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Subduction initiation: spontaneous and induced

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Abstract

The sinking of lithosphere at subduction zones couples Earth's exterior with its interior, spawns continental crust and powers a tectonic regime that is unique to our planet. In spite of its importance, it is unclear how subduction is initiated. Two general mechanisms are recognized: induced and spontaneous nucleation of subduction zones. Induced nucleation (INSZ) responds to continuing plate convergence following jamming of a subduction zone by buoyant crust. This results in regional compression, uplift and underthrusting that may yield a new subduction zone. Two subclasses of INSZ, transference and polarity reversal, are distinguished. Transference INSZ moves the new subduction zone outboard of the failed one. The Mussau Trench and the continuing development of a plate boundary SW of India in response to Indo–Asian collision are the best Cenozoic examples of transference INSZ processes. Polarity reversal INSZ also follows collision, but continued convergence in this case results in a new subduction zone forming behind the magmatic arc; the response of the Solomon convergent margin following collision with the Ontong Java Plateau is the best example of this mode. Spontaneous nucleation (SNSZ) results from gravitational instability of oceanic lithosphere and is required to begin the modern regime of plate tectonics. Lithospheric collapse initiates SNSZ, either at a passive margin or at a transform/fracture zone, in a fashion similar to lithospheric delamination. The hypothesis of SNSZ predicts that seafloor spreading will occur in the location that becomes the forearc, as asthenosphere wells up to replace sunken lithosphere, and that seafloor spreading predates plate convergence. This is the origin of most boninites and ophiolites. Passive margin collapse is a corollary of the Wilson cycle but no Cenozoic examples are known; furthermore, the expected strength of the lithosphere makes this mode unlikely. Transform collapse SNSZ appears to have engendered new subduction zones along the western edge of the Pacific plate during the Eocene. Development of self-sustaining subduction in the case of SNSZ is signaled by the beginning of down-dip slab motion, causing chilling of the forearc mantle and retreat of the magmatic arc to a position that is 100–200 km from the trench. INSZ may affect only part of a plate margin, but SNSZ affects the entire margin in the new direction of convergence. INSZ and SNSZ can be distinguished by the record left on the upper plates: INSZ begins with strong compression and uplift, whereas SNSZ begins with rifting and seafloor spreading. Understanding conditions leading to SNSZ and how hinged subsidence of lithosphere changes to true subduction promise to be exciting and fruitful areas of future research.

Keywords: subduction; ophiolites; collision; tectonics

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Jargon Box

Asthenosphere: weak, convecting upper mantle
ARM: amagmatic rifted margin, formed in the absence of extensive volcanism during continental rifting
DSDP: Deep Sea Drilling Project.
IBM: Izu-Bonin-Mariana arc system.
INSZ: induced nucleation of subduction zones.
Lithosphere: crust and uppermost mantle, cooled by conduction. Comprises the plate of plate tectonics.
MORB: mid-ocean ridge basalt.
MRC: Macquarie Ridge Complex.
NUVEL-1: model for estimating plate velocities over 1–5 million-year timescales.
ODP: Ocean Drilling Project.
OJP: Ontong Java Plateau.
Ophiolite: a fragment of oceanic (sensu lato) lithosphere.
SNSZ: spontaneous nucleation of subduction zones.
SSZ: supra-subduction zone (broadly, a convergent margin tectonic setting).
VRM: volcanic rifted margin, formed by volcanism during continental rifting.
Wadati-Benioff Zone: inclined zone of seismicity that marks the descending slab, named after the Japanese and US scientists who first recognized these features.

1. Earth: the subduction planet

Earth is a spectacularly unusual planet and one of its most remarkable features is the plate tectonic system. Missions to other planets reveal that ours is the only planet in the solar system with subduction zones and plate tectonics [1]. The unique nature of plate tectonics on Earth is equivalent to saying that only Earth has subduction zones [2]. In spite of this singularity, there are fundamental misconceptions that concern aspects of plate tectonics and mantle convection. Not only are these wrong, they are deeply embedded prejudices of many earth scientists that continue to be taught to students. The most important misconception is that mantle convection moves the lithosphere (see ‘Jargon Box’), dragging the plates as it moves. This is repeatedly shown in introductory textbooks. In fact, Earth’s mantle convects mostly because cold lithosphere sinks at subduction zones [3] with mantle plumes representing a ‘…clearly resolved but secondary mode of mantle convection’ ([4], p. 159). The base of the continents may be dragged by circulating mantle (continental undertow of [5]), but the pioneering conclusion of Forsyth and Uyeda [6] that the excess density of the lithosphere in subduction zones drives the plates continues to be supported by geodynamicists [7–9]. This negative buoyancy results from the small increase in density that silicates experience as temperature decreases, coupled with the fact that the thermal lithosphere thickens as it ages. Lithosphere becomes denser than the underlying asthenosphere within ~20–50 million years after it forms [7,10]. Density and mass excess continue to increase after this time, and subduction is Earth’s way of returning to equilibrium by allowing great slabs of old, dense lithosphere to sink beneath underlying asthenosphere.

There is a consensus among geodynamicists that the sinking of cold, gravitationally unstable lithosphere drives the plates and indirectly causes mantle to well up beneath mid-ocean ridges. Some estimate that 90% of the force needed to drive the plates comes from the sinking of lithosphere in subduction zones, with another 10% coming from ridge push [11]. Cenozoic plate motions are well predicted from the distribution of age—and thus the mass excess—in subduction zones [9,12]. Mantle tomography shows that subducted lithosphere may sink through the 660-km discontinuity and into the deep mantle [13,14]—striking demonstration that the lithosphere as it ages and cools progressively develops a density excess that takes as long to dissipate as it does to develop. In recognition of the fact that plate motions are passive responses to sinking of the lithosphere at subduction zones, it is more accurate to describe Earth's geodynamic regime as one of 'subduction tectonics' rather than 'plate tectonics'.

In spite—or, perhaps, because—of the seity of Earth's regime of subduction tectonics, we have much to learn about the physics that allow this remarkable mode of planetary convection [1] and when this began. The intention here is to approach the problem by examining Cenozoic examples of subduction initiation. The hope is that a focus on simple models and the clearest examples will encourage teams of geologists, geophysicists and geodynamicists to attack the problem from new perspectives.

2. When did subduction begin?

To understand how new subduction zones form today, we must also consider when and why this tectonic style was established on Earth. It is widely acknowledged that the early Earth was hotter and that there was more mixing of the mantle [1], but this does not require plate tectonics. If the plates ultimately move because they are dense enough to sink in subduction zones and they are sufficiently dense because they are cold, then the much higher heat flow in the Archean required correspondingly more time for the lithosphere to cool, thicken and become gravitationally unstable. The relatively buoyant nature expected for Archean oceanic lithosphere is amplified by the fact that higher heatflow at this time should have resulted in increased melting and thicker oceanic crust [15]. Pre-1.0 Ga oceanic crust is inferred to have been much thicker than modern oceanic crust [16]. Crust is much less dense than asthenosphere, so thicker crust strongly counteracts density excesses produced by lithospheric mantle. Both of these conclusions mitigate against the operation of plate tectonics on the early Earth [17]. Such logic is supported by the absence of plate tectonics on Venus, which may be a good analogue for tectonics on the Earth during the Archean.

When plate tectonics began can also be examined using the ophiolite record, which is an unequivocal index of plate tectonic activity. Ophiolites are fragments of oceanic lithosphere tectonically emplaced on continental crust. These assemblages testify to subduction and plate tectonics in two ways: their formation requires seafloor spreading and their emplacement requires plate convergence. The age distribution of ophiolites has been interpreted to suggest that seafloor spreading began in the late Archean [18], but evidence for Archean ophiolites is sparse and often controversial. Unequivocal ophiolites of this age are rare and the abundance of well-preserved Archean supracrustal sequences does not support contentions that the paucity of such ancient ophiolites is a preservation problem. There is evidence for generation and emplacement of ophiolites at about 1.95–2.0 Ga [19,20], but there was a long period after that for which little evidence for ophiolite formation and emplacement is preserved. It was not until Neoproterozoic time, ~900 Ma, that unequivocal ophiolites were produced, emplaced and abundantly preserved [21], and since that time ophiolites are ubiquitous in the geologic record. Thus, the ophiolite record suggests that seafloor spreading and creation of oceanic lithosphere followed by convergence of oceanic lithosphere may have occurred for brief periods at the end of the Archean and in the Paleoproterozoic, but most of the pre-Neoproterozoic geologic record lacks evidence for plate tectonics and subduction. Davies [7] argued from simple physics that the Earth did not cool sufficiently to allow the lithosphere to subduct until about 1 Ga. It is not clear what Earth's tectonic style was before the modern episode of plate tectonics, but Sleep [15] recognized three modes for convection in silicate planets: magma ocean, plate tectonics and stagnant lid. He noted that plate tectonics requires that a planet be in a delicate balance of thermal states. Plate/subduction tectonics can be shut down either by ridge lock, when the Earth's mantle is too cool to produce melt by adiabatic decompression, or by trench lock, which happens when Earth's interior is too hot and oceanic crust is too thick to subduct. The ophiolite record is thus consistent with geodynamic arguments that our planet wasn't cold enough for subduction to be securely established as the dominant tectonic mode until relatively recently.

3. Induced nucleation of subduction zones

Once we recognize that plate tectonics is the surface expression of a planet with subduction zones, and that these two linked phenomena are unusual among planetary bodies as well as through Earth history, it is easier to appreciate the difficulty of starting subduction zones. This essay focuses on the two ways that subduction must be able to start: spontaneous and induced (Fig. 2). In the case of spontaneous nucleation of a subduction zone, gravitationally unstable lithosphere collapses into the asthenosphere, whereas in the case of induced nucleation of a subduction zone, existing plate motions cause compression and lithospheric rupture. If it is true that the modern episode of plate tectonics began sometime in the Precambrian, and that the excess density of lithosphere drives plate motions, then spontaneous nucleation of at least one subduction zone is required to first set the plates in motion.

There is no question but that realistic numerical models are needed to understand subduction initiation. Still, much can be learned by focusing on what has and has not happened in the recent past, particularly since about half of the active subduction zones began in Cenozoic times [22]. The following discussion returns to these themes, emphasizing Cenozoic examples of subduction zone formation (locations of examples shown in Fig. 1). These examples are used because the geometric relationships of lithospheres, relevant plate motions and sequences of events can be reconstructed with relative confidence. Pre-Cenozoic examples are avoided because the evidence for subduction initiation is more obscure and reconstructions more conjectural the farther back in the geologic record we look.

Mode of initiation cannot be inferred from a subduction zone once it becomes ‘self-sustaining’ [22], but the two modes of subduction initiation should have distinctly different beginnings. For induced nucleation of a subduction zone (INSZ), the plates are already converging before the subduction zone forms. In one of its simplest forms, INSZ results from continued plate convergence after a subduction zone fails due to collision, the attempted subduction of buoyant crust [23]. A variation on this theme is when a component of convergence begins along a transform plate boundary as a result of a change in location of the associated plates’ Euler pole. Fig. 2 shows two ways that continued plate convergence may yield a new subduction zone. In the case labelled ‘Transference’, a buoyant crustal block enters the subduction zone and causes it to fail, and in the process is sutured to the original hanging wall of the subduction zone. Evidence of this event is preserved as a tectonic suture and an accreted terrane. Plate convergence may continue as a result of lithosphere sinking elsewhere along strike of the convergent plate margin, with the result that the oceanic lithosphere outboard of the collision zone ultimately ruptures to become a new subduction zone. The new site of subduction is transferred away from the collision site. A good example of this process can be seen in the formation of the Mussau Trench in the Western Pacific (Fig. 1), which may be forming in response to collision of thick crust of the Caroline Ridge with the Yap Trench farther west [24]. The Pacific plate continues to move west relative to the Philippine Sea plate and this convergence is partially accommodated by thrusting of the Pacific plate over the Caroline plate. Hegarty et al. [25] infer that about 10 km of Caroline lithosphere has been thrust beneath the Pacific plate. There is seismicity and structural evidence for thrust faulting but no arc-related igneous activity. Subduction-related igneous activity may commence if shortening continues and the Caroline plate descends to asthenospheric depths (magmatic arcs are associated with subduction zones beneath them at a depth of 65–130 km [26]).

A more problematic example of transference comes from the collision of India with Eurasia. India first collided with Eurasia in Eocene time but continues to converge. This caused widespread deformation, including thrusting in the Himalayas, crustal thickening beneath Tibet and ‘tectonic escape’ of portions of Eurasia to the southeast of the collision zone. A new subduction zone outboard of the colliding blocks has not formed but appears to be in the making. There has been a broad zone of deformation and seismicity in the Indian Ocean between India and Australia since late Miocene time (8.0–7.5 Ma) [27] (stippled zone in Fig. 1). In contrast to most intraplate regions, many earthquakes of magnitude 6 and 7 occur there, and that region may someday evolve into a new subduction zone.

The fact that a new subduction zone has not yet formed outboard of the Indo–Asian collision emphasizes the difficulty of nucleating subduction zones in old, cold oceanic lithosphere. It takes less work to deform large parts of continental Asia than it does to

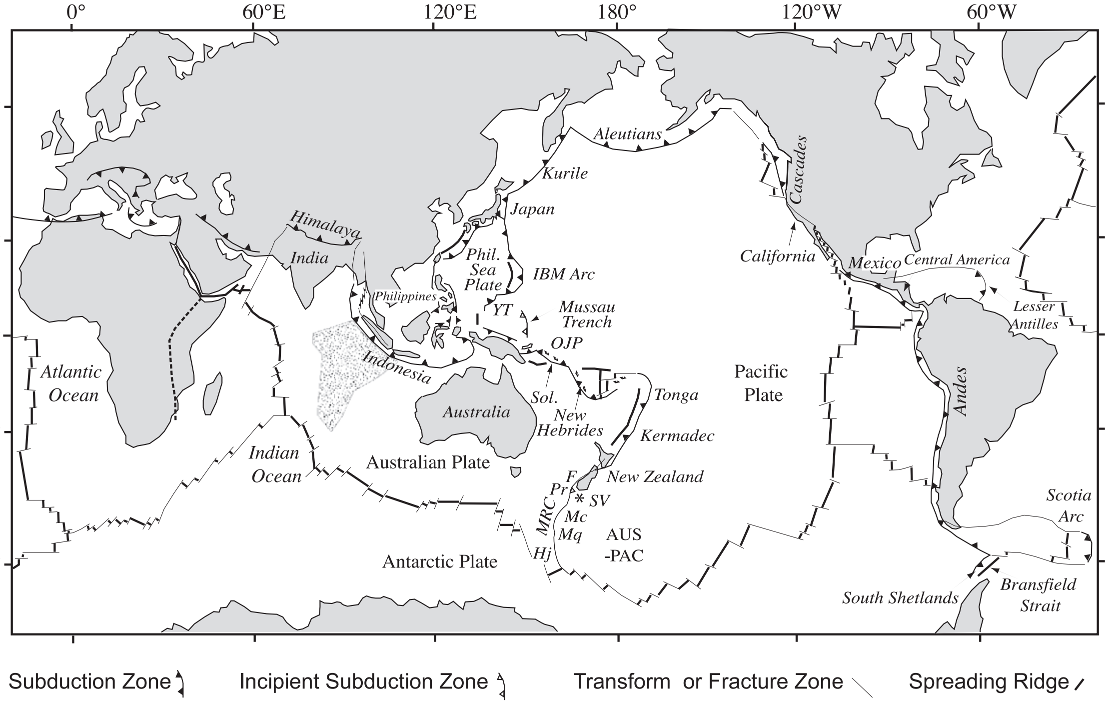

Fig. 1. Map showing location of subduction zones and where new subduction zones described in text have formed or are forming. YT=Yap Trench, Sol.=Solomon Arc, OJP=Ontong-Java Plateau. Stippled area in Indian Ocean between India and Australia is diffuse plate boundary of [88]. Area south of New Zealand is part of the MRC and Fiordland (F), including associated deeps: Puysgur (Pr), McDougall (Mc), Macquarie (Mq) and Hjort (Hj). General location of the pole of rotation for the Australian and Pacific plates is also shown (AUS–PAC).

start a new subduction zone in the Indian Plate. This is consistent with geodynamic models [28,29], which indicate the difficulty of rupturing intact oceanic lithosphere by compression. Successful nucleation of a subduction zone seems to require a pre-existing fault or other zone of lithospheric weakness. Note

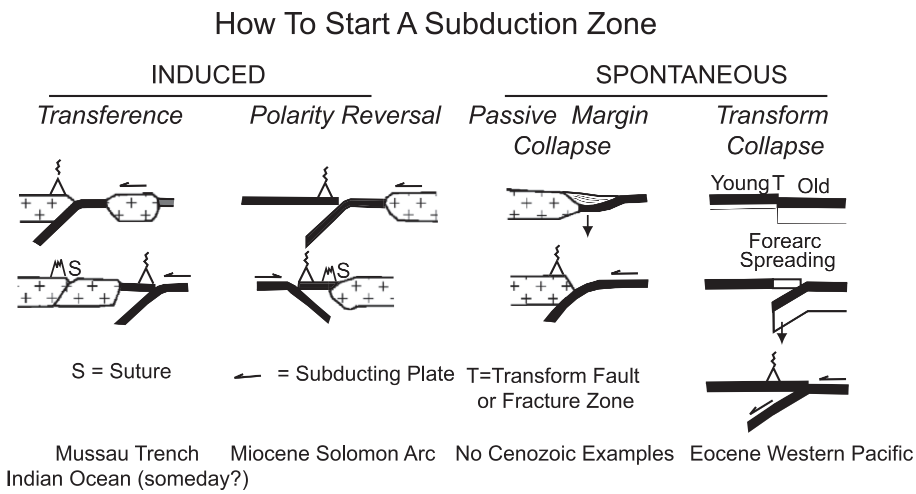

Fig. 2. General classes, subclasses and examples of how subduction zones form. See text for discussion.

that these models are solved in two dimensions, where orientation of the pre-existing fault is not important; in nature, the orientation of the fault with respect to the principal stress axis is very important. This may be the reason that a new subduction zone has not yet formed SE of India: no pre-existing faults have the proper orientation.

Toth and Gurnis [30] re-examined the issue of how subduction zones are born. Their numerical model imposed a convergence velocity of 2 cm/year on oceanic lithosphere with a fault that dipped 30°. Unless one plate is forced down at a rate of 1 cm/year or more, subduction is unlikely to become self-sustaining, because slower rates allow thermally induced density effects in the downgoing slab to dissipate. They also found that INSZ was accompanied by large amounts of uplift in the forearc, with the surface farther to the rear being strongly depressed. These predictions—early compression and uplift—are characteristics of INSZ. This response is also seen in physical models [31].

Another type of INSZ is polarity reversal (Fig. 2). Like transference, polarity reversal is triggered by buoyant crust entering the trench, but differs from transference, in that the new subduction zone nucleates in the originally overriding plate. The term 'polarity reversal' refers not only to the fact that the subduction zone dip changes direction, but also that the overriding and subducting plates exchange roles. The best Cenozoic example of INSZ polarity reversal is the Solomon arc, the result of the attempted subduction of the Ontong Java Plateau (OJP). OJP is the largest of the world's large igneous provinces, being the size of Alaska and with crust averaging 36 km thick [32]. This is much thicker than the ~20 km expected to result in subduction zone failure [32,33]. The OJP arrived at the Vitiaz Trench on the north side of the Solomon arc between 4 and 10 Ma ago [34,35], causing this south-dipping subduction zone to fail. Continued convergence between the Australian and Pacific plates forced a new subduction zone along the south flank of the Solomon Islands ~4 Ma ago [34,36].

The geodynamics of INSZ polarity reversal for arcs have not been quantitatively modeled, but upper plate responses are likely to be similar to those modelled for INSZ transference. Early stages should show upper plate uplift, thrusting and other expressions of compression. The major difference may be that, because back-arc regions have elevated thermal regimes as a result of subduction-related advection of hot asthenosphere, this lithosphere should be relatively thin and weak. Consequently, polarity reversal INSZ should require less compression than transference INSZ because young, thin lithosphere of back-arc regions should rupture more readily than the cold, thick and strong oceanic lithosphere outboard of the trench. This suggestion is supported by the fact that subduction on the south side of the Solomons began almost immediately upon arrival of the OJP at the Vitiaz Trench, whereas the Mussau Trough and mid-Indian plate deformation zones are still in relatively early stages of subduction zone formation.

The continuing reorganization of the Pacific–Australian plate boundary south of New Zealand is a good example of INSZ. The tectonics of this boundary are similar to that discussed by Casey and Dewey [37], whereby relatively minor adjustments in relative plate motion convert a transform fault into a convergent plate boundary. The effect is magnified the closer the transform is to the original pole of rotation, which in the case of the Macquarie Ridge Complex (MRC; discussed below) is only 1500 km or so distant (Fig. 1). With this geometry, small changes in the location of the Euler pole have large effects on the tectonic expression of the plate boundary. The reasons for this reorganization are not known, but it may be a far-field response to collisions along the northern margin of the Australian plate.

The Australian and Pacific plates are now obliquely converging. In the Puysegur Ridge region, just south of New Zealand (Fig. 1), the NUVEL-1 plate motion model predicts ~3–3.5 cm/year relative plate motion trending N50–60°E. This is 30–45° oblique to the N15–20°E trending plate boundary [38]. The plate boundary is locally changing from a dextral strike-slip transform fault into a subduction zone [39]. A region of deformation, the MRC, parallels the plate boundary. The MRC consists of a shallow ridge and adjacent deeps (Puysegur, McDougall, Macquarie and Hjort deeps) and is continuous from Fiordland of southernmost New Zealand (Fig. 1) to the triple junction in the south. Earthquake focal mechanisms and epicenters give strike-slip solutions except under Fiordland and the adjacent Puysegur Ridge region, which have thrust-type earthquakes [40].

The Puysegur Ridge in particular preserves good evidence for a transpressional strike-slip environment that is evolving into a subduction zone. There is a trench in the west, where the Australian plate underthrusts the Pacific plate [41]. The Puysegur Trench has an average depth of 5500 m, with a maximum of 6300 m. The trench shallows progressively southwards, terminating at 49°50′ S. Fifty kilometers or so to the east, Puysegur Bank is cut by an elongate trough containing the strike-slip Puysegur Fault. Thrust and strike-slip systems partition boundary-parallel and boundary-normal components of oblique convergence. Beneath Fiordland, intermediate-depth earthquakes define a Wadati-Benioff zone that dips steeply east [42]. The deepest earthquakes at 150 km mark the northern end of the Puysegur–Fiordland subduction zone. Solander volcano on the northern Puysegur Ridge may be the magmatic expression of the subducted Australian plate descending below the ~125-km threshold depth for arc magmatism. At least 350 km of the Australian plate has been subducted at the Puysegur Trench [43]. Given this slab length and assuming that convergence was constant at 3–3.5 cm/year, subduction began at about 10–12 Ma; this is consistent with the region's history of uplift [44]. The Puysegur subduction system may be propagating southward from a mature subduction zone near New Zealand to an incipient stage farther south [45].

The sequence of events associated with INSZ along the Puysegur Ridge is well preserved and includes a record of uplift and wave-cut planation of the overriding plate followed by subsidence to present water depths of down to 1.5 km. This sequence qualitatively agrees with geodynamic models for INSZ, which predict the sequence compression-uplift-subsidence as lithospheric thrusting progresses to true subduction [22].

4. Spontaneous nucleation of subduction zones

A recent comprehensive overview of subduction initiation from the perspective of numerical modelling by Gurnis et al. [22] concludes that subduction initiation appears to have been forced in all known Cenozoic cases. These arguments depend heavily on a variety of assumptions, most importantly for lithospheric rigidity. The validity of this conclusion for pre-Cenozoic subduction zones is suspect from the perspective of the chicken and the egg: subduction must have begun spontaneously at least once without a priori plate convergence in order to begin the plate tectonic regime. The following discussion focuses on geologic evidence in support of the spontaneous nucleation of subduction zones (SNSZ). SNSZ begins when old, dense lithosphere spontaneously sinks into the underlying asthenosphere. Lithospheric collapse at first does not lead to a change in plate motion, but at some point sinking lithosphere develops a down-dip component of motion that pulls the plate towards the subduction zone; this is when true subduction begins and plate motion changes. In contrast to INSZ, SNSZ affects the entire plate margin. Because oceanic lithosphere gets stronger as well as denser with age, SNSZ requires a lithospheric weakness—such as a fracture zone—to overcome lithospheric strength and allow collapse.

SNSZ can best be distinguished from INSZ in the rock record by the sequence of early events preserved in the upper plate. Whereas early INSZ events are strongly compressional, early SNSZ events are strongly extensional. This dichotomy is dictated by the distinctly different ways in which subduction zones nucleate in the two general cases: INSZ by plate convergence leading to lithospheric failure and SNSZ by lithospheric failure leading to plate convergence.

Geodynamic models are not yet able to dynamically model SNSZ but provide important constraints and perspectives. The following paragraphs are intended to demonstrate the likelihood that SNSZ is an unusual yet critical process. This includes presenting the geologic and geodynamic evidence against passive margin collapse, emphasizing similarities between lithospheric delamination and SNSZ, and exploring the significance of ophiolites. Finally, we revisit the formation of the Izu-Bonin-Mariana (IBM) and Tonga-Kermadec convergent margins in mid-Eocene time as the best example of SNSZ in the Cenozoic record.

4.1. Passive margin collapse

Conversion of passive margins into subduction zones is a hypothesis that is more widely accepted by the geoscientific community than the evidence warrants. There are no Cenozoic examples of a passive continental margin transforming into a convergent margin, in spite of the fact that seafloor adjacent to SE Africa, NW Africa and eastern N. America is about 170 Ma old. The broad—and largely uncritical—acceptance that this hypothesis enjoys mostly reflects the fact that the conversion of a passive margin into an active margin is required for the closing phase of the 'Wilson cycle', the name given to the repeated opening and closing of ocean basins [46,47]. It will be shown later that, if the third dimension is considered, the location of the subduction zone responsible for closing the ocean can be quite different than suggested by 2D tectonic cartoons.

Because of the central role that passive margin collapse is thought to play in the Wilson cycle, how this could happen is of considerable interest to geodynamicists. Models for this entail both INSZ and SNSZ, although the reversal of plate motion implicit in the Wilson cycle concept would seem to require SNSZ. This is also difficult to reconcile with our understanding of lithospheric strength. Cloetingh et al. [48] concluded that transformation of passive continental margins into active plate margins was unlikely because by the time that the adjacent oceanic lithosphere becomes dense enough to founder, it is too strong to fail. Their models indicated that the overall strength of old oceanic lithosphere at passive margins is too great to be overcome, even with the additional stresses of sediment loading. This conclusion must be modified because water weakens lithosphere [49]. Situations in which the strength of old, dense passive margin lithosphere could be overcome were examined by Erickson [50]. Using elastic-plate bending theory, he concluded that sediment loading could cause the oceanic lithosphere to collapse if a previous zone of weakness such as a fault was reactivated, possibly as a result of a change in plate motion. Another way to overcome lithospheric strength was modeled by Kemp and Stevenson [51], who noted that lithosphere is significantly weaker under tension than compression. They argued that oceanic and continental lithospheres could be stretched even in an environment of moderate compression (due to ridge push) as a result of greater subsidence of oceanic lithosphere just outboard of the continent-ocean lithosphere boundary. Tensional stress could lead to rifting between continental and oceanic lithospheres to allow old, dense oceanic lithosphere to sink asymmetrically and asthenosphere to well up beneath the rift and flood over the sinking oceanic lithosphere. Note that both of these studies [50,51] emphasize early sinking, early extension or both to overcome plate strength.

These models need refinement because we now know that 'transitional crust' between continent and ocean can be quite variable. Transitional crust can range in composition and strength from 'volcanic rifted margins' (VRM) with unusually thick sections of strong basaltic crust [52], to amagmatic rifted margins (ARM) with abundant serpentinite dominating a weak crust [53]. It is likely that ARM lithosphere is weak due to abundant faults and serpentinites, whereas VRM lithosphere should strongly weld oceanic and continental lithospheres. The differing magma budgets of these two crustal types indicate that associated lithosphere will also vary from depleted mantle beneath VRMs and undepleted mantle beneath ARMs. It seems likely that the diminished lithospheric strength and denser nature of undepleted (Fe-rich) mantle beneath ARMs favor these over VRMs as sites of passive margin collapse. Similar comments apply to the "oceanic plateau model" for subduction initiation [54]. Regardless of these subtleties, extreme skepticism is warranted regarding the efficacy of passive margin collapse SNSZ, in light of geodynamic considerations and the absence of Cenozoic examples.

4.2. Similarities between lithospheric delamination and SNSZ

The SNSZ models of Erickson [50] and Kemp and Stevenson [51] have strong similarities to the hypothesis of lithospheric 'delamination' as first advanced to explain uplift of the Colorado Plateau [55]. Bird argued that subcontinental mantle lithosphere was gravitationally unstable relative to underlying asthenosphere. He concluded that delamination occurs when asthenospheric mantle can rise into and above the level of the lithosphere. Once this happens, the lithosphere will peel away and sink, resulting in '. . .wholesale replacement of cold mantle by hot in a geologically short time. . .' ([55], p. 7561). In spite of uncertainties, the overall concept is broadly accepted, mostly because it explains so many otherwise inexplicable observations. These include uplift of the Colorado and Tibetan plateaux [56,57] and the Altiplano [58], inclined seismic zones beneath continental interiors that are not associated with active subduction [59,60], and changes in melt source from lithospheric mantle to asthenosphere accompanying rapid uplift [61].

The concept of delamination continues to evolve, with one mechanism involving the 'peeling off' and sinking of lithospheric mantle as envisaged by Bird and another the sinking of gravitationally unstable mantle and lower crust by Rayleigh-Taylor instability [62]. Nevertheless, the general process whereby denser subcontinental lithosphere can sink into underlying asthenosphere provides a useful analogy for SNSZ because it supports the idea that gravitationally unstable lithosphere can collapse into underlying asthenosphere.

4.3. Ophiolites, the Izu-Bonin-Mariana forearc and SNSZ

Consensus exists that ophiolites are fragments of oceanic lithosphere obducted onto continental crust, but the tectonic setting in which they form is often controversial. In the late 1960s and early 1970s, ophiolites were thought to have originated at mid-ocean spreading ridges but geochemical evidence increasingly compels the conclusion that most ophiolites form above subduction zones [63,64]. Until recently, such ophiolites were generally ascribed to back-arc basin settings, principally in order to reconcile 'supra-subduction zone' (SSZ) chemical compositions with crustal structure indicating seafloor spreading. This combination is found among active systems solely in back-arc basins. However, the following discussion argues that most—if not all—well-preserved ophiolites formed in forearcs during the early stage of SNSZ, and explores the implications of this conclusion for understanding SNSZ.

The evolution of thinking about SSZ ophiolites is summarized by Pearce [64], who coined the term and who has long played a key role in understanding them. This somewhat non-specific term acknowledges that the chemical characteristics of these igneous rocks indicate subduction zone magmatism but does not specify the precise tectonic setting. The term is defined [65] thus: "supra-subduction zone (SSZ) ophiolites have the geochemical characteristics of island arcs but the structure of oceanic crust and are thought to have formed by sea-floor spreading directly above subducted oceanic lithosphere. They differ from 'MORB' ophiolites in their mantle sequences, the more common presence of podiform chromite deposits, and the crystallization of clinopyroxene before plagioclase which is reflected in the high abundance of wehrlite relative to troctolite in the cumulate sequence. Most of the best-preserved ophiolites in orogenic belts are of this type". Subsequent studies confirm and expand on these observations, such as the fact that boninites—high-Mg andesites formed by hydrous melting of harzburgite—are common in some ophiolites, but are unknown from modern spreading ridges or older seafloor and do not erupt today.

The IBM arc system in the western Pacific is increasingly recognized as key for understanding how SSZ ophiolites form. This reflects the fact that the early record of this convergent margin is well preserved in IBM forearc crust, and that this crust is well exposed with limited sediment cover and no accretionary prism [66]. Forearc crust is readily studied on island exposures as well as in submerged realms by drilling, dredging and diving on the inner wall of the trench; serpentinite mud volcanoes bring up fragments of forearc and crust [67]. DSDP and ODP drilling in the 1970s and 1980s encouraged confidence that the limited exposures of pillowed tholeiites and boninites of Eocene age on islands in the Bonin and S. Mariana arc segments along with peridotite-gabbro-metavolcanic exposures in the IBM trench are representative of lithospheric structure for the entire IBM forearc [68] (although minor amounts of trapped older crust have been identified [69]). A consensus exists that the IBM forearc was created in a SSZ environment about the time that subduction began along the IBM trench [68]. Numerical models reveal the sequence of events leading to upper plate spreading as subduction begins [22,70]. A number of important controversies persist, including whether or not a transform fault was in the proper position and orientation to be converted into a subduction zone [70–72]. The bend in the Emperor–Hawaii seamount chain, which formed at about 43 Ma was an important if indirect argument for IBM SNSZ [12], but this may also record southward movement of the Hawaiian hotspot [73]. It was also suggested that boninites are abundant in the IBM forearc because subduction initiation occurred on the flanks of a mantle plume [74]. Geodynamic modelling results led to conclusions that IBM subduction was induced, not spontaneous [22,70].

In spite of these and other uncertainties, the IBM arc system is the best example of how subduction zones form by a process that entails abundant igneous activity in a broad region of the forearc. It is important to note that limited data indicate that the Tonga-Kermadec convergent margin began about the same time as the IBM arc system, about 45–50 Ma [75], satisfying a critical test of the SNSZ model: that subduction begins all along the 'downstream' margins of the affected plate (in this case, the Pacific plate) at about the same time. It should be noted that this does not preclude the possibility of INSZ for these systems, which could also cause plate-scale initiation of subduction.

The point to be stressed is that the IBM and Tonga-Kermadec forearcs represent ophiolites waiting to be obducted. These 'all-but' ophiolites have the structure, crystallization sequence and chemical composition observed in well-preserved ophiolites, and are in a tectonic setting that favors obduction of coherent lithospheric slices once a continent or other tract of buoyant crust clogs the subduction zone (Fig. 3). I extrapolate from this that most well-preserved SSZ ophiolites represent forearc lithosphere that formed during SNSZ events and which are often emplaced during collision events. Shervais [76] explicitly identified the classic Mesozoic ophiolites of Cyprus, Oman and California as representing forearcs that formed during episodes of subduction

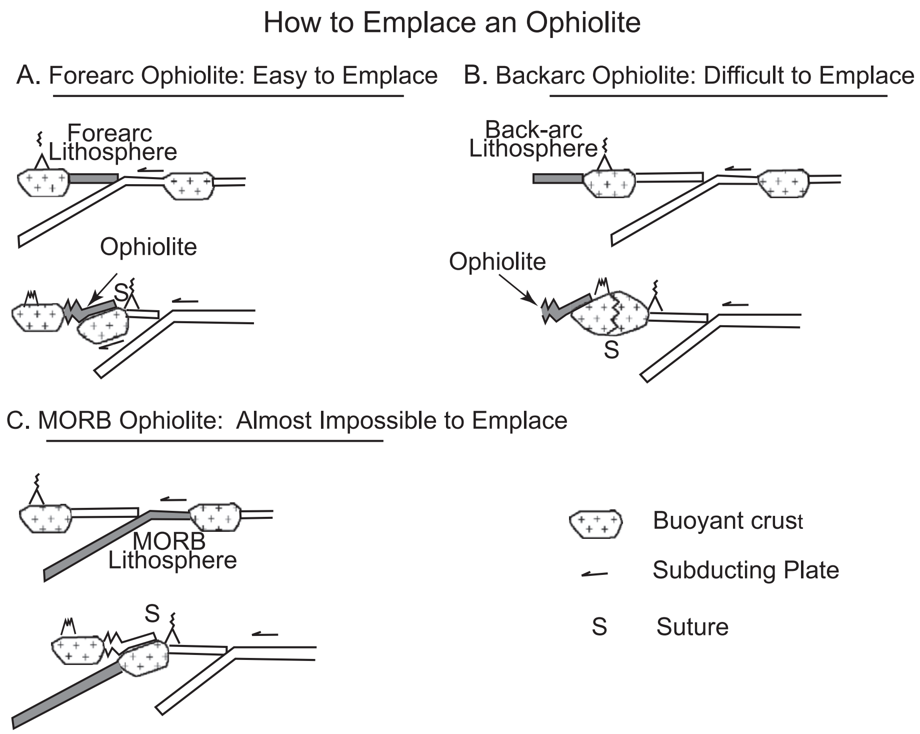

Fig. 3. Cartoon showing the relative feasibility of emplacing oceanic lithospheres created in forearc, back-arc basin and mid-ocean ridge settings. (A) It is relatively easy to emplace oceanic lithosphere of forearcs. Subduction of buoyant material leads to failure of subduction zone. Isostatic rebound of buoyant material follows, with ophiolite on top. (B) It is difficult to emplace back-arc basin oceanic lithosphere. Compression and shortening across the arc will lead to uplift of the arc. (C) It is almost impossible to emplace true MORB crust at a convergent plate boundary. Sediments and fragments of seamounts may be scraped off of the downgoing plate, but decollements do not cut deeply into the subducting lithosphere.

initiation. He argued that these and most—but not all—ophiolites form as outlined in the 'subduction zone infancy' model of [77]. This is shown in the right panel of Fig. 2 and details are outlined in Fig. 4. Some object to Tethyan ophiolites being infant arc products because they are not associated with a magmatic arc [78]. However, the portion of the Tethys being subducted was relatively narrow, so that collision occurred and subduction was arrested before the sinking slab reached the threshold depth (60–135 km) for mature arc magmagenesis.

4.4. Lithospheric collapse along transform faults and SNSZ

The model for SNSZ of the IBM arc system was first suggested by Natland and Tarney [79] and Hawkins et al. [80] and developed more fully by Stern and Bloomer [77]. Two lithospheres of differing density are juxtaposed across a lithospheric weakness, either a transform fault or fracture zone. Differences in density, elevation of the seafloor and depth to asthenosphere for the juxtaposed lithospheres are

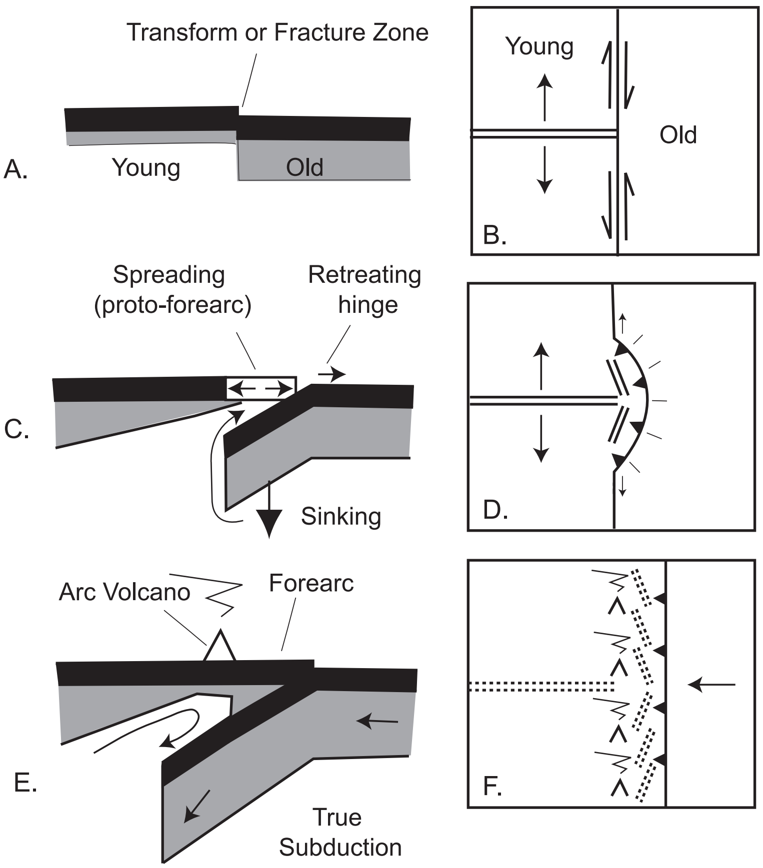

Fig. 4. Subduction infancy model of [77], modified to show the third dimension. Left panels are sections perpendicular to the plate boundary (parallel to spreading ridge) and right panels are map views. (A) and (B) show the initial configuration. Two lithospheres of differing density are juxtaposed across a transform fault or fracture zone. (C, D) Old, dense lithosphere sinks asymmetrically, with maximum subsidence nearest fault. Asthenosphere migrates over the sinking lithosphere and propagates in directions that are both orthogonal to the original trend of the transform/fracture zone as well as in both directions parallel to it. Strong extension in the region above the sinking lithosphere is accommodated by seafloor spreading, forming infant arc crust of the proto-forearc. (E, F) Beginning of down-dip component motion in sinking lithosphere marks the beginning of true subduction. Strong extension above the sunken lithosphere ends, which also stops the advection of asthenosphere into this region, allowing it to cool and become forearc lithophere. The locus of igneous activity retreats to the region where asthenospheric advection continues, forming a magmatic arc.

maximized where a spreading ridge intersects the transform, and this is where the process shown in Fig. 4C and D is most likely to begin. Sinking of old, dense lithosphere is accomplished by asymmetric or 'hinged' subsidence, accelerated by mass wasting of shallower crust adjacent to the transform as structural relief across this feature increases. Recent geodynamic modeling of this process requires a small amount of convergence (~2 cm/year) to overcome plate strength to cause subduction [70]. Propagation of the spreading ridge across the transform fault may accelerate depression of dense lithosphere adjacent to the fault. When the top of the sinking lithosphere is depressed beneath the level of the asthenosphere, the latter floods over the former, further accelerating the process. The pool of asthenosphere spreads laterally as it advances over the sinking lithosphere in directions that are both orthogonal to the original trend of the transform/fracture zone as well as parallel to it (Fig. 4C,D). The rate of propagation of the asthenospheric wedge is unknown but may be on the order of propagation rates of ~5 cm/year, as suggested for lithospheric delamination [55].

Geodynamic models show that asthenosphere wells up as the lithosphere continues to sink [70]. Upwelling asthenosphere can melt due to decompression alone, and this will be stimulated by water from the sinking lithosphere. Increasing pressure and temperature on the sinking plate will result in a massive expulsion of water, setting up conditions for unusually high extents of melting. Boninites (harzburgite melts) and unusually depleted arc tholeiites (lherzolite melts) and their fractionates are the hallmarks of this phase, although normal MORB-like tholeiites may also be produced in drier regions of upwelling mantle. Hinged subsidence of sinking lithosphere results in an environment of crust formation in the hanging wall that is indistinguishable from seafloor-spreading at mid-ocean ridges. Melting will be unusually extensive during this phase, and residual mantle will be unusually depleted harzburgite, characterized by Cr-rich spinel, such as that now found in peridotite exposures of inner trench walls [81] and recovered from serpentine mud volcanoes of the IBM forearc [82]. Similar lava and peridotite compositions are characteristic of the great, well-preserved SSZ ophiolites, such as Cyprus, Oman and California [75].

The phase of hinged subsidence of dense lithosphere beneath an expanding wedge of upwelling asthenosphere associated with seafloor spreading defines the infant arc phase. The distribution, orientation and length of spreading ridges associated with the 'proto-forearc' are speculatively shown in Fig. 4D aligned obliquely to the evolving trench and in an en-echelon pattern to account for along-strike expansion of the infant arc or proto-forearc. The infant arc phase may continue for 5 or 10 million years. Strong extension will exist in the region above the sinking lithosphere as long as the hinge continues to retreat rapidly. This will continue until the sinking lithosphere develops a significant component of down-dip motion, when true subduction begins (Fig. 4E,F). It is not clear what causes the change from hinged subsidence to true subduction; Stern and Bloomer [77] slab speculated that this may reflect the increasing difficulty of transferring asthenosphere from beneath and around the subsiding lithosphere, with down-dip slab motion possibly aided by transformation of basaltic crust into eclogite [83].

Once true subduction begins, the hingeline stops retreating (or 'rolls back' more slowly, typically at ~10% of the plate convergence rate [84]) and extensional forces in the forearc diminish (Fig. 4E,F). Asthenosphere is no longer advected beneath the proto-forearc and seafloor spreading ceases there. The subjacent mantle is conductively cooled by the subducting slab and becomes lithosphere. Flow of asthenosphere must reorganize during this transition. Migration of asthenosphere from beneath the sinking lithosphere is cut off and is supplied instead from convection induced beneath the overriding plate. The magmatic axis migrates away from the trench to form a fixed (relative to the trench) magmatic arc and vertical edifice building replaces spreading. The mantle beneath the forearc quickly cools following the establishment of a true subduction regime.

The general model outlined above explains the origin of the IBM subduction zone and reconciles the structure, composition and emplacement of most ophiolites. The model suffers from the fact that this style of subduction initiation is not now occurring and so cannot be studied directly (although the eastern part of the Mendocino Fracture Zone may be in the early stages of this evolution). The model also suffers because much of the geodynamic community thinks the oceanic lithosphere is too strong to flex as called for in the model, and the concept of the 'retreating hingeline' is especially difficult to reconcile with our understanding of lithospheric strength. Numerical models for subduction initiation require plate convergence in order to overcome the resistance of the plate to bending [22,70]. The objection based on lithospheric strength may not be fatal, because oceanic lithosphere may be weakened as a result of serpentinization accompanying reactivation of deep faults by down-flexing [85,86]. Normal faulting and concave-downward flexure of the lithosphere open cracks, which must be filled with seawater, promoting serpentinization of lithospheric peridotites and further weakening the lithosphere. Evidence for this process is seen in deep earthquakes in the outer rise associated with several circum-Pacific trenches and in associated double Benioff Zones [87]. Repetition and acceleration of the sequence flexure-fracture-serpentinization may so weaken the lithosphere that it is much easier to bend than present estimates of plate strength allow.

5. Implications for reconstructing subduction initiation in the geologic record

The forgoing discussion contrasts two fundamentally different ways that subduction zones form, which can be distinguished by whether or not the pertinent plates were converging prior to nucleation of the subduction zone. Alternatively, if the interior of a single plate ruptures (as is the case in transference INSZ), the question is whether or not strong compressional stresses preceded subduction initiation. In the case of INSZ, the plate convergence or intra-plate compression precedes subduction nucleation, so that subduction nucleation localizes shortening where lithospheric strength can be overcome. SNSZ contrasts with INSZ in that plate convergence may not precede the initiation of a subduction zone. Rather, the lithosphere descends many kilometers into the mantle before the pertinent plates begin to converge. In order to understand these fundamentally different modes of formation, fundamentally different geodynamic models must be created, and these models must be tested against our understanding of the early history of a convergent margin.

Unfortunately, it is difficult to determine relative plate motions before subduction nucleation from the geologic record. The continuing controversy about W. Pacific plate paleogeography and the significance of the bend in the Emperor–Hawaii Seamount chain are good examples of such uncertainties in the relatively robust Cenozoic record. These examples do not encourage confidence that we can be sure enough about past relative plate motion and plate margin orientations to allow us to distinguish ancient INSZ from SNSZ. We may need to rely instead on the expected consequences of the two subduction initiation modes preserved in the upper plate. Because INSZ results from compression, this mode results in early reverse faulting, folding, and uplift of the overriding plate as one plate is thrust over the other. This is predicted from geodynamic numerical experiments as well as physical modelling and is observed in late Cenozoic examples such as the MRC. As a starting point, all examples of subduction initiation where the early record in the overriding plate is one of compression, thrust faulting, folding and/or uplift should be interpreted as reflecting INSZ. In contrast, because SNSZ results from gravitational instability, its earliest expressions should reflect subsidence of the descending lithospheric plate. Again, as a starting point, all examples of subduction initiation where the early record in the overriding plate is one of extension, normal faulting, seafloor spreading and/or subsidence should be interpreted as manifesting SNSZ.

Two additional points need to be emphasized. First, most large, well-preserved ophiolites represent lithosphere produced during the infant arc phase of SNSZ. This is the simplest way to reconcile ophiolite chemistries with seafloor spreading structures in a setting that favors obduction. The classic ophiolites of Cyprus and Oman in particular should be regarded as infant arc lithosphere, which were emplaced when continental crust of Africa and Arabia slid underneath Eurasia [76]. Second, the identification of an ophiolite as manifesting SNSZ does not distinguish whether the lithosphere collapsed beneath a passive margin or along a transform fault. This requires an understanding of paleogeography that is often lacking. On a global scale, however, episodes of SNSZ due to passive margin collapse should result in Wilson cycles where the paths of the plates and continents during the closing phase parallel (in reverse) their paths during the opening phase. In contrast, SNSZ due to transform margin collapse should result in closing phases orthogonal to the opening phase. Examination of the first-order tectonic changes that the Earth has witnessed over the past 120 million years suggests that an overall N–S motion of the continents around an opening and closing Tethys has been supplanted by an overall E–W motion today (Fig. 5). This suggests that the most recent mode of major SNSZ was lithospheric collapse along transform margins.

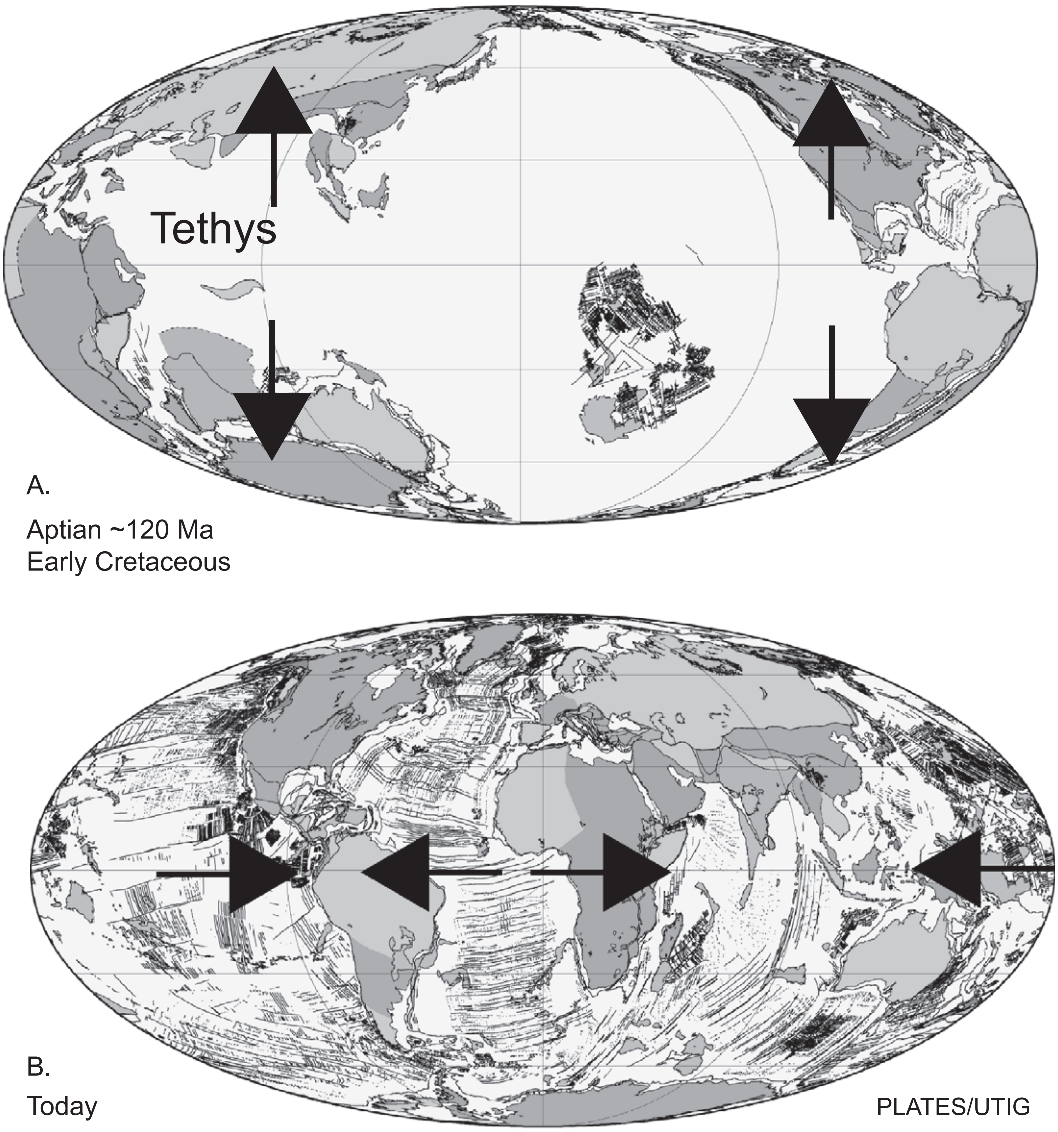

Fig. 5. Contrasting motions of the plates, early Cretaceous and now. (A) Aptian–Cretaceous (120 Ma). Plates are to a first approximation moving N–S, associated with the opening of the Tethys seaway. (B) Present. Plates are to a first approximation moving E–W. Figure courtesy of Lisa Gahagan, UT Institute of Geophysics.

6. Conclusions

We have made great advances towards understanding how INSZ occurs and what are the responses in the upper plate. Quantitative understanding of SNSZ is needed, particularly to constrain the conditions under which it can occur and how early stages of lithospheric foundering evolve to become true subduction. Such studies present great opportunities for understanding the significance of ophiolites, the history of plate motions and how the solid Earth system operates.

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