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Notes

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Abstract: Accretionary orogens form at intraoceanic and continental margin convergent plate boundaries. They include the supra-subduction zone forearc, magmatic arc and back-arc components. Accretionary orogens can be grouped into retreating and advancing types, based on their kinematic framework and resulting geological character. Retreating orogens (e.g. modern western Pacific) are undergoing long-term extension in response to the site of subduction of the lower plate retreating with respect to the overriding plate and are characterized by back-arc basins. Advancing orogens (e.g. Andes) develop in an environment in which the overriding plate is advancing towards the downgoing plate, resulting in the development of foreland fold and thrust belts and crustal thickening. Cratonization of accretionary orogens occurs during continuing plate convergence and requires transient coupling across the plate boundary with strain concentrated in zones of mechanical and thermal weakening such as the magmatic arc and back-arc region. Potential driving mechanisms for coupling include accretion of buoyant lithosphere (terrane accretion), flat-slab subduction, and rapid absolute upper plate motion overriding the downgoing plate. Accretionary orogens have been active throughout Earth history, extending back until at least 3.2 Ga, and potentially earlier, and provide an important constraint on the initiation of horizontal motion of lithospheric plates on Earth. They have been responsible for major growth of the continental lithosphere through the addition of juvenile magmatic products but are also major sites of consumption and reworking of continental crust through time, through sediment subduction and subduction erosion. It is probable that the rates of crustal growth and destruction are roughly equal, implying that net growth since the Archaean is effectively zero.

Classic models of orogens involve a Wilson cycle of ocean opening and closing with orogenesis related to continent–continent collision. These imply that mountain building occurs at the end of a cycle of ocean opening and closing, and marks the termination of subduction, and that the mountain belt should occupy an internal location within an assembled continent (supercontinent). The modern Alpine–Himalayan chain exemplifies the features of this model, lying between the Eurasian and colliding African and Indian plates (Fig. 1). The Palaeozoic Appalachian–Caledonian orogen (Wilson 1966; Dewey 1969), the Mesoproterozoic Grenville orogen (Gower *et al.* 1990; Hoffman 1991; Gower 1996), and the Palaeoproterozoic Trans-Hudson (Ansdell 2005), Ketilidian (Garde *et al.* 2002), Capricorn (Cawood & Tyler 2004) and Limpopo (Kröner *et al.* 1999) orogens are inferred ancient examples. Such models, however, do not explain the geological history of a significant number of orogenic belts throughout the world. Such belts lie at plate margins in which deformation, metamorphism and crustal growth took place in an environment of continuing subduction and accretion. These belts are termed accretionary orogens but have also been referred to as non-collisional or exterior orogens, Cordilleran-, Pacific-, Andean-, Miyashiro- and Altaid-type orogens, or zones of type-B subduction (Matsuda & Uyeda 1971; Crook 1974; Bally 1981; Murphy & Nance 1991; Windley 1992; Şengör 1993; Şengör & Natal'in 1996; Maruyama 1997; Ernst 2005). Accretionary

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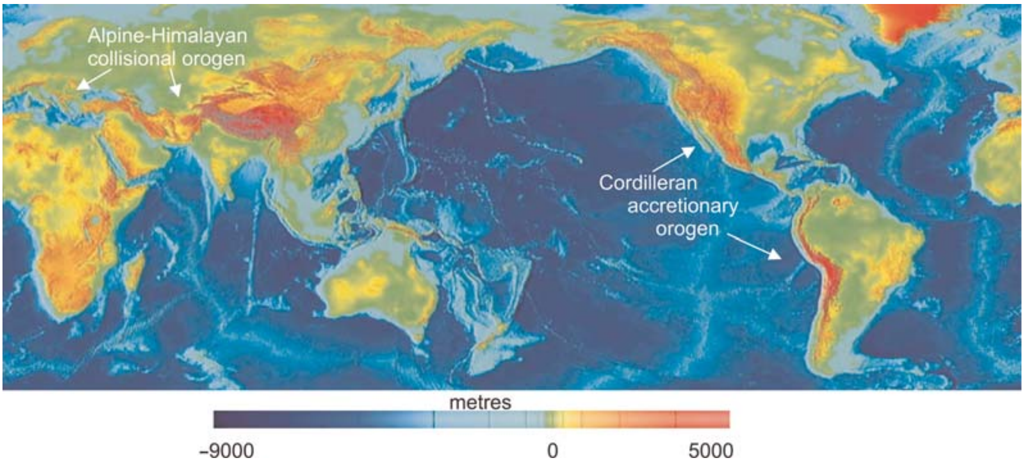

Fig. 1. Global elevation showing the collisional Alpine–Himalayan orogen (which occupies an internal location between the Eurasian and colliding African and Indian plates) and the accretionary North and South American Cordilleran orogens (which lie at plate margins involving continuing subduction). Image modified from NOAA, National Geophysical Data Center.

orogens appear to have been active throughout much of Earth history and constitute major sites of continental growth (Cawood *et al.* 2006). The accretionary orogens of the western and northern Pacific extending from Indonesia via the Philippines and Japan to Alaska, and the North and South American Cordillera are archetypical modern examples, with ancient examples represented by the Phanerozoic Terra Australis and Central Asian orogens, the Proterozoic orogens of the Avalon–Cadomian belt of the North Atlantic borderlands, Birimian of West Africa, Svecofennian of Finland and Sweden, Cadomian of western Europe, Mazatzal–Yavapai in southwestern USA, and the Arabian–Nubian Shield, and Archaean greenstone terranes (Windley 1992; Kusky & Polat 1999; Karlstrom *et al.* 2001; Johnson & Woldehaimanot 2003; Cawood 2005; Kröner *et al.* 2008; Murphy & Nance 1991).

Accretionary orogens form at sites of subduction of oceanic lithosphere. They include accretionary wedges, containing material accreted from the down-going plate and eroded from the upper plate, island arcs, back-arcs, dismembered ophiolites, oceanic plateaux, old continental blocks, post-accretion granitic rocks and metamorphic products up to the granulite facies, exhumed high- or ultrahigh-pressure metamorphic rocks, and clastic sedimentary basins. Accretionary orogens contain significant mineral deposits (Groves & Bierlein 2007), and thus provide the mineralization potential of many countries such as Australia, Canada, Chile, Ghana, Zimbabwe, Saudi Arabia, China, Kazakhstan, Mongolia and Indonesia. All accretionary orogens are ultimately involved in a collisional phase when oceans close and plate subduction ceases, and this may lead to significant structural modification of the accreted material and to partial or complete obliteration as a result of thrusting and extensive crustal shortening (e.g. Central Asian Orogenic Belt; India–Asia collision).

Our understanding of the processes for the initiation and development of accretionary orogens is moderately well established in modern orogens such as the Andes, Japan, Indonesia and Alaska, the broad structure and evolution of which are constrained by plate kinematics, seismic profiles, tomography, field mapping, palaeontology, and isotope geochemistry and geochronology (e.g. Oncken *et al.* 2006*a*; Fuis *et al.* 2008). However, the processes responsible for cratonization and incorporation of accretionary orogens into continental nuclei and the mechanisms of formation of most pre-Mesozoic accretionary orogens are less well understood. In a uniformitarian sense many of the features and processes of formation of modern accretionary orogens have been rarely applied to pre-Mesozoic orogens, and hence to elucidating Earth evolution.

Our aim is to outline the broad features of accretionary orogens and discuss their implications for understanding models of crustal growth. We believe that future research into accretionary orogens will increase our understanding of tectonic processes and crustal evolution just as work on geosynclines, plate tectonics and mountain belts, terranes, and supercontinents provided a stimulus to orogenic and geological research in past decades (Kay 1951; Aubouin 1965; Wilson 1966; Dewey 1969; Coney *et al.* 1980; Dalziel 1991; Hoffman 1991; Moores 1991).

Classification of orogens in space and time

Orogens are major linear zones of the Earth's crust that contain variably deformed rocks that accumulated over a long period, dominantly in a marine environment, and that show distinctive distribution of sedimentary facies, deformational styles and metamorphic patterns often aligned approximately parallel to the belt. They are preserved as mountain belts through crustal thickening, magmatism and metamorphism during one or more tectonothermal events (orogenies), which are generally of short duration with respect to the overall age range of the orogen, and which ultimately stabilize and cratonize the orogen.

Codifying orogens is fraught with difficulty as each has unique characteristics. However, we believe they can be grouped within a spectrum of three end-member types: collisional, accretionary and intracratonic (Figs 2 and 3). Collisional orogens form through collision of continental lithospheric fragments, accretionary orogens form at sites of continuing oceanic plate subduction, and intracratonic orogens lie within a continent, away from an active plate margin.

Plate-tectonic models of orogenic cycles have been dominated by work on collisional orogens involving a Wilson cycle (Dewey & Spall 1975) of opening and closing of an ocean basin with deformation and metamorphism related to subduction followed by collision of continental blocks to generate mountain belts (Wilson 1966; Dewey 1969; Brown 2009). This in part reflects the historical focus, and resultant development of geological ideas, on the classic orogens of Europe and eastern North American, which all formed in a collisional setting; the Appalachian–Caledonian, Hercynian and Alpine–Himalayan systems (see Miyashiro *et al.* 1982). However, it has also long been recognized that the Palaeozoic to Recent history of the Circum-Pacific region, where orogeny is in progress, does not readily fit such a model and that alternative mechanisms for this type of orogenesis are required (Matsuda & Uyeda 1971; Coney 1973; Crook 1974; Packham & Leitch 1974).

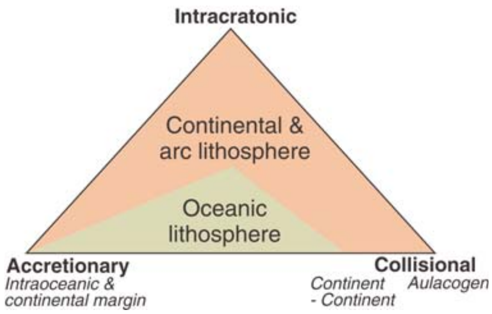

Fig. 2. Classification of orogen types into three interrelated end-members: collisional, accretionary and intracratonic. Accretionary orogens include those formed at continental margins and in intraoceanic settings, and collisional orogens include those formed from failed rifts (aulacogens) and through continent–continent collision. Accretionary and continent–continent collisional orogens lie at plate margins and form through the subduction of oceanic lithosphere, with the former forming at sites of continuing subduction and the latter at the termination of subduction.

Accretionary and collisional orogens (excluding aulacogens) form at sites of subduction of oceanic lithosphere and are end-members of a spectrum of orogen types (Figs 2 and 3). An early stage is represented by island arc accretion in, for example, Japan (Isozaki 1996; Maruyama 1997) and Alaska (Sisson *et al.* 2003). Such offshore arcs may accrete to one another and to an active continental margin, where they are incorporated into an Andean-type batholith and orogen; for example, the Coastal batholith of Peru, which engulfed the Casma volcanic arc (Petford & Atherton 1995), and the Peninsular batholith of Southern California, and elsewhere along the Cordillera of North and South America (Lee *et al.* 2007). Such arc-generated orogens as old as Neoarchaean have been recognized (Windley & Smith 1976; Windley & Garde 2009).

Final continental collision and termination of subduction within collisional orogens is generally preceded by an accretionary phase of subduction-related activity linked to ocean closure. Examples include a series of magmatic arcs developed within and along the margins of the Iapetus ocean of the Appalachian–Caledonian orogen (Cawood & Suhr 1992; van Staal *et al.* 1998) and the accreted Kohistan island arc in the western Himalaya orogen that was intruded by the Andean-style Kangdese batholith before final collision between India and Asia and closure of the Tethys ocean (Bignold & Treloar 2003); also, in the Palaeoproterozoic the Trans-Hudson orogen similarly formed during ocean closure and arc accretion events prior to collision of cratons (Lucas *et al.* 1999; St-Onge *et al.* 2009). The Indonesian island arc is currently in transition from a simple system involving underthrusting of oceanic lithosphere in the west to collision with Australian continental lithosphere in the east (Hamilton 1979; Snyder *et al.* 1996*b*). Conversely, accretionary orogens may contain accreted continental fragments such as in the Central Asian Orogenic Belt (Badarch *et al.* 2002; Kröner *et al.* 2007) and the Abas, Afif and Al-Mahfid terranes in the Arabian–Nubian Shield (Windley *et al.* 1996; Johnson & Woldehaimanot

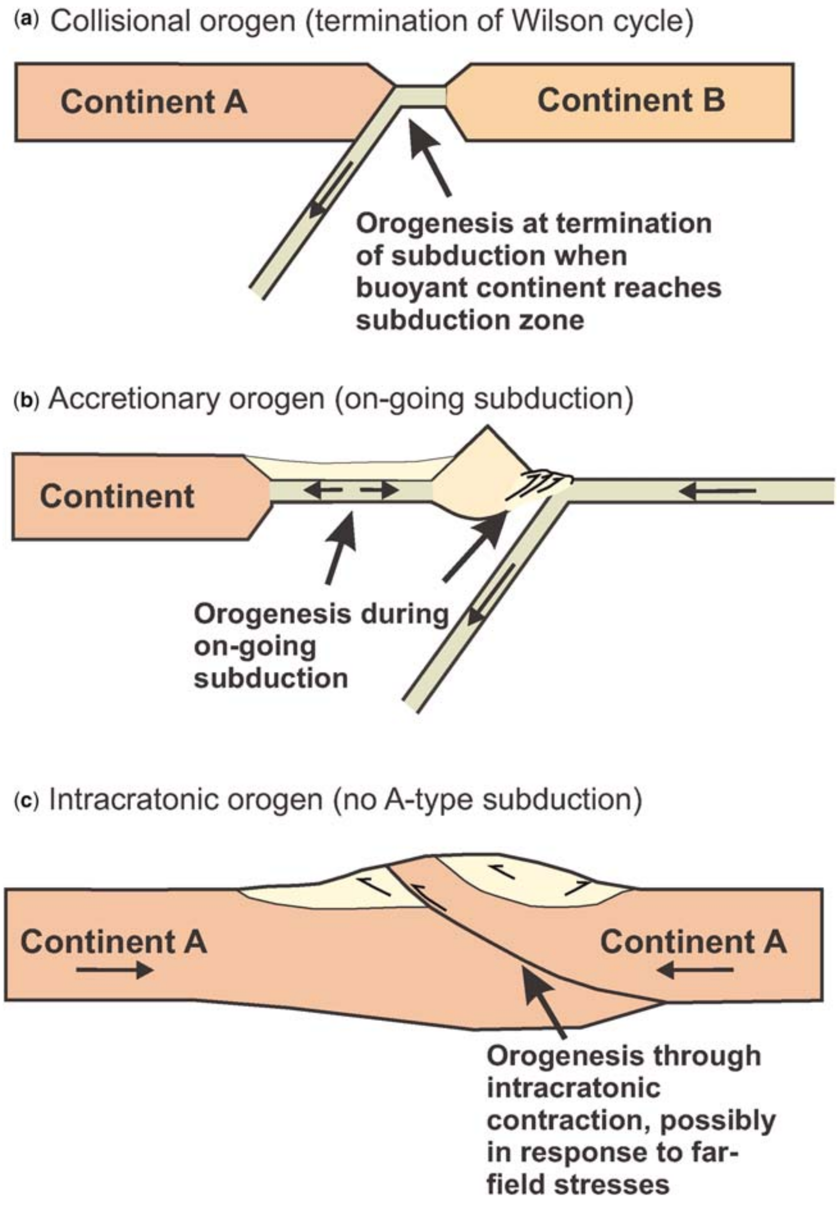

Fig. 3. Schematic cross-sections through (**a**) collisional, (**b**) accretionary and (**c**) intracratonic orogens.

2003; Stern *et al.* 2004). The long-lived accretionary Central Asian Orogenic Belt completed its history with a Himalayan-style collisional orogen in northern China (Xiao *et al.* 2003, 2004; Windley *et al.* 2007). Nevertheless, accretionary orogens stand out as an integral, well-defined group of orogens that are further characterized by significant crustal growth (Samson & Patchett 1991; Şengör & Natal'in 1996; Jahn *et al.* 2000*b*; Wu *et al.* 2000; Jahn 2004).

Continental extension that fails to lead to ocean opening and subsequently undergoes compression can occur at failed arms of ocean basins (aulacogens) and at intracontinental settings isolated from plate margins (Figs 2 and 3). The former represents a specific subset of collisional-type orogens that lack any evidence for the production and subsequent subduction of oceanic lithosphere, and the resultant converging continental fragments are the same as those that underwent initial extension (Hoffman

1973; Hoffman *et al.* 1974; Burke 1977; Şengör *et al.* 1978; Şengör 1995). The location of aulacogens adjacent to sites of successful ocean opening means that they are linked to subsequent sites of collisional or accretionary orogens; for example, the Oklahoma aulacogen, SE Laurentia, lies marginal to the Appalachian–Ouachita orogen. The degree of deformation and metamorphism during compressional reactivation of aulacogens is generally minimal (Hoffman *et al.* 1974). Aulacogens or rifts may also form at a high angle to the orogenic trend during collision (Şengör 1976; Şengör *et al.* 1978).

Sites of intracontinental subsidence are sites of thermal and/or rheological weakening that can be reactivated during compression, often in response to far-field stresses (see Sandiford *et al.* 2001). Examples include the late Mesoproterozoic to early Neoproterozoic successions in the North Atlantic (Cawood *et al.* 2004, 2007), and the Neoproterozoic Centralian Supergroup of Central Australia, which was deformed during the late Neoproterozoic Peterman and Palaeozoic Alice Springs orogenies (Collins & Teyssier 1989; Walter *et al.* 1995; Sandiford & Hand 1998; Hand & Sandiford 1999; Cawood & Korsch 2008). These examples are associated with the transformation of Rodinia into Gondwana, and the deformation and metamorphism of the sedimentary successions overlaps with events at the plate margin edges of these supercontinents (e.g. Cawood *et al.* 2004).

Some sites of intracratonic orogenesis are ultimately related to continental margin subduction. For example, intracratonic orogenic activity up to 1300 km inboard of the inferred plate margin occurred in the mid-Palaeozoic and Permo-Triassic in the Terra Australis orogen in eastern Australian and South African segments, respectively (Trouw & De Wit 1999; Collins 2002*a*), in China in the Mesozoic (Kusky *et al.* 2007*b*; Li & Li 2007) and along South America in the Tertiary (e.g. Kay & Mpodozis 2002). This activity parallels the plate margin and is related to strain localization in rheologically weak back-arc lithosphere (see Hyndman *et al.* 2005), possibly associated with flat-slab subduction and, hence, is part of the accretionary orogen deformation cycle. Other sites of intracratonic deformation occur inboard of zones of continent–continent collision (Fig. 1, Tianshan) and relate to stress transmission through weak quartz-dominated continental rheologies (see Dewey *et al.* 1986).

Characteristics of accretionary orogens

Accretionary orogens comprise a range of mafic to silicic igneous rocks and their sedimentary derivatives that develop on oceanic (e.g. West Pacific) or continental (e.g. Andes and Japan) lithospheric substrates during continuing plate convergence. Magmatic arc activity is characteristically calc-alkaline in composition but ranges from low-K tholeiite to shoshonitic, depending in part on the nature of the interaction of the magma with the arc substrate (e.g. Tatsumi & Eggins 1995). Magmatic activity occurs when the subducting, dehydrating slab of oceanic lithosphere interacts with mantle in the wedge of the overriding plate, and generally initiates at a depth of around 100 km or more above the downgoing slab (Tatsumi 2005). Importantly, and in contrast to magmatic activity at mid-ocean ridges or within-plate settings, arc magmas contain up to a few per cent water and other volatile phases and, as a consequence, are highly explosive (e.g. Krakatoa). Thus the extrusive products of arc volcanism are often reworked as pyroclastic and volcaniclastic deposits into intra-arc and arc flanking basins of the adjoining forearc and back-arc. Water also influenced the crystallization of the plutonic sections of arcs producing amphibole-bearing gabbros and ultramafic rocks, as demonstrated experimentally by Müntener *et al.* (2001) and geologically by Claeson & Meurer (2004) in the Proterozoic Trans-Scandinavian arc belt of Sweden.

Accretionary orogens are variably deformed and metamorphosed by tectonothermal events, commonly in dual, parallel, high-T and high-P regimes up to granulite and eclogite facies (Miyashiro 1973*a*; Ernst 2005; Brown 2006, 2009). Deformational features include structures formed in extensional and compressive environments during steady-state convergence (arc or back-arc v. accretionary prism) that are overprinted by short regional compressive orogenic events (Kusky & Bradley 1999; Collins 2002*a*).

Still-evolving accretionary orogens, such as those around the Pacific, have long, narrow aspect ratios, but completed orogens may be as broad as long (e.g. Central Asian Orogenic Belt, Arabian–Nubian Shield, and Superior and Yilgarn provinces). However, at least with the Superior and Yilgarn provinces, this appears to reflect the subsequent tectonic events such that the original linear extent of these bodies is unknown.

Lithotectonic elements of convergent plate margin systems include an accretionary prism incorporating accreted and tectonically dismembered ocean plate strata, forearc basin and its substrate, magmatic arc and back-arc basin. Some may also incorporate accreted arcs and oceanic plateaux and slices of convergent margin assemblages that have moved along the margin through strike-slip activity.

Crustal structure in accretionary orogens

Geophysical studies of accretionary orogens, including seismic reflection, refraction, seismic tomography and teleseismic data, have provided fundamental insights into structures of these orogens and their formation. Because of pronounced lateral variations across oceanic trenches and subduction zones, studies that combine multiple seismic and non-seismic data have been the most successful at determining the deep structure (e.g. Wannamaker *et al.* 1989).

Tonanki and Nankai subduction zones (Japan)

Japanese subduction zones are among the best-studied circum-Pacific accretionary complexes, with a lithospheric structure that is considered typical of many convergent plate margins. Two crustal models derived from seismic refraction–wide-angle reflection and gravity data collected across the Nankai Trough show the geometry of the subducting oceanic crust as it descends beneath the Japan volcanic arc (Fig. 4; Sagiya & Thatcher 1999; Kodaira *et al.* 2000; Nakanishi *et al.* 2002; Wells *et al.* 2003). These models define the geometry of the thick sedimentary basins that are located between the Nankai Trough and continental Japan. The subducting lower plate is divisible into three crustal layers separated from the upper mantle by a marked velocity jump. The wedge-shaped upper plate consists of a low-seismic velocity sedimentary package and the higher velocity igneous crust of the

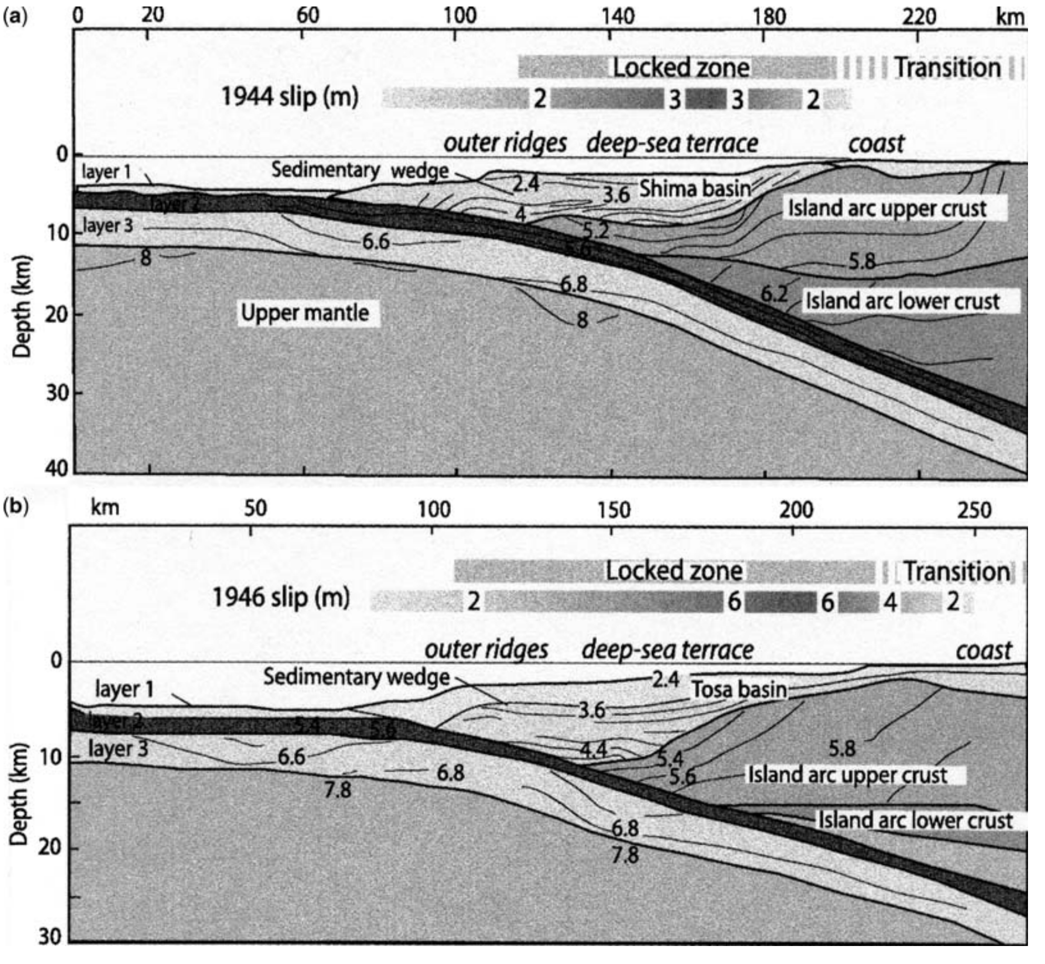

Fig. 4. Seismic velocity (km s^{-1}) structure of (**a**) the Tonanki subduction zone, site of the 1944 earthquake, and (**b**) the Nankai subduction zone, site of the 1946 earthquake. The relationship between the crustal structure and the locked zone is shown (Hyndman *et al.* 1995; Wang *et al.* 1995; Sagiya & Thatcher 1999; Kodaira *et al.* 2000; Nakanishi *et al.* 2002). The crustal structure is typical of many subduction zone complexes and includes a prominent low-seismic velocity sedimentary wedge and the higher-velocity igneous crust of the island arc (after Wells *et al.* 2003).

island arc, which is internally divisible into upper and lower crust on the basis of differences in seismic velocity.

Western margin of North America

Western North America is composed of a series of accreted oceanic domains (Coney *et al.* 1980; Samson *et al.* 1989; Fuis & Mooney 1990; Fuis 1998; Fuis *et al.* 2008). A detailed seismic transect from the active plate boundary at the Aleutian Trench in the Gulf of Alaska to the orogenic foreland fold-and-thrust belt on the margin of the Arctic Ocean shows a history of continental growth through magmatism, accretion and underplating (Fuis & Plafker 1991; Fuis *et al.* 2008; Fig. 5a). The edge of the Pacific plate (labelled 'A' in Fig. 5a and b) has velocities of 6.9 km s^{-1} and is covered by a thin upper layer of the lower oceanic crust of the Yakutat terrane with velocities of 6.1–6.4 km s^{-1}. This difference results in a structurally induced crustal doubling and contrasts with the situation inboard where the subducting oceanic crust has a normal 5–10 km thickness (Fig. 5b). Previously accreted oceanic lithosphere (B1, B2; Fig. 5b) is Mesozoic to early Cenozoic in age and contains magnetic, intermediate-velocity rocks of the Peninsular terrane, as well as interpreted regions of the Kula plate, which have velocities of 5.6–7.7 km s^{-1} at depth. The Cenozoic accretionary prism (C; Fig. 5a and b) is the Prince William terrane and the Mesozoic accretionary prism (D′; Fig. 5b) is a tectonic wedge, and includes the Chugach terrane and Border Ranges ultramafic–mafic assemblage (BRUMA; Kusky *et al.* 2007*a*). The backstop to the Mesozoic prism (E; Fig. 5b) is composed of the Peninsular and Wrangellia terranes. Near the Arctic margin the Brooks Range reveals crustal thickening attributed to the development of a foreland fold-and-thrust belt that overlies a tectonic wedge of North Slope lithosphere (Fuis *et al.* 1997). Crustal underplating in southern Alaska and crustal thrusting in northern Alaska overlapped in the Palaeogene and can be related to an orogenic float model in which a décollement extended northward from the subduction zone in the south to the Brooks Range in the north (Oldow *et al.* 1990; Fuis *et al.* 2008).

Southern Vancouver Island in British Columbia (Fig. 6a) contains material accreted from the subducting Juan de Fuca plate (A) (Clowes *et al.* 1987, 1995, 1997; Hyndman *et al.* 1990; Fuis & Clowes 1993). The Crescent–Siletz terrane, which is similar to the Yakutat terrane in Alaska, as well as unidentified tectonically underplated rocks and possible fragments were accreted in the Cenozoic era (B2, B3; Fig. 6a), together with a possible remnant of an oceanic plateau from the Palaeocene–Eocene. A thick wedge of Cenozoic accretionary prism lies near the toe of the overriding plate (C; Fig. 6a), whereas the Pacific Rim terrane is the Mesozoic accretionary prism (D; Fig. 6a). The Wrangellia terrane, the West Coast Plutonic Complex, and Nanaimo sediments in Georgia Strait form the backstop to the Mesozoic prism (E; Fig. 6a). An undivided lower crust is not defined, and other Cenozoic rocks include sedimentary strata on the continental shelf and Pacific Ocean basin (G; Fig. 6a). In contrast, the seismic transects in northern British Columbia and the Yukon show that the accreted Mesozoic and younger terranes are thin-skinned and constrained to the upper crust, and are underlain by a wedge of Proterozoic sedimentary rocks derived from cratonic Laurentia (Snyder *et al.* 2002, 2009). The deeper portions of the accreted blocks must have detached or underthrust the wedge during accretion. MacKenzie *et al.* (2005) reported evidence for ultrahigh-pressure garnet peridotite in the Canadian Cordillera that indicated that the thickness of the Proterozoic lithosphere was >100 km (and possibly up to 150 km).

A transect from Santa Cruz to Modesto in the San Francisco Bay region (Fig. 6b) is representative of accreted regions in northern and central California (Fuis & Mooney 1990; Page & Brocher 1993). There is no actively subducting oceanic crust, but oceanic lithosphere that accreted in the Cenozoic (B3; Fig. 6b) is preserved. The accreted material is interpreted as a 5–18 km layer at the base of the crust that can be traced seaward to the Pacific oceanic crust, containing intermediate- to high-velocity rocks. Other transects from California contain lithosphere accreted in the Mesozoic, which includes ophiolite complexes from the Great Valley, characterized as magnetic, dense, intermediate- to high-velocity rocks. The Cenozoic accretionary prism (Fig. 6b) is present as unidentified Cenozoic sedimentary rocks interfaulted with the Franciscan assemblage. The Mesozoic accretionary orogen (D′, E and G; Fig. 6b) comprises the Franciscan terranes, Coast Range ophiolite, the Great Valley sequence and Sierran foothills.

Songpan–Ganzi terrane (China)

The triangular-shaped Songpan–Ganzi terrane in the central Tibetan plateau lies between the Qinling–Qilian orogen to the north and the Qiangtang terrane to the south (Fig. 7). The crust consists of a vast tract of highly deformed and locally metamorphosed Triassic deep marine sedimentary rocks interpreted as the fill of a diachronously closing remnant ocean basin (Nie *et al.* 1994; Ingersoll *et al.* 1995; Zhou & Graham 1996). This terrane formed during Jurassic deformation and

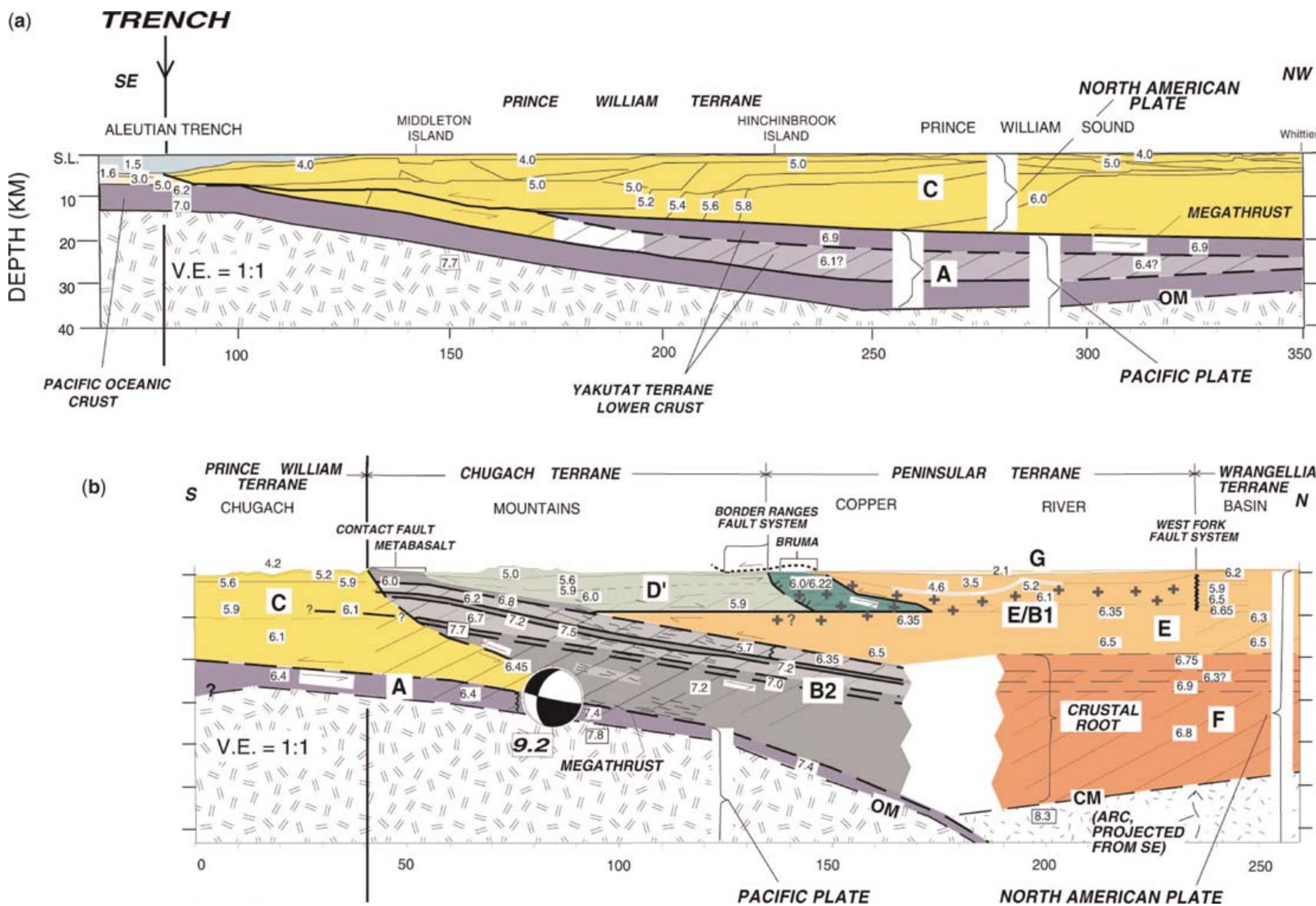

Fig. 5. (*Continued*).

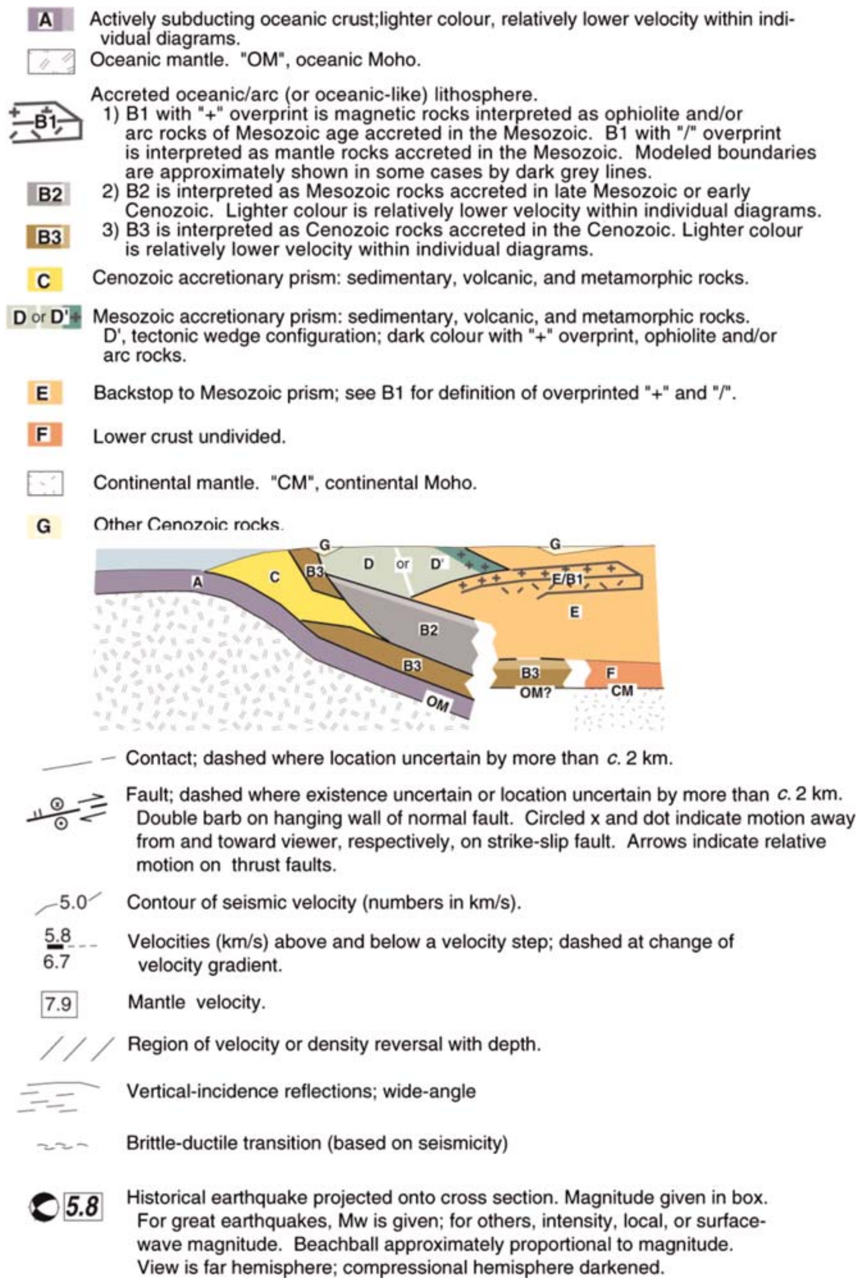

Fig. 5. Deep structure of the accretionary margin of southern Alaska, North American Cordillera (Fuis 1998). (**a**) The Pacific plate subducts at a shallow angle beneath metasedimentary rocks of the Prince William terrane (C) and underplated oceanic crust (Yakutat terrane lower crust). This active subduction zone maintains a low angle beneath the 250 km wide accretionary terrane. (**b**) The transition from the Prince William terrane (C) to the Chugach terrane (D) is marked by a pronounced transition from metasedimentary rocks to thick imbricated sheets of igneous oceanic crust. North of the Chugach terrane, the crust thickens to at least 50 km beneath the arc-related Peninsular terrane. This composite accretionary margin is highly diverse in terms of the crustal lithology. Section (b) is offset 150 km across strike from section (a).

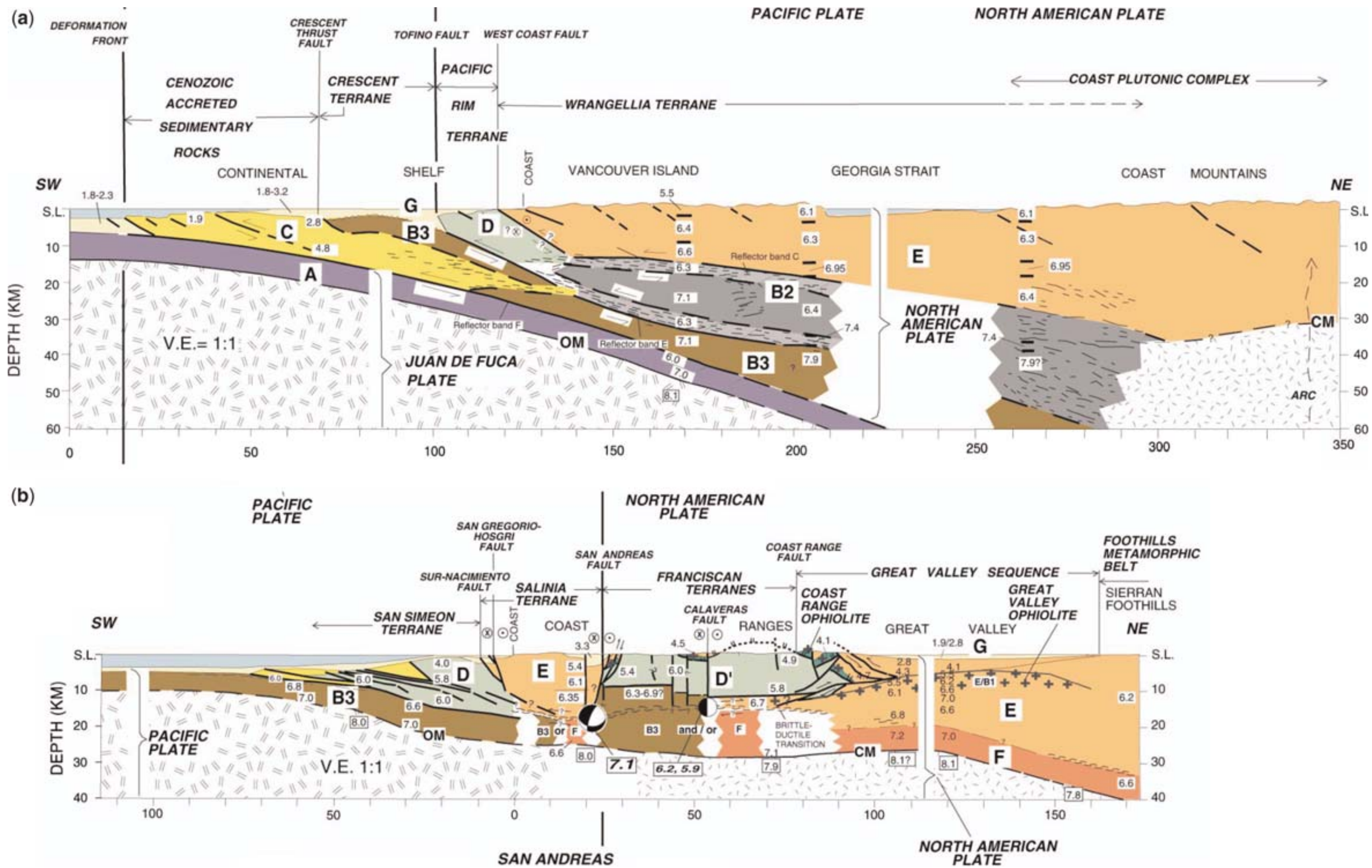

Fig. 6. (*Continued*).

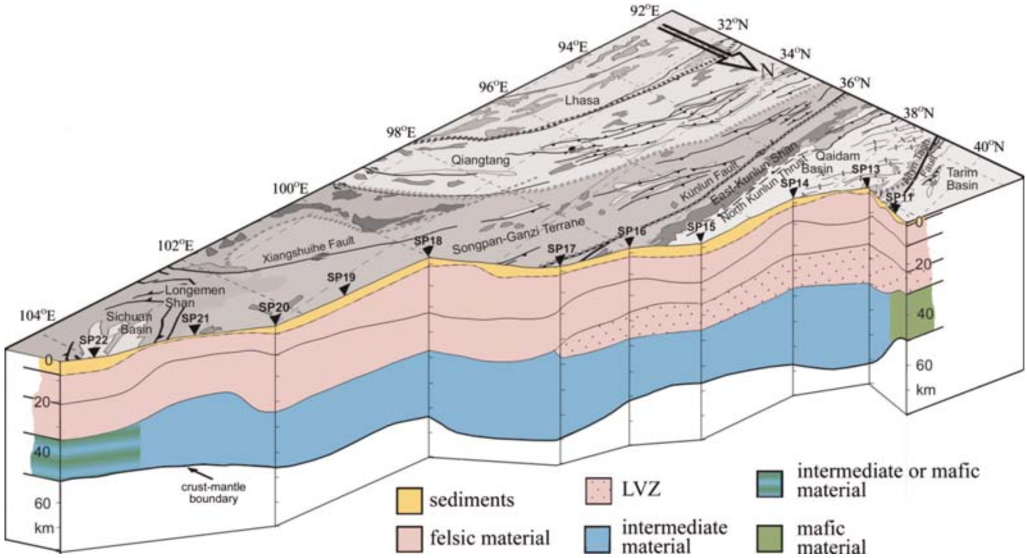

Fig. 7. Three-dimensional perspective view of the crustal structure beneath the Songpan–Ganzi accretionary orogen within a recent deep seismic transect from the southern Tarim basin to the Sichuan basin (Wang *et al.* in press). Colouring within the different layers indicates composition as derived from Poisson's ratio and P-wave velocity. Major faults and thrusts transected by the profile are qualitatively extrapolated at depth. The accreted upper and middle crust reaches a maximum thickness of 40 km beneath the Songpan–Ganzi terrane. The lower crust is 20–30 km thick and may consist of underthrust crystalline crust of continental affinity.

greenschist-facies metamorphism (Ratschbacher *et al.* 1996; Xiao *et al.* 1998) and was elevated above sea level at *c.* 20 Ma (e.g. Tapponnier *et al.* 2001). The terrane is inferred to be underlain by continental crust of the South China Block (Luo 1991). Recent seismic measurements show that the felsic upper crust (flysch? $V_p = 5.95\ \mathrm{km\ s^{-1}}$) is at least 10–20 km thick, and seismic velocities remain low ($V_p = 6.25\ \mathrm{km\ s^{-1}}$) to 40 km depth (Fig. 7; Wang *et al.* 2009). This implies that accreted material may attain a thickness of 40 km. Furthermore, the total crustal thickness is *c.* 70 km beneath the northern Songpan–Ganzi terrane (Fig. 7). The origin of the 20–30 km thick lower crust beneath this terrane is enigmatic, but the lack of surficial volcanic rocks indicates that an igneous origin is unlikely. A more plausible model would involve underthrusting of crystalline continental crust from the north and east. We note that the thickness of the lower crust is three to five times greater than the 7 km thickness of typical oceanic crust. Crustal thickness decreases towards the eastern border of the Tibetan plateau and reaches *c.* 48 km beneath the Sichuan basin (SP22, Fig. 7). Despite more than 14 km of crustal thinning, the topography remains constant across the Songpan–Ganzi terrane at an elevation of *c.* 4 km and then abruptly drops by 3.4 km from the elevated Longmen Shan into the low-lying Sichuan basin (elevation 0.6 km). Thinning of the crust along this portion of the profile is therefore mainly caused by thinning of the upper crust ($V_p = 5.95\ \mathrm{km\ s^{-1}}$).

Fig. 6. Cross-sections of North American Cordillera (Fuis 1998). (**a**) Accreted margin across Vancouver Island, British Columbia, showing some features similar to southern Alaska (Fig. 5b), particularly with respect to the diversity of crustal lithologies, including accreted sedimentary (C) and volcanic rocks (B3 and D) and tectonically imbricated sheets of igneous oceanic crust (B2). The total volume of igneous rocks at this margin exceeds that of sedimentary rocks. The crust thickens to about 50 km some 200 km east of the deformation front. (**b**) Central California, southern San Francisco Bay region: Santa Cruz to Sierra Nevada foothills. This accretionary margin has undergone hundreds of kilometres of lateral displacement along several faults, most prominently the San Andreas Fault. Some elements in common with the margins of southern Alaska (Fig. 5) and British Columbia (a) are the presence of (1) accreted sedimentary rocks (San Simeon and Franciscan terranes), (2) an arc terrane (Salina) and (3) underplated oceanic crust (B3). Deep geophysical data across these and other accretionary systems provide strong evidence for pronounced lithological diversity within the crust. Legend as for Figure 5.

Seismic data from Precambrian accretionary orogens

Seismic traverses across the Palaeoproterozoic accretionary orogens of the Svecofennian domain of NW Europe and the Hottah terrane of NW Canada have delineated mantle reflections dipping at about 30° from the Moho to about 100 km depths that are interpreted as fossil subduction zones (BABEL Working Group 1993*b*; Cook *et al.* 1999). Seismic reflection, refraction and geoelectric data imply the Svecofennian to be a collage of microcontinental blocks with intervening basins (Korja *et al.* 1993; Korja & Heikkinen 2005; Lahtinen *et al.* 2005, 2009). The reflection seismic data (BABEL, FIRE) revealed well-preserved pre-, syn- to post-collisional structures (e.g. a fossilized arc margin with an attached accretionary prism; BABEL Working Group, 1993*a*, *b*), whereas geochemical and petrogenetic studies suggest that the juxtaposed pieces were of Palaeoproterozoic origin. Comparative seismic reflection studies integrated with geological data for the Palaeoproterozoic Svecofennian, Scottish and Trans-Hudson orogens demonstrated in each case that 1.9–1.8 Ga lithosphere was wedged into crustal flakes that overrode Archaean margins (Snyder *et al.* 1996*a*). The Svecofennian accretionary orogen could serve as an analogue of the future accretionary-turned-collisional orogen that will be preserved when the Indonesian archipelago, with its variable size and age, is squeezed between Eurasia and Australia.

Crustal sections (Figs 4–7) cover a spectrum from active convergent plate margins to cratonized equivalents preserved in an accretionary orogen and reveal a range of processes in the development of continental crust. Actively subducting margins (Figs 4, 5 and 6a) reveal a coherent downgoing plate and an overlying forearc sedimentary wedge developed on an igneous arc basement, which at accreting margins form a backstop to offscraped sedimentary slivers. Stabilizing of the arc system occurs through underplating and accretion of oceanic material (Fig. 5, Yakutat terrane) and the progressive oceanward progression of the plate margin through accretion of trench sediments and of older arc systems (Figs 5 and 6). Termination of subduction as a result of changing plate kinematics (Fig. 6b) or continental collision and extensive crustal thickening (Fig. 7) results in final cratonization of the arc system.

Accretionary orogen types

Accretionary orogens can be grouped into two end-member types (Fig. 3), namely retreating and advancing (see Royden 1993*b*), based on their contrasting geological character, and modern examples from the eastern and western Pacific reflect a gross long-term kinematic framework with respect to an asthenospheric reference frame (Uyeda & Kanamori 1979; Dewey 1980; Lallemand *et al.* 2008). Retreating orogens are undergoing long-term extension in response to lower plate retreat (trench rollback), with respect to the overriding plate (Royden 1993*a*), resulting in upper plate extension, including back-arc basin opening as exemplified by the Tertiary history of the western Pacific (Taylor & Karner 1983; Leitch 1984; Schellart *et al.* 2006). Advancing orogens develop in an environment in which the overriding plate is advancing towards the downgoing plate at a rate equal to, or greater than, the rate of lower plate slab retreat, and this results in overall upper plate compression (Lallemand *et al.* 2005, 2008). For the eastern Pacific this corresponds to westward motion of the North and South American plates (Russo & Silver 1996*a*; Silver *et al.* 1998; Oncken *et al.* 2006*a*). This resulted in accretion (and strike-slip motion) of previously rifted arc and microcontinental ribbons, and the development of extensive retro-arc fold-and-thrust belts (e.g. Johnston 2001). Husson *et al.* (2008) argued that trench advance in South America is driven by high Andean topography and that this westward push is strong enough to shear the entire Pacific upper mantle with a surface velocity of 30 mm a^{-1}.

Advancing and retreating settings of accretionary orogens are simplified 2D representations of what is likely to be a more complex response to an overall environment of oblique convergence. Oblique accretion has played an important role in the assembly of the Cordillera in western North America (Johnston 2001; Colpron & Nelson 2006, 2009; Colpron *et al.* 2007), and probably also in many other orogens.

Retreating orogens

Retreating plate margins develop where the rate of rollback of the downgoing plate exceeds the rate of advance of the overriding plate, resulting in crustal extension in the latter. Rollback is driven by the negative buoyancy of the downgoing slab with respect to the underlying mantle, which in turn induces a backward sinking of the slab and retreat of the slab hinge, causing the overriding plate to extend (Elsasser 1971; Schellart & Lister 2004).

Upper plate extension leads to the development of intra-arc and arc-flanking basins culminating in rifting of the arc and development of back-arc basins (Dickinson 1995; Marsaglia 1995; Smith & Landis 1995). Many retreating orogens have multiple back-arc basins that generally, but not always, young outboard, towards the retreating plate margin. The preservation and incorporation of such basin fills within a retreating accretionary orogen is dependent on features or processes active during continuing subduction (e.g. thickness of sediment cover on the downgoing plate, rate of rollback) and the character of tectonothermal events that deform and stabilize the orogen in the rock record.

The process of rollback of the downgoing plate and the consequent development of back-arc basins is well developed in the SW Pacific. Between 82 and 52 Ma east- and NE-directed rollback of the Pacific plate by some 750 km was accommodated by opening of the New Caledonia, South Loyalty, Coral Sea and Pocklington back-arc basins (Schellart *et al.* 2006). Change in the relative motion of Pacific–Australia at 50 Ma resulted in subduction of the South Loyalty and Pocklington basins. This subduction was followed by two additional phases of rollback of the Pacific slab of some 650 and 400 km during opening of the South Fiji and Norfolk basins between 25 and 15 Ma and the Lau Basin from 5 to 0 Ma, respectively (Schellart *et al.* 2006). Slab rollback and back-arc basin extension is also argued to have played a fundamental role in the development and subsequent cratonization of the arc systems in the Terra Australis orogen in eastern Australia (Collins 2002*a*; Foster *et al.* 2009). The eastern third of Australia is composed of arc systems that developed along, and were accreted to, the rifted margin of East Gondwana following the initiation of subduction in the late Neoproterozoic (Cawood 2005). Subduction is inferred to have commenced at or near the continent–ocean boundary of East Gondwana. The width of eastern Australia and New Zealand prior to opening of the Tasman Sea was some 2000 km. Since that time, this region has undergone overall orogenic foreshortening of the order of 50%, but local foreshortening may have been considerably higher; for example, the western Lachlan segment of the Terra Australis orogen has a current width of 330 km and restored original width of between 800 and 1200 km (Gray & Foster 2006; Foster & Gray 2007). Thus, overall rollback of the proto-Pacific plate since the start of subduction towards the end of the Neoproterozoic until now is of the order of 6000 km, comprising some 4000 km during the Palaeozoic and Mesozoic that is preserved in the geological record of Australia and New Zealand and another 1800 km in the Cenozoic as documented by Schellart *et al*. (2006) in the SW Pacific. Rollback has not been continuous throughout this time frame and was undoubtedly interspersed with periods when rollback was either stationary or, with respect to the overriding plate, was advancing and driving periods of orogenesis (Collins 2002*a*). Extension was accommodated by both back-arc opening (Coney *et al.* 1990; Coney 1992; Fergusson & Coney 1992) and offscraping and accretion of material from the downgoing plate (Cawood 1982; Fergusson 1985).

Advancing orogens

Advancing orogens are characterized by widespread crustal shortening and uplift (Fig. 8), including the development of retroarc fold-and-thrust belts. The modern South American Cordillera is an example where deformation patterns can be placed into a plate-tectonic framework. Oncken *et al.* (2006*b*) provided a quantitative analysis of the spatial and temporal distribution of deformation in the Central Andes, and concluded that the amount and rate of shortening of the upper plate as well as lateral variability at the leading edge of the plate is primarily controlled by the difference between the upper plate velocity and the oceanic plate slab rollback velocity (see Russo & Silver 1996*a*; Silver *et al.* 1998; Schellart 2008). This allows the upper plate to override the downgoing plate, resulting in coupling and deformation across the plate boundary, including subduction erosion (Scholl & von Huene

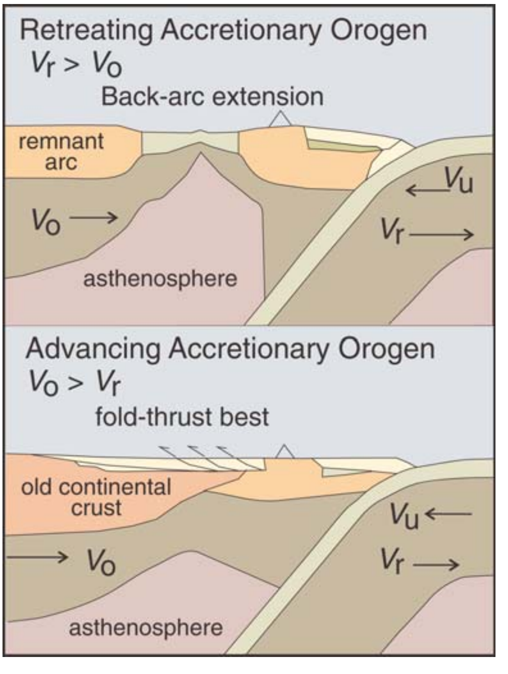

Fig. 8. Accretionary orogen types (after Cawood & Buchan 2007). For the retreating type the velocity of slab retreat (V_r) for the underriding plate (V_u) is greater than that of the overriding plate (V_o), whereas for the advancing orogen the velocity of the overriding plate is greater.

2009). Furthermore, Oncken *et al.* showed that the influence of differential trench–upper plate velocity is modulated by factors such as a phase of high trench sediment accumulation that reduced coupling across the plate boundary between about 45 and 33 Ma and a stage of reduced slab dip between 33 and 20 Ma that accelerated shortening. The magnitudes of post-Eocene crustal shortening across the Andes reach a maximum value of some 250–275 km in the central Andes and decrease to the north and south (Oncken *et al.* 2006*b*). The analysis of Schellart *et al.* (2007; see also Schellart 2008) suggests that this lateral variation in shortening may be related to along-strike variations in the rate of trench rollback, relative to the westward motion of the overriding South American plate. Schellart *et al.* (2007) used a global analysis of subduction zone width to show that trench migration rate is inversely related to slab width and depends on proximity to a lateral slab edge. Thus, for the 7000 km long South American trench, slab retreat is greatest at the northern and southern ends of the trench and least in the Central Andes in the vicinity of the Bolivian orocline.

The Central Andes has remained essentially stationary for the last 25 Ma and is the site of greatest crustal thickening. However, based on palaeomagnetic data, formation of the Bolivian orocline occurred in the last 10 Ma (Rousse *et al.* 2003), and not over the 25–50 Ma timeframe required by the model of Schellart *et al.* (2007). Russo & Silver (1996*b*) proposed that lateral variations in rollback are probably accommodated by trench-parallel flow in the mantle of the subducting slab. Seismic anisotropy beneath the Nazca Plate suggests trench-parallel mantle flow in the north and south but negligible flow in the central region, below the Bolivian orocline. Alternatively, Iaffaldano & Bunge (2008) argued that topographic load of mountain belts leads to increased frictional forces between the downgoing and overriding plates and that, specifically, uplift of the Andes over the last 10 Ma was linked to the slowdown in convergence between the Nazca and South American plates over this timeframe (see Iaffaldano *et al.* 2006).

Foreland fold-and-thrust belts, located inboard of the magmatic arc (Jordan 1995), are well developed in advancing orogens as a result of horizontal shortening, crustal thickening and resultant loading. They are well developed along the continental interior of the North and South American cordilleras.

Tectonic switching

Accretionary plate margins and orogenic systems can switch between phases of advance and retreat. Collins (2002*a*) proposed that episodes of orogenesis in the Lachlan segment of the Terra Australis orogen was driven by periodic advance of the downgoing plate through flat-slab subduction of an ocean plateau, which was otherwise undergoing long-term retreat with respect to the overriding plate. Lister *et al.* (2001) suggested that an accretionary orogen can undergo multiple cycles of tectonic mode switching (see also Beltrando *et al.* 2007). They proposed that accretion of a continental ribbon at a convergent plate margin results in shortening and burial of the overriding plate followed by a stepping out of the subduction zone beyond the accreted terrane and a new phase of rollback of the downgoing plate causing extension and exhumation in the overriding plate. This concept was extended by Lister & Forster (2006) into two types of tectonic mode switches: pull–push cycle, in which a retreating margin changes to an advancing margin, and the opposing push–pull cycle, in which the margin changes from advancing to retreating.

Sedimentary successions and accretionary orogens

Accretionary prisms

Structures formed during steady-state subduction are focused at the interface between the overriding and downgoing plates and are associated with the offscraping and underplating of material from the downgoing plate to form a subduction complex–accretionary prism (Fig. 9). Material accreted to the overriding plate can be subsequently removed and carried into the mantle through subduction erosion along the subduction channel (Scholl *et al.* 1980; Scholl & von Huene 2007). The subduction channel is the boundary zone between the upper and lower plates (Shreve & Cloos 1986; Beaumont *et al.* 1999). Variations in the strength and width of the subduction channel, which in part reflect the strength and thickness of material on the downgoing plate, can affect the behaviour of the overriding plate (De Franco *et al.* 2008). The interplay of advancing v. retreating accretionary plate margin with either the offscraping of material from the downgoing plate and its incorporation into an accretionary prism or the subduction (or erosion) of this material leads to the recognition of four types of plate margins (De Franco *et al.* 2008): accretionary prism with back-arc compression (e.g. Alaska, Sumatra, Nankai margin of Japan); erosive margin with back-arc extension (e.g. Central America, Marianas, Tonga); accretionary prism with back-arc extension (e.g. Lesser Antilles, Aegean, Makran); erosive margin with back-arc compression (e.g. Peru, Honshu margin of Japan, Kurile).

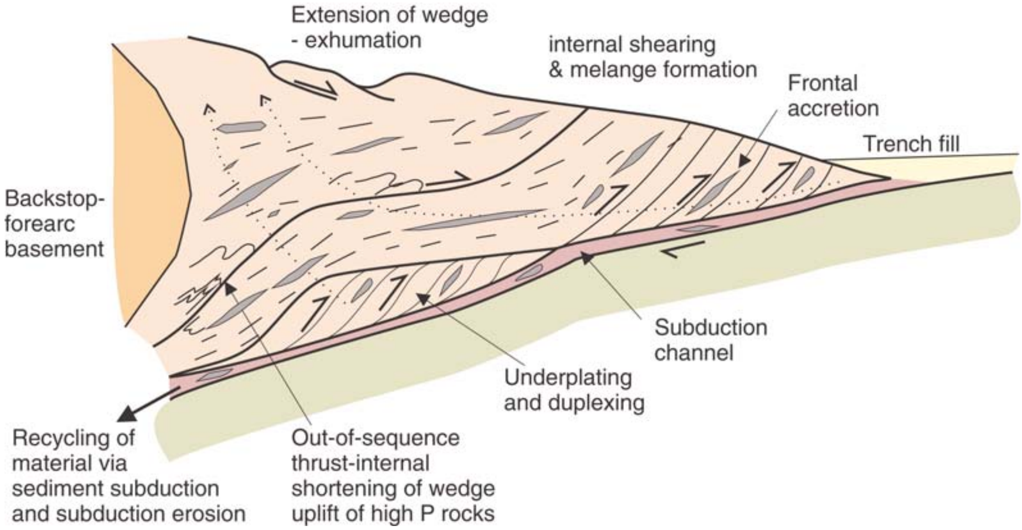

Fig. 9. Schematic section of accretionary prism showing sediment accretion through frontal accretion and basal accretion, and internal deformation of the wedge through contraction and extension (adapted from Kusky *et al.* 1997*b*). Particle paths are simplified net vectors and are drawn relative to backstop (see van Gool & Cawood 1994). Backstop is arc basement, and is composed of magmatic arc igneous rock or ophiolitic material.

Accretion of material results in the retreat of the trench axis seaward, away from the margin, and the active widening of the margin with time through progressive accretion from the downgoing plate (e.g. Seeley *et al.* 1974). According to Scholl & von Huene (2007), of the *c*. 42 000 km of active convergent margin subduction zones only some 30% have well-developed subduction complexes, and these are commonly characterized by a trench floor with a well-developed turbidite sequence and an orthogonal convergence rate of less than 40–50 km Ma^{-1}. Examples include Nankai, East Alaska, Cascadia and Sumatra–Andaman margins. Some high-latitude convergent margins (southern Chile and Alaska), contain small, young prisms developed through enhanced input of glacial age turbidites to trenches (Scholl & von Huene 2009). In southern Chile, the long-term processes on the margin are non-accreting or erosive (Bangs & Cande 1997).

A range of physical conditions extends from the trench floor to deep within the subduction complex. High fluid pressure and shearing lead to a spectrum of structures within the prism ranging from discrete thrust imbrication of relatively coherent sedimentary packages to chaotic mélange formation. Where least disrupted, the sequence displays a distinctive ocean plate stratigraphy consisting, from bottom to top, of a succession of mid-ocean ridge basalt (MORB), chert, hemipelagic mudstone, turbidite or sandstone and conglomerate (Fig. 10). This sequence records the history of sedimentation on the ocean floor as it travels from a mid-ocean ridge spreading centre to a trench. The biostratigraphy, structure, and geochemistry of this offscraped sequence has been studied in, for example, Phanerozoic circum-Pacific orogens in Japan (Isozaki *et al.* 1990; Matsuda & Isozaki 1991; Kimura & Hori 1993; Kato *et al.* 2002), California (Cowan & Page 1975; Sedlock & Isozaki 1990; Isozaki & Blake 1994), Alaska (Kusky & Bradley 1999; Kusky & Young 1999), Eastern Australia (Cawood 1982, 1984; Fergusson 1985) and New Zealand (Coombs *et al.* 1976; Mortimer 2004). Imbricated ocean plate stratigraphy is also increasingly recognized in Precambrian orogens; for example, in the 600 Ma Mona Complex of Anglesey, North Wales (Kawai *et al.* 2006, 2007; Maruyama *et al.* in press), the 2.7 Ga Point Lake greenstone belt (Kusky 1991), and possibly the 3.5 Ga chert–clastic sequence in the Archaean Pilbara craton (Kato *et al.* 1998; Kato & Nakamura 2003) and the 3.8 Ga Isua greenstone belt, West Greenland (Komiya *et al.* 1999). However, the validity of the Pilbara successions as an imbricated ocean-plate has been questioned by Van Kranendonk *et al.* (2007), who favoured a plume-related intracontinental setting. These rocks are interlayered with felsic volcanic rocks, and Williams & Collins (1990) have pointed out that they are commonly intruded by granites of the same age.

Systematic disruption of the ocean plate and trench sequence results in the production of broken

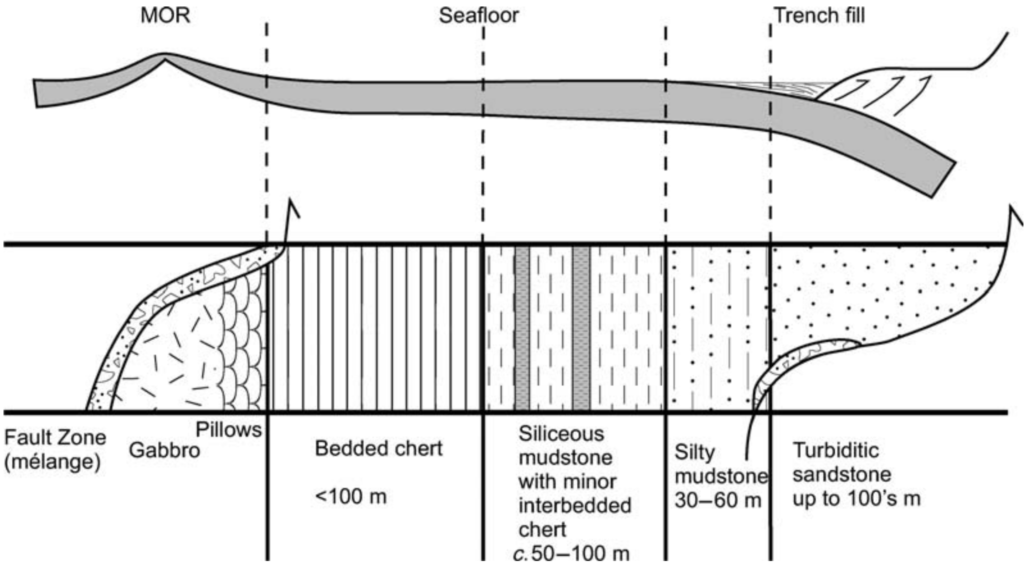

Fig. 10. Schematic section through a coherent thrust slice of offscraped ocean plate showing lithotectonic associations that develop on the plate as it moves from a mid-ocean ridge spreading centre to a trench.

formation and mélange (Greenly 1919; Hsü 1971). Results from southern Alaska and eastern Australia have shown that controls on whether chaotic mélange or coherent flysch are accreted at convergent margins may lie in the thickness of the sedimentary pile being subducted on the downgoing plate (Fergusson 1984; Kusky *et al.* 1997*b*; Kusky & Bradley 1999; Sisson *et al.* 2003). Downgoing plates with a thin sediment cover tend to concentrate shear strain in a thin zone that includes oceanic basement structural highs, whereas plates with thick sediment cover tend to disperse shear strains through a thick stratigraphic section with resulting less obvious deformation.

Bathymetric highs on the downgoing oceanic plate, such as guyots, may be offscraped into, and disrupt, the accretionary prism. Modern examples of seamount subduction appear to be associated with both sedimentary and tectonic disruption of the accretionary wedge (Ballance *et al.* 1989; Cawood 1990). Inferred seamount material, including alkali basalt and oceanic reef limestones, occurs in late Palaeozoic to Mesozoic accretionary prisms in Japan (Isozaki *et al.* 1990; Tatsumi *et al.* 1990; Isozaki 1997), including the huge late Jurassic Sorachi oceanic plateau, which was accreted in the early Cretaceous (Kimura *et al.* 1994). The eclogitic slab on the Sanbagawa mountains was derived from part of an oceanic plateau that was accreted, subducted and exhumed (Maruyama, pers. comm.). The Izu–Bonin arc collided with the Honshu arc in the late Cenozoic to give rise to spectacular indentation and curvature of the whole of central Japan (Soh *et al.* 1998).

Differentiating sedimentary successions within the accretionary orogen

Continuing convergent plate margin processes (subduction, magmatism, accretion, tectonic erosion), as well as subsequent processes involved in the incorporation and stabilization of convergent plate margin elements within continental lithosphere, destroy the original geometry so that accretionary orogens rarely contain a continuous and idealized distribution of lithotectonic elements. This structural complication may lead to uncertainty in ascertaining original settings and affinities for these elements, particularly in relation to their previous tectonothermal evolution, location of the original arc, apparent lack of accretionary prisms, occasional large dimension, and the composition and tectonic setting of the ophiolite slivers.

Many accretionary orogens, particularly the larger, less well-understood varieties, contain vast accumulations of deep-water turbidites that are tectonically intercalated with arc terranes and intruded by post-tectonic granites. The critical question here is whether these turbidites are accretionary wedge material, with ophiolitic slivers interpreted as ocean-floor lithosphere derived from the subducted plate, or whether they are back-arc basin fills, interlayered with the ophiolitic slivers representing remnants of the back-arc basin (see Foster & Gray 2000; Collins 2002*a*; Foster *et al.* 2009).

One of the major uncertainties has been the origin and timing of the heat budget within such turbidite assemblages, as many display a high-temperature (T), low-pressure (P) secondary mineral assemblage. If the turbidites are interpreted as offscraped subduction-related sequences then the mineral assemblage requires the migration of the magmatic arc, the inferred source of the heat, into the subduction complex (see Matsuda & Uyeda 1971) during oceanward propagation of the plate boundary during slab retreat. In this situation, the models that envisage accretionary orogens simply as subduction–accretion complexes require that the high-T regimes associated with arc magmatism should be superimposed upon low-T high-P regimes, including blueschist-facies terranes related to subduction accretion. Such overprints have been documented in the Chugach complex in Alaska and the New England segment of the Terra Australis orogen (Dirks *et al.* 1992, 1993), but many high-T, low-P metamorphic terranes bear no record of such overprinting. It is possible that the high-T metamorphism was sufficient to destroy all high-P evidence, but equally, many extensive turbidite complexes preserve original bedding and stratigraphic continuity over large areas, so it can be demonstrated that they never experienced high-P metamorphism associated with an evolving prism. In such instances, such turbidite piles could have filled back-arc basins, with the first metamorphism being associated with emplacement of igneous rocks, within either an arc (intra-arc basin) or a back-arc setting.

Another determinant of original setting of accretionary orogen turbidite assemblages is sedimentary lithotype. Accretionary prisms typically receive detritus from the adjacent arc and hence, commonly, are lithic-rich, whereas back-arc basins are likely to include detritus shed from the adjacent continental interior, and are likely to be more quartzose (Dickinson & Suczek 1979; Dickinson & Valloni 1980; Cawood 1983, 1990, 1991*a*, *b*; Dickinson 1985). Accordingly, arc–trench sandstones (forearc basins–accretionary prisms) also should contain a much higher proportion of young arc-derived zircons than old cratonic grains (Cawood *et al.* 1999; Cawood & Nemchin 2001). None the less, this is not always definitive, as far-travelled continental detritus can be deposited in accretionary prisms, such as the Barbados Ridge in the Lesser Antilles, which largely consists of Andean detritus shed via the Amazon (Parra *et al.* 1997).

The composition and structural relations of ophiolites are also probably discriminants between accretionary prisms and back-arc basins. Greenstones of accretionary prisms are typically fault-bound slivers ranging from those at the base of a relatively coherent thrust sheet (Fig. 10) to dismembered tectonic lozenges in a mélange. They can show either MORB or ocean island basalt (OIB) geochemical signatures (e.g. Cawood 1984), with the former rocks being detached from their base and occurring at the bottom of an ocean-plate sedimentary sequence (Fig. 10), where stratigraphic relations are preserved. Ocean island basalts can occur interstratified within the sedimentary sequence and may, if originating as guyots, be overlain or associated with limestone lenses. In contrast, back-arc basins are likely to contain sills and flows of basalts, which typically preserve original contact relations. Moreover, these basalts are MORB-like, but with a subtle subducted slab flux component, evident as elevated large ion lithophile element (LILE) abundances, which form the typical spiked pattern of subduction-related arc basalts on spidergrams, although this pattern is more subdued than that of arc basalts (Jenner *et al.* 1987; Hawkins 1994; Collins 2002*b*). In the 1970s and 1980s, such subtle LILE additions were commonly perceived as metasomatic effects from metamorphism, or as the products of melting mantle lithosphere enriched during a previous subduction event. However, through the Ocean Drilling Program, it became evident that these basalts were the dominant type of oceanic back-arc basins (e.g. Smellie 1994). These basalts also happen to be the most common type in retreating accretionary orogens.

The presence or absence of silicic tuff horizons is another possible discriminant of sedimentary successions in supra-subduction zone settings. Back-arc basins reflect the transitional tectonic stage between extending arc and formation of oceanic back-arc basins. They commonly receive the volcanic products associated with crustal melting during the initial stages of back-arc extension, and the explosive nature of arc magmas means that pyroclastic and volcaniclastic detritus is easily redistributed into the back-arc, and directly overlies the oceanic lithosphere and may be interstratified with, and dispersed within, any hemi-pelagic successions. In contrast, the ocean plate sequence incorporated into accretionary prisms, commonly located at least several hundred kilometres outboard from the prism, is less likely to contain volcanic layers, particularly ash flow ignimbrites. Accordingly, the presence of silicic volcanic rocks (with or without mafic counterparts), particularly in the deeper part of the stratigraphic succession, and intermittent silicic tuff horizons higher in the turbidite pile, are indicators of back-arc basin environments. It should be noted, however, that tuffs occur interbedded with cherts as part of an ocean plate stratigraphy in the Ordovician Ballantrae ophiolite in Scotland, which were then imbricated into an interpreted forearc accretionary environment (Sawaki *et al.* in press).

Metamorphic patterns in accretionary orogens

Miyashiro (1961, 1972, 1973*b*) highlighted that convergent plate margins are characterized by two regional, paired metamorphic belts of inferred similar age but contrasting mineral assemblages representing discrete P–T regimes. These are a high-P belt formed under a low geothermal gradient, which lies on the oceanward side of a low-P belt formed under a high geothermal gradient (Oxburgh & Turcotte 1970, 1971). The high-P belt was equated with the zone of the accretionary prism and the low-P belt with the magmatic arc and back-arc (Miyashiro 1972, 1973*b*). This duality of metamorphic belts is a characteristic feature of accretionary plate margins (Ernst 2005). The age equivalence and primary across-strike setting of paired belts was, however, critically investigated by Brown (1998*b*) for part of the Japanese arc. He proposed that the inboard, low-P–high-T Ryoke Belt and the outboard high-P–low-T Sambagawa Belt originally lay along strike and were juxtaposed by sinistral strike-slip motion along the Median Tectonic Line. The high-T metamorphism in belts such as the Ryoke and Abukuma in Japan are, according to Brown (1998*a*), the result of ridge subduction and slab window formation just inboard of the trench and downgoing plate (Bradley *et al.* 2003). This represents an alternative mechanism to the magmatic arc and back-arc for producing high-T metamorphism. The distribution of low-P, high-T metamorphic assemblages and any associated magmatism will be limited to the site of ridge subduction, and age relationships will be diachronous, reflecting ridge–trench migration.

'Back-arc basin' orogeny

One of the more enigmatic features of accretionary orogens is the presence of peak (high-T) metamorphic assemblages during contraction (Thompson *et al.* 2001; Collins 2002*a*), which is impossible to reproduce during the structural evolution of an accretionary prism (Jamieson *et al.* 1998). Hyndman *et al.* (2005) realized that back-arcs were always regions of high heat flow, irrespective of whether the orogen was advancing or retreating. They showed that around the entire Pacific Rim, elevated heat flow and thin crust are normal in back-arcs. Even in the North American Cordillera, where high mountains, fold-and-thrust belts, and foreland basins attest to long-term crustal shortening associated with an advancing orogen, the crust of the Cordillera is only 30–35 km thick and heat flow (*c.* 75 mW m^{-2}) is almost twice that of the adjacent craton (*c.* 40 mW m^{-2}). The high heat flow exists because the entire lithosphere is only 50–60 km thick (see also Currie & Hyndman 2006; Currie *et al.* 2008). This hot, thin zone of lithospheric weakness becomes the focus of shortening during periods of increased compressive stress, and the heat is a natural consequence of shallow convection in the hydrous mantle wedge above the subducting plate (Hyndman *et al.* 2005). As a result, compressional features in accretionary orogens, which form above the subducting plate, develop up to 1000 km away from the accretionary prism.

Furthermore, the high heat flow and corresponding rheological weakness of the back-arc region make it the likely site for the focusing of deformation within the accretionary orogen system. Deformation in the eastern Myanmar–western Thailand region of SE Asia (Shan-Thai block) is focused within a pre-existing back-arc basin subjected to oblique strain related to the India–Asia collision (Morley 2009). Strain partitioning is characteristic of this region and is heterogeneous, with adjoining regions of cold lithosphere, corresponding to a forearc basin setting (e.g. Central Basin in Myanmar), remaining undeformed and the site of continuing sedimentation.

Cratonization and driving mechanisms of orogenesis in accretionary orogens

Conversion of convergent plate margins into stable continental crust typically involves deformation and crustal thickening during one or more tectonothermal events. This conversion occurs in an environment of continuing plate convergence such that orogenesis involves transfer and concentration of stress in the upper plate through transitory coupling across the plate boundary (Cawood & Buchan 2007). Potential mechanisms of coupling include: (1) subduction of buoyant oceanic lithosphere (flat-slab subduction); (2) accretion of buoyant lithosphere (terrane accretion); (3) plate reorganization causing an increase in convergence across the boundary (Fig. 11). The effects of flat-slab subduction (e.g. Ramos *et al.* 2002) and suspect terrane accretion (e.g. Maxson & Tikoff 1996) should be spatially limited to the region of either the flat slab or the accretion zone, which in turn should result in short-lived orogenesis and/or diachronous events that migrate along the convergent margin in harmony with the subducted plate movement vector. These are local mechanisms in which the effect (e.g. orogenesis) is directly linked to the cause (buoyant slab causing coupling). These mechanisms can be observed in modern orogens (Kay & Mpodozis 2001; Mann & Taira 2004) and are commonly invoked as a mechanism in the geological record (e.g. Holm *et al.* 2005; St-Onge *et al.*

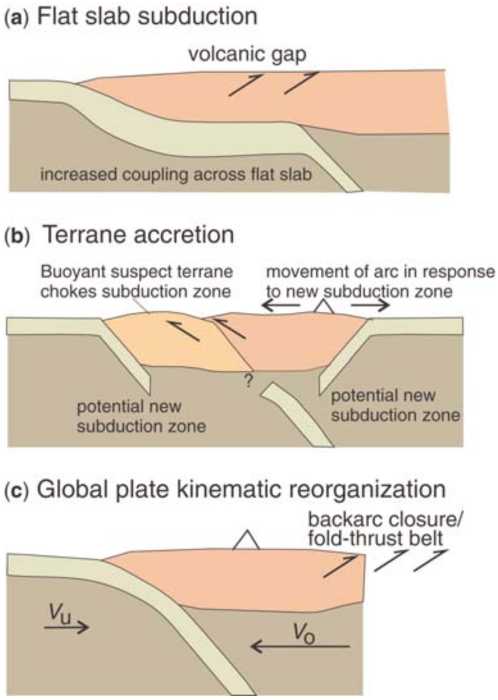

Fig. 11. Three possible modes of plate coupling that can drive orogenesis in an accretionary orogen: (**a**) flat-slab subduction; (**b**) terrane accretion; (**c**) plate reorganization leading to increased convergence.

2009). Accretion of arc and continental fragments has played a major role in the growth of SE Asia and the North American Cordillera (Coney *et al.* 1980; Metcalfe 1996*a*, *b*, 2002; Hall 2002, 2009). In contrast, the effects of plate reorganization are broader in scale and should extend synchronously along an orogen or plate boundary, albeit with variable effects, and reflect widespread and possibly long-term changes in orogenic character. Plate reorganization may traverse plate boundaries and be inter-orogen in extent. This reorganization is a regional mechanism in which the cause (plate reorganization) is not directly preserved at the site of its effect (orogenesis).

Crucial in establishing the potential contribution of these different coupling mechanisms to orogenesis is a detailed understanding of the spatial and temporal distribution of the tectonothermal effects of orogenic events in accretionary orogens. Synchroneity and cyclicity of accretionary orogenesis on an intra- and inter-orogen scale would suggest plate reorganization as the possible driver for orogenesis. Diachroneity, and/or orogenic events or belts restricted within an orogen would favour local events associated with terrane accretion or flat-slab subduction, or ridge subduction.

Subduction of buoyant oceanic lithosphere (flat-slab subduction)

Subduction of buoyant oceanic lithosphere will induce a flattening of the slab and can result from either the migration of young lithosphere associated with a spreading ridge or the thickened lithosphere of a hotspot (Fig. 11a; Gutscher *et al.* 2000; Gutscher 2002). Flat-slab subduction is currently occurring in a number of regions around the world, notably in southern Japan and South America (Ramos *et al.* 2002; van Hunen *et al.* 2002). Orogenesis driven by flat-slab subduction should be spatially limited to the region above the buoyant subducting lithosphere and will be relatively short-lived and diachronous, moving in concert with the subducting plate movement vector. Kay & Mpodozis (2001, 2002) argued that the thermal consequences of changing slab dip, combined with subduction of the Juan Fernandez Ridge hotspot track, have left a predictable magmatic and mineralization record in the Andes. Murphy *et al.* (1998) suggested that plume subduction led to flattening of the downgoing slab, generating ‘plume-modified orogeny’ (see Murphy *et al.* 1999, 2003; Dalziel *et al.* 2000). Flat-slab subduction and the resultant transitory plate coupling has been invoked as an important mechanism of orogenesis in the accretionary Lachlan orogen (Collins 2002*a*), in the North American Cordillera (Dickinson & Snyder 1978; Saleeby 2003), and in development of the Japanese accretionary orogen (Osozawa 1988; Underwood 1993; Isozaki 1996; Maeda & Kagami 1996; Brown 1998*a*). The mechanism for increased buoyancy with flat-slab subduction depends on the nature and rate of input of the thermal anomaly (Bradley *et al.* 2003; Kusky *et al.* 2003). Ridge subduction will be associated with a progressive increase in buoyancy, whereas plateau or hotspot subduction will induce a rapid change in crustal thickness and, hence, buoyancy. Subduction of plateaux and ridges has been proposed as a mechanism of orogenic growth in the Palaeoproterozoic Birimian terranes (Abouchami *et al.* 1990), and in the Archaean Zimbabwe craton (Kusky & Kidd 1992).

Ridge subduction is a diachronous process that typically involves a major change in plate convergence vectors between the upper plate and two subducting plates, with the change in plate convergence vectors across the subducting spreading ridge separated by a period of heating and igneous intrusion in the forearc and accretionary prism (above the slab window; Fig. 12). Such ridge–trench interaction played a major role in the development of the Tertiary North American Cordillera from Kodiak Island, Alaska, to Vancouver Island, British Columbia (Bradley *et al.* 2003; Sisson *et al.* 2003).

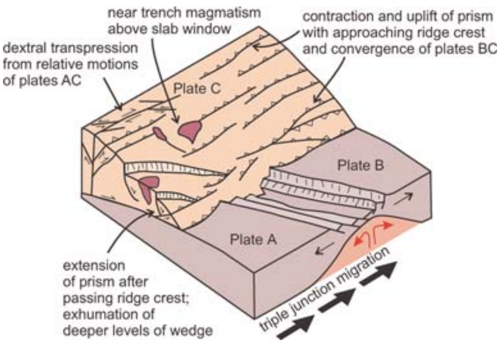

Fig. 12. Oblique view of hypothetical ridge–trench–trench triple junction, showing how structures in the upper plate will change with the passage of the triple junction, reflecting kinematic coupling between plates B and C before ridge subduction, and plates A and C after ridge subduction (adapted from Kusky *et al.* 1997*a*).

Deformation can be intense and is related not only to the plate convergence vectors (surface forces) but also to a change in dip of the subducting lithosphere as the ridge migrates along the trench (Kusky *et al.* 1997*a*; Haeussler *et al.* 2003; Pavlis *et al.* 2003; Roeske *et al.* 2003).

As a spreading ridge approaches a trench, progressively younger packages of offscraped sediments and volcanic rocks (Fig. 10) will be accreted to the overriding plate, and after a ridge is subducted, the accreted slabs will include progressively older packages of sediment and basalt along with the young sediments. Deformation becomes younger towards the trench and is superimposed on this complex pattern of ages of accreted sediments.

Accretion of buoyant lithosphere (terrane accretion)

If the lithosphere (oceanic or continental) is relatively thick (and buoyant) it may result in a temporary interruption to the subduction process through choking of the subduction zone, leading to the stepping out or flipping of the subduction zone (e.g. Ontong–Java Plateau; Petterson *et al.* 1999; Lister *et al.* 2001; Mann & Taira 2004). This flip will probably be associated with an interruption and/or migration of the magmatic arc (Fig. 11b). Terrane accretion was adopted by Coney *et al.* (1980) to explain the faulted juxtaposition of oceanic and convergent plate margin tectonostratigraphic units within the North American Cordillera. It is considered by many to constitute the main (sole) driving force for convergent margin orogenesis in that 'eventually a downgoing plate will carry continental or island arc crust into a subduction zone' (Moores & Twiss 1995, p. 212) to induce arc–arc or arc–continent collision, or terrane accretion (Dickinson 1977; Coney *et al.* 1980). Maxson & Tikoff (1996) argued that Cordilleran terrane accretion was the driving mechanism for the Laramide orogeny. Recent work in the Cordillera has emphasized that the number of terranes in the Cordillera is considerably less than originally envisaged by Coney and colleagues, and that many of the remaining terranes may not be suspect but are upper plate fragments that represent arcs and continental ribbons that lay outboard of, and along strike from, the Cordilleran margin (Monger & Knokleberg 1996; Johnston 2001; Colpron & Nelson 2006, 2009; Colpron *et al.* 2007). Seismic data across the northern Cordilleran orogen suggest that at least some of the accreted terranes are superficial with no deep crustal roots (Snyder *et al.* 2002, 2009) and may not be major impactors that drove orogenic events (Cawood & Buchan 2007); the terrane accretion model of orogenesis may therefore be 'suspect'.

Plate reorganization

Tectonic plate reorganization resulting from a change in the position and angular motion of Euler poles, perhaps related to termination of plate boundaries through collision or an increased spreading rate, will lead to a global readjustment in plate interactions and has been invoked as a potential cause of accretionary orogenesis (Colblentz & Richardson 1996). Vaughan (1995; see also Vaughan & Livermore 2005) proposed that pan-Pacific margin tectonic and metamorphic effects were a response to major plate reorganization associated with an increased spreading rate in the Pacific during the mid-Cretaceous (Sutherland & Hollis 2001). Cawood & Buchan (2007) highlighted evidence for deformation, mountain building and resultant crustal growth in accretionary orogens during phases of supercontinent assembly (Boger & Miller 2004; Foden *et al.* 2006). They undertook a detailed analysis of the timing of collisional orogenesis associated with supercontinent assembly compared with that for accretionary orogenesis along the margins of a supercontinent. They showed that age relations for assembly of Gondwana and Pangaea indicate that the timing of collisional orogenesis within the interior of the supercontinents was synchronous with subduction initiation and contractional orogenesis within the marginal Terra Australis orogen, which extended along the palaeo-Pacific margin of the these supercontinents.

Final assembly of Gondwana occurred at the end of the Neoproterozoic to early Palaeozoic, between about 590 and 510 Ma. This was coeval with a switch along the Pacific margin of the supercontinent from passive to convergent margin activity, followed by the Delamerian–Ross–Pampean orogenesis. Similarly, the final stages of assembly of the Pangaean supercontinent occurred during the end-Palaeozoic to early Mesozoic, between *c*. 320 and 250 Ma, and involved the accretion of Gondwana, Laurasia and Siberia. This phase of major plate boundary reorganization was accompanied by regional orogenesis along the Pacific margin of Gondwana–Pangaea (Gondwanide orogeny). The correspondence of this transitory coupling with, or immediately following, plate boundary reorganization, suggests that it may reflect plate kinematic readjustment involving increased relative convergence across the plate boundary. Cawood & Buchan (2007; see also Murphy & Nance 1991) suggested that this relationship probably reflects the global plate kinematic budget where termination of convergence during supercontinent assembly is compensated by subduction initiation and/or increased convergence along the exterior of the supercontinent. Transitory coupling across the plate boundary during subduction possibly accounts for the deformation and metamorphic pulses that develop in the accretionary orogens.

The analysis of Oncken *et al.* (2006*b*) suggests that Cenozoic orogenesis in the Andes is a response to global kinematic adjustment, in this case driven by opening of the Atlantic, resulting in an increase in westward drift of the South American plate relative to the Nazca plate (see Silver *et al.* 1998). Recently, Silver & Behn (2008) proposed that supercontinent collision may lead to a global loss of subduction. The data of Cawood & Buchan (2007), however, show that this concept is unlikely, at least in association with Gondwana and Pangaea assembly.

Accretionary orogens and plate tectonics; when did plate tectonics begin?

Because accretionary orogens require convergent plate margins, their appearance in the geological record heralds the initiation of horizontal plate interactions on the Earth. The question of when plate tectonics began, what criteria can be used to recognize it in the rock record and, once established, whether it was continuous, episodic and/or alternated with some alternative process are much debated (Sleep 2000; Hamilton 2003; Stern 2005; O'Neill *et al.* 2007; Condie & Kröner 2008). The consensus of opinion, however, is that convergent plate interaction and the recycling and subduction of material from the Earth's surface into the mantle has been active since at least 3.2–3.0 Ga (Cawood *et al.* 2006; Dewey 2007; Condie & Kröner 2008; Windley & Garde 2009) and possibly considerably earlier (Komiya *et al.* 1999; Harrison *et al.* 2005; Nemchin *et al.* 2008; Nutman *et al.* 2009; Polat *et al.* 2009). Well-constrained palaeomagnetic data demonstrate differential horizontal movements of continents in both Palaeoproterozoic and Archaean times, consistent with the lateral motion of lithospheric plates at divergent and convergent plate boundaries (Cawood *et al.* 2006). Well-preserved and unambiguous ophiolites associated with juvenile island-arc assemblages and modern-style accretion tectonics occur in the Palaeoproterozoic Trans-Hudson orogen of the Canadian shield (2.0 Ga Purtuniq ophiolite, Scott *et al.* 1992), the Svecofennian orogen of the Baltic shield (1.95 Ga Jormua ophiolite, Peltonen & Kontinen 2004), and in the Mazatzal–Yavapai orogens of southeastern Laurentia (1.73 Ga Payson ophiolite, Dann 1997). Furnes *et al.* (2007) proposed that ophiolite-related sheeted dykes and pillow basalts occur in the 3.8 Ga Isua greenstone belt in SW Greenland, but this has been disputed by Nutman & Friend (2007). Late Archaean mafic–ultramafic complexes in the North China craton have also been interpreted as ophiolites (Kusky *et al.* 2001; Kusky 2004), but have also been disputed (Zhao *et al.* 2007). Paired high-P–low-T and high-T–low-P tectonothermal environments (Miyashiro 1961, 1972, 1973*b*), requiring plate subduction, have been recognized as far back as the Neoarchaean (Brown 2006, 2009).

Inferred arc-related assemblages (magmatic arc, intra-arc basins and back-arc basins), indicative of subduction and convergent plate interaction occur within greenstone sequences in many Archaean cratons including Yilgarn, Pilbara, Superior, North China, Slave, and southern Africa. The lithological association in these greenstones, including calc-alkaline volcanic rocks, and locally boninite, shoshonite and high-Mg andesite, along with the associated geochemical signatures, are almost identical to those found in rocks of modern convergent plate margin arcs (Condie & Harrison 1976; Hallberg *et al.* 1976; de Wit & Ashwal 1997; Bai & Dai 1998; Polat & Kerrich 1999, 2004; Cousens 2000; Percival & Helmstaedt 2004; Smithies *et al.* 2004, 2005; Kerrich & Polat 2006; Polat *et al.* 2009).

Styles of Archaean and Proterozoic mineralization resemble Phanerozoic deposits related to subduction environments (Sawkins 1990; Kerrich *et al.* 2005), including a Palaeoarchaean porphyry Cu deposit (Barley 1992) and Archaean and Palaeoproterozoic volcanogenic massive sulphide Cu–Zn deposits (Barley 1992; Allen *et al.* 1996; Syme *et al.* 1999; Wyman *et al.* 1999*a*, *b*).

Condie & Kröner (2008) listed several distinctive petrotectonic assemblages such as accretionary prisms as well as arc–back-arc–forearc associations that argue for the existence of accretionary orogens since the early Archaean. For instance, the 3.2 Ga Fig Tree greywacke–shale sequence of the Barberton greenstone belt in South Africa has long been interpreted in terms of an accretionary prism (e.g. Lowe & Byerly 2007, and references therein), and there is geochronological, structural and geophysical evidence for terrane accretion in the late Archaean Abitibi greenstone belt, Superior Province, Canada through convergent plate interaction (Percival & Helmstaedt 2004, and references therein). The recognition of ocean plate stratigraphy in an orogen's rock record is a key indicator of both mid-ocean ridge spreading, required for its generation, and subduction accretion, necessary for its preservation. As such, it provides a key indicator for plate tectonics in the rock record. The proposal that an ocean plate stratigraphy is preserved in the Marble Bar greenstone belt in the Pilbara craton, supported by trace element geochemical data (Kato & Nakamura 2003), and in the Isua greenstone belt in Greenland (Komiya *et al.* 1999), suggests that ridge–trench movements and therefore plate tectonics were in operation in the Palaeoarchaean.

Deep seismic reflection profiling across a number of late Archaean and Palaeoproterozoic belts has identified dipping reflectors, in some cases extending into the mantle, which underlie arc assemblages in the preserved accretionary orogen and are interpreted as a frozen subduction surface (Calvert *et al.* 1995; Cook *et al.* 1999; Cook & Erdmer 2005; Korja & Heikkinen 2005; Percival *et al.* 2006; Lahtinen *et al.* 2009).

Nutman *et al.* (2009) and Polat *et al.* (2009) have presented data in support of convergent plate margin processes within the Eoarchaean accretionary orogens of Isua, Greenland and Anshan, China (*c*. 3.8–3.6 Ga). They showed that the lithotectonic assemblages in these regions and their geochemistry are similar to those in Phanerozoic convergent plate margins involving the subduction of young, hot lithosphere. Harrison *et al.* (2005) inferred that subduction may extend back to the earliest phases of Earth evolution. They suggested that the isotopic systematics of Jack Hills zircons, northern Yilgarn, indicate formation in a continental environment characterized by calc-alkaline magmatism and crustal anatexis, features seen in the modern Earth in convergent margin settings and implying that subduction was established by 4.4 Ga ago.

Accretionary orogens and continental growth

Continental growth involves the addition of mantle-derived (juvenile) material to the crust. Arc magmatism within accretionary orogens is invoked as the major source of this material but with additional input derived from mantle plumes. Geochemical and isotopic data have shown that the composition of continental crust resembles subduction-related igneous rocks and suggest that that there has been progressive growth of continental crust through time (Fig. 13; Taylor 1967; Taylor & McClennan 1985; McCulloch & Bennett 1994; Arculus 1999). Thus, accretionary orogens, with their subduction-related plate margins, are seen as the sites of net continental growth, rather than collisional orogens, which are envisaged as sites of crustal reworking (Dewey *et al.* 1986). Geochemical and isotopic studies from Neoproterozoic to Phanerozoic accretionary orogens in the Arabian–Nubian Shield, the Canadian Cordillera and the Central Asian Orogenic Belt indicate massive addition of juvenile crust during the period of 900–100 Ma (Samson *et al.* 1989; Kovalenko *et al.* 1996; Jahn *et al.* 2000*a*; Wu *et al.* 2000; Jahn 2004; Stern in press). However, recent whole-rock Nd and zircon Hf(–O) isotope data imply that continental crust formation was episodic, with significant pulses of juvenile magmatism and crustal growth in the Archaean and Palaeoproterozoic, and with no significant addition in the Phanerozoic (Fig. 13; Condie 1998, 2004; Hawkesworth & Kemp 2006; Kemp *et al.* 2006). Punctuated crustal growth may be related to mantle plume activity (Stein & Hofmann 1993; Condie 1998), and creates a paradox of global proportions, because plate tectonics can account for *c*. 90% of Earth's current heat loss with the remainder lost through plume activity (Davies 1999), and thus magmatic arcs should be the major site of continental growth. The isotopic age data outlined in Figure 13 are based on an analysis of the preserved rock record and the assumption that it is representative of the record of growth. In addition, improved microanalytical techniques are suggesting that the contribution of subduction processes to continental growth may have been masked. Hf–O isotopic analysis of zircons from the classic I-type granites of eastern Australia show that these rocks were formed by reworking of sedimentary materials through mantle-derived magmas rather than by melting of igneous rocks (Kemp *et al.* 2007) and thus are critical components of continental growth. Bulk rock isotopes will mask this component, and this suggests that the component of such material in continental growth may have been underestimated in the past.

The volume of continental crust that is added through time via juvenile magma addition appears to be effectively compensated by the return of continental and island arc crust to the mantle. Clift *et al.* (2009) and Scholl & von Huene (2009) have estimated that the long-term global average rate of arc magma additions is 2.8–3.0 $km^3\ a^{-1}$. Crustal

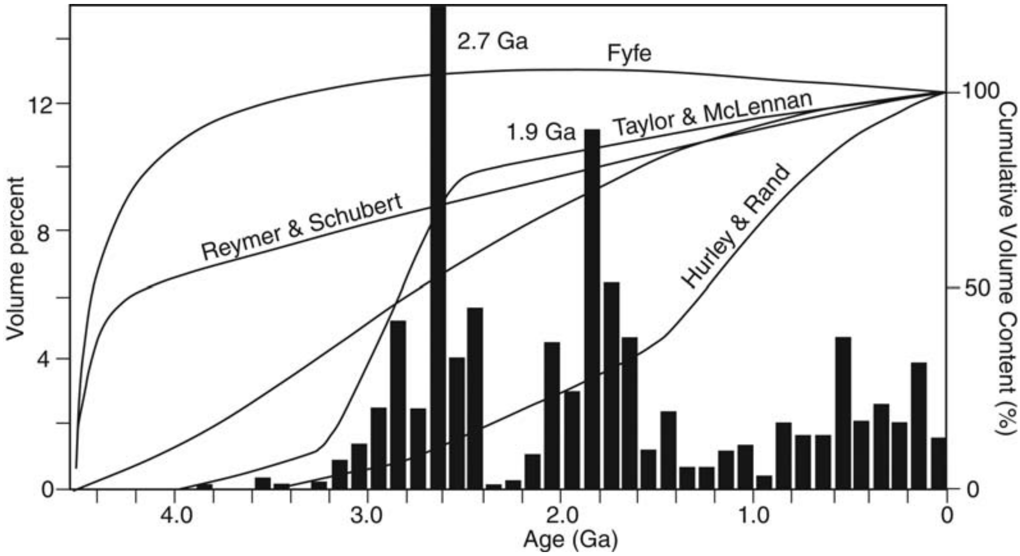

Fig. 13. Histogram of volume distribution of juvenile continental crust based on a compilation of U–Pb zircon ages integrated with Nd isotopic ages and lithological associations. Also plotted are models of continental growth, with 100% representing the present-day cumulative volume of crust (adapted from Condie 2005). Early models suggested slow initial growth followed by more rapid recent growth and were based on the geographical distribution of Rb–Sr and K–Ar isotope ages (Hurley & Rand 1969), which we now realize are probably reset by younger orogenic events. Some models suggested rapid early growth of continental crust, slowed by recycling of continental material and therefore resulting in a slower rate of growth or even a decrease in continental volume (Fyfe 1978; Reymer & Schubert 1984; Armstrong 1991). Other models have taken an intermediate approach to growth and require a more linear growth (unlabelled) or rapid growth during the late Archaean followed by steady-state growth driven by island arc magmatism (Taylor & McClennan 1985).

addition rates vary within arc systems over time, with the initial phase of arc construction (first 10 Ma) marked by rates that can be up to an order of magnitude larger than the long-term rate for an arc (Stern & Bloomer 1992; Stern 2004; Jicha *et al.* 2008).

Recycling of continental crust at convergent plate margins occurs by sediment subduction, subduction erosion and detachment of deeply underthrust crust (Scholl *et al.* 1980; Scholl & von Huene 2007, 2009; Clift *et al.* 2009). Sediment subduction entails the movement of lower plate sediment beneath the arc along the subduction channel (Fig. 9; Cloos & Shreve 1988*a*, *b*). Arc material may also be transported into the trench and then carried into the subduction channel. Material in the channel is carried beneath the frontal arc and if not underplated is carried into the mantle on the downgoing plate. Subduction erosion involves the transfer of material from the upper plate into the subduction channel and downward into the mantle. The loss of continental and arc crust through sediment subduction and subduction erosion has been estimated by Scholl & von Huene (2009) to be around 2.5 $km^3\ a^{-1}$ of which some 60% is due to erosion. Scholl & von Huene (2009) noted that continent and island arc crust is also carried into the mantle, where it can be detached, and is lost during final ocean closure and collision. They estimated that an additional 0.7 $km^3\ a^{-1}$ is recycled into the mantle by this process. Thus, the total volume of crustal material moved into the mantle at subduction zones is around 3.2 $km^3\ a^{-1}$. This rate is sufficient that if plate tectonics has been operating since around 3.0 Ga (see Cawood *et al.* 2006) then a volume equal to the total current volume of continental crust would have been recycled into the mantle (Scholl & von Huene 2009).

Given uncertainties in these estimates for both the addition of crust and its removal from convergent plate margins, the net growth of continental crust is effectively zero, with crustal growth through magma addition effectively counterbalanced by removal of material. Thus, plate tectonics in general and convergent plate margins in particular, as represented by accretionary orogens, are not the sites of continental growth through time but rather sites of crustal reworking. Any single arc system can, however, show net addition or removal of material, hence allowing its preservation or removal from the rock record. For example, the South American margin has been undergoing long-term crustal loss such that the trench has migrated landward with respect to the upper plate with time, resulting in the magmatic arc younging away from the trench and Jurassic arc magmas forming the most seaward land outcrops in the current forearc (Stern 1991; Franz *et al.* 2006; Glodny *et al.* 2006; Kukowski & Oncken 2006). Areas of rapid accretion of material, either through arc magmatism during the early stages of arc development or through the accretion of already assembled continental (e.g. arc fragments of the North American Cordillera) and thickened oceanic (e.g. Ontong–Java) crustal fragments, are more likely to survive the effects of crustal reworking and are more likely to be preserved in the geological record. This may therefore lead to selective preservation of periods of continental growth in the rock record as exemplified in Figure 13. In addition, the thickened crust of oceanic islands and plateaux, once incorporated into convergent plate margins (such as Wrangellia, Sorachi, Sanbagawa and some Archaean greenstones), are selectively preserved even during periods of subduction erosion.

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