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Initiation of Subduction Zones as a Consequence of Lateral Compositional Buoyancy Contrast within the Lithosphere: a Petrological Perspective

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Tonga and Mariana fore-arc peridotites, inferred to represent their respective sub-arc mantle lithospheres, are compositionally highly depleted (low Fe/Mg) and thus physically buoyant relative to abyssal peridotites representing normal oceanic lithosphere (high Fe/Mg) formed at ocean ridges. The observation that the depletion of these fore-arc lithospheres is unrelated to, and pre-dates, the inception of present-day western Pacific subduction zones demonstrates the pre-existence of compositional buoyancy contrast at the sites of these subduction zones. These observations allow us to suggest that lateral compositional buoyancy contrast within the oceanic lithosphere creates the favoured and necessary condition for subduction initiation. Edges of buoyant oceanic plateaux, for example, mark a compositional buoyancy contrast within the oceanic lithosphere. These edges under deviatoric compression (e.g. ridge push) could develop reverse faults with combined forces in excess of the oceanic lithosphere strength, allowing the dense normal oceanic lithosphere to sink into the asthenosphere beneath the buoyant overriding oceanic plateaux, i.e. the initiation of subduction zones. We term this concept the 'oceanic plateau model'. This model explains many other observations and offers testable hypotheses on important geodynamic problems on a global scale. These include (1) the origin of the 43 Ma bend along the Hawaii–Emperor Seamount Chain in the Pacific, (2) mechanisms of ophiolite emplacement, (3) continental accretion, etc. Subduction initiation is not unique to oceanic plateaux, but the plateau model well illustrates the importance of the compositional buoyancy contrast within the lithosphere for subduction initiation. Most portions of passive continental margins, such as in the Atlantic where large compositional buoyancy contrast exists, are the loci of future subduction zones.

KEY WORDS: *subduction initiation; compositional buoyancy contrast; oceanic lithosphere; plate tectonics; mantle plumes; hotspots; oceanic plateaux; passive continental margins; continental accretion; mantle peridotites; ophiolites*

INTRODUCTION

The advent of plate tectonic theory over 30 years ago has revolutionized Earth science thinking, and provided a solid framework for understanding how the Earth works. Many forces may contribute to plate motions (e.g. Forsyth & Uyeda, 1975; Turcotte & Schubert, 1982; Cox & Hart, 1986), but pull by the subducting slab as a result of its negative thermal buoyancy, further enhanced by changes to denser minerals with depth, is widely accepted as the major driving force for plate motion and plate tectonics (e.g. Forsyth & Uyeda, 1975; Turcotte & Schubert, 1982; Cox & Hart, 1986; Davies & Richards, 1992; Stein & Stein, 1996; Davies, 1998; Richards *et al.*, 2000). It follows that there would be no plate tectonics if there were no subduction zones. Yet how a subduction zone begins remains poorly understood (e.g. Vlaar & Wortel, 1976; McKenzie, 1977; Cloetingh *et al.*, 1982; Hynes, 1982; Karig, 1982; Casey & Dewey, 1984; Ellis, 1988; Mueller & Phillips, 1991; Erickson, 1993; Kemp & Stevenson, 1996; Toth & Gurnis, 1998). In this paper, we develop a new concept that the initiation of subduction zones is a consequence of lateral

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compositional buoyancy contrast within the lithosphere. This concept differs from previous models in the literature, but is consistent with observations, and makes testable predictions on important geodynamic problems.

THE NEW CONCEPT AND A HISTORICAL PERSPECTIVE

The principal vector of motion of a subducting slab is vertical, hence the driving force of the slab is gravitational attraction, i.e. its negative buoyancy with respect to the asthenosphere. Therefore, it would be physically optimal if one part of the lithosphere experienced a greater gravitational attraction than its adjacent neighbour before or during the initiation of a subduction zone. This requires the pre-existence of a density contrast within the lithosphere. If the lithosphere in question is thermally uniform (e.g. for lithosphere >80 Myr old), the density contrast must result from a compositional contrast. The density difference as a result of compositional difference in different parts of the crust or lithosphere as an observation, a concept, and a tectonic force has long been recognized, from Wegener's continental drift (Wegener, 1912) to Holmes' mantle convection (Holmes, 1945), to Hess' continental 'freeboard' (Hess, 1962), and to more recent concepts of lithological stratification (O'Hara, 1973; Oxburgh & Parmentier, 1977; Jordan, 1988) and lithological 'icebergs' (Abbott *et al*., 1997; Herzberg, 1999). However, the potential effect of compositional buoyancy variation on subduction initiation has been largely overlooked. The statement 'since buoyancy forces are produced by *temperature differences* . . . ' (McKenzie, 1977) ruled out the effect of compositional buoyancy contrast as a force for subduction initiation. Models that suggest transforms or fracture zones (e.g. Casey & Dewey, 1984) and spreading ridges (e.g. Casey & Dewey, 1984) as sites of subduction initiation have difficulties unless a significant buoyancy contrast exists across these features. Significant compositional buoyancy contrast across a ridge is unlikely. Although thermal buoyancy contrast across a ridge is possible as a result of asymmetric spreading (e.g. Stein *et al*., 1997; Forsyth *et al*., 1998), this effect becomes negligible when the lithosphere is sufficiently old and cold under conditions more 'optimal' for subduction initiation. A small thermal contrast across a transform is possible in young lithosphere, but the effect on density difference is minimal because of the small thermal expansion ($\sim 3 \times 10^{-5}\,K^{-1}$). The Romanche Transform in the equatorial Atlantic may be the only transform on Earth across which the ridge encounters old lithosphere of up to ~75 Ma. There a large thermal buoyancy contrast must exist, but the lack of transform-perpendicular compression (see below) does not allow the development of reverse faults—the precursors of subduction zones. Therefore, ridges, transforms and fracture zones are unlikely to be active sites of subduction initiation although they could be reactivated subsequently as zones of weakness (e.g. Casey & Dewey, 1984; Clift & Dixon, 1998; Toth & Gurnis, 1998). Although this postulated reactivation is attractive, materials on both sides of a ridge or transform should be broadly similar as they are produced by similar processes in similar environments. Hence, it is unlikely that compositional buoyancy contrast across these weak zones would develop throughout their evolutionary histories. Therefore, it is physically implausible why one side of an old ridge, or transform or fracture zone prefers to sink while the other side chooses to rise under any deviatoric stresses.

We discuss below the concept that compositional buoyancy contrast within the lithosphere creates the favoured and necessary condition for the initiation of subduction zones. This applies not only to intra-oceanic subduction initiation but also to subduction initiation along passive continental margins (Dewey & Bird, 1970; Burke *et al*., 1976) despite the difficulties inherent in modelling the latter (Cloetingh *et al*., 1982; Erickson, 1983; Mueller & Phillips, 1991; Kemp & Stevenson, 1996) and the rarity of present-day examples. The initiation and southward propagation of the Ryukyu subduction zone, developed on the Chinese continental margin since the Miocene, is perhaps an excellent modern example for subduction initiation at passive continental margins (Lallemand *et al*., 2001).

OBSERVATIONS THAT LEAD TO THE NEW CONCEPT

The logical question is why subduction begins where it does. The lithospheric material differences on both sides of an active subduction zone may provide a clue to this question. Subduction of the Nazca Plate beneath the Andes is illustrative—dense oceanic lithosphere subducts into the asthenosphere beneath less dense continental lithosphere, where compositional buoyancy contrast across the trench is indisputable. Such compositional buoyancy contrast must pre-date the onset of the subduction because the bulk of the South American continent is Precambrian in age, whereas the oldest, already subducted Nazca (or Farallon) Plate portion is probably <300 Myr old. Likewise, a compositional buoyancy contrast also exists and pre-dates the inception of intra-oceanic subduction zones in the Tonga and Mariana arcs in the western Pacific (Fig. 1)—two of the important intra-oceanic subduction zones on Earth.

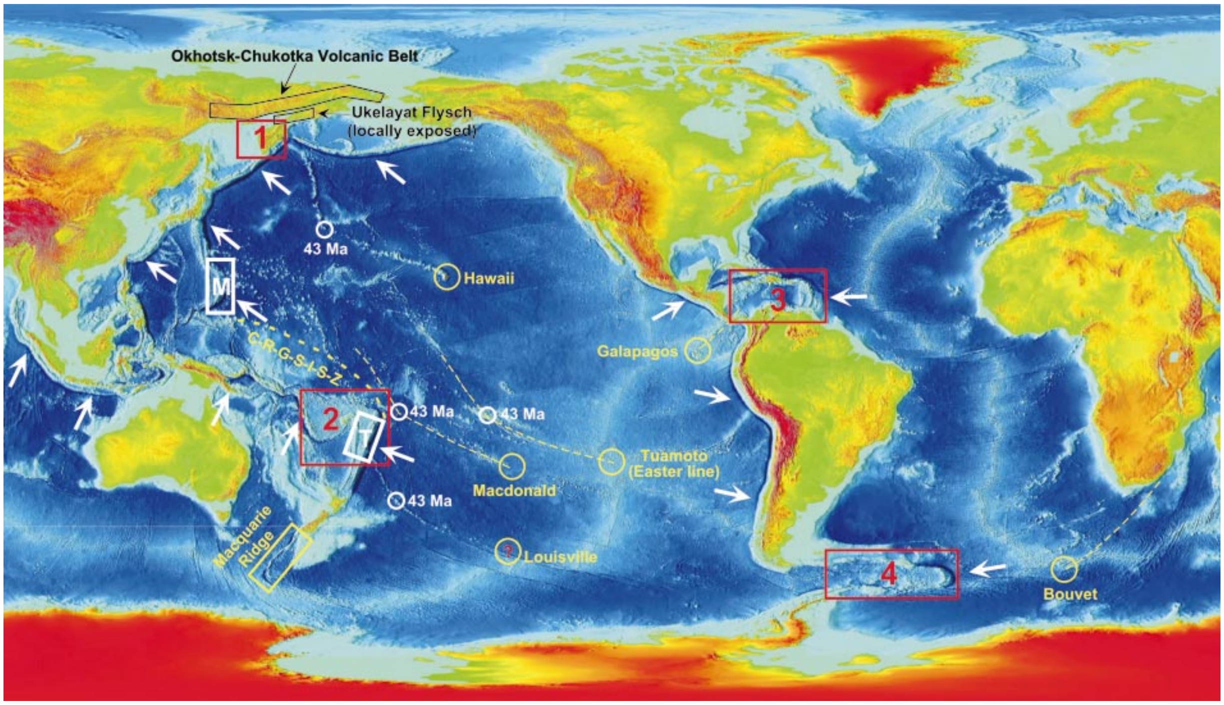

Fig. 1. Portion of world map (from http://www.ngdc.noaa.gov/mgg/image/) showing (1) Mariana and Tonga (white rectangles 'M' and 'T') island arc–subduction systems in the western Pacific from which the fore-arc peridotites studied here were sampled; (2) red rectangles numbered 1–4 are hypothesized or known oceanic plateaux representing mantle plume heads: 1, Kamchatka–Okhotsk Sea, Hawaiian plume head; 2, Tonga–Fiji plateau, Louisville plume head; 3, Caribbean–Colombian plateau, Galapagos plume head; 4, Scotia Plate, Bouvet plume head. White arrows are vectors of motion of oceanic plates that are subducting into the nearby subduction zones. Highlighted are locations of some present-day hotspots, and the 43 Ma bends along hotspot tracks on the Pacific Plate. The thick yellow-dashed line labelled C-R-G-S-I-S-Z is the zone of unusual intraplate seismic activity extending from Samoa through the Gilbert Islands and the Ralik fracture zone to the Caroline Islands (Okal *et al.*, 1986). The Macquarie Ridge is labelled with a yellow rectangle. The Okhotsk–Chukotka continental arc and the Ukelayat Flysch (Garver *et al.*, 2000) are indicated with black outlines.

Figure 2 compares the mineral compositions of Mariana and Tonga fore-arc peridotites with those of abyssal peridotites formed at mid-ocean ridges. Peridotite melting studies (e.g. Jacques & Green, 1980) show that Cr/(Cr + Al) in residual spinel and $Mg/(Mg + Fe^{2+})$ in residual olivine, orthopyroxene and clinopyroxene increase whereas Al_2O_3 contents in these minerals decrease with increasing extents of melting (Fig. 2). Abyssal peridotites are residues of mantle melting that creates the ocean crust at ocean ridges (Dick *et al.*, 1984; Niu *et al.*, 1997). The extent of melting and depletion increases with increasing plate spreading rate (Niu & Hékinian, 1997) and increases when ridges approach hotspots (Dick *et al.*, 1984; Niu *et al.*, 1997). Figure 2 demonstrates that the Mariana and Tonga fore-arc peridotites are on average much more depleted than abyssal peridotites, pointing to greater extents of melting and depletion. The greater extent of melting requires either a hot mantle (e.g. mantle hotspots) or wet mantle (e.g. hydrous mantle above subduction zones). In either case, the key point, illustrated in Table 1, is that melting residues are progressively less dense with increasing extent of melting and depletion because of decreasing Fe/Mg ratios in the whole rock and constituent minerals (e.g. O'Hara, 1973; Jordan, 1988; Niu & Batiza, 1991; Herzberg, 1999); that is, the Mariana and Tonga fore-arc peridotites are less dense than abyssal peridotites. It should be noted also that although Al_2O_3 is a minor constituent in depleted melting residues, its concentration determines the amount of dense garnet phase that may form at high pressures. Therefore, the more depleted (low Fe/Mg and low Al_2O_3) residues (e.g. fore-arc peridotites) will form less garnet, and thus are even more buoyant than less depleted residues (e.g. abyssal peridotites) at higher pressures (or greater depths).

The great depletion of fore-arc peridotites has been widely interpreted as resulting from slab-dehydration-induced high-degree melting of mantle wedge peridotites to provide arc magmatism (e.g. Dick & Bullen, 1984; Bonatti & Michael, 1989). This interpretation, which assumes that fore-arc peridotites are residues of such mantle wedge melting, has never been validated (Niu *et al.*, 2001*a*, 2001*b*). On the contrary, several lines of evidence indicate otherwise: (1) the accepted depth

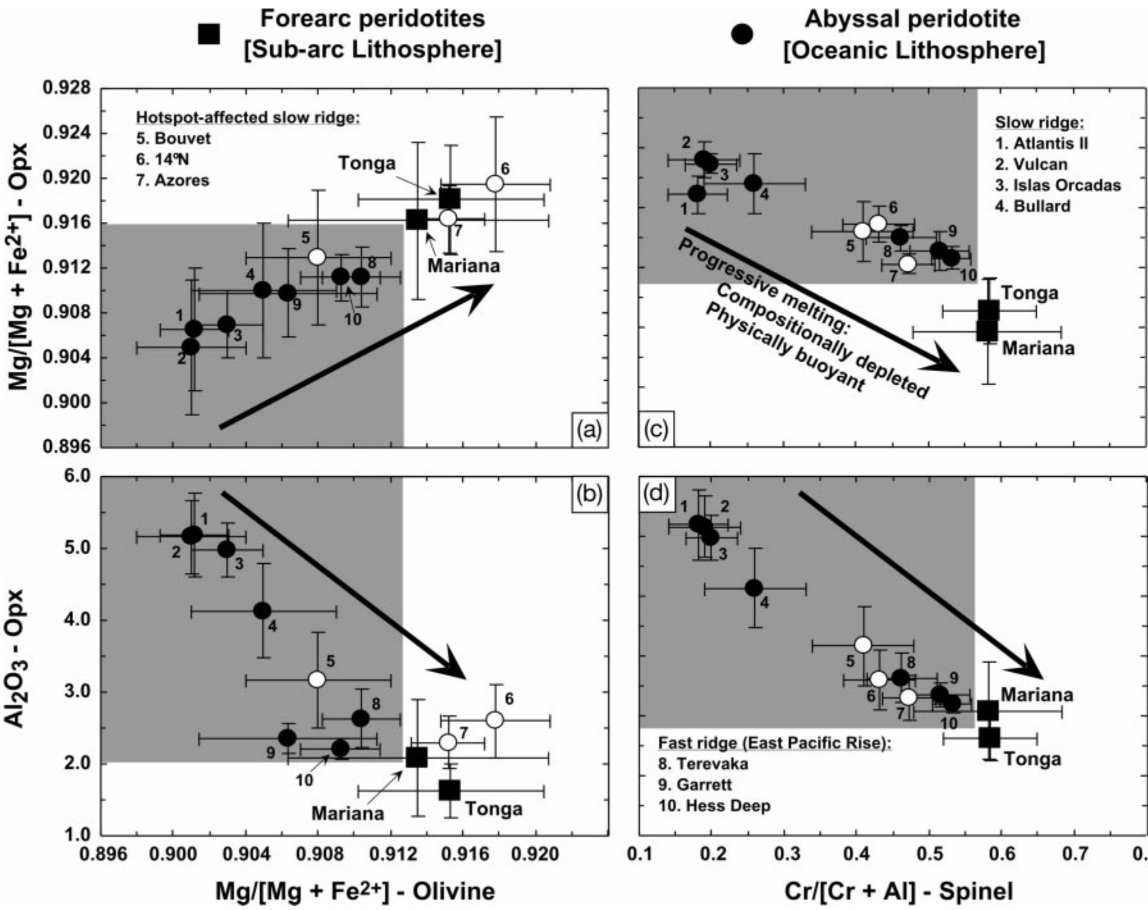

Fig. 2. Comparison of Tonga and Mariana fore-arc peridotites (squares) with abyssal peridotites (circles; Niu & Hékinian, 1997) in mineral compositional spaces (a)–(d). Black arrows point to the direction of increasing extent of melting. With increasing extent of melting, residual peridotites become more depleted (lower Fe/Mg and Al_2O_3), and thus physically more buoyant. It should be noted that fore-arc peridotites are on average far more depleted, thus physically more buoyant, than abyssal peridotites, which is particularly clear in (c) and (d). Also, although Al_2O_3 is a minor constituent in depleted melting residues, its concentration determines the amount of dense garnet phase that may form at high pressures. Therefore, at higher pressures, the more depleted residues (lower Al_2O_3) are even more buoyant than less depleted residues. Open circles are samples from slow-spreading ridges affected by hotspots/wetspots (they point to 5, 6 and 7 in the inset of the upper left panel).

range (~70–150 km) of mantle wedge melting and the predominantly subducting-slab-induced flow field in the mantle wedge (e.g. Davies & Stevenson, 1992) make the exhumation of arc melting residues to the fore-arc level improbable because the melting residues move downward induced by the down-going slab, not upward against mantle flow (e.g. England, 2001). (2) An Os-isotope study (Parkinson *et al.*, 1998) has revealed that some Mariana fore-arc peridotites are ancient (820–1230 Ma), suggesting that these rocks originate from an ancient melting and depletion episode. (3) The primitive Tonga fore-arc cumulates overlying the peridotites are dominated by troctolites (Niu *et al.*, 2001*a*), which crystallized from dry melts. If the parental magmas were wet arc melts, the primitive cumulates would be wehrlites—olivine–clinopyroxene rocks (Gaetani *et al.*, 1993). Therefore, Mariana and Tonga fore-arc peridotites are not residues of modern mantle wedge melting, but represent their respective sub-arc lithospheres whose depletion pre-dates the inception of present-day subduction zones in the western Pacific some 50 Myr ago (Moberly, 1972; Taylor, 1993). In other words, before the initiation of Mariana and Tonga subduction zones, a compositional buoyancy contrast within the lithosphere already existed, which satisfies the favoured and necessary condition for subduction initiation identified above.

It follows that if the fore-arc peridotites and abyssal peridotites represent sub-arc and oceanic lithospheres respectively, then subduction of the dense oceanic lithosphere beneath the less dense sub-arc lithosphere is physically straightforward. It should be noted that the calculated density contrast of ~0·7% is equivalent to a temperature difference of ~230°C for a commonly accepted thermal expansion coefficient of $\alpha = 3 \times 10^{-5}\,K^{-1}$ for mantle minerals (e.g. Stein & Stein,

Table 1: Differences in average mineral composition and density between abyssal peridotites (oceanic lithosphere) and Tonga and Mariana fore-arc peridotites (sub-arc lithosphere)

	Sub-arc lithosphere Mariana and Tonga fore-arcs Fore-arc peridotites[1]		Oceanic lithosphere Mid-ocean ridges Abyssal peridotites[2]	
Compositionally[3]		*More depleted*	*Less depleted*	
Mg-no.-Ol	[4]$n = 194$	0·915 ± 0·006	> 0·898 ± 0·082	$n = 117$
Mg-no.-Opx	$n = 247$	0·917 ± 0·006	> 0·908 ± 0·006	$n = 154$
Mg-no.-Cpx	$n = 146$	0·929 ± 0·021	> 0·917 ± 0·011	$n = 152$
Cr-no.-Spinel	$n = 204$	0·584 ± 0·084	> 0·307 ± 0·134	$n = 149$
Al_2O_3-Opx (wt. %)	$n = 247$	1·809 ± 0·647	< 4·103 ± 1·161	$n = 154$
Al_2O_3-Cpx (wt. %)	$n = 146$	2·497 ± 1·146	< 5·028 ± 1·493	$n = 152$
Al_2O_3-WR (wt. %)	$n = 31$	0·546 ± 0·453	< 1·611 ± 0·727	$n = 148$
Mg-no.-WR	$n = 31$	0·923 ± 0·007	> 0·902 ± 0·003	$n = 148$
Extent of melting[5]		> 25%		≤20%
Physically[6]		*Less dense*	*More dense*	
ρ(25°C, 1 bar)-Ol	$n = 194$	3·323 ± 0·007	< 3·335 ± 0·007	$n = 117$
ρ(25°C, 1 bar)-Opx	$n = 247$	3·274 ± 0·014	< 3·281 ± 0·004	$n = 154$
ρ(25°C, 1 bar)-Cpx	$n = 146$	3·301 ± 0·011	< 3·308 ± 0·004	$n = 152$
ρ(25°C, 1 bar)-Spinel	$n = 204$	4·316 ± 0·120	> 4·005 ± 0·104	$n = 149$
		Model weight fraction:[7]	Weight fraction:[8]	
ρ(25°C, 1 bar)-WR[9]		3·308 ± 0·008	< 3·330 ± 0·007	$\Delta\rho \sim 0{\cdot}7\%$
		Ol = 0·681	Ol = 0·750	
		Opx = 0·316	Opx = 0·185	
		Cpx = 0·003	Cpx = 0·055	
		Spinel = 0·001	Spinel = 0·010	

[1]Tonga and Mariana fore-arc peridotite samples were provided by R. L. Fisher and J. W. Hawkins of Scripps Institution of Oceanography.
[2]Abyssal peridotite data used are from Dick *et al*. (1984), Niu (1997), Niu & Hékinian (1997) and Niu *et al*. (1997).
[3]Abbreviations: Ol, olivine; Opx, orthopyroxene; Cpx, clinopyroxene; WR, whole-rock analysis. Mineral compositions for fore-arc peridotites were analysed by Y. Niu using a JEOL 8800 Super Probe at the University of Queensland. Whole-rock major elements for the same samples were analysed by Sharon Price by inductively coupled plasma optical emission spectrometry at the Queensland University of Technology.
[4]For fore-arc peridotites, n refers to the total number of analyses of the 31 samples studied plus mineral data published by Parkinson & Pearce (1998) for Mariana; for abyssal peridotites, n refers to individual samples (see footnote 2).
[5]The extent of melting is estimated from Mg-number of the constituent minerals and whole-rock analyses (Niu, 1997).
[6]Densities, expressed as g/cm^3, are calculated for all minerals or samples at the standard state (25°C, 1 bar) for simplicity (Niu & Batiza, 1991).
[7]Model weight fractions of fore-arc peridotites were derived from the extent of melting (Niu, 1997) in combination with least-squares solutions using mineral compositions and whole-rock analyses reconstituted by assuming that both Si and Mg are mobile but Al and Fe are immobile during serpentinization and that unserpentinized peridotites should lie on the mantle depletion array. It should be noted that the high Opx/Ol ratio in fore-arc peridotites (vs abyssal peridotites) is consistent with melting at greater pressures (Herzberg, 1992).
[8]Average weight fractions of abyssal peridotites are from Niu (1997) and Niu *et al*. (1997).
[9]The calculated density difference of ~0·7% is equivalent to a temperature effect of ~233°C for two peridotites with identical compositions by using a thermal expansion coefficient of $\alpha = 3 \times 10^{-5}\,K^{-1}$.

1996). This temperature difference is similar to, or greater than, the postulated thermal contrast between a hot mantle plume and its ambient asthenosphere. It should be noted also that abyssal peridotites are residues that have risen to the shallowest level, and experienced the greatest degrees of decompression melting and depletion (Dick *et al*., 1984; Niu, 1997; Niu *et al*., 1997). They thus represent the least dense portion of the oceanic lithospheric mantle. Therefore, the calculated density of 3·33 g/cm^3 is the minimum value of the mean oceanic lithosphere and the fore-arc lithosphere should be $>0{\cdot}7\%$ less dense than mean oceanic lithosphere.

The old apparent age of the Mariana sub-arc lithosphere suggests that it may be, at least partly, a remnant of ancient subcontinental lithosphere (e.g. as beneath Japan arcs) although the Mariana arc system may have been complicated by the evolution of the Philippine Sea Plate and by the involvement of mantle plumes in its early history (Macpherson & Hall, 2001). This would suggest the existence of a passive continental margin (e.g. eastern Asia) before the inception of present-day Pacific Plate subduction.

The excessive depletion of the Tonga sub-arc lithosphere, plus the existence of dry basaltic melts inferred from the troctolitic cumulates, points to partial melting of a hot mantle. As the Louisville Seamount Chain (LSC) in the southern Pacific is a typical hotspot track (e.g. Hawkins *et al.*, 1987; Lonsdale, 1998; Norton, 2000), and is currently subducting into the Tonga and Kermadec Trenches, a plume-head product might be anticipated (Campbell & Griffiths, 1990)—a less dense, buoyant lithosphere such as an unsubductable oceanic plateau (Ben-Avraham *et al.*, 1981; Abbott *et al.*, 1997). Seamounts, seamount chains and ridges are subductable, but mantle plume heads with thickened lithospheric roots [e.g. ~300 km of the Ontong Java Plateau (OJP) (Klosko *et al.*, 2001)] are not (Ben-Avraham *et al.*, 1981; Abbott *et al.*, 1997). Therefore, the plume-head material of the Louisville hotspot should be preserved. The Tonga lithospheric root, by its excessive depletion, meets the requirement. The LSC isotope signatures in some Tonga arc lavas (Ewart *et al.*, 1998) may reflect the presence of the underlying plateau lithosphere. As Tonga lithosphere is volumetrically small considering the long-lived history of the LSC, it is possible that much of the Fiji Plate and the Tonga lithosphere are parts of the same plateau system, but modified by complex tectonic activities in the SW Pacific since the Eocene. The Fiji Plate is as shallow as the OJP and Tonga arc–backarc system, and thus is similarly buoyant. The plateau (vs micro-continent) nature of the Fiji Plate is also suggested by its underlying asthenosphere reflected by MORB lavas along the recently (~12 Ma) opened North Fiji Basin spreading centre (e.g. Eissen *et al.*, 1994). Some enriched or ocean-island basalt (OIB)-like lavas (Eissen *et al.*, 1994) may be assimilated plateau material itself or may be associated with recent asthenospheric flow of the Samoa mantle plume. The northwestward subduction of the Pacific Plate beneath Tonga and the northeastward subduction of the Australian oceanic plate beneath the New Hebrides is consistent with the Fiji Plate being a buoyant unsubductable oceanic plateau. Importantly, plate reconstructions of the SW Pacific (Yan & Kroenke, 1993) locate the Louisville hotspot beneath the Tonga–Fiji plateau between 100 and 65 Ma (the oldest seamount near the Tonga trench). This, plus the fact that no seamounts north of the Fiji plateau are older than 100 Ma, argues that the Tonga–Fiji plateau is indeed the best candidate for the plume head of the Louisville hotspot. Both geochemical studies (Mahoney *et al.*, 1993) and tectonic reconstructions (Yan & Kroenke, 1993) indicate that the OJP is unrelated to the Louisville hotspot.

The foregoing discussion and the fact that the dense Solomon Plate subducts northeastward beneath the OJP demonstrates the general significance of the concept that compositional buoyancy contrast within the lithosphere creates the favoured and necessary condition for subduction initiation.

ILLUSTRATION AND QUANTITATIVE EXAMINATION OF THE NEW CONCEPT

Our concept of compositional buoyancy contrast within the lithosphere as favoured sites for subduction initiation is of general significance on a global scale. This concept, however, is best illustrated, as an example, by the presence of such a contrast at the ‘edges’ of oceanic plateaux on the seafloor.

Oceanic plateau model

Figure 3 illustrates the evolution of an oceanic plateau into a supra-subduction system such as the Tonga–Fiji plateau. An oceanic plate is initiated at ocean ridges, and thickens with time as it cools and moves away from the ridge. When sufficiently cold and dense, it returns to the mantle eventually through subduction zones (e.g. Forsyth & Uyeda, 1975; Davies & Richards, 1992; Stein & Stein, 1996). A mantle plume initiates at a basal thermal boundary layer in the mantle and rises because of thermal buoyancy and because of the formation and growth of a more buoyant plume head (Campbell & Griffiths, 1990; Hill *et al.*, 1992; Davies, 1998). When the plume head reaches a shallow level, it melts by decompression, and produces thick basaltic crust and thickened, highly depleted residues (Campbell & Griffiths, 1990; Hill *et al.*, 1992; Herzberg & O’Hara, 1998; Herzberg, 1999). This whole assemblage is less dense than the underlying asthenosphere, thus forming the buoyant oceanic plateau (Burke *et al.*, 1978; Ben-Avraham *et al.*, 1981; Abbott *et al.*, 1997). This plateau moves with the plate over the hotspot, leaving a track on the younger seafloor. When this buoyant plateau reaches a subduction zone (Fig. 3b): (1) it is too buoyant to subduct, thus becoming part of

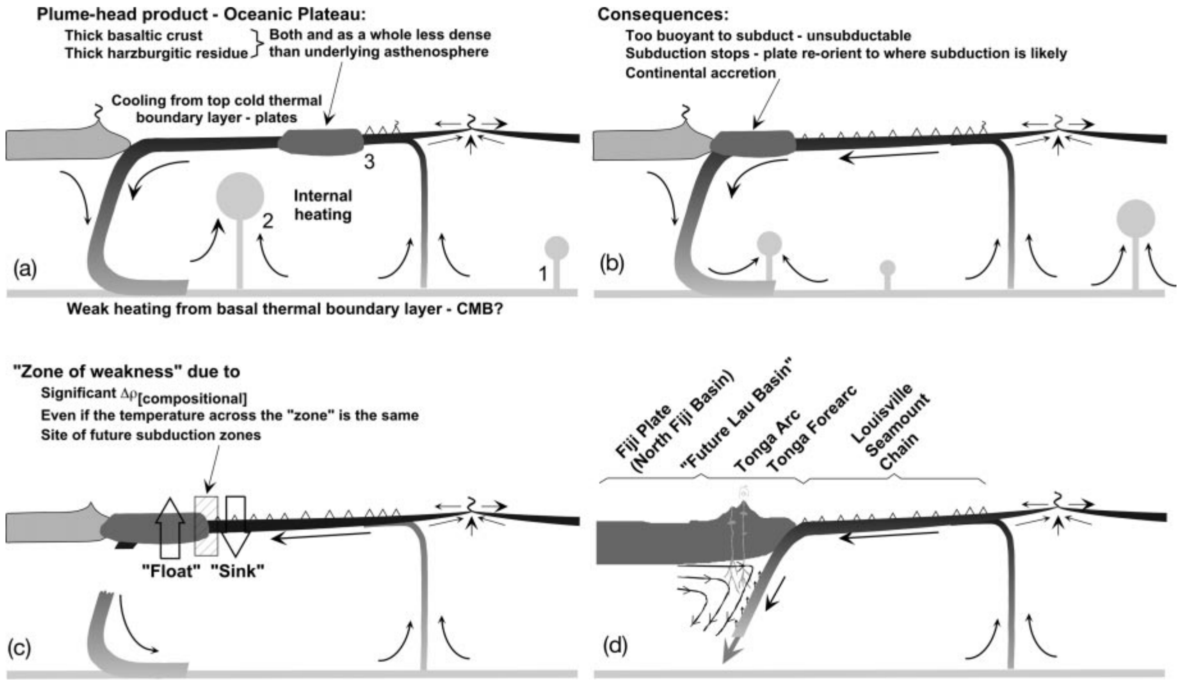

Fig. 3. Schematic illustrations of the concept of subduction initiation when an oceanic plateau reaches an existing trench. (a) Initiation, thickening and subduction of oceanic lithosphere. Initiation and rise of a mantle plume from a basal thermal boundary layer (1), development of plume head (2), and formation of oceanic plateau by decompression melting of plume head (3). (b) The plateau moves with the plate and a hotspot track is generated on the younger seafloor as the plate moves over the plume stem. This plateau, when it reaches the trench, has important consequences as indicated. (c) A large compositional buoyancy contrast at the plateau edge becomes the focus of the stress within the plate, establishing the favoured and necessary condition for the initiation of a new subduction zone. (d) Initiation and subduction of the dense oceanic lithosphere soon leads to dehydration-induced mantle wedge melting for arc magmatism. It should be noted that (d) is meant to illustrate the concept, simplified and exaggerated to include the Fiji Plate.

newly accreted continent (e.g. Ben-Avraham *et al.*, 1981; Abbott *et al.*, 1997; Herzberg, 1999); (2) subduction stops (or 'trench jams') at least momentarily; (3) the plate reorients its motion to where subduction is easier. Importantly, a large compositional buoyancy contrast develops at the edge of the plateau (Fig. 3c). The normal oceanic lithosphere is cold and dense, and will 'sink', but the compositionally buoyant plateau tends to 'float'. This compositional buoyancy contrast marks the 'zone of weakness' for the initiation of new subduction zones. We hypothesize that the Tonga–Fiji plateau is the buoyant mantle plume head of the Louisville hotspot. We also hypothesize that Kamchatka arc lithosphere and the northern part of the Okhotsk Sea lithosphere are the plume-head products of the Hawaiian hotspot (Fig. 1).

The Caribbean and Scotia Seas are the only small plates beneath which active subduction is occurring in the Atlantic. The Caribbean–Colombian terrane is considered to be the plume head of the Galapagos hotspot (Burke *et al.*, 1978; Duncan & Hargraves, 1984; Hill *et al.*, 1992). We further hypothesize that the Scotia Plate is the plume head of the Bouvet hotspot. The existence of Bouvet-like geochemical signatures observed in lavas from the Scotia Sea spreading centre (Pearce *et al.*, 2001) may be due to contamination with the underlying plume-head lithosphere. It should be noted that the accepted Bouvet hotspot track extends from Bouvet Island towards the SE margin of Africa (e.g. Georgen *et al.*, 2001) (Fig. 1). Indeed, the Scotia Plate–plateau was located at the postulated birthplace of the Bouvet hotspot offshore southern Africa before the opening of the South Atlantic at ~130 Ma (e.g. Scotese *et al.*, 1988).

Quantitative consideration

Figures 4 and 5 develop our concept quantitatively. Focal mechanisms show that oceanic lithosphere older than 35 Ma is always under compression, mostly because of ridge push (F_{RP}, ~4 × 10^{12} N/m) (e.g. Wiens & Stein, 1983) (Fig. 4b). Such compression will lead to the development of two groups (A and B) of conjugate faults if the lithosphere is physically uniform (Fig. 4b). If there exists a compositional buoyancy contrast within the lithosphere, i.e. between the less dense oceanic plateau (which tends to 'float') and dense normal oceanic lithosphere (which tends to 'sink') (Fig. 4c), this buoyancy contrast will induce secondary extension and compression, making Group A faults (transtensional) weaker than Group B faults (transpressional). A reverse fault (Fig. 4d) will thus develop along one of the weakest Group A fault planes at or near the lithological contact to release stress.

The relevant forces associated with this reverse fault can be evaluated by assuming an idealized geometry as in Fig. 4d. This assumption is reasonable for the purpose because the dip direction is correct (Fig. 4c) although the actual geometry of the lithological contact is unknown (but could be complex). To initiate subduction is equivalent to initiating the displacement along the 'reverse' fault plane. The theory (Turcotte & Schubert, 1982) is that a shear stress with magnitude of $\tau = f_s\sigma_n$ or $\tau = f_s(\sigma_n - p_w)$ (if pore pressure p_w is present; where the friction coefficient is $f_s = 0{\cdot}85$ for most rocks) must be applied parallel to the fault to cause sliding when the two sides of the fault are pressed together by the normal stress σ_n if the lithosphere under consideration is compositionally uniform. For compositionally heterogeneous lithosphere (Fig. 4d), the fault plane is already weakened. This is because the normal oceanic lithosphere (NOL) is dense; it tends to sink and pulls away from the less dense oceanic plateau lithosphere (OPL). The OPL, which tends to float (Fig. 4d), is arguably more coherent than the NOL because of its simple genesis relative to that of the NOL near ocean ridges. Thus, the loading effect of the OPL in response to the sinking NOL should be relatively small and can thus be neglected for this purpose. The physical consequence of the incipient sinking of the NOL can be evaluated by considering the triangular area relevant to, and beneath, the fault plane. The potential body force (negative buoyancy) of the incipient sinking by a 1 m thick sliver of the NOL is $F_B = [\rho\alpha\Delta T/2][\{\tfrac{1}{2}\} T_{OL}^2 \tan(\pi/2 - \beta)]g$, where the first factor in the brackets is the density contrast between

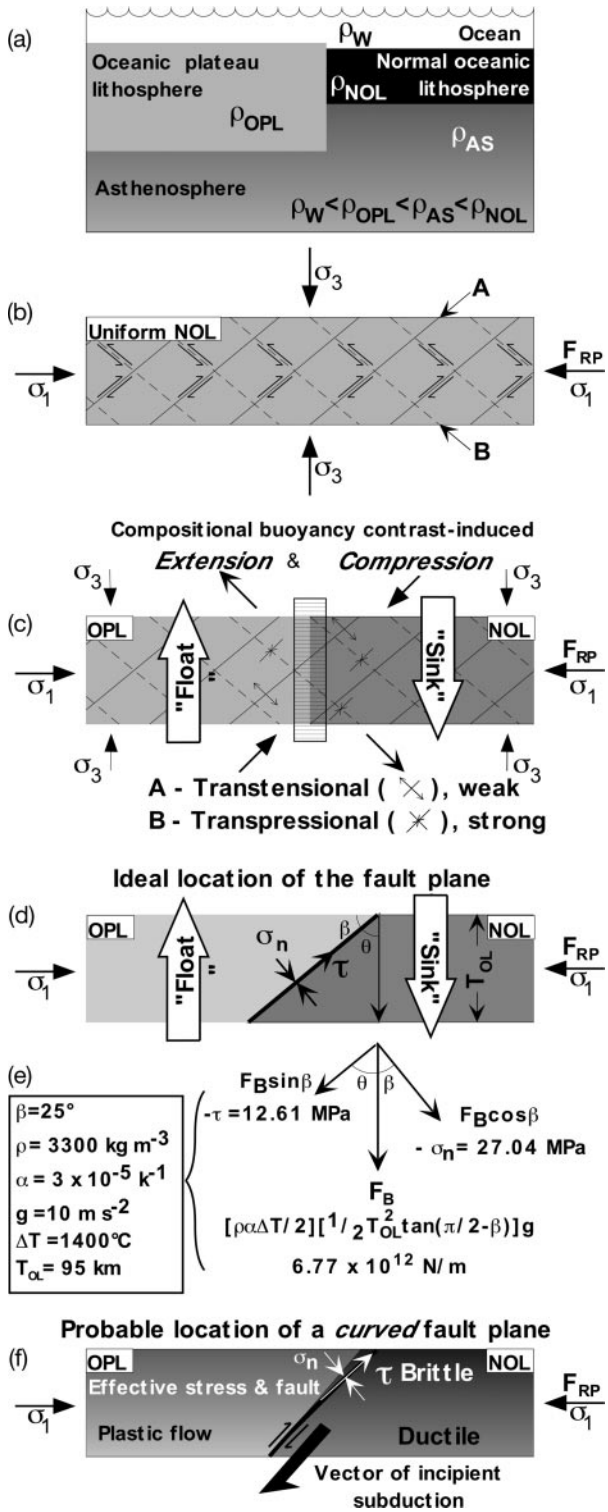

Fig. 4. (a) The ocean–lithosphere–asthenosphere system illustrating density differences as a result of thermal or compositional changes. (b) Two groups (A and B) of conjugate faults develop in the normal oceanic lithosphere as a result of ridge push (F_{RP}) or other lateral deviatoric compression. (c) Compositional buoyancy contrast exists between the less dense oceanic plateau (OPL, which tends to 'float') and dense normal oceanic lithosphere (NOL, which tends to 'sink'). This buoyancy contrast makes Group A faults transtensional and Group B faults transpressional, thus leading to the development of a transtensional reverse fault along an A-fault plane at or near the lithological contact. (d) Idealized geometry and location of the reverse fault plane for quantitative evaluation of the relevant forces and stresses associated with it. (e) The physical consequences of the incipient sinking of the NOL are evaluated by considering the volume of lithosphere of 1 m thickness defined by the triangular area beneath the fault plane. The parameters used are taken from Stein & Stein (1996). (f) Schematic illustration showing that the reverse fault for subduction initiation may be developed on the NOL side owing to its weakness (vs OPL, which is arguably stronger because of rapid within-plate emplacement) during its early history of development near ocean ridges where the lithosphere is highly faulted. This suggests that slivers of normal oceanic lithosphere as rare genuine ophiolites formed at ocean ridges could be incorporated in the fore-arc region during subduction initiation.

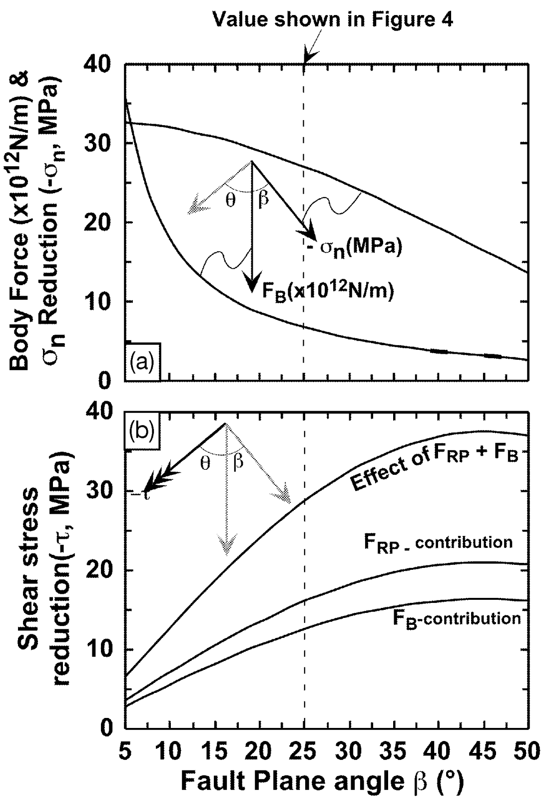

Fig. 5. Calculated forces and stresses as a function of the dip angle β of the fault plane across a plate of 95 km thickness. (a) Potential buoyancy force (F_B) at all practical dip angles of 20–45° is similar to or significantly greater than ridge-push force ($\sim 4 \times 10^{12}$ N/m), which by itself reduces normal stress by 15–0 MPa, effectively 'opening' the fault plane and creating resistance-free sliding. (b) Both F_{RP} and F_B together reduce along-fault-plane shear resistance by 25–37 MPa for realistic dip angles (20–45°). This shear-stress reduction is significantly greater than the shear strength of the oceanic lithosphere—only a few MPa (Kanamori & Anderson, 1975; Wiens & Stein, 1983)—and is also greater than the assumed lithosphere shear strength of ~10–20 MPa in theoretical models (McKenzie, 1977; Hynes, 1982). It should be noted that the shear-stress reduction would be >50–70 MPa if only the upper elastic portion (~50 km) is considered, and would be still greater if the effect of water is included.

the NOL and the underlying asthenosphere due to temperature differences. The second factor is the volume of that triangle per metre length. ρ is the mean density of the NOL at room temperature (i.e. 25°C), α thermal expansion (3×10^{-5} K^{-1}), g acceleration due to gravity (10 m/s^2), T_{OL} the thickness of mature NOL (95 km), and $\Delta T = 1400$°C is the temperature difference between the base of the NOL and the seafloor, with $\Delta T/2$ assumed to be the mean temperature of the NOL. All these parameters are taken from the GDH1 model (Stein & Stein, 1996). It should be noted that the results are actually model independent. For a dip angle of $\beta = 25°$, for example, the potential sinking force, $F_B = 6{\cdot}77 \times 10^{12}$ N/m, is of the order of F_{RP} ($\sim 4 \times 10^{12}$ N/m). This sinking force tends to pull away the NOL from the fault plane by reducing the normal stress (σ_n) by ~27 MPa, and tends to slide the NOL along the fault plane downward by reducing the shear resistance (τ) by ~13 MPa. The net effect is to ease sliding along the fault plane—aiding subduction initiation.

Figure 5a shows that for a range of reasonable dip angles, say 20–45°, the potential sinking force (F_B) is in the range of (~4–8) $\times 10^{12}$ N/m, which is similar to or greater than F_{RP} ($\sim 4 \times 10^{12}$ N/m), and which by itself reduces normal stress (σ_n) by 15–30 MPa, effectively 'opening' the fault plane and creating resistance-free sliding. Figure 5b shows that the combined effect of F_{RP} and F_B reduces along-fault-plane shear resistance by 25–37 MPa for dip angles of 20–45°. This shear-stress reduction alone (without considering the huge reduction of σ_n) is already significantly greater than the shear strength of oceanic lithosphere, which is only a few MPa inferred from intra-plate earthquake stress drops (Kanamori & Anderson, 1975; Wiens & Stein, 1983), and is also greater than assumed lithosphere shear strength of ~10–20 MPa in theoretical models (McKenzie, 1977; Hynes, 1982). Therefore, compositional buoyancy contrast within the oceanic lithosphere under compression creates adequately the conditions for lithosphere break-up and subduction initiation.

It should be noted that the net effect of 'compositional buoyancy contrast within the oceanic lithosphere under compression' in favour of subduction initiation is even greater than the above estimation. This is because the above shear-stress reduction (1) does not consider the effect of normal stress reduction that tends to pull apart the fault plane, which weakens the fault plane tremendously, and (2) does not consider the effect of water that permeates along and lubricates the fault plane. Importantly, the calculation is based on a lithosphere of 95 km thickness, but if only the upper ~50 km elastic portion of the lithosphere (Stein & Stein, 1996) is considered, the effective shear-stress reduction would double the values in Fig. 5, of the order of >50–70 MPa.

THE NEW CONCEPT VS EXISTING MODELS

Our concept of compositional buoyancy contrast within the lithosphere for subduction initiation has general significance, and differs from previous models in the literature that have difficulties in explaining many observations. For example, the minimum deviatoric compression of $\sim 3 \times 10^{13}$ N/m needed to initiate subduction at passive margins (Mueller & Phillips, 1991) is unavailable unless there is an exceptional situation. The proposed tensional vs compressional stress field at passive margins (Kemp & Stevenson, 1996) also represents special vs general situations. All these special conditions are inconsistent with the fact that plate tectonics is a general consequence of mantle convection. Sediment accumulation at passive margins both as an excess load and as a thermal blanket to weaken the lithosphere is physically plausible (Erickson, 1993; Kemp & Stevenson, 1996), but its effect alone on subduction initiation is limited. Younger oceanic lithosphere is relatively weak and requires less shear stress for rupture (Cloetingh *et al.*, 1982), but there is no evidence of such rupture and resulting subduction initiation throughout the world's ocean basins. The suggestion that subduction initiates within continents (Ellis, 1988) is physically unlikely to result in world-wide subduction of oceanic lithosphere beneath continental and oceanic lithospheres. The Macquarie Ridge (Fig. 1), marking the Australian–Pacific plate boundary in the southern ocean, is considered to be the most convincing case of incipient initiation of a subduction zone (Ruff *et al.*, 1989), but the absence of a buoyancy difference across the plate boundary makes it physically difficult. It is this difficulty that has led to the conflicting views in the literature on which plate subducts beneath which. The fact that the diffuse compressional boundary in the central Indian Ocean has not developed into a subduction zone (Wiens *et al.*, 1985) is consistent with our concept because there is no compositional buoyancy contrast across this diffuse boundary. The unusual intraplate seismic intensity along the Samoa–Gilbert–Ralik–Caroline line in the SW Pacific (Fig. 1) allowed Okal *et al.* (1986) to speculate on an incipient subduction initiation. This observation is important because this line marks in a broad sense a compositional buoyancy contrast in the lithosphere between the buoyant Tonga–Fiji plateau and OJP to the SW and the old, cold, and dense Pacific Plate to the NE.

THE EFFICACY OF THE NEW CONCEPT AND PREDICTIONS

The concept of lateral compositional buoyancy contrast or oceanic plateau model that we present here for subduction initiation may provide insights into many fundamentally important, yet unresolved, problems, as follows.

The origin of the 43 Ma bend along the Hawaii–Emperor Seamount Chain

The Hawaii–Emperor Seamount Chain (H-ESC) on the Pacific Plate is the best defined hotspot track on the Earth. Its volcanic age progression away from Hawaii not only confirms plate tectonic theory, but also verifies the hotspot hypothesis (Morgan, 1971). If hotspots are the surface expressions of deep, fixed mantle plumes, the hotspot tracks record the direction, absolute velocity, and possible changes of the plate motion. This would suggest that the ~43 Ma bend along the H-ESC reflects a sudden change in Pacific Plate motion direction by ~60°. The lack of an apparent mechanism for such change (Norton, 1995), plus the argument that hotspots sources are not fixed (Norton, 1995, 2000; Tarduno & Cottrell, 1997; Koppers *et al.*, 2001), has led to speculation that the ~43 Ma bend may be caused by the southward drift of the Hawaiian hotspot source before ~43 Ma (Norton, 1995). Although less well defined, the 43 Ma bend is apparent along other hotspot tracks on the Pacific plates (Morgan, 1971; Norton, 1995, 2000; Lonsdale, 1998). This would require a simultaneous multi-hotspot source swing in the Pacific deep mantle. It is indeed difficult to imagine that all these hotspot sources move at the same time, in the same direction and by the same amount. It is, however, physically straightforward to have the Pacific Plate, a single unit, reorient itself.

We hypothesize (Fig. 3b) that collision of the Hawaii plume-head plateau with an existing subduction zone—the Cretaceous–Eocene Andean-type Okhotsk–Chukotka active continental margin (Fig. 1) (Garver *et al.*, 2000)—some 43 Myr ago stopped Pacific Plate subduction momentarily. The Pacific Plate then reoriented its motion in the direction where subduction was likely—the present-day western Pacific, where subduction zones had formed ~7 Myr earlier (Moberly, 1972; Taylor, 1993). This readily explains the 43 Ma bend along the H-ESC, and why this 43 Ma bend is less sharp along hotspot tracks farther to the south such as the Tuamoto, Macdonald and Louisville seamount chains on the Pacific Plate (Norton, 2000) (Fig. 1). This prediction is strongly supported by the recent work of Garver *et al.* (2000). These workers established that ~44 Ma is the youngest fission-track grain age (FTGA) of primary igneous zircons in the far-east Kamchatka fore-arc Ukelayat Flysch sandstones derived from the Okhotsk–Chukotka continental arc (Garver *et al.*, 2000) (Figs 1 and 6a). This minimum age strongly suggests the termination of the Chukotka continental arc volcanism—the provenance of the FTGA zircons. This termination of the arc volcanism as a result of subduction cessation is probably the consequence of the collision of the Hawaiian mantle plume head against the Okhotsk–Chukotka continental arc. We believe that the collision at ~44 Ma is the actual cause of, not coincidental with, the sudden reorientation of the Pacific Plate at ~43 Ma marked by the bend along the H-ESC (Fig. 6b). Assuming the oldest Meiji Seamount along the H-ESC is ~82 Ma, the

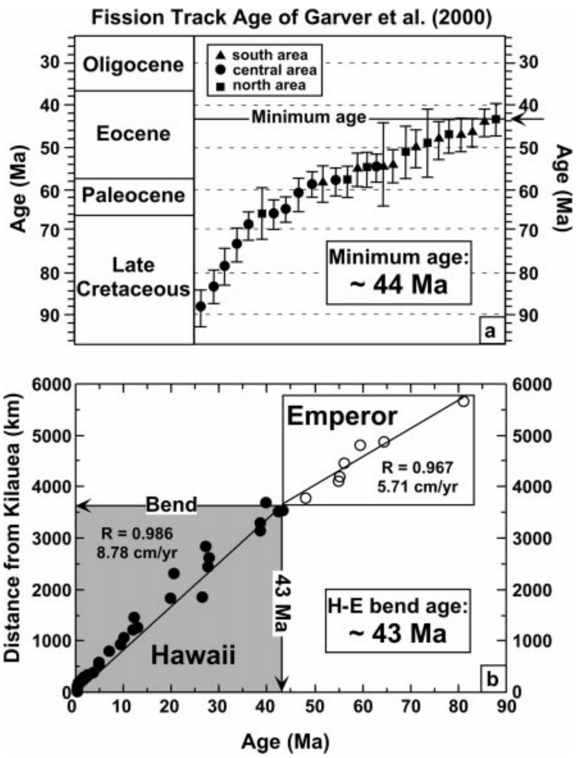

Fig. 6. (a) Fission-track grain age (FTGA) of primary igneous zircons in the Cretaceous–Tertiary Ukelayat Flysch, simplified from Garver *et al.* (2000). The provenance of the zircons is the Andean-type Okhotsk–Chukotka continental magmatic arcs (Fig. 1). It should be noted that the minimum zircon age of ~44 Ma (Garver *et al.*, 2000) is essentially the same as the age of the ~43 Ma bend along the Hawaii–Emperor Seamount Chain (H-ESC). (b) Volcanic K–Ar age data of H-ESC for all seamounts by Clague & Dalrymple (1989) except the oldest data point by Ar–Ar from Detroit seamount (Keller *et al.*, 1995) plotted against distance from Kilauea (Clague & Dalrymple, 1989). It should be noted that, contrary to the previous notion, there is a statistically significant (at >99% confidence levels) change in Pacific Plate velocity from ~5·71 cm/yr before 43 Ma to ~8·78 cm/yr after 43 Ma with respect to the hotspot.

Hawaiian mantle plume would have an age of ~125 Ma (i.e. 43 Ma + 82 Ma = 125 Ma), which is not significantly different from the ~122 Ma of the first phase of OJP volcanism (Mahoney *et al.*, 1993). It has been suggested that there is no obvious change in spreading rate of the Pacific Plate at 43 Ma with respect to Pacific hotspots (Clague & Dalrymple, 1989; Keller *et al.*, 1995; Tarduno *et al.*, 2001), but Fig. 6b shows that spreading rate change associated with the 43 Ma bend, from ~5·71 cm/yr to ~8·78 cm/yr, is significant.

Ophiolite emplacement, genesis of boninites, and TTGs in arc–fore-arc settings

Ophiolites were originally thought to represent normal oceanic lithosphere generated at ocean ridges, but their emplacement onto land by obduction against gravity raises difficulties. Subsequent studies have shown that lavas from most ophiolite sequences have geochemical signatures typical of arc magmas [see review by Bloomer *et al.* (1995)]. Figure 3d shows that these supra-subduction zone ophiolites (i.e. arc–fore-arc volcanics) may represent magmatic assemblages developed at edges of oceanic plateaux during early stages of subduction initiation. Figure 4f suggests that slivers of normal oceanic lithosphere, as rare genuine ophiolites, formed at ocean ridges could also be incorporated in the fore-arc region during subduction initiation without invoking any particular anti-gravity obduction. In this context, formation of boninite melts by subducting-slab dehydration-induced melting of highly depleted harzburgitic residues (oceanic plateau or sub-continental lithospheric roots) is credible, which is supported by highly depleted incompatible element abundances yet enriched isotopic signatures seen in northern Tonga boninites (Ewart *et al.*, 1998). Dehydration-induced melting of the plateau crust can also readily produce the tonalite–trondhjemite–granodiorite (TTG) assemblage (and equivalent volcanics) as seen on the Fiji plateau and in the Tonga–Kermadec fore-arc regions (e.g. Robin *et al.*, 1993; Worthington *et al.*, 1999).

Implications for continental accretion

Our oceanic plateau model predicts the formation of the 'first subduction zones' on Earth, i.e. the inception of plate tectonics, probably in the Archaean. This is also relevant to continental crust growth, about which there has been much discussion over the last 30 years. Although the 'island-arc andesite model' for continental crust accretion (Taylor, 1967) has been popular, there is mounting evidence in support of an oceanic plateau model for continental crust accretion (Ben-Avraham *et al.*, 1981; Abbott & Mooney, 1995; Abbott *et al.*, 1997; Albarède, 1998). Because continental crust cannot be sustained without support from its buoyant lithospheric roots (Jordan, 1988), and because of the coupling between the crust and its underlying lithosphere (Griffin *et al.*, 1999; Pearson, 1999), a unified model is required to explain the accretion of both components. The highly depleted (low Fe/Mg) Archaean cratonic lithosphere, and the progressively less depleted continental lithospheric roots around Archaean cratons through to Proterozoic to Phanerozic orogenic belts (e.g. Boyd, 1989; Herzberg, 1993; Griffin *et al.*, 1999) are consistent with decreasing extent of melting and depletion by mantle plumes as a result of Earth's secular cooling (Richter, 1988). All these observations, plus the more mafic nature of bulk continental crust than andesites (Ben-Avraham *et al.*, 1981; Abbott & Mooney, 1995; Rudnick, 1995; Abbott *et al.*, 1997; Albarède, 1998) argue for the volumetric significance of continental growth by oceanic plateaux. Although the role of arc magmatism for continental crust growth is apparent (see below), the volumetric significance of this role is debated (e.g. Reymer & Schubert, 1984; Abbott & Mooney, 1995; Albarède, 1998; Dimalanta *et al.*, 2002). The question is why the average continental crust has arc-like geochemical signatures (Taylor, 1967; Rudnick, 1995), i.e. depletion of high-field strength elements (e.g. Nb, Ta, Ti), which mantle plume or plateau volcanic rocks do not have. This can be resolved by the oceanic plateau model for subduction initiation (this study) and for continental crust growth (Ben-Avraham *et al.*, 1981; Rudnick, 1995; Abbott *et al.*, 1997; Albarède, 1998).

The Earth's early cooling may have been achieved by mantle plumes to produce 'proto-continents' with komatiite crust and highly depleted (low Fe/Mg), buoyant 'cratonic' lithospheric residues (Boyd, 1989; Herzberg, 1993, 1999; Griffin *et al.*, 1999). These 'proto-continents' are physically buoyant with positive topography, thus allowing the development of 'first subduction zones' at their edges. The subduction-related volcanism would add 'arc' geochemical signatures to the existing crust. The dehydration melting of the proto-crust in these arc settings also produces more silicic TTG assemblages—an important aspect of continental crust accretion over Earth's history, particularly in the Archaean (e.g. see Condie, 1997). Addition of new oceanic plateaux with arc assemblages (Fig. 3d) to the existing continents by collision preserves these arc rocks within the continents. We consider this is a repetitive process over Earth's history. In other words, the bulk continental lithosphere (including crust) is the aggregate of plumes–plateaux with subduction–arc terranes between them. This is consistent with geological observations that all the continents consist of cratonic blocks joined together by orogenic belts (Abbott & Mooney, 1995) with typical subduction-zone magmatic assemblages. This explains why average continental crust has arc-like geochemical signatures. In this context, it is noteworthy that island arcs have positive topography, whereas ocean floor has negative topography. Consequently, it is easy to incorporate arc ophiolites (abundant) in continents during collision, but physically difficult to transport true ophiolites (rare) from ocean floor onto land against gravity.

Passive continental margins as loci of future subduction zones

Passive continental margins mark the largest lateral compositional buoyancy contrast within the lithosphere. These margins must, therefore, be potential loci of subduction zones, which would explain many aspects of the geological record as elegantly described by the Wilson cycle (Dewey & Bird, 1970; Burke *et al.*, 1976), and the fact that no ancient (>200 Ma) ocean floor has survived recycling. The rejection of passive margins as sites of subduction initiation comes from the lack of present-day examples along passive margins in the Atlantic and Indian Oceans. This has led to the suggestion that stress in excess of 400 MPa is needed to overcome the model strength of oceanic lithosphere (Cloetingh *et al.*, 1982; Mueller & Phillips, 1991), but such model strength is inconsistent with the observation that stress drops in oceanic intra-plate earthquakes average only a few MPa (Kanamori & Anderson, 1975; Wiens & Stein, 1983). This inconsistency, plus the conflicting views on whether the stress field at passive margins is compressional (e.g. Cloetingh *et al.*, 1982; Mueller & Phillips, 1991) or extensional (Kemp & Stevenson, 1996), suggests that it has not been understood why no obvious subduction occurs along passive margins in the Atlantic. The two locations where subduction occurs in the Atlantic are of intra-oceanic subduction type associated with oceanic plateaux such as the South Sandwich and Lesser Antilles subduction systems (see above; also Fig. 1).

In fact, subduction does initiate at passive continental margins such as the Ryukyu arc–subduction system developed on the Chinese continental margin. This arc–subduction system, which is in its transition to an active continental margin, started in early Miocene from the north and has propagated southward since then. The southern Ryukyu did not become a subduction zone until 8 Ma (Lallemand *et al.*, 2001). We consider that the apparent paradox that no obvious subduction initiation occurs along the passive margins in the Atlantic and Indian Oceans may result from an overlooked fact that portions of many of the passive margins in the Atlantic, around Africa and Australia (perhaps also the Antarctic) are characterized by 'seaward-dipping reflectors' (SWDRs)—the thick sequences of basalts and intrusives associated with mantle plumes during continental break-up (Eldholm & Coffin, 2000). Such voluminous magmatic constructions emplaced in short time periods are arguably coherent and difficult to break. Importantly, the lithospheric roots beneath the SWDRs are thickened, highly depleted residues of mantle plume melting. They are thus too buoyant to sink. Consequently, subduction will not preferentially initiate at the sites of the SWDRs, but at sites away from the SWDRs. Upon initiation at the non-SWDR sites, the subduction zone will propagate laterally as is the case of the Ryukyu arc at the Chinese continental margin (see above). We predict that subduction initiation along the American passive margins in the Atlantic will start through lateral propagation away from the existing Lesser Antilles and South Sandwich subduction zones where the required stress is minimal. When the lateral propagation reaches the sites of SWDRs, subduction may develop, particularly when deep portions of the intrusives become metamorphosed to dense eclogites.

FUTURE WORK

The new concept for subduction initiation we present here explains many geological observations. Given the global tectonic significance of subduction initiation, it is our intention that this contribution will stimulate future research within the geoscience community. For example, we hypothesize, with indirect evidence, that the Fiji–Tonga lithosphere represents the plume head of the Louisville hotspot, the Scotia Sea Plate is the plume head of the Bouvet hotspot, and the Kamchatka–northern Okhotsk Sea lithosphere is the plume head of the Hawaiian hotspot. However, these hypotheses must be seriously tested. Although geophysical surveys such as geomagnetic, gravity and seismic experiments are important, we think that the ultimate test will rely on ocean drilling.

We predict that drilling at appropriate sites in the North Fiji Basin will reveal its oceanic plateau nature and its genetic link with the Tonga lithosphere and the Louisville hotspots (Fig. 1). We predict that drilling at key sites in the northern Okhotsk Sea lithosphere and Scotia Sea will provide a test on whether they are plume-head materials of the Hawaiian and Bouvet hotspots respectively. The plume-head and arc materials are distinct petrologically, geochemically and geochronologically. Drilling into the active arc–fore-arc systems in these locations will prove difficult because of the thickened arc volcanic constructions. It should be noted that oceanic plateau crust is often thought to be thick (up to 30 km for OJP), but it is difficult to resolve seismically between mantle residues and ultramafic cumulates without actual drilling. Also, without drilling, it is difficult to know to what extent the plateau crust may have actually been eroded or modified in the arc–fore-arc settings by crustal melting—a process that is likely to have produced silicic rocks (e.g. TTGs) from the Archaean to the present.

SUMMARY

(1) We propose that lateral compositional buoyancy contrast within the lithosphere creates the favoured and necessary condition for subduction initiation. This concept is of general significance, but it is best illustrated by the presence of compositional buoyancy contrast at the edges of oceanic plateaux, which are compositionally highly depleted and physically buoyant relative to the neighbouring normal oceanic lithosphere.

(2) Such heterogeneous lithosphere under deviatoric compression (e.g. ridge push or other equivalent forces) develops reverse fault zones with combined forces in excess of the oceanic lithosphere strength, allowing the cold, dense normal oceanic lithosphere to sink into the hot asthenospheric mantle beneath the buoyant overriding oceanic plateaux, i.e. the initiation of subduction zones. We term this concept the 'oceanic plateau model'.

(3) This plateau model explains the origin of the $\sim$43 Ma bend along the Hawaii–Emperor Seamount Chain on the Pacific Plate, which suggests that the Hawaii hotspot is probably $\sim$125 Myr old.

(4) The plateau model also explains ophiolite emplacement, genesis of boninites and TTG assemblages in arc–fore-arc settings, and favours the proposal that continental growth is achieved largely by mantle plumes and oceanic plateaux. The model also explains the arc-like geochemical signatures of average continental crust.

(5) Passive continental margins are loci of future subduction zones. Portions of many passive margins are mechanically strong with thickened volcanic constructions or SWDRs and buoyant lithospheric roots associated with the activity of mantle plumes during continental breakup. Subduction will not preferentially initiate at these sites, but in places away from these sites. Following initiation, the subduction zone will propagate laterally. We predict that subduction initiation along the American passive margins in the Atlantic will start through lateral propagation away from the existing Lesser Antilles and South Sandwich subduction zones where the required stress is minimal. When the lateral propagation reaches the sites of SWDRs, subduction may develop, particularly when deep portions of the intrusives are metamorphosed to dense eclogites.

(6) Ocean drilling is needed to verify the hypothesis that Fiji–Tonga lithosphere represents the plume head of the Louisville hotspot, Scotia Sea Plate the plume head of the Bouvet hotspot, and Kamchatka–northern Okhotsk Sea the plume head of the Hawaiian hotspot.

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