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Catastrophic initiation of subduction following forced convergence across fracture zones¹

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Abstract

Although the formation of subduction zones plays a central role in plate evolution, the processes and geological settings that lead to the initiation of subduction are poorly understood. Using a visco-elastoplastic model, we show that a fracture zone could be converted into a self-sustaining subduction zone after approximately 100 km of convergence. Modeled initiation is accompanied by rapid extension of the over-riding plate and explains the inferred catastrophic boninitic volcanism associated with Eocene initiation of the Izu-Bonin-Mariana (IBM) subduction zone. Using global plate reconstructions, we suggest that IBM nucleation was associated with a change in plate motion between 55 and 45 Ma. We estimate that the forces resisting IBM subduction initiation were substantially smaller than available driving forces.

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1. Introduction

The formation of new subduction zones is a fundamental, yet poorly understood, process in the normal evolution of tectonic plates. Processes that lead to the demise of subduction zones, such as the subduction of buoyant material (i.e. continental or young oceanic lithosphere), must ultimately result in the formation of new subduction zones. Several questions regarding subduction initiation are unanswered: what are the kinematic and dynamic relations between plate motions and the formation of new subduction zones? Do changes in plate motion precede or follow incipient subduction? What tectonic settings are most favorable for the nucleation of new trenches? What are the short term tectonic and dynamic consequences of incipient subduction?

In looking for answers to these questions, there are two stumbling blocks which have inhibited progress, one theoretical and one empirical. First, while events such as the opening and closure of

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ocean basins [1] suggest that subduction initiation is common, theoretical models suggest it should be quite difficult. For plate convergence to proceed at a subduction zone, the sum of the driving forces (plate tectonic forces, F_{pt} , and the slab's negative buoyancy, F_s) must exceed the net of resisting forces (fault friction, F_f , and elastic plate bending, F_{el}) [2], i.e.:

$$F_{\rm s} + F_{\rm pt} > F_{\rm f} + F_{\rm el} \tag{1}$$

Subduction becomes 'self-sustaining' when F_s alone is sufficient to overcome the resisting forces.

Using this approach, estimates of the force required to create a new subduction zone typically exceed the force available from ridge push [3]. Alternatively, assuming weaker faults one concludes that resisting forces are comparable to those driving plate motions [4]. The second factor impeding progress is that examples of the formation of subduction zones in the recent geological past are poorly exposed, and typically overprinted. For example, the Macquarie Ridge Complex (MRC) south of New Zealand may represent Late Miocene to present incipient subduction



Fig. 1. Tectonic setting of the Philippine Sea Plate, with free air gravity from [39]. Key to abbreviations: I = Izu arc, B = Bonin arc, M = Mariana arc, WPB = West Philippine Basin, KPR = Kyushu-Palau Ridge, CBCS = Central Basin Spreading Center, DR = Daito Ridge, ODR = Oki-Daito Ridge.

[5,6]. Some have argued that the northernmost segment once represented a nascent subduction zone but that it may have failed to transform into a self-sustaining system [7].

Although older than MRC, the Izu-Bonin-Mariana (IBM) trench (Fig. 1) overcomes these ambiguities as a type-locality for deciphering the causes and consequences of subduction initiation. First, there is little doubt that IBM is a self-sustaining subduction system: it represents deep mantle penetration of oceanic lithosphere [8] which is amongst the oldest [9] and potentially most negatively buoyant currently found. Second, the IBM trench has a distinct beginning in the Middle Eocene [10] at 49–48 Ma [11] and a clearly defined subsequent history [12,13]. Magnetic lineations of the West Philippine Basin (WPB) show that it is mostly composed of oceanic lithosphere which formed between 60 and 33 Ma at the now extinct Central Basin Spreading Center (CBSC) [14,15]. The Kyushu-Palau Ridge (KPR) separates older crust of the WPB from younger crust which formed during initiation and subsequent back-arc extension and arc volcanism of the IBM trench. The perpendicular orientation of the KPR to magnetic lineations formed at the CBSC suggests that these formed a ridge-transform intersection [16]. Although there are ambiguities, conversion of a transform boundary or fracture zone into an oceanic trench is the simplest explanation for the early history of the IBM system [17].

Two competing hypotheses have been formulated for the initiation of the IBM subduction zone. According to the first, changes in relative plate motion across transform faults offsetting the CBSC forced the formation of a new subduction zone [18]. According to the second, old oceanic lithosphere juxtaposing younger lithosphere at a fracture zone undergoes spontaneous convective instability driven by the localized difference in negative thermal buoyancy [13,19]. Theories for the origin of the IBM subduction zone often take into account the putative change in the direction of the Pacific Plate with respect to the Hawaiian hot-spot (i.e. the bend in the Hawaiian-Emperor island chain) at 43 Ma; Stern and Bloomer [13] specifically argue that the self-nucleation of the IBM arc caused the change in the Pacific Plate. Consequently, according to the self-nucleation hypothesis, changes in plate motion should post date initiation of subduction while according to the forced convergence hypothesis changes in plate motion precede initiation.

A limitation for testing these hypotheses has been the lack of realistic models for incipient subduction which simultaneously account for elastic plate flexure, viscous flow in the underlying mantle, and the growth of faults. For example, viscoelastic models for forced subduction initiation incorporated initially dipping faults [4], not vertical ones typical of fracture zones. Moreover, models for self-nucleation have assumed viscous rheologies with inadequate viscosity contrasts across the lithosphere and mantle [19], and also neglected the important role of elastic flexure [2]. Recent work incorporating some of these effects [20] focused on the development of a lithospheric scale shear zone in the early stages of subduction initiation at a continental margin. In contrast, we focus on hypotheses for incipient intra-oceanic subduction and have followed the evolution of a fracture zone to a self-sustaining subduction zone.

2. Modeling the evolution of fracture zones

2.1. Initial and boundary conditions

The models in this study are two dimensional and track the evolution of a fracture zone in cross section, where an offset in lithospheric age places a young plate against well-developed thermal lithosphere (Fig. 2). The numerical domains are 200– 300 km deep and initially 800–1000 km wide in most cases (Table 1), with the lithospheric age offset initially centered 300 km from the left side. The models have variable resolution, with a maximum of 1 km×1 km resolution in a 145 km wide by 60 km deep region centered on the initial location of the age offset. Minimum resolution is 2.25 km wide×2 km deep in the lower (>60 km deep) left and right corners of the domain.

The models' initial temperature conditions are



Fig. 2. Initial and boundary conditions of fracture zone models. Top: map view schematic showing where the cross section of the model (dashed line) lies with respect to a mid-ocean ridge offset. Middle: initial topography across a 40 Myr lithospheric age offset. Bottom: initial thermal conditions and location of the initial weak zone (box), when included.

given by the solutions for conductively cooled halfspaces. The two adjacent halfspaces are linearly smoothed over a few nodes at their interface. The mesh is perturbed near the free surface so that the initial topography of the plates is set by isostatic equilibrium. In some models, a weakened fault zone is initially placed between the two plates (Table 1).

We employ two sets of mechanical boundary conditions, depending on whether or not we are imposing relative motion between our model plates. In all models, the left and bottom boundaries have zero normal velocities and free tangential velocities, and the top boundary is a free surface. For models with no imposed relative plate motions, conditions for the right boundary are identical to the left. For imposed motion cases, the right boundary has zero vertical velocities and horizontal velocities are uniform but of opposite sign over the top- and bottom-most 50 km, with a cosine taper in between. This boundary condition drives a large scale counterclockwise circulation throughout the right half of the model. Therefore, we must demonstrate which model cases develop flow fields that become dominated locally by slab buoyancy forces (F_s), and in which cases flow fields remain dominated by boundary conditions.

2.2. Rheology

Domains of viscous, elastic, and plastic deformation are determined self-consistently and vary spatially and temporally due to the temperature, stress, and strain rate dependencies of each process. When deformation is visco-elastic, the total deviatoric strain rate is described by Maxwell visco-elasticity [21]: Table 1

List of numerical experiments, giving numerical grid dimensions, imposed convergence rates, and fault dimensions and plate ages at the beginning of each experiment

Case	Grid	Convergence rate	Age offset	Fault friction	Initial fault dimensions	Stable or sustaining subduction
	(width \times depth, km)	(cm yr^{-1})	(Myr)	$(\angle \phi)$	(km)	
1	1000×400	0	0-40	0	12×20	Stable 15 Myr
2	1000×400	0	0-40	0	0×0	Stable 7.5 Myr
3	900×600	0	0-40	0	65×20	Stable, plug delaminates
4	800×200	1	0–40	10	0×0	Stable to 5.5 Myr; young plate buckles
5	800×200	1	0-40	0	5×15	Sustained 140 km
6	800×200	2	0–40	0	5×15	Sustained 115 km
7	800×200	1	0-50	5	5×15	Stable to 165 km
8	800×200	1	0-20	0	5×15	Sustained 160 km
9	800×200	1	0–40	0	5×15	Stable to 170 km
10	1000×300	0	0-120	0	7×50	Stable to 5.5 Myr
11	1000×300	1	0-120	0	7×50	Sustained 125 km
12	800×200	1	0-40	0	0×0	Stable to 140 km
13	1000×300	0	0-40	0	6×32	Stable to 30 Myr
14	1000×300	1	0-20	0	6×14	Sustained 120 km
15	1000×300	2	0-40	0	6×14	Sustained 107 km
16	1000×300	2	0-80	0	6×14	Sustained 100 km
17	1000×300	2	0-40	5	6×14	Stable to 105 km
18	1000×300	2	0–20	0	6×14	Sustained 107 km
19	1000×300	0	0-60	0	6×32	Stable to 20 Myr
20	1000×300	2	10-40	0	6×14	Sustained 109 km
21	1000×300	2	20-40	0	6×14	
22	1000×300	2	10-20	0	6×14	Sustained 105 km
23	1000×300	2	0-40	0	5×15	Reference viscosity increased
						$\times 10$; stable to 200 km
24	1000×300	2	0–40	0	5×15	Fault cohesion 10 MPa; sustained 160 km
25	1000×300	2	0–40	0	5×15	Fault cohesion 20 MPa; stable to 200 km
26	1000×300	2	0-40	2	5×15	Stable to 172 km

Fault friction angle (ϕ) is defined in Section 2.2. Last column lists the convergence needed for self-sustaining subduction, or if stable, the duration of the numerical experiment.

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$$\dot{\boldsymbol{\varepsilon}}_{ij}^{\mathrm{d}} = \frac{1}{G} \dot{\boldsymbol{S}}_{ij} + \frac{s_{ij}}{2\eta} \tag{2}$$

where s_{ij} is the deviatoric stress tensor, *G* is the shear modulus, and η is viscosity (Table 2). The isotropic (non-deviatoric) deformation is modeled with a constitutive law using the total stress and strain tensors:

$$\varepsilon_{ii} = \frac{\sigma_{ii}}{3K} \tag{3}$$

where K is the bulk modulus (Table 2). Viscous deformation is incompressible and based on non-Newtonian temperature-dependent creep of olivine [22], with viscosity given by:

Table 2		
Rheological	model	parameters

Symbol	Description	Model value				
η_0	Reference viscosity	5.0×10 ¹⁹ Pa s				
$\dot{oldsymbol{arepsilon}}_0$	Reference strain rate	10^{-15} s^{-1}				
N	Strain rate exponent	3.5				
T_0	Reference temperature	1400°C				
H	Activation energy	5.4 kJ mol^{-1}				
R	Gas constant	8.31 J mol ⁻¹				
		K^{-1}				
ϕ	Friction angle, unstrained	30°				
C	Cohesion	44 MPa				
C_{f}	Fault cohesion	4 MPa				
Κ	Bulk modulus	50 GPa				
G	Shear modulus	30 GPa				
v	Poisson's ratio	0.25				

Definitions are in Section 2.2.

$$\eta = \eta_0 \left(\frac{\dot{\varepsilon}_0}{\dot{\varepsilon}^d}\right)^{(1/n)-1} \exp\left[\frac{H}{R}\left(\frac{1}{T} - \frac{1}{T_0}\right)\right]$$
(4)

where $\dot{\boldsymbol{\varepsilon}}^d$ is the second invariant of the deviatoric strain rate tensor (Table 2).

The onset of irreversible plastic yielding is determined by expressing a yield potential f in terms of the major (σ_1) and minor (σ_3) principal stresses [23], leading to:

$$f = \frac{1}{2}(\sigma_3 - \sigma_1) + \frac{1}{2}(\sigma_3 + \sigma_1)\sin(\phi) - C\cos(\phi)$$
 (5)

where ϕ is the angle of friction, *C* is cohesive strength, and tensional stresses are positive. (For Mohr-Coulomb plasticity, employed here, the coefficient of friction, μ , is equal to the tangent of the friction angle, ϕ .) Deformation remains elastic for f < 0. Plastic strain rates are determined as the gradient of a separate flow potential:

$$\dot{\boldsymbol{\varepsilon}}_{i}^{p} = \lambda \frac{\partial g}{\partial \sigma_{i}} \tag{6}$$

where λ is a rate constant. The plastic flow potential is given by:

$$g = \frac{1}{2}(\sigma_3 - \sigma_1) + \frac{1}{2}(\sigma_3 + \sigma_1)\sin(\psi)$$
(7)

where ψ is the angle of dilatancy. In all of the cases presented here, we model incompressible plastic flow with ψ set to zero. Plastic strain results in a linear reduction of cohesion and friction angle up to a cumulative plastic strain ε_{ps} , at which point faults do not weaken further.

The numerical model is based on the FLAC (Fast Lagrangian Analysis of Continua) method [24]. The method's main strength is that it follows an explicit formulation in which complex non-linear rheologies are easily incorporated, and is described in detail elsewhere [25,26].

3. Plate boundary evolution at transform boundaries

3.1. Models with no applied plate motion

We explored the conditions critical for the development of self-sustaining subduction (Table 1). All of our models without imposed convergence fail to develop a lithospheric-wide instability, suggesting that it is highly unlikely that the entire lithosphere at a fracture zone will spontaneously founder. While pressure gradients cause delamination of the lowest viscosity portions of the thermal boundary layer, they do not supply enough force to bend the plate, even if the fracture zone is the location of a deep weak fault. The pressure gradients that act to deflect the plate are proportional to lateral temperature gradients at the plate boundaries. Both gradients are largest at the beginning of the numerical experiments and decay with time for two reasons: the difference in thermal boundary layer thicknesses creating the pressure gradients becomes smaller as the plates age,

Temperature (C)



Fig. 3. Evolution of a model fracture zone initially having a 65 km wide weak zone at (a) 0 Myr, (b) 0.2 Myr, and (c) 0.4 Myr. Black lines mark boundaries of the weakened material. Initial ages of the plates are 0 and 40 Myr.

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and advection and diffusion laterally diminish the gradients. In one study indicating successful self-nucleation [19], the plates' viscosities were unreasonably small ($\sim 10^{22}$ Pa s) and density was compositionally, not thermally controlled. Models with non-diffusing (compositional) buoyancy forces enhance the likelihood of instability but will not produce the decay of pressure gradients with time characteristic of thermal boundary layers.

No combination of fault rheology or geometry

produced self-sustaining subduction without applied convergence. For example, if a weak zone 65 km wide (Fig. 3a, Case 3) is present in the older plate at a fracture zone (as may occur for two closely spaced, serpentinized fracture zones), the entire weak section is initially pushed down and to the right by pressure gradients, and resembles incipient subduction (Fig. 3b). However, the weak section rapidly delaminates from the surface without pulling the stronger portions of the lithosphere down (Fig. 3c). After delamination of the



Fig. 4. Evolution of a model fracture zone with a 40 Myr age offset and imposed convergence of 2 cm yr⁻¹ at four instances: (a) 0.8 Myr, (b) 5.4 Myr, (c) 6.0 Myr, and (d) 6.8 Myr. Each instance shows temperature, instantaneous flow lines (white), and current (solid) and initial (dashed) topography. The red 'x' shows the uplift and subsidence experienced by one unit of surface rocks. Panels c and d also show the horizontal component of surface velocity during the rapid extensional phase in the over-riding plate. See also the supplementary online animation; more animations are available at http://www.gps.caltech.edu/~gurnis/ Movies/movies-more.html.

weak plug, the model resembles the initial conditions of cases with thinner weak zones, none of which evolved towards subduction.

3.2. Models with applied compression

Subduction can occur if the fracture zone is moderately compressed and generally becomes self-sustaining after approximately 100 km of convergence. Our models show that the transition from forced to self-sustaining subduction results in an abrupt change in plate motions. During the initial stages of convergence (<100 km) the older plate slowly thrusts under the younger, but as slab buoyancy becomes the dominant force the slab begins to fall vertically into the mantle at a high velocity (three to four times the far-field velocity imposed at the right boundary). During vertical descent of the old plate, a wide region of surface extension forms as the trench rapidly migrates oceanward. To illustrate this transition, we describe a typical case (#6, Table 1) in which subduction becomes self-sustaining after ~ 115 km of convergence for an offset in lithospheric age of 40 Myr and an imposed convergence rate of 2 cm yr^{-1} (Fig. 4, supplementary online animation). Our models generally require less net convergence (110-140 km) for self-sustained subduction than predicted analytically (130-180 km) [2], most likely due to the analytical estimate's use of an excessively large elastic plate thickness [3].

The force required to converge at a constant velocity increases during the first 15 km of convergence (Fig. 5). This increase in resistance arises since work is needed to bend the plate as it enters the trench, with a commensurate flexural bulge on the old plate (Fig. 4a,b). After 15 km of convergence the fault extends to a depth of ~25 km and the compressive force reaches a maximum of 1.0×10^{12} N m⁻¹. Subsequently, the necessary external force continuously decays as slab buoyancy increasingly accounts for a larger portion of the driving force.

With further convergence, negative buoyancy of the subducting slab eventually surpasses the elastic bending resistance of the plate. Subduction becomes self-sustaining after ~ 115 km of convergence and the motion of the older plate abruptly changes from horizontal to vertical, as seen in the instantaneous flow lines (Fig. 4c,d). At the trench, the slab's vertical velocity greatly exceeds its horizontal convergence velocity, causing the surface location of the trench to move in the direction opposite to the imposed far-field convergence (trench rollback). As trench rollback occurs, the young over-riding plate rapidly thins by extension and hot mantle infiltrates the region above the slab as its dip becomes steeper.

Extension of the over-riding plate at the onset of subduction is discontinuous, not diffuse, and the surface velocities are plate-like on the overriding side (Fig. 4c,d). The trench recedes ~ 140 km during the first 1.4 Myr of self-sustained subduction, an average extension rate of 10 cm yr⁻¹. The rollback rate accelerates continuously during this time, so that instantaneous rollback rates at the termination of the model exceed the million year average (Fig. 4d).

3.3. Critical aspects for initiating subduction

We found that the creation of self-sustaining subduction zones occurs under restricted conditions. The evolution of the required external force as a function of net convergence is largely independent of convergence rate (Fig. 5c), implying that the force balance is dominated by the state of elastic flexure of the plate, and that rate-dependent (viscous) resistance to lithospheric motion is negligible. For convergence at 1 cm yr^{-1} , the resistance is actually slightly larger than that of the reference case of 2 cm yr^{-1} because the transition from viscous, strain rate-dependent creep to elastic deformation occurs at a higher temperature for lower strain rates. Therefore, reducing velocity (i.e. strain rate) results in a minor increase of the elastic thickness, so that the elastic bending resistance is increased. The counterintuitive decrease in force as velocity increases is a result of velocity boundary conditions; for example, if forces were imposed as boundary conditions, then larger velocities would result from increasing force [4]. Self-sustaining subduction is not achieved until the net convergence surpasses 135 km for this slower convergence rate because conduction of heat has more time to warm the slab.



Fig. 5. Tectonic force required to maintain a constant convergence rate as a function of total convergence. Except where noted, applied convergence rate is 2 cm yr⁻¹, C=4 MPa, $\mu=0$, and other rheological parameters are as in Table 2. (a) Influence of subducting plate age. (b) Influence of over-riding plate age. (c) Force evolution for convergence rates of either 1 or 2 cm yr⁻¹, with over-riding and subducting plate ages initially zero and 40 Myr. (d) Influence of plate and fault rheology. Case labeled ' $\eta \times 10$ ' has increased reference viscosity by a factor of 10.

For older plates, the buoyancy force available to drive subduction initiation is larger but elastic resistance is also greater. These effects nearly cancel, so that plates with ages of 20–80 Myr all reach comparable states of extension in the over-riding plate at nearly equal amounts of convergence. Furthermore, since the maximum required tectonic force increases with age of the subducting plate (Fig. 5a), an extremely old subducting plate may be more detrimental than beneficial for nucleating subduction.

The maximum resistance is more sensitive to the age of the over-riding plate (Fig. 5b). For given initial fault dimensions, plates are initially coupled to greater depths as the starting age of the over-riding plate increases, and the fault is therefore able to support large stresses to greater depths. Once a fault develops to sufficient depths to decouple the plates, typically within the first 10 km of convergence, the over-riding plate's age has little influence on the evolution of forces (Fig. 5b). After ~40 km of convergence, there is a slight reduction in the required tectonic force because the downgoing plate is kept cooler when subduction occurs beneath a colder over-riding plate.

The rheological properties of the plates and the fault zone can strongly influence the character of early subduction. For example, increasing the reference viscosity η_0 by a factor of 10 (Fig. 5d) effectively increases the elastic thickness of the downgoing plate, so that the necessary tectonic force is higher and the extensional rollback phase is not reached in the first 200 km of convergence. Increasing either the cohesion or coefficient of friction of the fault zone deters subduction initiation. The average stress supported by a fault zone of depth z_f is:

$$\overline{\sigma} \approx C + \frac{1}{2} \,\mu g \rho z_{\rm f} \tag{8}$$

neglecting geometrical effects for non-vertical faults. Even for coefficients of friction much less than that inferred for any real rock (e.g. $\mu = 0.035$, Fig. 5d), the fault zone supports average stresses exceeding 30 MPa over a depth of 50 km, and subduction is far from self-sustained. Our models suggest that self-sustained subduction requires that average fault zone stresses are below ~20 MPa. An average fault zone stress in the range

of a few tens of MPa agrees with estimates from force balances in back-arc basins (10–30 MPa [27]) and inferences of the amount of frictional heating required to match heat flow profiles (10– 60 MPa [28]). Since even weak rocks such as serpentine ($\mu \approx 0.4$ [29]) would support fault-averaged stresses exceeding 290 MPa, these low inferred stress levels would require that effective fault pressures are reduced by high pore pressure fluids.

A weak fault zone as an *initial* condition is also crucial for nucleating subduction. Without a weak fault (cases 4 and 12), initial convergence across a fracture zone is readily transmitted to the younger plate. In this situation, convergence is accommodated first by buckling the younger plate until a fault is formed. This fault, however, has an initial dip where the *younger* plate is underthrusted, a situation unfavorable for prolonged subduction. Much more convergence will be needed for systems without an initial weak zone to evolve to self-sustaining subduction.

4. Synthesizing geodynamic and geologic constraints

4.1. Extension and volcanism in nascent subduction zones

Our results can be compared with the regional evolution of incipient IBM subduction. The geological evolution of IBM during its infancy, a 10-15 Myr interval from \sim 50 to \sim 35 Ma, was distinctly different from both its subsequent evolution and that of other ancient and modern arcs [13]. The early arc was dominated by boninitic volcanism, which requires a high degree of partial melting of a clinopyroxene-poor source that is depleted in basaltic components but enriched in volatiles, presumably with an abnormally high geothermal gradient [13,17]. When the Mariana fore-arc was palinspastically restored with the KPR (Fig. 1), Stern and Bloomer [13] found that initial volcanism was exceptionally broad (~ 200 km) compared to present arcs (~ 50 km in width), magma production rates were substantially higher than found in fully developed arcs (with rates in the early IBM comparable to midocean ridges), and the tectonic environment was strongly extensional. This extensional signature is apparently at odds with previous ideas in which incipient subduction is favored in compressional settings [2,18]. However, our models, which develop zones of localized extension even though they are regionally compressive, provide a compelling match to IBM evolution, and would allow large amounts of mantle to melt over a wide extensional zone.

The models have a wide zone of extended lithosphere directly above the shallow dipping slab in which hot mantle is brought directly in contact with the top of the slab at shallow depths (Fig. 4d). Geotherms (Fig. 6) directly above the slab are likely to be extremely steep in comparison to those of fully developed subduction zones or those expected in modern intra-oceanic subduction zones undergoing back-arc extension. Because the slab is shallow dipping while falling vertically (Fig. 4d), these nascent subduction zones are different from present subduction zones, such as Tonga and the Marianas, where back-arc spreading centers lie above more steeply dipping slabs. Since the slab is expected to dehydrate at shallow depths [13,17], the dehydration in these models would release water into hot upwelling mantle, allowing extreme amounts of partial melting. The combination of rapid upwelling to shal-



Fig. 6. Vertical geotherms for the model shown in Fig. 4d. Inset shows temperature field at 6.8 Myr, with a contour interval of 200°C up to 1000°C. Vertical lines on the inset mark location of geotherms.

low depths and an influx of volatiles would presumably favor abundant boninite production. Additional melting-related buoyancy from melt retention or iron depletion in the residual mantle could further enhance melt production rates during this phase of rapid extension [30].

4.2. Vertical plate motions

The over-riding plates of the WPB underwent rapid vertical motions simultaneous with the initiation of IBM subduction. On the WPB, north of the CBSC, a set of arcs (Daito Ridge and Oki Daito Ridge, Fig. 1) are found which were active in the Mesozoic but subsequently subsided to deep water by the Paleocene [31]. However, cob-

bles in an Eocene conglomerate recovered from a deep drill hole adjacent to the now flat-topped Daito Ridge show rapid uplift in the Eocene [31], synchronous with the initiation of subduction. Daito Ridge subsequently subsided to a depth of 1.5 km from the Eocene to Quaternary [32]. Rapid Eocene uplift and subsidence were associated with volcanism [33], presumably from the adjacent KPR. Rapid uplift is expected where older plates are underthrusted, such as the rapid uplift synchronous with subduction initiation on the northern MRC [6]. Away from the spreading center, our models (cases 20 and 22) show that a 10 Myr old over-riding plate would buckle during underthrusting and give rise to uplift on the order of 1 km. In contrast, our models suggest that for



Fig. 7. Paleo-age grids of the Pacific–Eurasia–Philippine plate region based on a global isochron data-set and plate circuit closure. Red lines indicate subduction zones, purple lines indicate spreading ridges, black lines are strike-slip boundaries. White dotted lines mark the trace of PAC–IZA fracture zones. Blue patches represent the Izu-Bonin and KPRs while light gray patches represent the Oki-Daito Complex. Black, red, and pink arrows are the convergence vectors between Pacific and Eurasia at 55, 45, and 35 Ma, respectively. Abbreviations: ANT = Antarctic Plate, AUS = Australian Plate, CBSC = Central Basin Spreading Centre, EUR = Eurasian Plate, IZA = Izanagi Plate, OD = Oki-Daito Complex, PAC = Pacific Plate, PH = Philippine Plate.

underthrusting beneath younger lithosphere near the active spreading center, little uplift would be expected.

4.3. Reconstructing the origin of the IBM subduction zone

Our models for transforming a fracture zone into a self-sustaining subduction zone require approximately 100 km of compression before the slab begins its rapid vertical descent. As indicated by volcanic ages [11], rapid subduction may have begun in IBM as early as 49 Ma, and possibly is a driving mechanism for the change in Pacific Plate motion. However, the bend in the Hawaiian-Emperor hot-spot track may reflect changes in the motion of the hot-spot and not a change in absolute plate motion [34]. Moreover, the WPB and IBM have undergone substantial clockwise rotation since the Eocene [35]. Due to these complexities, we have used a new set of reconstructions based on a global plate circuit [36] incorporating a fixed Atlantic hot-spot reference frame together with a revised interpretation for WPB opening [37] to determine if any changes in Pacific Plate motions could have forced convergence across the Eocene IBM margin.

The reconstructions show that in the Eocene, the orientation of the proto-IBM boundary is parallel to the orientation of predicted Izanagi



Fig. 8. Ages of over-riding and downgoing plates, and the maximum resistance to convergence along the transform boundary that becomes the IBM subduction zone in the reconstructions. Vectors give plate motions relative to the lower right plate. On the over-riding plate, dashed vector is for pre-55 Ma motion and the solid vector is for motions between 55 and 45 Ma. The large discontinuity in seafloor age near the spreading center is due to an inferred ridge jump that had occurred just prior to 55 Ma.

(IZA)–Pacific (PAC) fracture zones (Fig. 7). From this geometry, the Pacific Plate's rate and direction of motion changed substantially from near parallel to fracture zones at 55 Ma to obliquely convergent between 55 Ma and 45 Ma. An upper limit of $\sim 2-3$ cm yr⁻¹ convergence rate across the IZA–PAC fracture zone is found by assuming IZA remained fixed with respect to EUR, such that 100 km of convergence could have occurred within ~ 5 Myr. Our models suggest that this fracture zone could therefore have been forced to the stage of rapid rollback and self-sustaining subduction.

The force required to create this new plate boundary can be estimated using the boundary evolution models and the paleogeographic reconstructions. From the paleogeographic reconstruction at 55 Ma (Fig. 7), we determine the ages of plates on both sides of the transforms and fracture zones bounding the Pacific and IZA. Using the force evolution curves in Fig. 5a,b (plus other cases), we derive an empirical relation for the maximum resisting force for converging plates of different ages.

We estimate that the integrated resisting force along 2500 km of the nascent IBM trench was $\sim 2 \times 10^{19}$ N (Fig. 8). The driving forces on the Pacific Plate at 55 Ma are unknown; however, the plate was relatively fast-moving (~ 6 cm yr⁻¹ with respect to EUR) and was connected to both a long-lived subduction zone in the west and an extensive ridge system in the east (Fig. 7). Present day driving forces are estimated to total 1.9×10^{21} N [38]. If the Pacific Plate had 10% of this force available at 55 Ma, then resisting forces along the incipient trench would only be 10% of the driving force. Consequently, changes in plate motion appear capable of providing the force required to nucleate a new subduction zone even if only a small amount of the available tectonic forces can be applied at a fracture zone of sufficient offset.

5. Conclusions

We have shown that oceanic fracture zones can be transformed into a self-sustaining subduction zone after at least ~ 100 km of convergence. Forces required to drive convergence never exceed expected driving forces ($< 4 \times 10^{12}$ N m⁻¹), so long as convergence is accommodated along a weak (supporting at most ~ 10 MPa average stress) fault zone. As the forces governing slab dynamics evolve and subduction transitions toward a self-sustaining state, vertical descent of the slab increases rapidly. This in turn leads to rapid trench rollback and extension in the overriding plate. The models show the first self-consistent development of zones of rapid, localized extension within areas that are undergoing regional compression, and therefore help to overcome the paradox between early extensional geological signatures preserved in fore-arc basements and the need for convergence to start subduction.

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