Geological Society, London, Special Publications

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Geological Society, London, Special Publications 2003; v. 218; p. 21-41
doi:10.1144/GSL.SP.2003.218.01.03

Arc–trench rollback and forearc accretion: 1. A collision-induced mantle flow model for Tethyan ophiolites

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Abstract: Tectonically active remnants of Neo-Tethys represented by Mediterranean and western Pacific marginal seas are characterized by rapidly propagating backarc extension episodes. These appear to be triggered by random subduction nucleation events, commonly signalled by the appearance of refractory boninites in volcanic 'proto-arcs'. As backarc basins evolve, active arcs separate from their 'proto-arc' remnants and may split again if more than one basin-opening episode occurs. Accreting arc–forearc terranes are therefore likely to incorporate proto-arc, backarc, and (in some cases) inherited continental fragments, as evidenced by their structural complexity and lithological diversity. Forearc complexes typically show positive Bouguer gravity anomalies and significant age discrepancies within and between their crustal and mantle components. Where exposed, their lower stratigraphic horizons may include boninite-bearing assemblages along with tectonized fragments of mid-ocean ridge basalt (MORB) basement and hydrated refractory peridotite. These are typically intruded by sodic plagiogranite (adakite) and high-temperature Mn-, Fe-rich hydrothermal veins ('epidosites'), further indications of subduction nucleation at, or close to a pre-existing spreading axis. Where the arc–trench rollback process is terminated by collision with an approaching continent, or with another retreating forearc complex, MORB-like backarc lithosphere is rapidly reconsumed, in some cases following a change in subduction polarity. In contrast, given their preponderance of ultra-refractory serpentinized peridotite, forearc complexes are relatively buoyant, resist subduction, and are prone to entrapment during early stages of an orogeny. The associated interplay of extension and compression offers a compelling scenario for resolving the so-called ophiolite 'conundrum' and explaining the near-ubiquity of ophiolites in orogenic belts. We propose that rapid arc–trench rollback pulses are driven largely by collision-induced mantle flow in addition to commonly cited 'slab pull' effects. This is supported by the evidence of isotopic mantle flow tracers, seismic tomography, and the coupled kinematics of marginal basins and continental escape. Model applications to some well-known Tethyan ophiolites are developed in a companion paper.

Enigmatic rock assemblages known as ophiolites are characteristic features of most orogenic mountain belts (Dilek *et al.*, 2000). For several decades ophiolitic rock assemblages have been regarded as 'obducted' fragments of the oceanic lithosphere, generated at the global mid-ocean ridge system (Gass 1968, 1989; Moores *et al.* 1968; Coleman 1977) or above intra-oceanic subduction zones (Miyashiro 1973, 1975, 1977; Pearce *et al.* 1981, 1984). The implicit paradox of these interpretations has been reconciled to some extent by ascribing ophiolites to subduction-related marginal basins (Miyashiro 1973, 1975, 1977; Pearce *et al.* 1981, 1984; Beccaluva *et al.* 1994; Shervais 2001). However, because modern marginal basins *per se*, along with mid-ocean ridges and supra-subduction volcanic arcs, lack several key features of orogenic ophiolites, the ophiolite conundrum persists (Dilek *et al.*, 2000: Moores *et al.* 2000). Are some ophiolites exclusively oceanic in character and others dominantly arc-like, and are refractory boninites and high-Mg andesites (HMAs), absent from 'normal' ocean ridges, volcanic arcs and backarc basins, ubiquitous in ophiolites? Also, why are their high-temperature metamorphic 'soles' invariably mid-ocean ridge basalt (MORB)-like and near equivalent in age, and what causes their inflected metamorphic pressure–temperature–time (P–T–t) histories? Finally, what is the significance of the apparent correspondence observed between ophiolite genesis and distal plate tectonic events, and why is the time lag (less than *c.* 10 Ma) between the inception of ophiolite formation and their emplacement in continental margins so short?

As part of a solution to the ophiolite conundrum, these questions may be largely resolved if

From: DILEK, Y. & ROBINSON P. T. (eds) 2003. *Ophiolites in Earth History*. Geological Society, London, Special Publications, **218**, 21–41. 0305-8719/03/$15

ophiolites are considered to represent entrapped intra-oceanic forearc assemblages, as proposed by, for example, Dewey & Casey (1979), Casey & Dewey (1984), and Stern & Bloomer (1992). This notion has been resisted on the grounds that any plausible, general explanation for such a linkage is lacking. If, on the other hand, the ophiolite–forearc analogue is valid, and possibly unique, questions concerning the genesis of forearcs and how they may be incorporated into 'classic' orogenic assemblages still pose an important problem. There is, in fact, substantial evidence in support of the ophiolite–forearc analogy. Both features are uniquely characterized by the presence of MORB-like and calc-alkaline sequences, boninites, HMA (e.g. Figs 1 and 2), sodic granitoids ('adakite'), and Fe-, Mn-rich hydrothermal deposits, and share structural attributes that record a progression from sea-floor spreading to subduction-related tectonic processes (e.g. Shervais 2001). Moreover, a preliminary synthesis of geochronological data for Tethyan ophiolites (Flower *et al.* in prep.) suggests that their inception may be connected to distal plate tectonic effects. It is reasonable, therefore, to suggest that the sum of shared structural, stratigraphic, and petrological features in ophiolites and forearcs is an essential basis for resolving the ophiolite conundrum.

The Tethyan tectonic belt is the Earth's most active locus of continental plate collisions and offers a unique opportunity to pursue this goal. Remnant Neo-Tethyan basins in the Mediterranean and western Pacific, for example, show examples of forearc accretion in response to repeated episodes of subduction nucleation and backarc basin opening (Royden 1993a, 1993b; Bloomer *et al.* 1995; Jolivet & Faccenna 2000; Wortel & Spakman 2000) (Fig. 3). Moreover, the correlation of

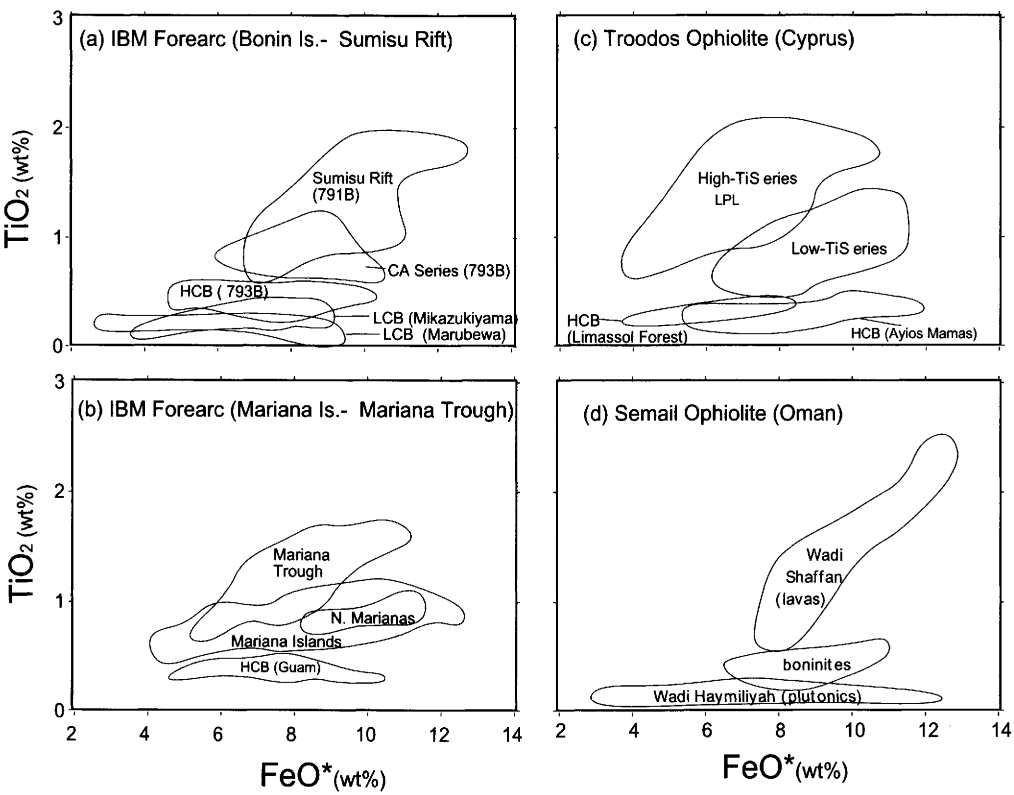

Fig. 1. Plots of TiO_2 vs. FeO^* (in wt%) for eruptive and intrusive lithologies sampled from typical Izu–Bonin–Mariana intra-oceanic forearcs and the Troodos (Cyprus) and Semail (Oman) ophiolites. (**a**) Bonin Islands and Sumisu Rift (Izu–Bonin–Mariana system) (Pearce *et al.* 1992a; Taylor *et al.* 1994). (**b**) Mariana Islands (Reagan & Meijer 1984; Stern *et al.* 1989) and Mariana Trough (Gribble *et al.* 1998). (**c**) Troodos ophiolite, Cyprus (Malpas *et al.* 1984; Flower & Levine 1987; Gibson *et al.* 1987; Rogers *et al.* 1989; Taylor *et al.* 1992; Bednarz & Schmincke 1994; Portnyagin *et al.* 1996, 1997). (**d**) Semail ophiolite, Oman (Umino *et al.* 1990; Lachize *et al.* 1996; Pezard *et al.* 2000; Ishikawa *et al.* 2002).

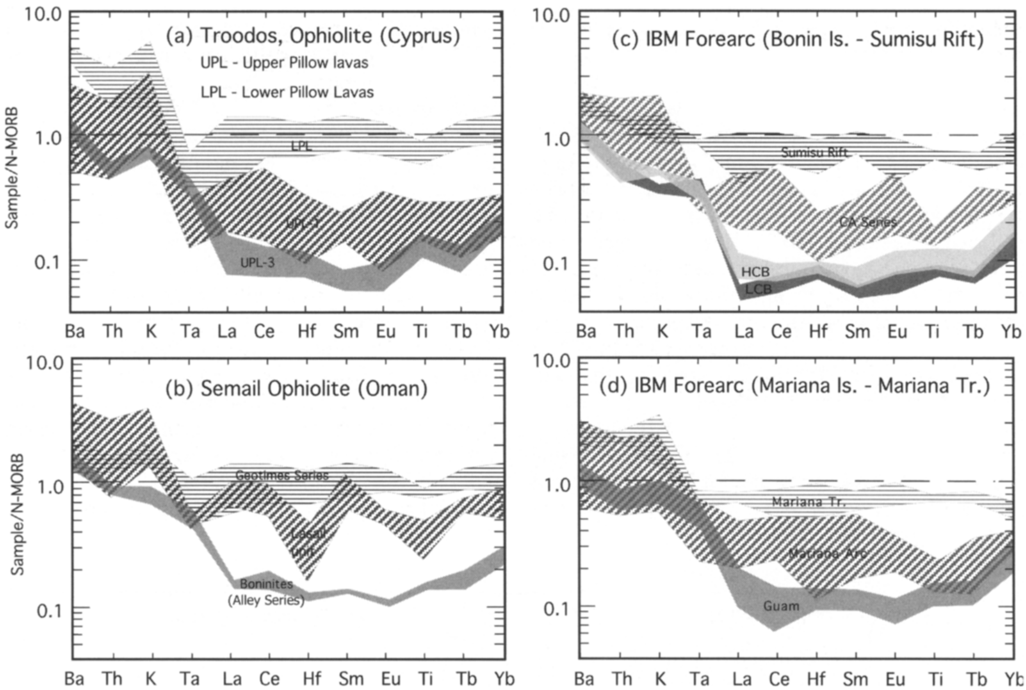

Fig. 2. MORB-normalized incompatible element distributions for eruptive lithologies sampled from typical forearcs and ophiolites. (**a**) Troodos ophiolite, Cyprus: Upper Pillow Lavas (series 1 and 3), and Lower Pillow Lavas (Flower & Levine 1987; Gibson *et al.* 1987; Rogers *et al.* 1989; Bednarz & Schmincke 1994; Portnyagin *et al.* 1996, 1997). (b) Semail ophiolite, Oman (Umino *et al.* 1990; Lachize *et al.* 1996; Pezard *et al.* 2000; Ishikawa *et al.* 2002). Alley Series volcanic rocks (calc-alkaline and boninitic), Geotimes Series volcanic rocks (MORB-type volcanic rocks). (**c**) Izu–Bonin arc–forearc, and backarc Sumisu Rift (Izu–Bonin–Mariana system) (Pearce *et al.* 1992a; Taylor *et al.* 1994). (d) Mariana arc–forearc (Reagan & Meijer 1984; Stern *et al.* 1989) and backarc Mariana Trough (Gribble *et al.* 1998).

such processes with distal plate kinematic changes suggests a fundamental connection between ophiolite genesis and global-scale plate tectonics (Flower *et al.* 2001; Flower 2003). The causes of arc–trench rollback may therefore be crucial to our understanding of ophiolites and their geodynamic significance in Earth history. Here, following the 'Tectonic Facies' approach of Hsü (1994) and Hsü (1997), we present an 'actualistic' model for ophiolites based on processes of forearc evolution in western Pacific and Mediterranean marginal basins. In a companion paper (Dilek & Flower this volume) the model is adopted as a template for interpreting three well-studied Tethyan ophiolites.

A brief history of Tethys

Tethyan orogens mark a succession of continental plate collisions that followed breakup of the Gondwana continent and repeated cycles of ocean basin opening and closure. Although the relevant plate kinematic reconstructions are controversial, there is a general agreement that the northward drift of Gondwana fragments involved three or more such cycles of opening (e.g. Dercourt *et al.* 1986; Audley-Charles & Harris 1990; Ustaomer & Robertson 1993; Metcalfe 1996; Stampfli & Borel 2002). These cycles were commenced with diachronous 'unzipping' of the northern margin of Gondwana, and produced Tethyan basins evolving as triangular inlets that propagated westward from the proto-Pacific Ocean. According to the majority of views, Palaeo-Tethys was initiated in the Late Devonian with the separation of continental blocks that later amalgamated as the North China, South China, Iran, Kazakhstan, Indochina, Qaidam, Tarim, and Hainan blocks (Audley-Charles & Harris 1990; Metcalfe *et al.*, 1999). Meso-Tethys probably began to open in the Early Permian with detachment of the Cimmerian and other micro-continents, and Neo-Tethys began opening between the Late Triassic and Late Jurassic with the separation of what later became the Pelagonia, Tauride–Anatolide, Lhasa, West Burma, and Woyla blocks (Dilek *et al.* 1999; Metcalfe *et al.*,1999).

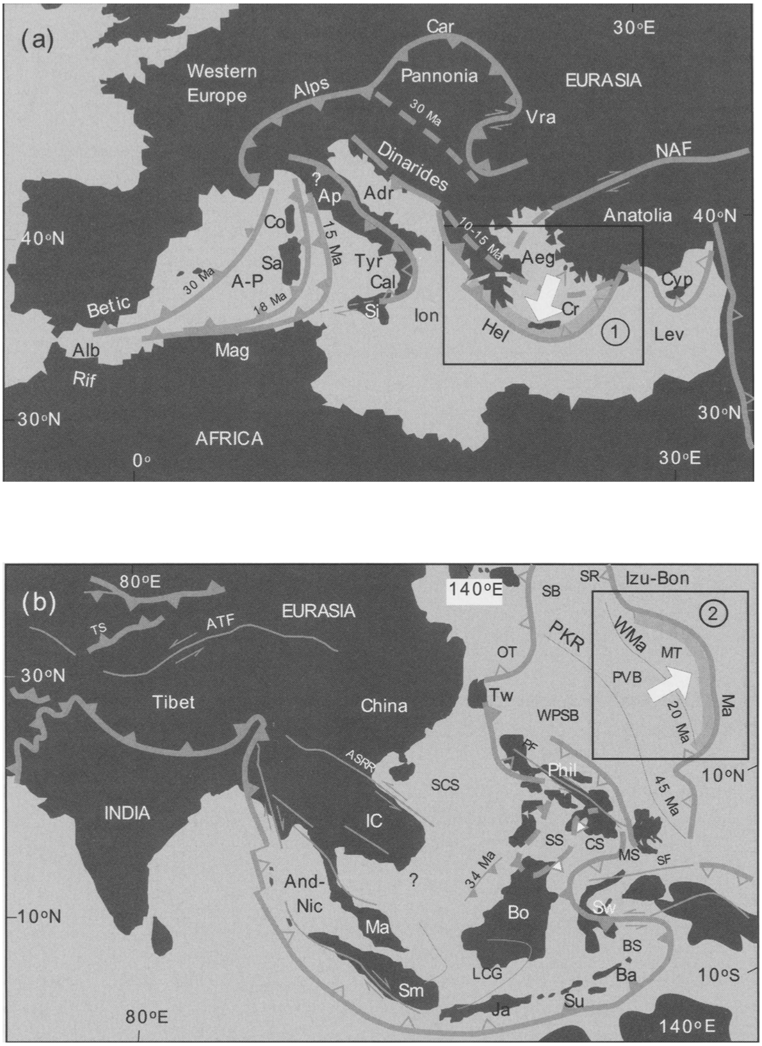

Fig. 3. Plate boundary evolution and arc–trench rollback in active Tethyan domains. (**a**) The circum-Mediterranean region (after Wortel & Spakman 2000). Arrows indicate directions of probable slab tearing beneath the Apennine–Calabrian, Hellenic, and Carpathian arcs. Adr, Adriatic Sea; Aeg, Aegean Sea; Alb, Alboran Sea; Ap, Apennines; A-P, Algero-Provencal Basin; Cal, Calabria; Car, Carpathians; Co, Corsica; Cr, Crete; Cyp, Cyprus; Hel, Hellenic arc–trench; Ion, Ionian Sea; Lev, Levantine Basin; Mag, Maghrebides (from the Rif to Sicily); NAF, North Anatolian Fault; Sa, Sardinia; Si, Sicily; Tyr, Tyrrhenian Sea. Barbs indicate subduction or thrusting vergence, black suggesting a continuous slab, and white, possible post-collision breakoff. (**b**) Western Pacific region (after Flower *et al.* 1998, 2001). Arrows indicate directions of probable slab tearing beneath the Himalayas, Sunda–Banda arcs, and Northern Luzon–Taiwan. TS, Tien Shan; ATF, Altyn Tagh Fault; And-Nic, Andaman–Nicobar Islands; IC, Indochina; Ma, Malay Peninsula; Sm, Sumatra; Bo, Borneo; Ja, Java; Su, Sunda; Ba, Banda; Sw, Sulawesi; Phil, Philippines; Tw, Taiwan; Izu-Bon, Izu–Bonin Islands; Ma, Mariana Islands; Wma, West Mariana arc; PKR, Palau–Kyushu ridge; Ru, Ryukyu Islands; SCS, South China Sea; SS, Sulu Sea; CS, Celebes Sea; MS, Molucca Sea; WPSB, West Philippine Sea Basin; BS, Banda Sea; SB, Shikoku Basin; SR, Sumisu Rift; PVB, Parece Vela Basin.

Palaeo-Tethys

The closure of Palaeo-Tethys and the corresponding inception of Meso-Tethys were marked by the accretion of Kunlun, Qaidam and Ala Shan Terranes to Kazakhstan–Siberia in the Early Permian, followed in the Late Permian to Early Triassic by suturing of Sibumasu and Qiangtang to Cathaysialand as Palaeo-Tethys was finally consumed by subduction (Metcalfe *et al.*, 1999). Meso-Tethys closure between the Late Triassic and Late Jurassic was accompanied in the east by diachronous accretion of the Lhasa, West Burma, and Woyla micro-continents (Metcalfe *et al.*, 1999), and in the west, Cimmeria, Iran, Pelagonia, and others (Dercourt *et al.* 1986; Ustaomer & Robertson 1993; Stampfli & Borel 2002) to Eurasia. Finally, as the African, Arabian, and Indian plates collided with Eurasia, and Australia collided with newly accreted Sunda, remnants of the Neo- and Meso-Tethyan lithosphere were being progressively consumed by subduction (Dercourt *et al.* 1986; Audley-Charles & Harris 1990; Metcalfe *et al.*, 1999; Stampfli & Borel 2002).

Although successive Tethyan basins were more or less separated by micro-continents throughout much of the Triassic and Jurassic, they may have remained connected at their western extremities, between the Mediterranean and Caucasus. This interpretation is supported by evidence suggesting that Mid-Jurassic remnant basins were being consumed by subduction as collisions between retreating arc–forearc complexes and continents prevented further extension (Stampfli & Borel 2002). For example, Paleocene closure of the Liguria–Piedmont basin was coeval with the Betic–Rif, western–northern Alpine, and Carpathian orogenies and, as younger basins collapsed, it was followed by the Neogene–Pleistocene Apennine, Maghrebe, Dinaride, and Hellenide orogenies (Faccenna *et al.* 1997; Jolivet *et al.* 1999; Jolivet & Faccenna 2000; Stampfli & Borel 2002).

Neo-Tethys

The closure of Neo-Tethys coincided with opening of the North Atlantic Ocean that began at *c.* 180 Ma and was followed by the separation of East and West Gondwana at *c.* 158 Ma. By *c.* 130 Ma East Gondwana (Africa–India–Seychelles–Madagascar–Australia–Antarctica–South America) had also begun to sunder as opening of the South Atlantic commenced at *c.* 110–100 Ma and the North Atlantic opening continued. By the Mid-Cretaceous, a block comprising Africa and India–Seychelles–Madagascar began to detach from Australia–Antarctica, followed shortly by the separation of Australia and initiation of seafloor spreading at the Southeast Indian Ridge (Metcalfe 1996). At *c.* 98 Ma, the India–Seychelles block separated from Madagascar and by the Late Cretaceous, along with Africa–Arabia and Australia, was moving rapidly northwards towards accreting Eurasia. Finally, following separation from the Seychelles at *c.* 65 Ma (Gnos *et al.* 1997), India collided with Eurasia between *c.* 50 and 45 Ma (Lee & Lawver 1994). After separating at *c.* 40 Ma, Arabia and Africa collided with accreting Eurasia at *c.* 30 Ma and 25 Ma, respectively (Dewey *et al.* 1989; Jolivet & Faccenna 2000).

The record of Gondwana disaggregation and (partial) reassembly offers a potential rationale for the timing and location of 'spontaneous' subduction nucleation and, in turn, the processes giving rise to ophiolites. However, the causes of subduction nucleation remain unclear. Are such events determined by global-scale plate kinematics (as suggested by Gnos *et al.* 1997) or do they reflect viscous mantle instabilities caused by density and thermal heterogeneities (e.g. Toth & Gurnis 1998; Faccenna *et al.* 1997)? These questions bear, in turn, on whether ophiolites represent a global-scale plate tectonic phenomenon (e.g. determining where they are initiated) or local phenomena related to an imminent plate collision (determining both where they are initiated and where they are emplaced). Although this latter question is beyond the scope of the present paper, we will attempt to provide a basis for its future consideration.

Towards an actualistic model

Today, Tethyan tectonic and magmatic activity is dominated by effects of the African, Arabian, Indian, and Australian collisions concomitant with continued basin opening in parts of the Mediterranean Sea and western Pacific (e.g. Fig. 3). On a global scale, subduction zones may remain static for lengthy periods and evolve as simple linear orogens. Others, notably in the regions discussed here, are observed to migrate oceanward at rates exceeding 100 m ma^{-1}, often developing into spectacular oroclines (Fig. 3). Such rapid 'arc–trench rollback' processes are expected to continue indefinitely unless they are terminated by collisions of their retreating arc–forearc complexes with mid-ocean ridges, continental plates, or other migrating subduction systems.

Forearc complexes as lithological 'high-tide marks'

In the Mediterranean, rollback cycles have mostly been interrupted by forearc collisions and the ensuing consumption (or 'collapse') by subduction of short-lived backarc basins (Dalziel 1989; Clift & Dixon 1998; Jolivet & Faccenna 2000; Robertson 2000). In contrast, arc–trench rollback in the western Pacific has been relatively unconstrained with continuing eastward propagation of forearcs, free from the 'jaws' of an impending plate collision (Karig 1971; Hussong *et al.* 1981; Karig *et al.* 1986; Jolivet *et al.* 1991b; Tamaki & Honza

1991). On the other hand, these regions show strong similarities, arc–trench rollback cycles in both cases being initiated by splitting of nascent volcanic 'proto-arcs' into active and remnant segments, the associated refractory magmas indicating unusual thermal conditions (Pe-Piper & Piper 1989, 1994; Stern & Bloomer 1992; Hawkins & Castillo 1998; Insergueix-Filippi *et al.* 1998, 2000). The net effect of this type of process, sometimes compounded by additional arc splitting events (Hussong *et al.* 1981; Ishii *et al.* 1995; Fassoulas 1999), is to produce an evolving forearc terrane that potentially includes the igneous and metamorphic products of successive 'proto-arc', arc, and backarc episodes (Bloomer 1983; Hickey-Vargas 1989; Johnson *et al.* 1991, 1992; Giaramita *et al.* 1992; Ishii *et al.* 1992, 1995; Marlow *et al.* 1992) along with characteristically high-temperature hydrothermal deposits (Banerjee *et al.* 2000; Fryer *et al.* 2000; Gillis & Banerjee 2000; Banerjee & Gillis 2001; Gillis 2002). The presence of allochthonous continental and oceanic lithosphere fragments may represent crustal features prior to the inception of rollback (Johnson *et al.* 1991; Giaramita *et al.* 1992; Parkinson *et al.* 1998; Parkinson & Arculus 1999). As they evolve, therefore, arc–forearc complexes progressively resemble lithological 'high tide marks' (HTMs), increasingly heterogeneous, accreted assemblages of proto-arc, arc, and backarc crust exhibiting significant internal age and structural discrepancies (Flower *et al.* 1998, 2001; Flower 2003).

Subduction nucleation

At least two lines of evidence highlight the anomalous thermal character of asthenospheric mantle associated with subduction nucleation events that appear to trigger rollback cycles: the presence of boninite and high-magnesium andesites (HMA) in proto-arcs (Casey & Dewey 1984; Stern *et al.* 1989; Stern & Bloomer 1992; Clift & Dixon 1998; Wallin & Metcalf 1998) and the inflected P–T–t histories of sub-ophiolitic metamorphic 'soles' interpreted from thermobarometric studies (Wakabayashi & Dilek 1988, 2000; Gjata *et al.* 1992; Insergueix-Filippi *et al.* 1998, 2000; Searle & Cox 1999; Bebien *et al.* 2000; Dimo-Lahitte *et al.* 2001).

Boninite–HMA volcanism is relatively rare in modern settings. A notable exception, however, is the Hunter Ridge, between the southernmost New Hebrides and Fiji islands, where boninite magmatism marks a locus of incipient subduction along an active transform fracture zone (Falloon & Crawford 1991; Danyushevsky *et al.*, 1995; Crawford *et al.* 1997). Here, the eastern extremity of the New Hebrides subduction system appears to be nucleating along a transform fault that is linked to the southward-propagating North Fiji spreading centre (Monzier *et al.* 1993a, 1993b, 1997). Examples of coeval, if now-extinct, boninite and HMA volcanism also characterize forearc–remnant arc pairs in the Mediterranean and western Pacific regions, confirming the unusual thermal character of subduction nucleation events. The best-documented example of subduction nucleation occurs in the Izu–Bonin–Mariana (IBM) 'subduction factory' (MARGINS) where Mid-Eocene subduction nucleation was followed by rapid backarc basin opening and rollback of the IBM arc–forearc terrane (Karig 1971; Hussong *et al.* 1981; Stern & Bloomer 1992; Bloomer *et al.* 1995; Hawkins & Castillo 1998).

The locus of subduction inception is marked by the Palau–Kyushu ridge, a boninite-bearing 'proto-arc' remnant that dissects the Philippine Sea Plate. Subduction probably began at *c.* 50 Ma, shortly before the India–Asia collision (*c.* 45–40 Ma) and reorientation of Pacific Plate motion (*c.* 43 Ma) (Bloomer *et al.* 1995; Hawkins & Castillo 1998) (Fig. 4) with underthrusting along a transform fracture zone of the West Philippine Sea Basin Plate, either by the Pacific Plate or a younger, hypothetical, North New Guinea Plate (Stern & Bloomer 1992). As subduction continued, splitting of the Palau–Kyushu proto-arc led to opening of the Parece Vela Basin (*c.* 40–25 Ma), with concomitant rollback of the newly active West Mariana arc. Splitting of the latter gave way to further rollback of the active Mariana arc with opening of the still-active Mariana Trough (Karig 1971; Hussong *et al.* 1981). Mariana Trough opening continues today accompanied by northward 'unzipping' of the West Mariana–Mariana arc, offering a diachronous, actualistic model for proto-ophiolite genesis.

According to such a model, the present-day Mariana forearc includes the accumulated products of West Philippine Sea sea-floor spreading, 'proto-arc' boninitic and calc-alkaline activity, and subsequent (West Mariana, Mariana) arc volcanism. Boninites dredged from the Palau–Kyushu ridge match those in lower horizons of the Mariana forearc, the latter feature conforming in these and other respects to an *in situ* 'proto-ophiolite' (Ishii *et al.* 1988; Ogawa & Taniguchi 1989). Mid-Miocene HMA volcanic rocks (*c.* 13 Ma) are likewise preserved in central and southern parts of the Ryukyu forearc and are matched by analogous activity in the Fujian–Taiwan region (Shinjo 1999), suggesting subduction nucleation, triggered by collision of the Luzon arc with Eurasia, prior to opening of the Okinawa Trough (Shinjo 1999). Other examples include Mid-Miocene (*c.* 14–15 Ma) HMA vol-

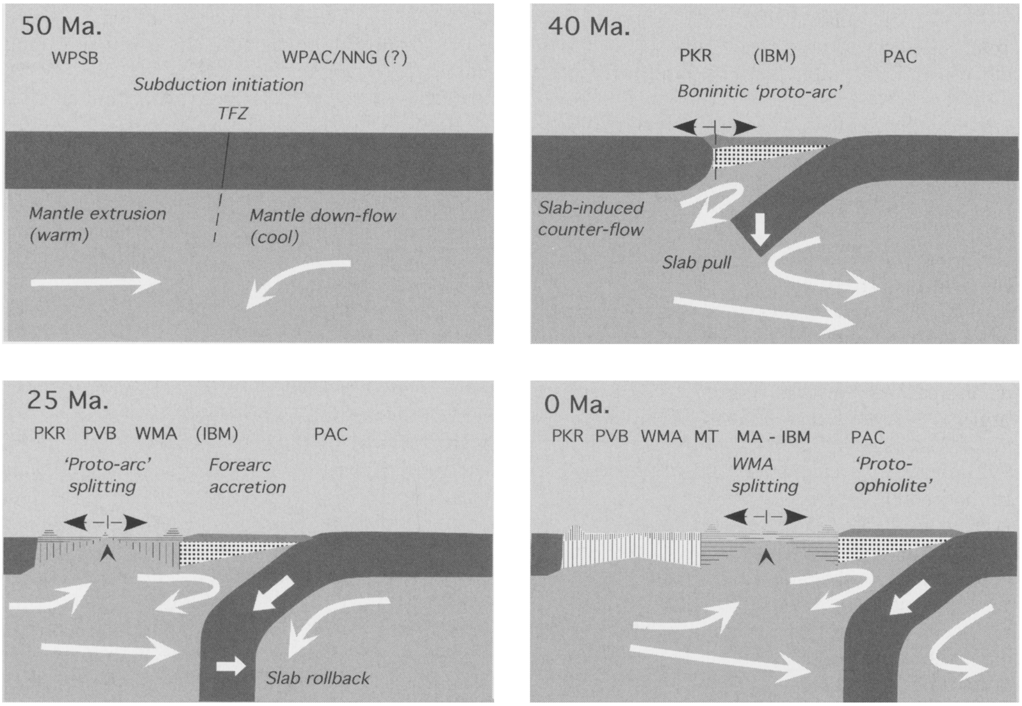

Fig. 4. Arc–trench rollback model, from Bloomer *et al.* (1995), based on evolution of the Izu–Bonin–Mariana forearc and eastern part of the Philippine Sea Plate (following Karig 1971; Hussong *et al.* 1981;Stern & Bloomer 1992). (**a**) 50–40 Ma. Subduction nucleation beneath the West Philippine Sea Basin Plate either by the Pacific Plate or (hypothetical) young North New Guinea Plate along a transform fracture zone in proto- West Philippine Sea Basin spreading centre. Boninite melt genesis accompanies early forearc development with inception of the calc-alkaline arc forming the Palau–Kyushu ridge. (**b**) 40–25 Ma. Continued subduction with slab steepening of the Pacific Plate, splitting of the Palau–Kyushu arc, Parece Vela Basin opening, and rollback of the active West Mariana arc. (**c**) 25–0 Ma. Continued subduction, Mariana Trough opening by splitting of the West Mariana arc, and rollback of the active Mariana arc. (**d**) 0 Ma. Subduction beneath the modern Mariana arc–trench system with continued Mariana Trough opening by 'unzipping' of the West Mariana–Mariana arc to the north (Iwo Jima). PKR, Palau–Kyushu ridge; PVB, Parece Vela Basin; WMA, West Mariana arc; MT, Mariana Trough; MA, Mariana arc; IBM, Izu–Bonin–Mariana forearc; PAC, Pacific Plate; WPSB, West Philippine Sea Basin; WPAC, Western Pacific; NNG, 'North New Guinea' Plate.

canism recorded from islands between the Hellenic forearc and Cycladean metamorphic core complexes (Smith & Spray 1984; Pe-Piper 1994; Forster & Lister 1999; Migiros *et al.* 2000) that corresponds to HMA-bearing ophiolite fragments in the Hellenic forearc (Fortuin *et al.* 1997; Clift 1998), and Oligo-Miocene HMA (*c.* 18 Ma) in Sardinia and Corsica, and in Calabrian forearc ophiolite fragments (Beccaluva 1982; Delaloye *et al.* 1984; Compagnoni *et al.* 1989; Beccaluva *et al.* 1994), which pre-date opening of the Tyrrhenian Sea (Morra *et al.* 1997; Padoa 1999).

Experimental studies of boninite melts suggest that they result from the combined effects of mantle decompression and slab-derived H_2O-rich fluid (Umino & Kushiro 1989; van der Laan *et al.* 1989; Falloon & Danyushevsky 2000), interpreted by some as an indication of ocean ridge spreading conditions (Gjata *et al.* 1992; Stern & Bloomer 1992; Peacock 1994; Peacock *et al.* 1995). However, comparison of experimental data for basalts, boninites, and variably fertile peridotites suggests that three additional conditions are required for boninites to form: (1) anomalous asthenospheric potential temperatures (T_p > *c.* 1400 °C); (2) significant lithospheric extension (stretching factors, β, > *c.* 3) (McKenzie & Bickle 1988; Latin & White 1990); (3) refractory (previously melt-depleted) peridotite sources (e.g. van der Laan *et al.* 1989; Hirose & Kawamoto 1995; Hirose 1997; Falloon & Danyushevsky 2000) (Fig. 5). In other words, boni-

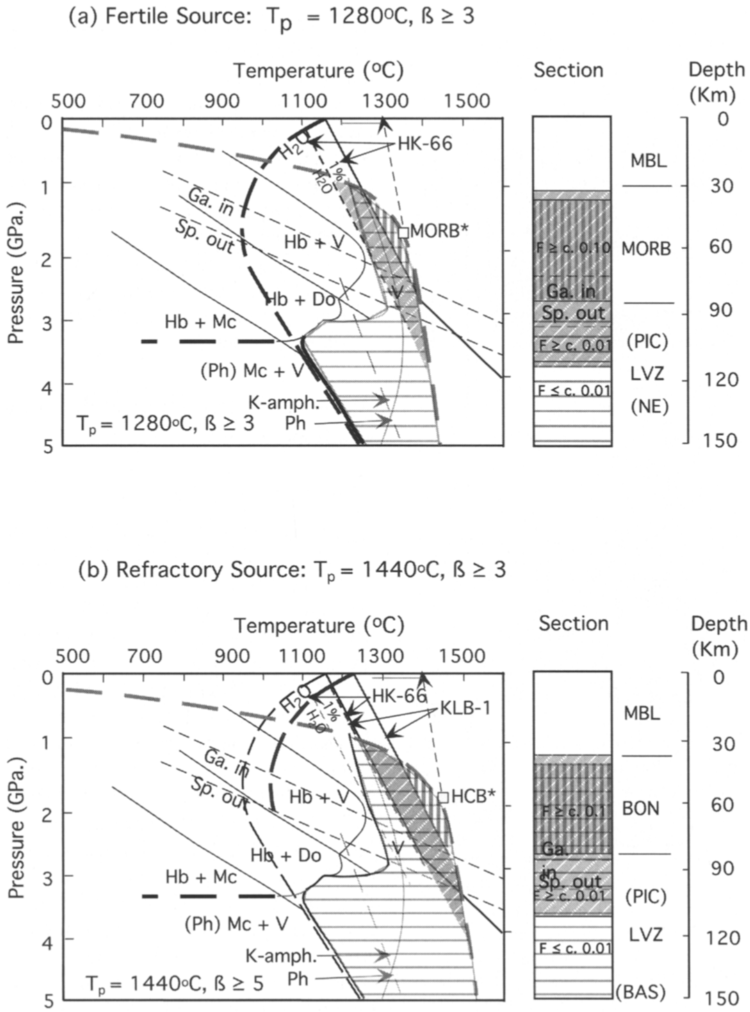

Fig. 5. Partial melting models for (**a**) fertile and (**b**) refractory peridotite in the presence of H_2O and CO_2 between pressures of 0 and 5 GPa, interpolated from published experimental data. Geotherms (bold dashed curves) are calculated for asthenospheric potential temperatures (T_p) of 1280 °C and 1440 °C, respectively, for stretching factors (β) of *c.* >3 and >5, assuming 'pure shear' lithosphere extension (McKenzie & Bickle 1988; Latin & White 1990). The solidus curve (bold line) for 'fertile' peridotite (pyrolite)– C–H–O is taken from Wyllie (1990) for the condition $X = CO_2/(CO_2 + H_2O) = 0.8$ along the bold grey curve marking equilibrium between amphibole, carbonate, peridotite, and vapor; based on sources given by Wyllie *et al.*, (1990). The anhydrous solidus for fertile peridotite (HK-66) is from Hirose & Kushiro (1993), with a hypothetical H_2O-undersaturated (1 wt% H_2O) solidus interpolated from hydrous experiments on refractory lherzolite (KLB-1) (Hirose 1997). Anhydrous, H_2O-saturated, and H_2O-undersaturated (1 wt% H_2O) solidi for the refractory lherzolite (KLB-1) are from Hirose & Kawamoto (1995). Predicted melt segregation conditions agree with those determined experimentally for primitive MORB and high-Ca boninite (HCB) (e.g. van der Laan *et al.* 1989; Falloon & Danyushevsky 2000). MBL, mechanical boundary layer; BON, boninite; LVZ, low-velocity zone; PIC, picrite; NE, nephelenite; BAS, basanite.

nite liquidus temperatures appear to exceed those of 'normal' mid-ocean ridge magmas by *c.* 150–200 °C and require a source that is significantly less fertile than that producing normal MORB (van der Laan *et al.* 1989; Falloon & Danyushevsky 2000).

These observations concur with predictions from 2D numerical models (Gjata *et al.* 1992; Insergueix-Filippi *et al.* 1998, 2000) and are also supported by anomalous P–T–t histories recorded from sub-ophiolitic metamorphic soles (Fig. 6). These appear to record patterns of cold thrusting

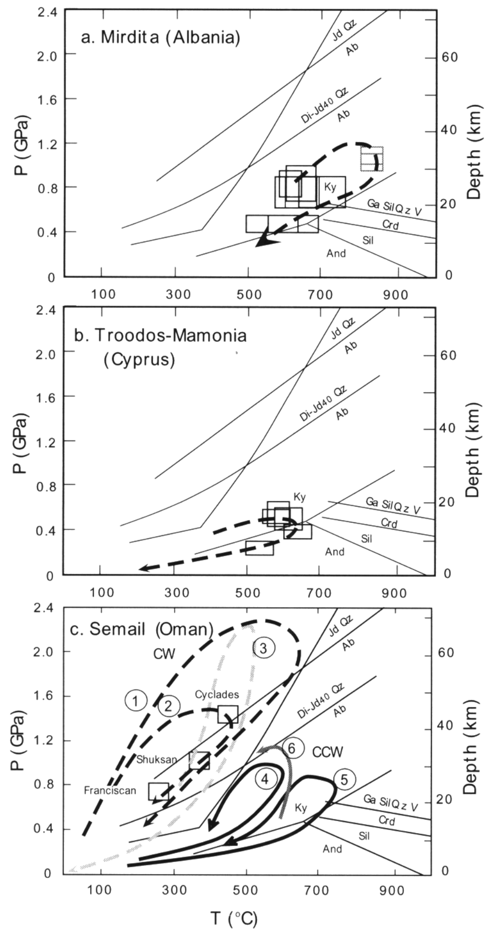

Fig. 6. Interpolated $P-T-t$ histories for sub-ophiolitic metamorphic soles and other metamorphic lithologies. (**a**) Mirdita ophiolite, Albania (Gjata *et al.* 1992; Robertson & Shallo 2000; Dimo-Lahitte *et al.* 2001). (**b**) Troodos–Mamonia complexes, Cyprus (Malpas *et al.* 1992; Bailey *et al.* 2000). (**c**) Semail ophiolite, Oman. 1 and 2, Saih Hatat; 3, Saih Hatat–Wadi Tayin; 4, Asimah–Bani Hamid; 5, Hacker & Gnos (1997); 6, As Sitah eclogites → Hulw blueschist (Searle *et al.* 1994; Searle & Cox 1999). Pressure-temperature equilibration conditions for ophiolitic blueschist–bearing units from the Shuksan, and Franciscan terranes and Cyclades (Aegean) remnant arc are also shown. Clockwise (CW) and counter-clockwise (CCW) $P-T-t$ paths appear to distinguish 'high temperature' from 'high pressure' sole types.

to at least *c.* 60 km depth with temperatures rising to well over 700 °C, a stage of near-isobaric cooling, and (eventual) exhumation (Wakabayashi & Dilek 1988, 2000; Hacker 1991; Gjata *et al.* 1992; Beccaluva *et al.* 1994; Searle *et al.* 1994; Wakabayashi & Unruh 1995; Dilek & Whitney 1997; Dilek *et al.* 1997; Hacker & Gnos 1997; Searle & Cox 1999; Dimo-Lahitte *et al.* 2001), suggesting that metamorphic 'soles' may be best interpreted as MORB-like relics of incipient subduction (e.g. Fig. 6; Wababayashi & Dilek, this volume). Although some ophiolites slightly post-date their high-temperature soles, most are coeval or slightly older, consistent with interpretations that sundered forearc and remnant 'proto-arc' components were single entities prior to splitting and the onset of rollback. Incipient subduction relics are thus preserved *a priori* in forearc 'proto-ophiolites' (Ishii 1989; Ogawa 1995).

Arc–trench rollback: endogenous vs. exogenous causes?

Plate kinematic effects

Although reliable age data for ophiolites are sometimes difficult to acquire and have proved contradictory, a clearer spatial–temporal picture is emerging of relations between active Tethyan forearc complexes and their respective remnant proto-arcs. For example, boninite-bearing 'proto-arcs' (*c.* 49–45 Ma), preserved in the IBM forearc and its remnants (Hawkins & Castillo 1998) and the Zambales (Philippines) ophiolite (Encarnacion 1997), are nearly coeval with the initiation of Celebes Sea opening (*c.* 48 Ma) (Beiersdorf *et al.* 1997), whereas they slightly predate the 'hard' collision of India with Eurasia (*c.* 45–40 Ma) and the corresponding change in Pacific Plate motion from NNW to NW. The Palawan and Mindoro ophiolites (*c.* 34 Ma) also appear to mark a change from sea-floor spreading to convergence, coeval with the inception of Alao Shan–Red River left-lateral shearing (*c.* 33 Ma) and sea-floor spreading in the South China Sea (*c.* 32 Ma). Both were terminated by a collision between the Sulu Ridge arc remnant and the West Philippine arc (*c.* 17 Ma) (Rangin *et al.* 1995), which triggered the initiation of Sulu Sea opening (*c.* 17 Ma) (Rangin *et al.* 1995; Yumul *et al.* 1998, 2001). New subduction that produced the Ryukyu proto-arc probably occurred between *c.* 21 and 18 Ma, preceding collisions of Taiwan (*c.* 15–12 Ma) (Chung *et al.* 1994), and other micro-continents with the Luzon arc, at *c.* 17–16 Ma (Rangin *et al.* 1985; Pubellier & Cobbold 1996; Pubellier *et al.* 1996) and 12–6 Ma (Sibuet & Hsü 1997).

In western Tethys, the Late Cretaceous Semail and Troodos ophiolites (*c.* 98–75 Ma) (Urquhart & Banner 1994; Hacker & Mosenfelder 1996; Hacker *et al.* 1996) and those exposed in the Zagros and Tauride belts (*c.* 98–75 Ma) (Dilek *et al.* 1999; Parlak & Delaloye 1999; Ghazi & Hassanipak 2000; Parlak *et al.* 2000; Babaie *et al.*

2001a, 2001b; Ghasemi *et al.* 2002) and Hellenic arc (*c.* 98–75 Ma) (Langosch *et al.* 1999, 2000), were formed shortly before collisions of the Iranian (*c.* 75–70 Ma), Apulian (*c.* 70–65 Ma), Pelagonian (*c.* 72 Ma), and other micro-continents (Robertson & Shallo 2000; Stampfli *et al.* 2001; Stampfli & Borel 2002) with accreting Eurasia. However, younger ophiolitic remnants (*c.* 12–10 Ma) in Crete and the Aegean Cyclades are near-coeval with both Aegean continental collapse (Lee *et al.* 1990) and the initiation of westward escape by Anatolia (*c.* 13–5 Ma) (Le Pichon 1982; Le Pichon *et al.* 1995). Thus, although rollback of Hellenic subduction may have continued since the Paleocene, as inferred from seismic tomography (Spakman *et al.* 1992; Spakman & Bijwaard 1998), it was probably interrupted by the effects of regional microplate collisions in the Late Cretaceous and Pliocene (Le Pichon *et al.* 1995).

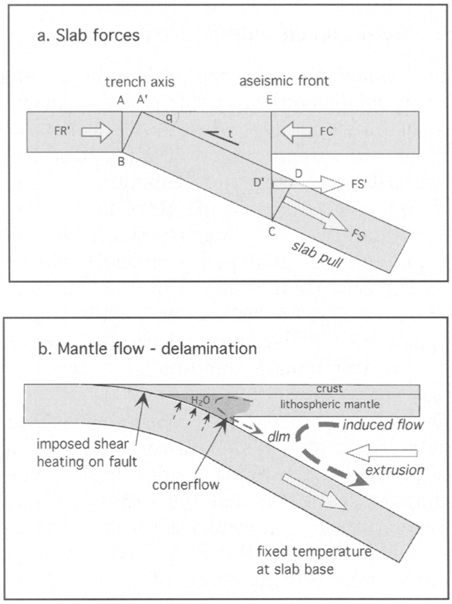

Fig. 7. Hypothetical effects of 'slab pull' and 'mantle extrusion'. (**a**) Slab force model, showing interaction between subducting slab and overriding plate (from Seno & Yamanaka 1998). Where backarc basin opening is a passive response to slab compression, slabs show 'down-dip compression'. However, down-dip tension accompanies backarc opening in some cases (e.g. the Mariana, Kyushu and Hellenic arcs) (Seno & Yamanaka 1998), suggesting slab retreat is driven by trenchward mantle flow. The dip of the subducting plate is *q*, AB is the trench axis and CE the aseismic front. PS′ is the effective ridge push, FS the slab pull, and PC the collision force. FS′ is the horizontal component of the traction on CD′. *t* is the shear stress at the thrust zone (see Seno & Yamanaka 1998). Where rollback is slow, FS′ is negative, in which case F_c could be positive or negative depending on whether P_s' is smaller or larger than FS′. If FS′ is negative and opposed to FP′, FC is likely to be very small, resulting in back-arc extension. (**b**) Two-dimensional mantle flow model (after van Keken *et al.* 2002); the slab is assumed to be subducting at constant speed. Two flow components are shown: (1) 'endogenous' (slab-induced) flow; (2) exogeneous (e.g. collision-induced) flow.

'Slab pull' and extrusion tectonics

Encarnacion *et al.* (2001) proposed a linkage between genesis of the South Palawan (Philippines) ophiolite and the coupled inception of South China Sea spreading and left-lateral motion on the Red River fault (Lee *et al.* 2000; Wang *et al.* 2000). According to the classic extrusion tectonics model (Tapponnier *et al.* 1982, 1986), marginal basin opening and arc–trench rollback are linked responses to collision-induced lithosphere 'escape', as interpreted for opening of the Aegean and South China Sea basins and seaward escape, respectively, of Anatolia and Indochina (Briais *et al.* 1993; Le Pichon *et al.* 1995; Lundgren *et al.* 1996). However, as a general explanation of marginal basin opening, extrusion tectonics seems unable, at least in these cases, to account for the observation that basin opening commenced prior to, and proceeded at a faster rate than, the escape of their respective conjugate blocks (Chung *et al.* 1997; Lee *et al.* 2000; Wang *et al.* 2000; Le Pichon *et al.* 2002).

Given the apparent linkage of continental escape and marginal basin opening, we need to consider the extent to which backarc basin opening is intrinsic to subduction and what, if any, exogenous factors play a role. As already noted, arc–trench rollback has usually been ascribed to slab buoyancy forces, assuming these to exceed those of the convecting asthenosphere (Isacks & Molnar 1971; Uyeda & Kanamori 1979). Most studies of the mechanical interactions between subducting and overriding plates suggest that where backarc basin opening is passive, the seismicity associated with subduction shows 'down-dip compression' (e.g. Fig. 7a). If horizontal traction is negative, backarc extension will be relatively modest (Seno & Yamanaka 1998). In some cases (e.g. the Mariana, Kyushu and Hellenic arcs), however, down-dip tension accompanies backarc opening, suggesting that slab retreat is more rapid, and driven by trenchward mantle flow (Seno & Yamanaka 1998). These relationships are shown in Figure 7b, indicating the possibility of two potential flow field components: 'endogenous' (slab-induced) and 'exogeneous' (mantle-driven).

Collision-induced mantle extrusion

Thus although slab pull can explain the dynamics of many subduction zones, it is probably unable to account for those cases, such as the Hellenic and Mariana arc systems (Seno & Yamanaka 1998), where slab steepening, arc bending, and accelerated basin opening coincide (McCabe & Uyeda 1983; Hynes & Mott 1985; Dvorkin *et al.* 1993; Bevis *et al.* 1995). Slab pull is probably subsidiary to exogenous mantle flow (Seno & Yamanaka 1998), a mechanism that can potentially reconcile marginal basin and continental escape kinematics with the accretionary build-up of forearc complexes.

During early stages of typical Wilson Cycles, plate motions are probably driven by a combination of mantle upwelling ('ridge push'), downwelling ('slab pull'), and the effects of lateral impingement (on continental cratonic keels) (Forsyth & Uyeda 1975; Russo & Silver 1996). For example, the correspondence of arc–trench rollback in the Caribbean and South Scotia Sea regions (Dalziel *et al.* 2001) to accelerated westward motion of South America (Russo *et al.* 1993; Russo & Silver 1996) may be a far-field mantle flow response to the 30–25 Ma Africa–Eurasia collision (Silver & Russo 1996). In response to the latter, westward migration of the Mid-Atlantic Ridge and the corresponding eastward offset of major Mid-Atlantic hotspots (Iceland, St. Helena, Tristan da Cunha, the Azores, and Bouvet) would have been immediately translated to South American plate motion. At later stages, asthenosphere flow is likely to be displaced by thick continental plates as they approach each other and, eventually, collide. For example, Tamaki (1995) suggested that lateral displacement of asthenosphere prior to and following the India–Asia collision led to rapid eastward propagation of Western Pacific marginal basins. Such a process, broadly consistent with the timing and kinematics of basin opening (Hall 2002), also offers a plausible explanation for widespread intra-plate volcanism that characterizes much of east and SE Asia, contamination of the upper mantle beneath eastern Eurasia and western Pacific basins (attributed to delamination of the Sino-Korean craton), and the sharp boundary separating DUPAL-like (contaminated) from N-MORB Pacific mantle.

Accordingly, if HTM ('high-tide mark') forearc assemblages (the accreted igneous and metamorphic products of arc–trench rollback) are a valid analogue for ophiolite, such features can be taken to represent distal mantle flow boundaries (Flower *et al.* 1998, 2001; Flower 2003). This is not to say that arc–trench rollback is exclusively triggered by lateral mantle flow produced by plate collisions. As already noted, mantle flow fields giving rise to arc–trench rollback and proto-ophiolite genesis may be contingent on other modes of differential plate motion. However, the notion of collision-induced mantle extrusion as the driver of Tethyan ophiolite genesis appears able to reconcile coeval continental escape (Armijo *et al.* 1989; Jolivet *et al.* 1991a, 1991b), post-collision lithosphere stretching (England & Molnar 1997a, 1997b; Ren *et al.* 2002), and arc–trench rollback (Hussong *et al.* 1981; Tamaki & Honza 1991) (Fig. 7), along with regional mantle attributes cited by Flower *et al.* (1998, 2001) and Flower (2003).

From proto-ophiolite to ophiolite: a preliminary Tethyan verdict

Studies of metamorphic soles record an unambiguous pattern of cold thrusting to at least *c.* 60 km depth, temperatures rising to well over 700 °C, a stage of near-isobaric cooling, and (eventual) exhumation (Wakabayashi & Dilek 1988, 2000, this volume; Hacker 1991; Gjata *et al.* 1992; Beccaluva *et al.* 1994; Searle *et al.* 1994; Wakabayashi & Unruh 1995; Dilek & Whitney 1997; Dilek *et al.* 1997; Hacker & Gnos 1997; Searle & Cox 1999; Dimo-Lahitte *et al.* 2001) (e.g. Fig. 5). The only serious alternative to subduction nucleation as a trigger for ophiolite formation is the proposal that ophiolites result from the consumption of recently active spreading centres at pre-existing subduction zones (e.g. Hacker & Mosenfelder 1996; Hacker *et al.* 1996). This rests on the assumption that newly formed (<10 Ma) oceanic lithosphere is too buoyant to be subducted (Cloos *et al.* 1998) and that high-temperature metamorphic soles represent mid-ocean ridge rather than 'normal' subduction conditions (Hacker & Mosenfelder 1996; Hacker *et al.* 1996). Although the two interpretations are not mutually exclusive, the question of which process dominates appears to hinge on: (1) the correspondence or otherwise of metamorphic sole and ophiolite compositions; (2) validity of the ophiolite–forearc analogue; (3) possible spatial–temporal correlations between ophiolites and regional plate tectonics. In general, 'ridge subduction' is the less appealing, given its absence from recent or active rollback cycles, whereas 'subduction nucleation', on the other hand, appears to be the rule rather than the exception (Dewey & Casey 1979; Claesson *et al.* 1984; Gjata *et al.* 1992; Stern *et al.* 1992; Monzier *et al.* 1993a; Beccaluva *et al.* 1994; Bloomer *et al.* 1995; Crawford *et al.* 1997; Clift & Dixon 1998; Insergueix-Filippi *et al.* 1998, 2000; Wakabayashi & Dilek, this volume).

Although some ophiolites slightly post-date their high-temperature soles, most are coeval or slightly older, consistent with interpretations that sundered forearc and remnant 'proto-arc' components were single entities prior to splitting and the onset of rollback. Incipient subduction relics are thus preserved *a priori* in forearc 'proto-ophiolites' (Ishii 1989; Ogawa 1995). This model is supported by thermochronological data for metamorphic soles from the Mirdita (Albania), Troodos (Cyprus), Tauride (Turkey), and Semail (Oman) ophiolites (e.g. Searle & Malpas 1980; Gnos 1998; Malpas *et al.* 1992; Searle *et al.* 1994; Hacker & Gnos 1997; Dilek *et al.* 1999; Searle & Cox 1999). The MORB-like compositional character, inflected thermal gradients, and counter-clockwise $P-T-t$ trajectories of many metamorphic soles (Searle & Malpas 1980; Ghent & Stout 1981; Malpas *et al.* 1992; Shallo 1992; Encarnacion & Mukasa 1995; Hacker & Mosenfelder 1996; Hacker *et al.* 1996; Gnos *et al.* 1997; Bebien *et al.* 1998; Searle & Cox 1999; Bebien *et al.* 2000; Dimo-Lahitte *et al.* 2001) are all consistent if these features represent relict slab fragments from a subduction nucleation event (e.g. Fig. 5). The association of boninites with fossil transform faults such as the Arakapas fault zone (Simonian & Gass 1978; Flower & Levine 1987; MacLeod & Murton 1993), and analogous features in the Semail ophiolite (Smewing *et al.* 1977; Boudier *et al.* 1988; MacLeod & Rothery 1992), reinforces the likelihood that 'hot' subduction nucleation is largely confined to near-ridge loci.

The available age data for Neo-Tethyan ophiolites allow for three broad conclusions that, we contend, are consistent with exogenous mantle flow as the driving force in their genesis. First, many, if not most, Tethyan ophiolites were emplaced after relatively short time intervals following subduction nucleation at oceanic spreading axes (Hacker & Mosenfelder 1996; Hacker *et al.* 1996; Dilek *et al.* 1999). This strongly suggests that 'oceanic' lithosphere consumed during processes of ophiolite development (although not necessarily their inception) was formed in marginal basin rather than major ocean basin settings (Pearce *et al.* 1984). Second, the significance of subduction nucleation, as a precursor to ophiolite genesis, in relation to distal plate tectonic events, needs further study (Gnos *et al.* 1997; Moores *et al.* 2000). Finally, given the near-ubiquitous presence of ophiolites in orogens, their formation is necessarily restricted to late stages of Wilson Cycles, immediately preceding continent–continent plate collisions.

In summary, our model for Tethyan ophiolite genesis is depicted as six hypothetical stages (illustrated in Fig. 8): (1) continental plates separate as Palaeo-Tethys begins opening; passive asthenosphere upwells beneath spreading axis to produce MORB-like oceanic lithosphere (Fig. 8a); (2) Laurasia blocks continued Palaeo-Tethys opening; new subduction is initiated at a weak (e.g. transform) zone, with boninite magmatism forming a 'proto-arc' on the overriding MORB plate, followed by 'normal' calc-alkaline arc volcanism; Neo-Tethyan rifting is initiated in Gondwana (Fig. 8b); (3) the 'Cimmeria' micro-continent detaches from Gondwana as Palaeo-Tethys continues subducting; MORB-like backarc extension and arc–forearc rollback occur in response to the compression of Tethyan asthenosphere beneath Laurasia and Cimmeria (Fig. 8c); (4) continued 'Cimmerian' micro-continent migration leads to oblique collision and diachronous breakoff of the Palaeo-Tethyan ocean slab, accompanied by enhanced continental sediment subduction; deflected asthenosphere flow field leads to shoshonite and potassic granite magmatism derived from continent-contaminated asthenosphere (Fig. 8d); (5) Neo-Tethys continues to open until relict MORB-like backarc lithosphere is completely subducted and detached from the overriding continental plate (Fig. 8e); (6) Neo-Tethys ceases opening as Palaeo-Tethyan basin collapse is completed, and new subduction is initiated; assemblage of relict arc–forearc–backarc (ophiolite) is entrapped in the ensuing continent–continent orogeny (Fig. 8f).

Petrological, structural, and stratigraphic data from Tethyan ophiolites consistently support the role of subduction nucleation and arc–trench rollback in their development and are reviewed by Dilek & Flower (this volume).

Conclusions

(1) A unique combination of features, rare or absent from mid-ocean ridge spreading systems and subduction zones (ancient or modern), characterize Tethyan ophiolites: MORB-like 'oceanic' basement and near-coeval high-temperature metamorphic soles, succeeded by boninitic 'proto-arcs', juxtaposed refractory peridotites, and anomalous high-temperature 'epidosites'.

(2) These features appear to preclude mid-ocean ridge, and 'normal' arc or backarc basin provenance, and are uniquely analogous to 'proto-ophiolites' currently forming in modern forearcs.

(3) Modern forearcs are generated following subduction nucleation (commonly at mid-ocean ridge transforms, signalled by sequences of boninitic 'proto-arc', 'normal' arc, and backarc activity, accompanied by high-temperature hydrothermal flow) and evolve in response to one or more episodes of arc splitting and basin opening produced in response to arc–trench rollback.

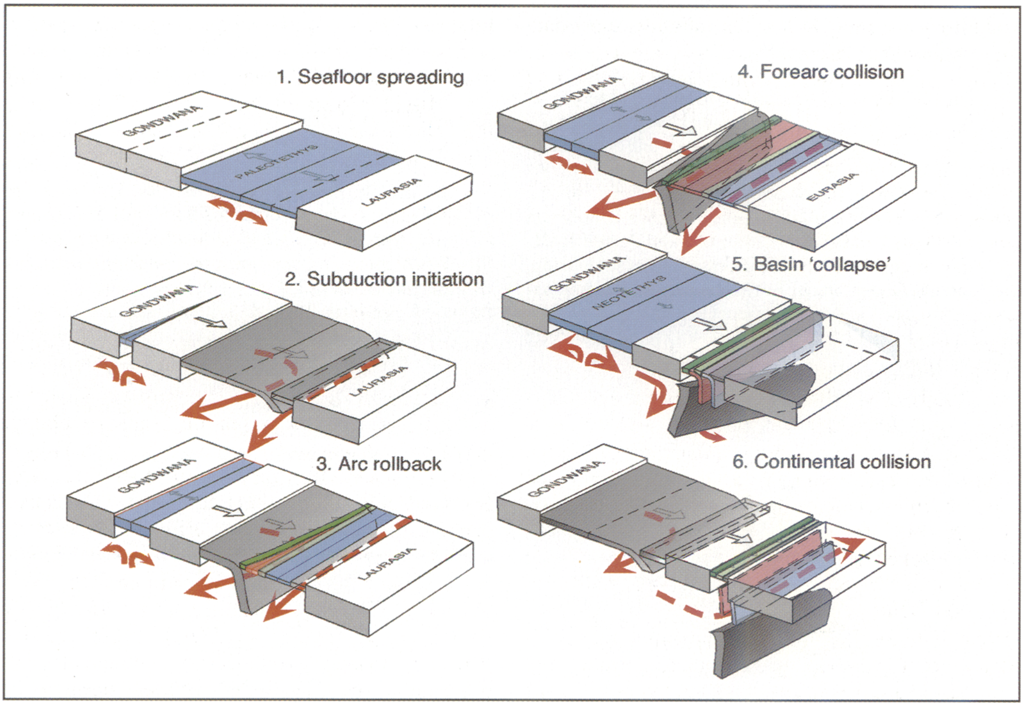

Fig. 8. Mantle-driven Tethyan 'sub-cycle' during late (pre-collision) stages of a Wilson Cycle. (**a**) Continental plates separate as Palaeo-Tethys begins to open; passive asthenosphere upwells beneath spreading axis to produce MORB-like oceanic lithosphere. (**b**) Laurasia blocks continued Palaeo-Tethys opening; new subduction is initiated at weak (e.g. transform) zone, with boninite magmatism forming a 'proto-arc' on the overriding MORB plate, followed by 'normal' calc-alkaline arc volcanism; Neo-Tethyan rifting is initiated in Gondwana. (**c**) 'Cimmerian' microcontinent detaches from Gondwana as Palaeo-Tethys continues subducting; MORB-like backarc extension and arc–forearc rollback occur in response to the compression of Tethyan asthenosphere beneath Laurasia and Cimmeria. (**d**) Continued 'Cimmerian' microcontinent migration leads to oblique collision and diachronous breakoff of the Palaeo-Tethyan ocean slab, accompanied by enhanced continental sediment subduction; deflected asthenosphere flow field leads to shoshonite and potassic granite magmatism derived from continent-contaminated asthenosphere. (**e**) Neo-Tethys continues to open until relict MORB-like backarc lithosphere is completely subducted and progressively detached from the overriding continental plate. (**f**) Neo-Tethys ceases opening as Palaeo-Tethyan basin collapse is completed, and new subduction is initiated; assemblage of relict arc–forearc–backarc (ophiolite) is entrapped in the ensuing continent–continent orogeny; asthenospheric flow contaminated by delaminated or subducted continental crust (not shown) provides sources for shoshonite, lamproite, and kamafugite ($\pm$ carbonatite) magmatism.

(4) Subduction nucleation and arc–trench rollback cycles in the Mediterranean and western Pacific appear to be triggered by the effects of subhorizontal mantle flow resulting either directly from collision-induced asthenosphere extrusion, or indirectly via collision-induced plate kinematic adjustments.

(5) If a rollback episode successfully evades orogenic entrapment, forearc accretion continues indefinitely, as appears to be the case in the western Pacific. On the other hand, if entrapped by a collision, as in the Mediterranean, forearc lithosphere tends to resist subduction (in contrast to backarc basin lithosphere) and is readily preserved as ophiolites.

(6) The apparent correspondence of modern and recent subduction nucleation events to 'hard' plate collisions and their respective plate kinematic responses may be discerned from Tethyan ophiolite age data.

(7) The ophiolite 'conundrum' is a false dichotomy if, as seems to be the case, traditional 'ocean ridge' and 'supra-subduction' models are 'correct' only when considered together in a unified arc–trench rollback model.

(8) Refractory mantle sources of the type yielding calc-alkaline and boninitic magmas at newly forming subduction systems are almost certainly not generic to mid-ocean ridges. Moreover, near-ubiquitous ophiolite components such as boninite

and HMA are exclusive to the initial 'hot subduction' stages of a rollback cycle.

We thank H. Furnes, J. Deng, V. Mocanu, E. Moores, P. Robinson, R. Russo, P. Thy, J. Wakabayashi and G. Zakariadze for discussions of the ideas presented here, and J. Encarnacion, P. Robinson and an anonymous referee for thorough and constructive reviews of the paper. Support from the US National Science Foundation (EAR- 0238416 and INT-0129492 to M.F.J.F.; and EAR-9796011 to Y.D.), UNESCO Earth Sciences Division, the National Geographical Society (to Y.D.), the International Union of Geological Sciences, and NATO (CRG-970263 and EST.CLG-97617 to Y.D.) is gratefully acknowledged.

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