from a mature orogen in late Pliocene time, are exposed in the Coast Ranges. (C) Conglomerates (most clasts are igneous) and (D) channelized sandstones of the early post-collisional Ordovician Rosroe Formation, South Mayo, Ireland. Provenance of clasts in the Rosroe Formation is dominantly from the metamorphosed continental passive margin, but with up to 20% influx from the accreted arc (Clift et al., 2009b). (E) and (F) show sedimentary basin deposits of the syn-collisional Jurassic Naknek Formation, southeast Alaska (photos in E and F courtesy of J.M. Trop): (E) Proximal fan-delta forearc strata (person at left for scale) dominated by poorly to moderately sorted, inversely graded conglomerate (with granitic clasts whose Middle Jurassic isotopic ages match those of nearby Talkeetna arc plutons) deposited by marine sediment gravity flows. (F) Distal pro-delta forearc strata (person for scale) characterized by thinly bedded, sharp-based, normally graded sandstone turbidites (resistant units) and massive, black mudstone deposited by suspension fallout. Sandstones are dominated by quartzofeldspathic detritus and granitic rock fragments. Mudstones yield Jurassic ammonites, radiolarian, and foraminifera fossils.
filling. Basin facies there include deep-water flysch overlying shallow marine reef carbonate atop volcanic basement, indicating formation by rapid subsidence. These intra-arc basins developed, filled, and were inverted rapidly with sedimentation lasting a mere 0.8–3.1 My.

**Backarc basins**

Basins commonly occur behind intra-oceanic arcs (Fig. 17.2), originating either from rifting and extension after arc development (as has occurred twice in the Mariana arc and twice in the Tonga arc; Hawkins, 1974; Taylor, 1992) or as an older ocean floor that pre-dates the arc and subduction zone (Karig, 1971; Taylor and Karner, 1983). Sedimentation in basins behind arcs is dominated by volcanic and volcaniclastic products of the arc, with facies indicating increasing water depths with distance from the arc. Where backarc basins have formed by arc rifting, products of silicic volcanism can be abundant in the backarc stratigraphy (e.g., Izu-Bonin arc; Nishimura et al., 1992; Iizasa et al., 1999; Fiske et al., 2001).

Backarc basins are unlikely to be preserved after arc-continent collision. Arc-passive margin collision (the type of collision shown in Fig. 17.4 in which the arc faces the continent) is usually 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 (Dewey and Ryan, 1990; Teng, 1990; Konstantinovskaia, 2001; van Staal et al., 2007; Dickinson, 2008). Where oceanic arcs collide with active margins via closure of a backarc basin, as occurred during accretion of the Jurassic–Cretaceous Alisitos arc terrane of Baja California (Busby, 2004; Busby et al., 2006) and the Jurassic Talkeetna-Chitina arc terrane of southern Alaska (e.g., Plafker and Berg, 1994; Fig. 17.5), the backarc region is necessarily subducted before collision even begins. Although much of the backarc basin and its sedimentary fill would thus be destroyed, remnants of the backarc may be preserved as deformed, offscraped fragments within the new suture zone. In the deforming backarc, oceanic arc plutons may be exposed, eroded, and material from them locally deposited as thick turbidites and conglomerates in slope fan complexes (e.g., Ashigara Basin of Japan; Soh et al., 1998). In such cases where arcs accrete onto active margins by backarc-basin closure, a situation that requires more than one subduction zone (or at least large-scale thrusting and convergence between continent and backarc, in addition to the principal subduction zone), there is substantial potential for large-scale subduction and recycling of part of the arc massif (e.g., Suyehiro et al., 1996; Busby, 2004).

Despite the generally low preservation potential of backarc basins, several accreted arc terranes do contain kilometers-thick backarc stratigraphy (e.g., Fig. 17.3A, B). In the Alisitos accreted arc of Baja California, the Jurassic Gran Canyon Formation comprises volcanic and volcaniclastic backarc material that directly overlies a supra-subduction-zone ophiolite (Kimbrough, 1984; Busby, 2004). With a hydrothermally altered base interpreted to reflect deposition on hot rifted arc crust, this backarc sequence consists of deep marine pyroclastic deposits dominated by lapilli tuff and tuff breccia; proximal dacitic pyroclastic flows formed graded beds tens of meters thick (Busby, 2004). Multiple episodes of arc rifting are inferred from these backarc sequences that would have isolated the backarc from the active arc front as the backarc topography broke up into a series of fault-bounded blocks and basins. Backarc deposition was capped by fine-grained epiclastic volcaniclastic and lithic sandstone and siltstone as the proximal sediment flux from the arc was interrupted by rifting (Busby, 2004).

Large accreted basins that had been situated behind a Jurassic arc system also occur in southern Alaska. Their sedimentary fill spans the change from a pre- to a post-collisional setting over tens of My (Trop and Ridgway, 2007). The Nutzotin and Wrangell Mountains Basins (Trop et al., 2002; Manuszak et al., 2007) were both situated behind the intra-oceanic Talkeetna-Chitina arc and received Middle to Late Jurassic arc-derived sediment; facies include marine volcaniclastic sandstone, mudstone, chert, and tuff. Similar depositional environments (distal, deep-water sedimentation behind the arc) were also inferred for Lower Jurassic Talkeetna Formation rocks exposed in the nearby Horn Mountains, although in this latter case the transition from intra-oceanic activity to collision is not well preserved (Clift et al., 2003). After the Talkeetna-Chitina arc was incorporated into the Wrangell Composite Terrane (WCT) and ceased activity (Late Jurassic time), the Wrangell Mountains basin remained active within the WCT and was positioned as a
forearc basin to the newly active Chisana arc (between the Chisana and extinct older arc) (Trop et al., 2002, 2005; Trop and Ridgway, 2007). Sedimentary rocks in the Wrangell Mountains Basin then recorded uplift and exhumation of the Talkeetna-Chitina arc and the WCT. Concurrently, the Nutzotin Basin remained active as a foreland basin behind the Chisana arc, accumulating ~3 km of upward-coarsening sediment that represents submarine fan and overlying marine shelf deposition (Manuszak et al., 2007). (Note that this basin predated the Chisana arc behind which it sat; a basin’s position behind any given arc does not require that that basin formed in the same way as extensional backarc basins in the Mariana and Tonga arcs or backarc basins of the Alisitos arc, Baja California, that were the products of arc rifting and seafloor spreading; Clift, 1995; Busby, 2004.) Sedimentation in basins both behind and in front of the Chisana arc continued to evolve through the Cretaceous; the Nutzotin and Wrangell Mountains basins were uplifted in Late Cretaceous to Tertiary time, and folded coevally with major dextral faulting (Trop and Ridgway, 2007).

**BASINS FORMED DURING ARC-CONTINENT COLLISION**

In addition to short-lived intra-arc basins forming in an active arc-continent collision zone (Huang et al., 1995), arc collisions cause foreland basins to form by crustal loading, much as they would in a continent-continent collision zone. These basins involve lithosphere that is generally some of the most thermally mature (i.e., cooler) and thus strongest in the oceans, with high flexural rigidity resulting in broad basins. The major distinction of arc-continent collision forelands is that because orogenesis and subsequent collapse generally occur rapidly (5–15 My; Abbott et al., 1997; Friedrich et al., 1999), the foreland basins form, fill quickly, and then may be inverted as the flexural load is reduced when an active continental margin becomes established outboard of the collision zone. This is because re-establishment of subduction is often accompanied by both extensional collapse of the collisional orogen and rapid erosion, both of which combine to reduce the load and lessen flexure. Foreland-basin formation is ongoing in the collision between Australia and Papua New Guinea, where the Leron Formation comprises a fan delta complex at the base of the Finisterre Range (Abbott et al., 1994). Similarly, the Timor-Aru Troughs represent young foreland basins formed by flexural loading of the Australian continental margin as it collides with the Banda arc (Snyder et al., 1996; Londoño and Lorenzo, 2004). Because the Australian continental margin formed in Jurassic time, it is relatively mature and rigid compared to arc collision zones involving younger crust resulting in a broad foreland basin.

Brown et al. (2006b) described the Southern Urals accretionary complex as a thrust stack that developed in a foreland basin on the continental side of the Devonian arc-continent collision zone in central Asia, which received sediment while at the same time the old forearc basin was filled from the same sources. Using Taiwan as the type example, it is apparent that flexural foreland basins reach their maximum extent and depth at the height of collision, after the inversion and closure of many (if not all) pre-existing arc basins. They are filled by sediment derived across strike from the orogen, and may contain both continental and shallow marine facies, with deep marine flysch in the earlier stages of mountain building, shallowing as the basin fills (Figs. 17.7 and 17.8).

As the active zone of oblique arc-continent collision moves along the margin (as in Taiwan), the post-collisional orogen progressively collapses during reversal of subduction polarity when the compressive tectonic forces of collision are removed. Collapse involves tectonic extension of weak, hot, orogenic crust, and forms basins where sedimentation can resume (Teng, 1996). Such basins tend to form near the suture zone and do not affect the broader foreland. However, the redistribution of mass away from a central edifice might be expected to lessen the load on the subducting passive margin and potentially allow a partial unflexing of the plate and inversion of the foreland basin. The clearest modern example of a post-collisional collapse basin is the Ilan Plain/Okinawa Trough of northern Taiwan (Fig. 17.4). In the Urals arc-continent suture, shallow-water carbonates unconformably overlying the arc units record post-collisional collapse of the orogen (Brown and Spadea, 1999). Likewise, after arc-continent collision in the Anti-Atlas orogen of Morocco, inferred post-orogenic collapse (580–550 Ma) led to extension forming faulted molasse basins in which acidic volcanic rocks and volcaniclastic sediment were deposited (the Ouazarzate Group; Thomas et al., 2002).
CASE STUDY: SYNCHRONOUS BASIN DESTRUCTION AND FORMATION DURING ARC-CONTINENT COLLISION AT TAIWAN

The very clear tectonic relationship in which the oceanic Luzon arc is in a state of progressive and migrating collision with the passive margin of China (Suppe, 1984; Teng, 1990) reveals the different styles of sedimentary basins associated with different phases of collision. It is important to note that the pre-, syn- and post-collisional basins are all filled by sediment eroded largely from the more metamorphosed high mountain range in the orogenic core. Taiwan is one of the greatest modern sediment-producing areas on Earth (Milliman and Syvitski, 1992), a result of its rapid rates of rock uplift, friable lithologies, and the erosive tropical climate. Sediment is transported southward along strike via the Manila Trench and the long axis of the forearc (North Luzon Trough), bringing orogenic sediment from the mountains into the pre-collisional arc region (Fig. 17.7). Similarly, sediment is transported along strike by the Lanyang River, itself guided by the Lishan Fault into the Okinawa Trough, which is interpreted as a post-collisional extensional collapse basin (Figs. 17.4 and 17.7; Clift et al., 2008). Thus, because of substantial along-strike sediment transport, the erosional history recorded by the basins around Taiwan (and presumably other arc-continent collisional orogens) is not generally synchronous with the stage of tectonic development in which the basin itself formed.

As described above, the pre-collisional basins have the lowest preservation potential, although some remnant of the North Luzon Trough does survive between the Central and Coast Ranges. Indeed, fluvial and alluvial fan sedimentation continues in the Longitudinal Valley of Taiwan between the two ranges, despite the rapid uplift on either side. Because the deepwater sediment of the North Luzon Trough is tectonically overthrust and buried, those deposits may survive collision and later be exhumed. Early syn-collisional sedimentary rocks of the Paliwan Formation (Fig. 17.8B), composed of proximal eroded metasedimentary rocks of late Pliocene age, are exposed in the Coast Ranges adjacent to the core of the mountains (Dorsey, 1992). The metamorphosed character of clasts in the Paliwan Formation shows that they were eroded from a mature orogenic belt during the late Pliocene. Because material effectively migrates north through the system as the collision progresses, we infer that these sedimentary rocks were deposited in a basin south of the highest ranges at that time, but, like the present system, contain material recycled from a proto-Central Range source located to the paleo-North. Since their deposition, the Paliwan conglomerates have been deformed and uplifted.

The largest sedimentary mass in the Taiwan collision zone lies beneath the Taiwan Straits: the foreland basin, where the Oligocene South China Sea rifts have been buried under ~2 km of sediment, locally >5 km thick (Lin et al., 2003). The flexural rigidity of the margin lithosphere is not particularly high ($T = 8–13$ km; Lin and Watts, 2002), resulting in a relatively narrow basin (~70 km), but one with a high preservation potential. In contrast, sediment eroded from the Coast Range that is deposited in the backarc basin of the Luzon Arc (i.e., the Huatung Basin of the Philippine Sea) is likely to be subducted and lost as subduction polarity reverses. The Huatung Basin (Fig. 17.7) receives significant sediment flux from the orogen but this crust is in turn subducted under the newly created continental margin of the Ryukyu arc. Although some material may be offscraped into the new accretionary prism of the continental margin, the majority will likely be lost and subducted to depth under the arc (Clift and Vannucchi, 2004).

SUMMARY

Arc-continent collisions occur commonly in the plate-tectonic cycle and result in rapidly formed and rapidly collapsing orogens, often spanning just 5–15 My. Accretion of new material onto continents via arc-continent collision (accompanied by siliceous magmatism and loss of dense, mafic lower crust) is thought to be a major process governing evolution of the continental crust through time. Rapid uplift rates at arc-continent collision zones generate substantial amounts of sediment, which is transported along the colliding margin in pre- and post-collisional basins, and in the foreland. Collisions of intra-oceanic arcs with passive continental margins (a situation in which the arc, on the upper plate, faces the continent) involve substantially different geometry than collisions of intra-oceanic arcs with active continental margins (a situation requiring more than one convergence zone and in which the arc, on the lower plate, backs
into the active continental margin), with variable preservation potential for basins in each case. In the latter situation, there is substantial potential for large-scale subduction and recycling of part of the arc massif.

Substantial differences are also expected between trench and forearc evolution (and associated sedimentation) in tectonically erosive versus tectonically accreting margins, both before and after collision. During intra-oceanic arc activity before collision, the greater water depths that typify erosive margins lead to slow formation of deeper-water sedimentary facies than occur in accretionary margins. The preservation potential of trench-slope basins is very low; in collision they are rapidly uplifted and eroded, and at erosive margins especially in intra-oceanic subduction they are progressively destroyed by subduction erosion. After arc-continent collision, preservation of deformed and, if underplated, metamorphosed trench sediment and trench-slope basins is biased toward margins that were tectonically accreting for a substantial length of time before collision.

Forearc basins in tectonically erosive oceanic arcs are usually floored by strong lithosphere and may well survive collision with a passive margin, sometimes continuing sedimentation throughout collision and orogeny, driven by eclogite formation in the underlying slab. Sedimentation rates in surviving forearc basins can be rapid; these basins contain material derived from the deformed and metamorphosed passive margin and exhumed accreted arc, and are characterized by turbiditic, mass-wasted, and sometimes fan facies. Forearc basins perched on weak, deformed, accretionary-complex basement are more likely to be destroyed in collision zones because they are more likely to be deformed, uplifted, and eroded. The low flexural rigidity of intra-arc basins makes them deep and, if preserved, potentially long records of arc and collisional tectonism. Whereas pre-collisional forearc and intra-arc basins may be filled and then survive the collision, backarc basins are typically subducted and the sedimentary cover either lost or preserved only as fragments in mélange sequences.

A substantial proportion of the sediment derived from collisional orogenesis ends up in a foreland basin that forms as a result of collision, and may be preserved largely undeformed. Compared to continent-continent collisional foreland basins (e.g., Zagros and Himalaya), arc-continent collisional basins are short-lived and may experience partial inversion after collision as the orogen collapses during the establishment of a new, active continental margin outboard of the collision zone.

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