

3D NUMERICAL MODELS FOR ALONG-AXIS VARIATIONS IN DIKING AT
MID-OCEAN RIDGES

by

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Dedication

I would like to dedicate this thesis to my mother, 田霞 (Tian, Xia). I wouldn't have a chance to experience this wonderful world without her giving birth to me. She rears me up by herself with her great love, optimism and peseverence. Without her guidance and support, I would not become who I am.

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“In the midst of winter, I found there was, within me, an invincible summer.”

—Albert Camus

“People have no higher calling than to strive for the greater good of humankind and society and that the future of humanity can be assured only when there is a balance between scientific development and the enrichment of the human spirit.”

—Kazuo Inamori

“不失其所者久，死而不亡者寿。”

—《道德经》

Abstract

Tian, Xiaochuan. M.S. The University of Memphis. May 2015 Master of Science.
3D Numerical Models for Along-axis Variations in Diking at Mid-Ocean Ridges. Major
Professor: Dr. Eunseo Choi.

Bathymetry reveals diverse morphologies at Mid-ocean Ridges (MORs). Previous studies show that the morphologies at slow spreading MORs are mainly controlled by the ratio (M) between rates of magma supply and plate extension. 2D models successfully explain many cross-sectional observations across the ridge axis. However, magma supply varies along the ridge and the interaction processes between the tectonics and magmatism at MORs are inevitably three dimensional. We investigate the consequences of this along-axis variation in diking in terms of faulting pattern and the associated structures with a 3D parallel geodynamic modeling code, SNAC. Many structural features observed are produced. We also proposed asynchronous faulting induced tensile failure as a new possibility for explaining corrugations. $\bar{M} = 0.6425$ is suggested as a boundary value for separating abyssal hills and oceanic core complexes (OCCs) formation. Previous inconsistency for OCCs formation between 2D model results ($M = 0.3 \sim 0.5$) and field observation that M can extends beyond the range is reconciled by our 3D along ridge coupling argument.

Table of Contents

Chapter	Page
1 Introduction	1
1.1 Review of Literature	1
1.2 Statement of Research Purpose	6
2 Methods	7
2.1 Method of approach	7
2.2 Model Setup	8
3 Results	13
3.1 Reference model	13
3.1.1 Constant M model M88ConT2	17
3.2 Main structural characteristics	18
3.2.1 Location of termination	18
3.2.2 Geometry of trough	19
3.2.3 Inward fault jump	19
3.2.4 Fault alternation	20
3.2.5 Mass wasting	20
3.2.6 Hourglass-shaped median valley	21
3.2.7 Corrugations and mullion structures	22
3.3 Effects of the functional forms of M variation	25
3.3.1 Models with M between 0.2 and 0.8 and Type 1 weakening	25
3.4 Effects of the weakening rate	29
3.5 Effects of the range of M variation	31
4 Discussion	32
4.0.1 Models with M between 0.5 and 0.8 and Type 2 weakening ^{EC} :M58(Lin,Sin,Sqrt)T2	36

4.1 Comparing model results with nature observation	37
4.1.1 Inward fault jump	37
4.1.2 Fault alternation	40
4.1.3 Mass wasting	42
4.1.4 Hourglass shape median valley	43
4.1.5 Mullion structure	45
4.1.6 Corrugations	46
5 Conclusions	49
Bibliography	51

List of Tables

1	List of 3D numerical experiments.	11
2	Summary of 3D Model Parameters	12
3	Mean M values of the models with ^{EC:} <u>Type 1 (or is it Type 2?)</u> weakening. ^{EC:} <u>20 km segment. (The value is calculated by integrating M along the ridge axis and divided by the length of the model domain in z-axis.)</u>	36

List of Figures

1	Profiles of bathymetry across MORs.	1
2	Two bathymetric profiles across the Mid-Atlantic Ridge around 30°N with vertical exaggeration of 10. A-A' is closer to the segment center while B-B' is at the tip of the segment.	2
3	Relationship between the maximum crustal thickness variations (ΔH_c) along a ridge segment and the segment length (L). The dashed line is the best-fit linear regression of the combined data [Chen and Lin, 1999].	3
4	Upper one: modeling result for fast spreading agrees well with the observation of East Pacific Rise. Lower one: modeling result for slow spreading ridges agrees well with the bathymetry of Mid Atlantic Ridge. Adapted from [Buck et al., 2005].	4
5	Snapshots of modeled fault behavior and seafloor morphology for values M = 0, 0.5, and 0.7; model allows thermal evolution. Structural interpretation is superimposed on modeled distribution of strain rate; model time is indicated in panels at lower right; dashed white line at bottom is 600 °C isotherm and approximates the brittle-ductile transition; dashed seafloor is original model seafloor, red seafloor is that formed dominantly by magmatic accretion, and solid bold seafloor is fault surface.[Tucholke et al., 2008, Whitney et al., 2012]	5
6	Essential components of the numerical method.	7
7	Model setup.	9

8	Evolution of plastic strain and surface topography of the reference model M28LinT1. Each snapshot shows plastic strain plotted on the model domain, the five times exaggerated topography and the time at which it is taken. The initial seafloor is at 0 km of elevation. The black dashed lines are the terminations of faults. The red dashed lines in g) and h) are the transfer faults that connect the ter- minations. The inset in f) plots plastic strain with opacity linearly proportional to plastic strain value.	13
9	Bird's eye view of a breakaway in the reference model at a) 85 kyr, b) 135 kyr, c) 265 kyr and d) 295 kyr. The breakaway is marked by green solid line in a). A narrow zone of depressed topography ("trough") is marked by red dashed line inside the medain valley (blue area) in a). Color scales are the same with those in Figure 8.	14
10	Illustration of the development of bending stress in lithosphere associated with faulting at the MOR. The blue line is the neutral plane where $\sigma_{xx} = 0$. Above the neutral plane is compression ($\sigma_{xx} < 0$) and beneath it is tension ($\sigma_{xx} >$ 0).	15
11	Bird's-eye view of velocity field with plastic strain plotted with opacity lin- early proportional to its value. (color scale is the same as Figure 8.a)) . . .	16
12	Bird's eye view of σ_{xz}	16
13	Evolution of plastic strain and surface topography of the model: M88ConT2 (Table 1). (color scale of topography is the same as Figure 8.a)	17
14	The second invariant of strain rates plotted on the reference model's vertical cross-sections along the ridge at 107.5 kyr. Terminations and breakaways are marked by black and red arrows.	18
15	The weak detachment fault tip reaches the pre-existing shear stress. Plastic strain is plotted with opacity linearly proportional to its value.	20

16	Velocity of the mass wasting hanging wall. Magnitudes of the velocity are shown by the colors of the arrow heads. Plastic strain is plotted with opacity linearly proportional to its value.	21
17	Plastic strain, topography and stresses evolution for M28SqrtT1.	22
18	Along ridge axis σ_{xz} (bird's-eye view of three layers of model domain) (positive(red) means clockwise). Plastic strain is plotted with opacity linearly proportional to its value.	23
19	M28SqrtT1 (Table 1). Bending stress drop (a versus b) at the lower M side due to the cut-back. No bending stress drop (c versus d) at the higher M side as a comparison.	24
20	New fault front chase after cut-back.	24
21	Bird's-eye view of the evolution of the hourglass shape median valley. It is generated by attaching the topography of the M28SqrtT1 model to its mirror reflection by assumming symmetrical M variation (0.2 to 0.8 to 0.2). . . .	25
22	Higher σ_{xx} shows inside the median valley on the negative x -axis direction of the ridge axis.	25
23	Bird's eye view of the evolution of the corrugations. Color scales for the topography is the same as Figure 8.	26
24	Evolution of mullion structures.	26
25	Three functional forms of M variation. M begins to exceed the M = 0.5 black line at Z=10 km, 7 km, 5 km for M28LinT1, M28SinT1 and M28SqrtT1 respectively.	27
26	Bird's-eye view for comparing the length and timing of inward fault jump. .	27
27	$\partial M / \partial Z$ comparision.	28
28	Bird's eye view of the topography. ^{EC:} <u>(without vertical exaggeration.)</u> [don't need to say. When you did it, you need to say so.] ^{EC:} [Say somewhere that you put the origianl and an mirror iamage together in this figure.]	29

29	M57SinT2 versus M57SinT1 (Table 1)	30
30	Faulting and stress evolution for M57SinT1.	31
31	Initiation and duration of main structural features. ^{EC} : [Labels of an individual plot are too small. Symbols can be bigger, too.]	32
32	Along-axis variations in total accumulated tectonic heave T in the past 1.86 Ma (since chron C2n), and consequent inferred magmatic component M (= 1 – T) as a proportion of total plate separation (blue line). Pink shaded areas delineate loci of the active or recently active OCCs. The relative contributions of tectonic strain from the western and eastern flanks of the axis that give rise to the total heave T are shown by red and green lines respectively. Adapted from [MacLeod et al., 2009].	35
33	Schematic development of a megamullion. No vertical exaggeration. (a)~(c) shows the detachment fault evolution during amagmatic phase. (d) Increased magma supply pushed the detachment fault away from ridge axis and forms a new normal fault near the ridge axis (“inward fault jump”). Adapted from [Tucholke et al., 1998].	38
34	Adapted from [Dick et al., 2008].	39
35	Adapted from [Xu et al., 2009].	40
36	Bathymetry from 12.8~14.2 °N Mid-Atlantic Ridge. Crossection A~A' and B~B' are 5 times vertical exaggerated. From GeoMapApp.	41
37	Trade-off between change in bending force ΔF_b and weakening in the fault interface ΔF_w . H is the thickness of the brittle crust and γ is the size of initial weak perturbation and A defines the maximum bending force change. Adapted from [Lavier et al., 2000]. (For more details, please refer to [Lavier et al., 2000])	42

38	Comparing bathymetry at 13°N Mid-Atlantic Ridge to the mass wasting behavior of M28SqrtT1. The model topography is a mirror symmetric flip according to the dash line, it shows the case of M varies in a square root functional form from 0.2 to 0.8 to 0.2. The bathymetry image is generated by GeoMapApp [Ryan et al., 2009].	43
39	Bathymetry and magnetization of 13~15 °N MAR. Adapted from [Smith et al., 2008].	44
40	Hourglass median valley at 22 ° 30 'N MAR. Image is generated from GeoMapApp.	44
41	Mullion structures on the surface of Kane OCC at 23 °N MAR. Image is generated from GeoMapApp.	45
42	Simpler 3D model with 4 km of seawater pressure on top. Seeds are partially added along the ridge segment. Color scale is the same as Figure 23.	46
43	Simpler 3D model with 20 km of seawater pressure on top. Color scale is the same as Figure 23.	46
44	Simpler 3D model with 4 km of seawater pressure on top. Seeds are added along the whole ridge segment. Color scale is the same as Figure 23.	47

1 Introduction

1.1 Review of Literature

Geodynamic modeling as well as a variety of geological, geophysical observations and lab experiments have provided insight into the processes occurring at the mid-ocean ridges (MORs) [e.g., Tucholke and Lin, 1994, Blackman et al., 2002, Behn et al., 2006, Behn and Ito, 2008, Ito and Behn, 2008, Baines et al., 2008, Escartín et al., 2008, Canales et al., 2008, Dick et al., 2008, Dannowski et al., 2010, Olive et al., 2010, Reston and Ranero, 2011, Reston and McDermott, 2011]. In particular, the advent of high-resolution multi-beam bathymetric data has made it possible to discover differences in axial topography between slow and fast spreading ridges and morphological transition from the center of a ridge segment to the tip of the ridge segment.

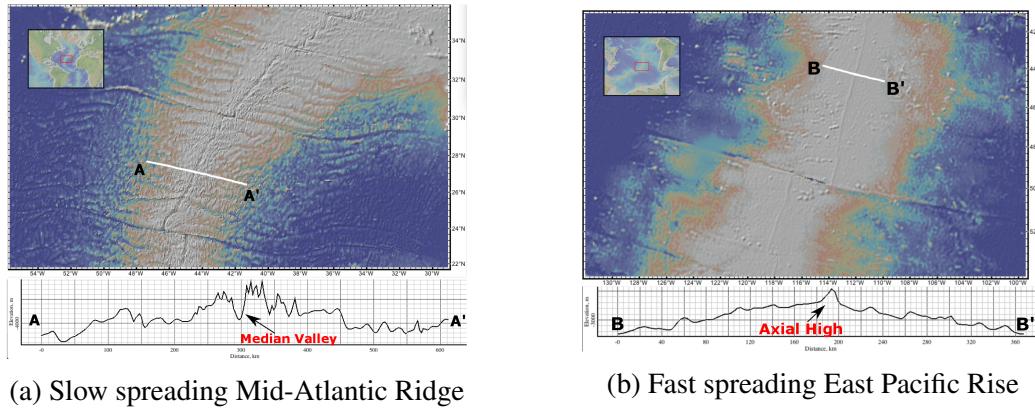


Figure 1: Profiles of bathymetry across MORs.

Variations in morphologies among different MORs are mainly controlled by four factors: magma supply, tectonic strain, hydrothermal circulation and spreading rate [Fowler, 2004]. Among them, the spreading rate shows the strongest correlation with the ridge morphology. Slow-to-intermediate spreading ridges (half spreading rate less than 4 cm/yr) produce median valleys that are typically 10~20 km wide and 1~2 km deep (e.g., Mid-

Atlantic Ridges, Figure 1a). Fast-spreading ridges (half spreading rate greater than 5 cm/yr) like the East Pacific Rise have axial highs that are 10~20 km wide, 0.3~0.5 km high (Figure 1b).

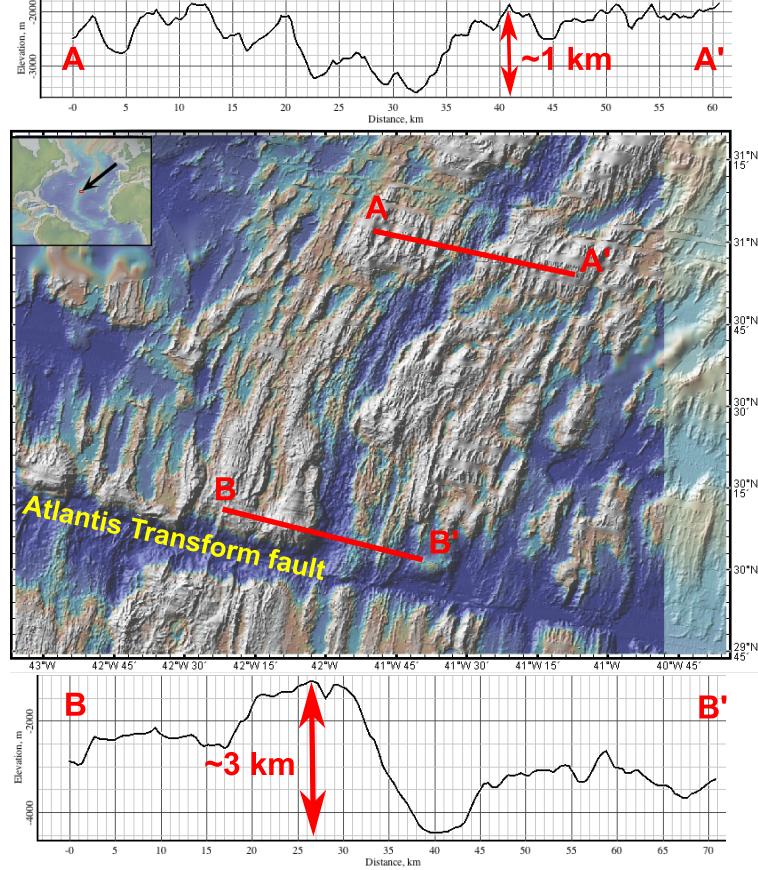


Figure 2: Two bathymetric profiles across the Mid-Atlantic Ridge around 30°N with vertical exaggeration of 10. A-A' is closer to the segment center while B-B' is at the tip of the segment.

Slow spreading ridges exhibit along-axis variations in off-axis morphology, the width and depth of median valleys and crustal thickness. Figure 2 shows that the topographic profile near to the center of the ridge segment (A-A') is rather symmetrical and has a higher frequency with a median valley \sim 12 km wide and \sim 1 km deep. In contrast, the near-tip profile (B-B') is asymmetrical and has a much lower frequency with a median valley wider than 30 km and shows a greater relief (\sim 3 km). The maximum along-axis

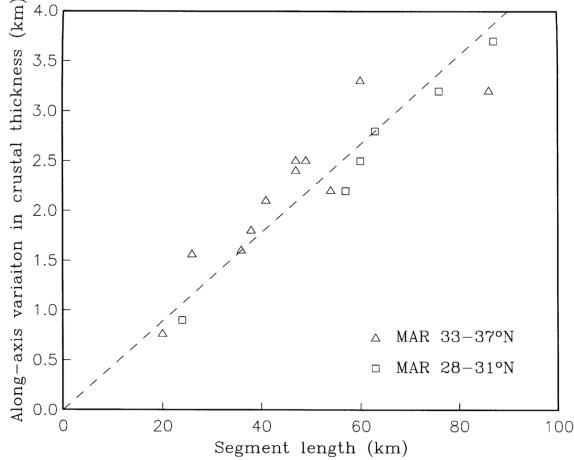


Figure 3: Relationship between the maximum crustal thickness variations (ΔH_c) along a ridge segment and the segment length (L). The dashed line is the best-fit linear regression of the combined data [Chen and Lin, 1999].

variation in crustal thickness ΔH_c is linearly increasing with segment length L [Chen and Lin, 1999] and the relationship is $\Delta H_c(L) = 0.0206L$ (Figure 3).

Magma supply at the MORs is mostly a passive process when no hot plume is present [Fowler, 2004]. Driven by both vertical pressure difference and buoyancy due to horizontal density difference, hot mantle rises up to fill the vacated room produced by the plate separation. Decompression of the upwelling hot mantle results in partial melting. The generated magma upwells to the upper crust and feeds the dikes at the ridge center.

The passive nature of the melting process at the MORs leads to the major difference between fast and slow spreading ridges. At fast spreading ridges, magma is generated at a higher rate than at slow spreading ridges. Thus, fast spreading ridges can sustain more frequent diking, which efficiently release stresses generated by far-field forces driving the plate motion. In contrast, diking is less frequent at slow spreading ridges and can only partially releases the stresses associated with plate motion. The plates associated with slow-spreading ridges experiences internal deformations (e.g. tectonics process like normal faulting) when the accumulated extensional stress exceeds the strength of the crust.

Buck et al. [2005] defined the ratio between the rate of plate separation by diking and spreading rate as $M = V_{dx}/2V_x$, where V_{dx} is the rate of opening by diking at the ridge

center and V_x is the half spreading rate of the MOR. According to this definition, $M = 1$ represents the case where dike injection is so frequent that magma supply is sufficient to release all the tensional stresses from plate separation. $M = 0$ corresponds to the case of no magma supply, in which diking does not account for any of the plate motion and therefore plates kinematics requires plates to go through internal deformations. As shown in Figure 4, an axial high forms at a fast spreading ridge ($M = 1$) due to buoyancy from lateral density difference across ridge axis but a median valley forms at a slow-spreading ridge ($M = 0.5$) due to near-axis normal faulting, which is in turn caused by the stretching of oceanic lithosphere.

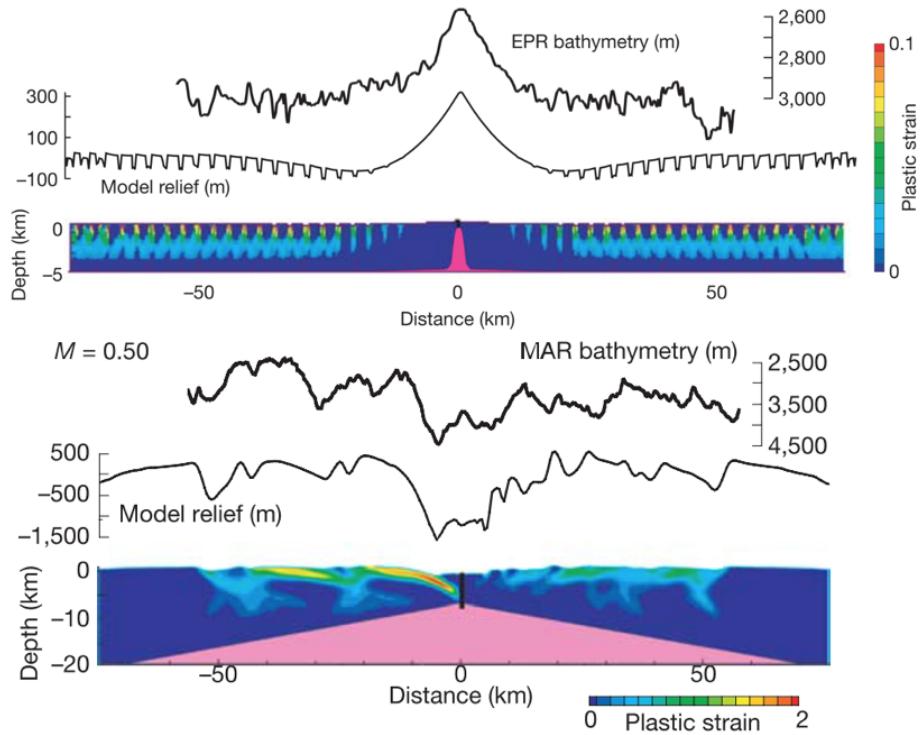


Figure 4: Upper one: modeling result for fast spreading agrees well with the observation of East Pacific Rise. Lower one: modeling result for slow spreading ridges agrees well with the bathymetry of Mid Atlantic Ridge. Adapted from [Buck et al., 2005].

Tucholke et al. [2008] expand the investigation on the role of M in the mid-ocean

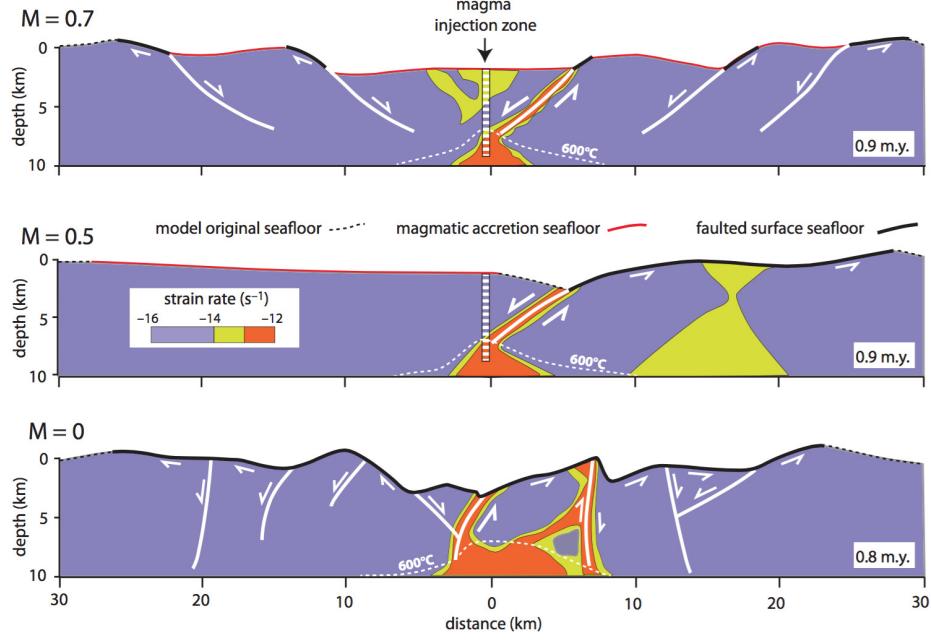


Figure 5: Snapshots of modeled fault behavior and seafloor morphology for values $M = 0, 0.5$, and 0.7 ; model allows thermal evolution. Structural interpretation is superimposed on modeled distribution of strain rate; model time is indicated in panels at lower right; dashed white line at bottom is $600\text{ }^{\circ}\text{C}$ isotherm and approximates the brittle-ductile transition; dashed seafloor is original model seafloor, red seafloor is that formed dominantly by magmatic accretion, and solid bold seafloor is fault surface.[[Tucholke et al., 2008](#), [Whitney et al., 2012](#)]

ridge mechanics. They focus on the faulting behaviors of slow spreading ridges and find that the OCCs are most likely to form when M varies from 0.3 to 0.5. When $M = 0.7$ (Fig. 5), repeated diking pushes faults that have formed at the spreading center away from the ridge axis. Since the thickness of the brittle layer increases away from the ridge axis due to cooling effects, frictional and bending energy for maintaining the fault also increases. When the energy needed for maintaining an existing fault exceeds the energy for breaking a new near-axis fault, the old fault is replaced by the new one and most of the extension is accommodated by the new fault. When $M = 0.3 \sim 0.5$, the normal fault remains active for a long time and rotates to a low angle normal fault (detachment fault), exhuming the lower crust and mantle materials to the seafloor. When $M < 0.3$, most of the tension is accommodated by intra-plate deformations rather than by diking and as a result, faulting pattern is more complicated and unsteady.

1.2 Statement of Research Purpose

The M-factor formulation used in the previous 2D models [e.g., Tucholke et al., 2008, Buck et al., 2005] successfully explained major features found in across-ridge profiles of seafloor bathymetry. However, 2D models have limitations in studying the along-ridge variations in morphology and faulting patterns. Magma supply at fast spreading ridges seems always sufficient for accommodating plate motions with little variation along the ridge axis. The relatively uniform topography along fast spreading ridges is considered to be consistent with the uniform abundance of the magma supply. However, along the slow spreading ridges, bathymetry, gravity anomaly and results from reflection and refraction seismology show strong correlation with variation in crustal thickness [e.g., Ryan et al., 2009, Chen and Lin, 1999, Lin et al., 1990, Tolstoy et al., 1993]. Because oceanic crust is mainly formed by upwelled magma at the ridge, variation in the thickness of the crust implies variation in magma supply. At slow spreading ridges, the degree of cooling by hydrothermal circulation, thermal structures and even local spreading rate [Baines et al., 2008] also varies both along and across the ridge axis and they appear interrelated. Thus, for slow-to-intermediate spreading ridges, the interactions between tectonics and magmatism at MORs are inevitably 3D processes and 3D numerical models are desirable for better understanding factors controlling both across- and along-ridge morphology variations.

The purpose of this thesis is to extend the M-factor formulation originally developed for 2D models to three dimensions (3D) by implementing it into a 3D numerical modeling code **SNAC** (**S**t**G**erma**I**N **A**nalysis of **C**ontinua) [Choi et al., 2008]. By systematically exploring the behaviors of the 3D models and comparing them with observations, we aim to better understand how the mid-ocean ridge magmatism and tectonic deformations interact.

2 Methods

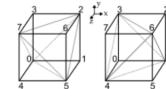
2.1 Method of approach

The numerical modeling code, SNAC (StGermaiN Analysis of Continua), is an explicit Lagrangian finite element code that solves the force and energy balance equations for elasto-visco-plastic materials. Figure 6 shows major components of SNAC.

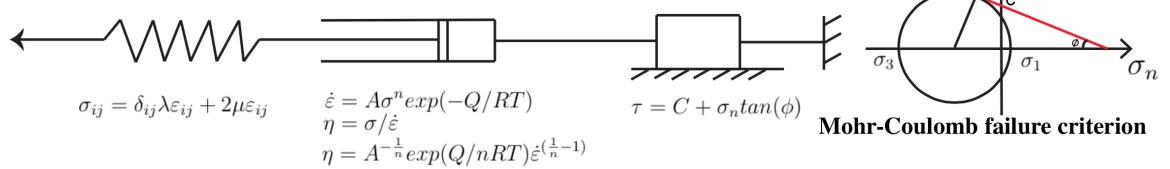
SNAC: a 3D, MPI parallelized, updated Lagrangian explicit finite difference code for modeling long-term tectonic evolution of the Earth's elasto-visco-plastic crust and mantle. (Choi et al., 2008)

Momentum Balance Equation: $\frac{\partial \sigma_{ij}}{\partial x_j} + \rho g_i = \rho \frac{Dv_i}{Dt}$

Spatial Decritization: A 3D domain is discretized into hexahedral elements, each of which is filled with two sets of 5 tetrahedra.



Elasto-Visco-Plastic (EVP) Rheology:



Diking M Formulation: $\mathbf{M} = \mathbf{Vdx} / 2\mathbf{Vx}$ (Vdx is the dike widening velocity in x direction, 2Vx is the full spreading velocity)

Stresses introduced by a dike accretion strain(dike widening) $\Delta\varepsilon_{xx}$ in each time step dt:

$$\Delta\sigma_{xx} = (\lambda + 2\mu)\Delta\varepsilon_{xx} \quad \Delta\sigma_{yy} = \lambda\Delta\varepsilon_{xx} \quad \Delta\sigma_{zz} = \lambda\Delta\varepsilon_{xx}$$

Figure 6: Essential components of the numerical method.

For each time step, strain and strain rates are updated based on the initial or previous velocity fields under the constraints from boundary conditions. A constitutive model returns updated stresses corresponding to these deformation measures. Internal forces are then calculated from the updated stresses, which is plugged into the momentum balance equation together with the body force term. Then, the damped net force divided by inertial mass yields acceleration at a node point, which is time-integrated to velocity and displacement.

A 3D domain is discretized into hexahedral elements, each of which is in turn divided into two sets of tetrahedra. This symmetric discretization prevents faulting from favoring

a specific direction or “mesh grains”.

Rheology for the oceanic lithosphere is assumed to be elasto-visco-plastic (EVP). When viscosity is high at low temperature, the EVP rheology implemented in SNAC essentially becomes the Mohr-Coulomb plasticity with strain softening that can create shear bands that behave like faults. Strain softening is realized by cohesion decreasing with increasing amount of permanent (i.e., plastic) strain. I assume this relationship is linear for simplicity. It is sufficient for a full description of such a linear strain weakening to define initial and final values of cohesion and a critical plastic strain at which cohesion becomes the final value. I define the rate of strain weakening as the cohesion difference divided by the critical plastic strain and use it as one of the model parameters. When temperature is high and viscosity is low, the rheology becomes the Maxwell viscoelasticity and can model creeping flow. This property of the EVP model makes it possible to set up a structure with a brittle lithosphere and a ductile asthenosphere through a proper temperature distribution. Rheological parameters are taken from previous studies that use a similar rheology [e.g., [Buck et al., 2005](#); [Tucholke et al., 2008](#)] or from lab experiments [e.g., [Kirby and Kronenberg, 1987](#)].

For 3D diking processs, the strain $\Delta\varepsilon_{xx}$ associated with diking leads to stresses changes, $\Delta\sigma_{xx}$, $\Delta\sigma_{yy}$ and $\Delta\sigma_{zz}$. These stress changes due to diking are computed according to the linear elastic constitutive equations $\sigma_{ij} = \lambda\varepsilon_{kk}\delta_{ij} + 2\mu\varepsilon_{ij}$.

2.2 Model Setup

The 3D models have a common geometry of $60\text{ km} \times 20\text{ km} \times 20\text{ km}$ in x , y and z axes respectively with a resolution (Δx) of 1 km (i.e., Δx is the size of each hexahedron element)(Fig. 7). The initial temperature field linearly increases from $0\text{ }^{\circ}\text{C}$ at the top surface to $240\text{ }^{\circ}\text{C}$ at the depth of 6 km, reflecting enhanced cooling due to hydrothermal circulation. Below 6 km, the temperature profile follows the semi-infinite half-space cooling model of moving plates [e.g., [Turcotte and Schubert, 2002](#)]. Two sides perpendicular

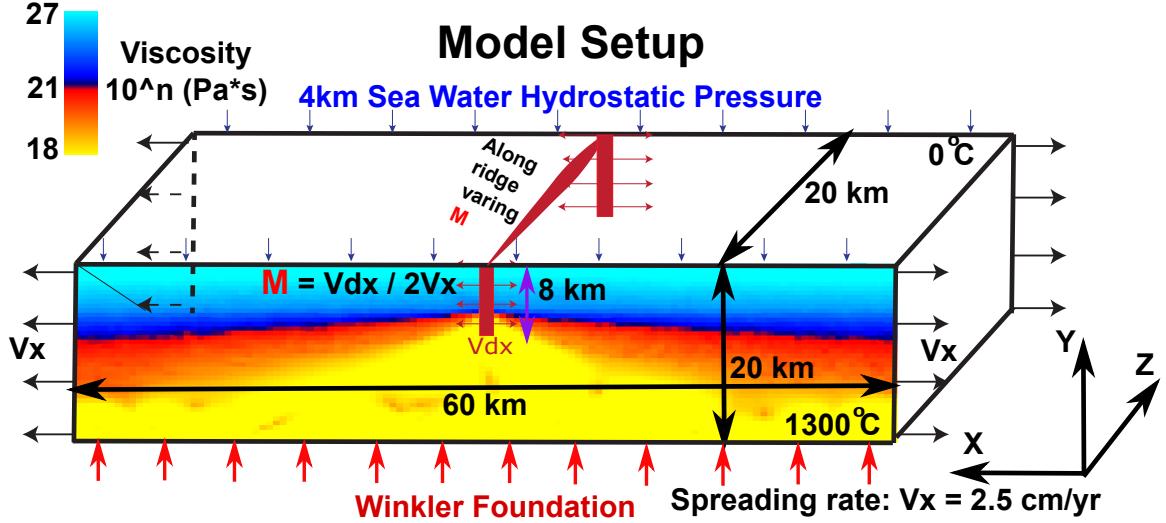


Figure 7: Model setup.

to the z coordinate axis are free-slip. The top surface has vertical tractions from water columns, of which heights are locally determined as $(4000 - h(x, z))$ m, where $h(x, z)$ is the topography at a location, (x, z) . The bottom surface is supported by the Winkler foundation. Temperature is fixed at 0 °C on the top surface and at 1300 °C on the bottom surface.

Diking, represented by the factor M as described above, is assumed to occur in the middle of the doamin (Fig. 7), where the lithosphere is the thinnest.

We adopt the linear isotropic elasticity, power-law viscosity of dry diabase [e.g., Kirby and Kronenberg, 1987, Buck et al., 2005] and the Mohr-Coulomb plastic model. The complete list of model parameters are given in Table 2.

Before running 3D models, I have run hundreds of psedudo-2D models for initial setup and benchmarking with previous studies [e.g., Buck et al., 2005, Tucholke et al., 2008]. Preliminary pseudo-2D results show that the model behavior in faulting pattern is sensitive to the rate of strain weakening. Two cases of strain weakening are tested in the 3D models. In one case (denoted as Type 1 weakening), cohesion linearly decreases from 44 MPa (denoted as C_i) to 4 MPa (C_e) for plastic strain accumulating from 0 ($\varepsilon_{p_i}^1$) to 0.1 ($\varepsilon_{p_e}^1$). It has a characteristic fault slip of 150 m for pseudo-2D models and 300 m for 3D

models. The other case (Type 2 weakening) assumes cohesion linearly decreasing from 44 MPa (C_i) to 4 MPa (C_e) for plastic strain accumulating from 0 ($\varepsilon_{p_i}^2$) to 0.33 ($\varepsilon_{p_e}^2$). In this case, the characteristic fault slip for pseudo-2D models is 500 m and for 3D models is 1 km. The characteristic fault slip is defined as $\Delta X_c = 3\Delta x \varepsilon_{p_e}$ where $3\Delta x$ represents the thickness of the shear bands which is usually 2 to 4 times Δx (size of a hexahedron element) [Lavier et al., 2000]. When ΔX_c amount of slip takes place at the fault interface, the cohesion of the material at the faulting interface decreases to C_e . In this way, under the same amount of ΔX_c , models with different resolution should produce the same faulting patterns.

Although how to estimate the M values from observations is a subject of on-going research, constraints are available from a large dataset of bathymetry, gravity and seismic surveys as well as geological drilling. Generally, at slow spreading ridges, magma supplies mostly at the center of the ridge segment and decreases towards the tip of the segment [e.g., Tolstoy et al., 1993, Chen and Lin, 1999, Carbotte et al., 2015]. There is also evidence for shorter wavelength of 10 to 20 km discrete focus of magma accretion along the ridge axis [Lin et al., 1990].

The numerical cost of a 3D model is non-trivial. For 2 Myr of model time, each model usually runs on 192 cores for about 48 hours (i.e., around 10^4 core-hours).

Based on these observational constraints and computational cost, I start considering a few scenarios of variations in M along the ridge axis. They are 1) three types of functional forms, linear, sinusoidal and square root; 2) three ranges of M variation along the ridge axis, 0.5 to 0.7, 0.5 to 0.8 and 0.2 to 0.8; and 3) two types of weakening rate, type 1 and type 2.

While 18 models are possible, I could run 15 models with the available computational resources. The complete list of the models is given in Table 1.

Table 1: List of 3D numerical experiments.

Model	M range	Functional Form	Type of weakening	For short
1	0.2-0.8	Linear	Type 1	M28LinT1
2	0.2-0.8	Sinusoidal	Type 1	M28SinT1
3	0.2-0.8	Square Root	Type 1	M28SqrtT1
4	0.5-0.7	Linear	Type 1	M57LinT1
5	0.5-0.7	Sinusoidal	Type 1	M57SinT1
6	0.5-0.7	Square Root	Type 1	M57SqrtT1
7	0.5-0.7	Linear	Type 2	M57LinT2
8	0.5-0.7	Sinusoidal	Type 2	M57SinT2
9	0.5-0.7	Square Root	Type 2	M57SqrtT2
10	0.5-0.8	Sinusoidal	Type 1	M58SinT1
11	0.5-0.8	Square Root	Type 1	M58SqrtT1
12	0.5-0.8	Linear	Type 2	M58LinT2
13	0.5-0.8	Sinusoidal	Type 2	M58SinT2
14	0.5-0.8	Square Root	Type 2	M58SqrtT2
15	0.8-0.8	Constant	Type 2	M88ConT2

Table 2: Summary of 3D Model Parameters

Number	Variable	Description	Value	Units
1	W_{dike}	Dike width	2	km
2	D_{dike}	Dike depth	8	km
3	H	Crustal thickness at dike	6	km
4	dT/dy	Crustal thermal gradient	40	K/km
5	T_1	Temperature at lower boundary of crust	240	°C
6	g	Gravity acceleration	10	m/s ²
7	$demf$	Dimensionless force damping factor	0.8	N/A
8	dt	Time step	1.5768e+07	second
9	$topokappa$	Parameter for topography smoothing	0	N/A
10	$shadowDepth$	Ghost elements for parallel computing	2	N/A
11	$meshI$	Mesh number in X direction	60	N/A
12	$meshJ$	Mesh number in Y direction	20	N/A
13	$meshK$	Mesh number in Z direction	20	N/A
14	L_I	Length in X direction	20	km
15	L_J	Length in Y direction	20	km
16	L_K	Length in Z direction	20	km
17	ρ	Density	3000	kg/m ³
18	λ	Lamé's constant	30	Gpa
19	μ	Shear modulus	30	Gpa
20	$refvisc$	Reference viscosity	0.125e-17	Pa ⁻ⁿ /s
21	$activationE$	Activation Energy	276.0e+3	kJ/mol
22	vis_{min}	viscosity minimum cutoff	1.0e+18	Pa * s
23	vis_{max}	viscosity maximum cutoff	1.0e+27	Pa * s
24	$srexponent$	Power of power law in viscosity	3.05	N/A
25	$\varepsilon_{p_i}^1$	initial plastic strain for piecewise Type 1 weakening	0	N/A
26	$\varepsilon_{p_i}^2$	initial plastic strain for piecewise Type 2 weakening	0	N/A
27	$\varepsilon_{p_e}^1$	end plastic strain for piecewise Type 1 weakening	0.1	N/A
28	$\varepsilon_{p_e}^2$	end plastic strain for piecewise Type 2 weakening	0.33	N/A
29	C_i	initial Cohesion for piecewise weakening	44	Mpa
30	C_e	end Cohesion for piecewise weakening	4	Mpa
31	ϕ	Friction angle	30	°
32	$remesh_{timestep}$	Remesh when timestep reach its value	400000	N/A
33	$remesh_{length}$	Remesh when the global minimum of the ratio of the volume of a tetrahedron to one of its surface area	0.6	N/A
34	$topTemp$	Surface temperature	0	°C
35	$bottomTemp$	Bottom temperature	1300	°C
36	V_x	Half spreading rate	7.9e-10	m/s

3 Results

3.1 Reference model

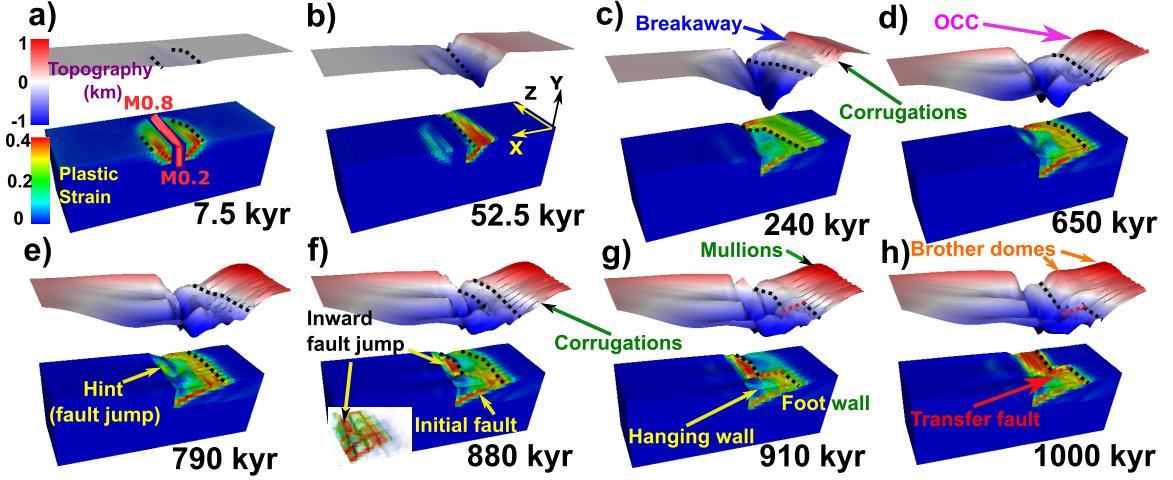


Figure 8: Evolution of plastic strain and surface topography of the reference model M28LinT1. Each snapshot shows plastic strain plotted on the model domain, the five times exaggerated topography and the time at which it is taken. The initial seafloor is at 0 km of elevation. The black dashed lines are the terminations of faults. The red dashed lines in g) and h) are the transfer faults that connect the terminations. The inset in f) plots plastic strain with opacity linearly proportional to plastic strain value.

I consider the model with M varying linearly from 0.2 to 0.8 along the ridge axis with type 1 weakening rate as the reference model and denote it as M28LinT1 following the shorthand notations given in Table 1. I describe the general model behaviors in this section and then provide details and mechanisms of the major structural features of the model in the next section.

For the first 7.5 kyr (Figure 8.a), normal faults, represented by localized plastic strain, begin to form near the ridge axis. Because stresses due to plate extension accumulate faster at the lower M side than at the higher M side, faults first initiate at the lower M side and then propagate to the higher M side. Such an asynchronous initiation of faults along the ridge axis creates offset in breakaway at later stages, i.e., the breakaway at the lower

M side moves further away from the ridge axis than that of the higher M side (Figure 9).

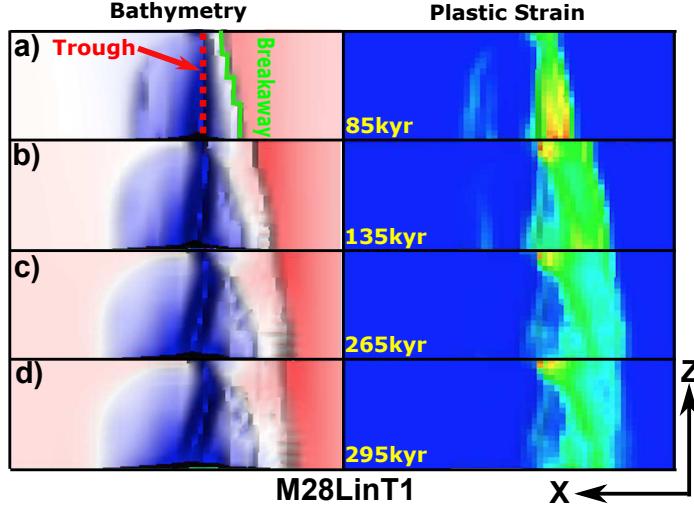


Figure 9: Bird's eye view of a breakaway in the reference model at a) 85 kyr, b) 135 kyr, c) 265 kyr and d) 295 kyr. The breakaway is marked by green solid line in a). A narrow zone of depressed topography (“trough”) is marked by red dashed line inside the median valley (blue area) in a). Color scales are the same with those in Figure 8.

Likewise, the model produces a median valley that widens and deepens with rates inversely proportional to the M value (i.e. rate of local magma supply) (Fig. 8a-c; Fig. 9a-d).

By 52.5 kyr (Figure 8.b), the normal fault on the right hand side of the ridge axis remains active while the one on the left becomes inactive. As the active fault slips, crust at the footwall bends in a clockwise rotation as is illustrated in Figure 10. The upper part of the active fault plane (shown as plastic strain in the model) is exposed to the seafloor.

The active normal fault on the right rotates to a lower dip of $\sim 30^\circ$ at the root of the fault and to $\sim 0^\circ$ at the exposed fault interface after about 240 kyrs (Fig. 8.c). However, the normal fault at the higher M side (especially for $M > 0.7$) experiences less fault rotation and the termination of the fault is closer to the ridge axis. The maximum relief between the breakaway and the trough inside the median valley becomes larger than 1 km. In addition, ~ 2 km wavelength corrugations begin to form between the breakaway and

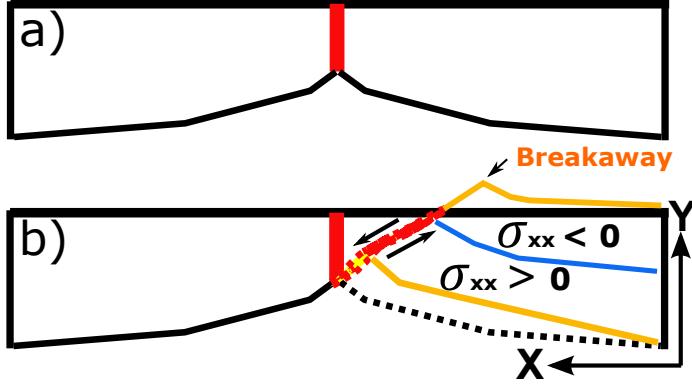


Figure 10: Illustration of the development of bending stress in lithosphere associated with faulting at the MOR. The blue line is the neutral plane where $\sigma_{xx} = 0$. Above the neutral plane is compression ($\sigma_{xx} < 0$) and beneath it is tension ($\sigma_{xx} > 0$).

termination at the lower M side ($M < 0.35$). I discuss the formation mechanism for the corrugations in Discussion.

By 650 kyr (Figure 8.d), the detachment fault reaches its lowest dip angle and its termination stops moving away from the ridge axis. The original breakaway of this detachment has already moved out of the model domain. The total fault offset at this point is greater than the thickness of the crust and thus would be sufficient for exhuming the upper mantle materials.

A new near-axis fault first appears at the center of the model domain with $M \in (0.5, 0.65)$ and then propagates in a positive z direction (Fig. 8.d,e). At this time, the initial detachment fault is still active and takes up most of the extension.

The new near-axis normal fault at the higher M side cuts through the hanging wall of the detachment fault at 880 kyr (Fig. 8.f). It coexists with the initial detachment fault and begins to accommodate most of the intra-plate extension. This event is called the “inward fault jump” [Tucholke et al., 1998; Dick et al., 2008].

By 910 kyr (Fig. 8.g), the inward fault jump completes in the $M > 0.5$ region: the new high-angle fault takes up all the extension and the initial detachment fault becomes completely inactive. The block that was previously a hanging wall to the detachment becomes a footwall of the new fault, passively moving with the plate to the negative x -axis direc-

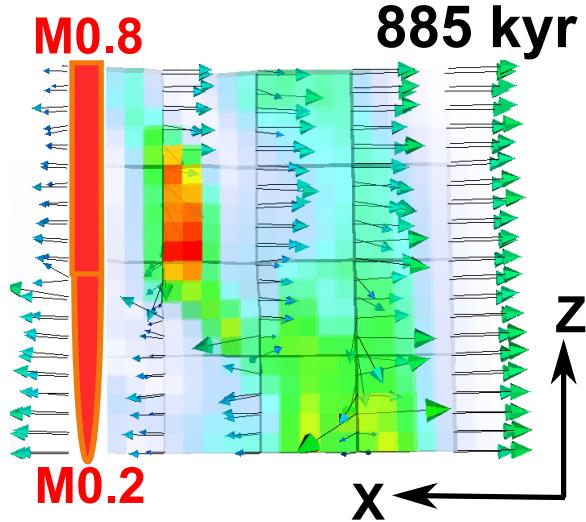


Figure 11: Bird's-eye view of velocity field with plastic strain plotted with opacity linearly proportional to its value. (color scale is the same as Figure 8.a))

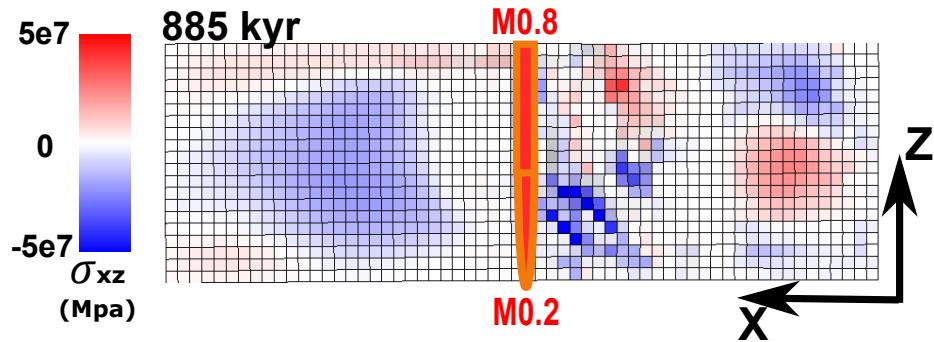


Figure 12: Bird's eye view of σ_{xz} .

tion. At the lower M side, the detachments is still active and the hanging wall continues to move toward the positive x -axis direction (Figure 11). This dextral sense of relative motions between the high and the low M side produces a region of sinistral shear stress σ_{xz} (Figure 12) and eventually creates a transfer fault (Fig. 8.h). As the inward jumped fault evolves, another dome ajacent to the initial OCC is produced at the higher M side by 1000 kyr.

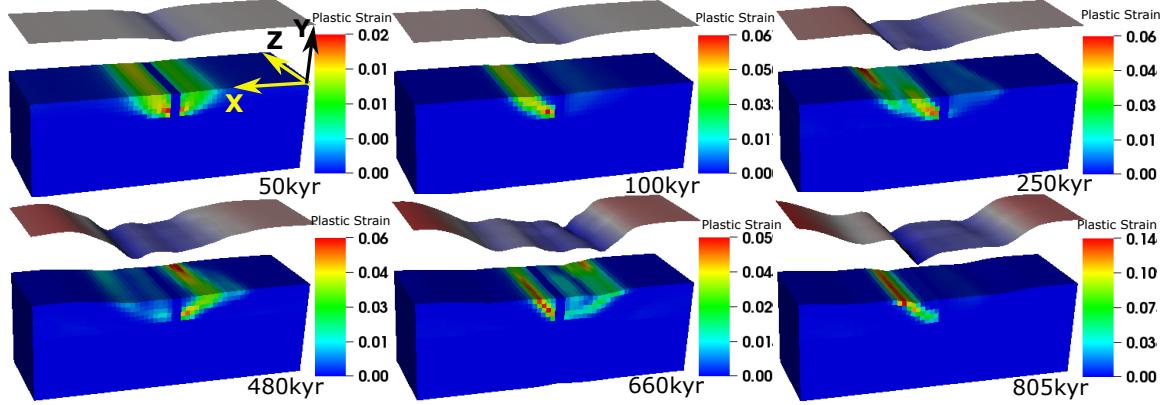


Figure 13: Evolution of plastic strain and surface topography of the model: M88ConT2 (Table 1). (color scale of topography is the same as Figure 8.a)

3.1.1 Constant M model M88ConT2

As a comparison to the varying M models, a constant M model is run. It shows similar behaviors with a corresponding 2D model from the previous studies and does not show along-ridge variations in terms of morphology and faulting.

As shown in Figure 13, model M88ConT2 produces a ~ 20 km wide and $1\sim 2$ km deep median valley, which is similar to the generally observation of the Mid-Atlantic Ridges. The width and depth of the median valley is almost constant along the ridge as contrast to the varying M models. The variation of the location of the breakaway and termination along the ridge that is mentioned in the reference model (M28LinT1) does not show up. Because the magma supply is constant along the ridge with $M = 0.8$, there is no stress perturbation along the ridge. Thus, the normal faults along the ridge initiate at the same time and the slipping rate of the fault is also constant along the ridge axis. The synchronized fault initiation results in no offset between breakaways and the constant slipping rate produces no along ridge axis variation in the position of the termination. In addition, neither corrugations nor mullion structures are generated. Normal faults alternate on each side of the ridge axis with a period of ~ 10 km. This fault alternation produces symmetrical high frequency abyssal hills.

3.2 Main structural characteristics

I describe seven structural characteristics of the reference model: Location of the termination, geometry of the trough, inward fault jump, fault alternation, mass wasting, hourglass-shaped median valley and corrugations and mullion structures. Since the details of these features differ among the models, they are useful for delineating and contrasting complicated model behaviors.

3.2.1 Location of termination

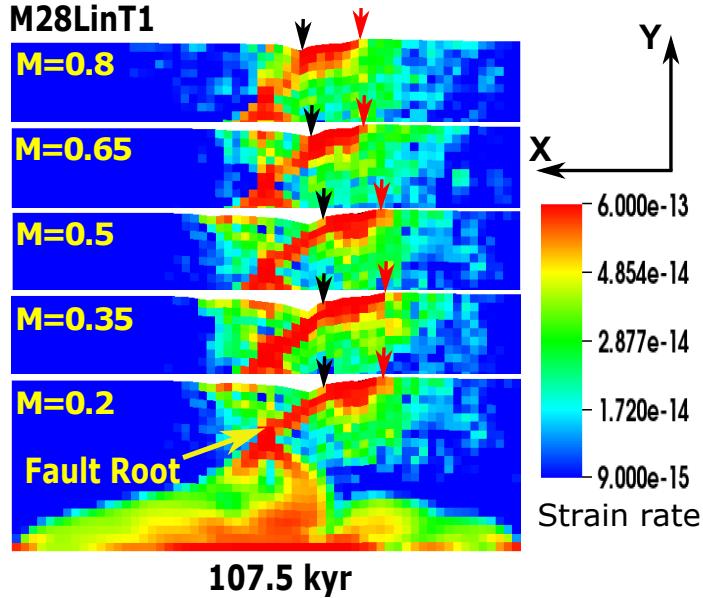


Figure 14: The second invariant of strain rates plotted on the reference model's vertical cross-sections along the ridge at 107.5 kyr. Terminations and breakaways are marked by black and red arrows.

Location of a fault termination varies along the ridge as indicated by black dashed lines in Figure 8 and black arrows in Figure 14). The distance between the termination and the ridge axis as well as the dip angle of the buried portion of the detachment do not show correlation with M values when $M < 0.5$. In contrast, when $M > 0.5$, the distance from the ridge axis to the termination decreases with increasing M (Figure 14). The crust

at the footwall experiences less bending and the detachment fault remains in a higher angle as well as that the terminations are closer to the ridge axis. Because the root of the fault is slowly pushed away from ridge axis while the termination of the faults are closer to ridge axis due to a later initiation of the fault. , One thing needs to be noted is that the trough at the higher M side correspond to the terminations but detached from the terminations at the low M side ($M < 0.5$) as shown in Figure 14.

3.2.2 Geometry of trough

The depressed narrow region that develops in the median valley is termed as “trough”. The reference model showed that its shape in the bird’s eye view evolves from a straight line parallel to the ridge axis to a line oblique to the ridge axis (Figure 9). Initially, the trough along the ridge corresponds to the termination. However, at the lower M side ($M < 0.5$), as the normal fault rotates to a lower dip at the lower M side, the trough is no longer coincident with the fault termination and is moving slowly to the left because the hanging wall is pulled by the conjugate plate. However, the trough on the higher M side ($M > 0.5$) is pushed away from the ridge axis [Tucholke et al., 2008]. But since the trough cannot bypass the termination, the trough at $M = 0.8$ is restricted closer to the ridge axis. Together it generates the curved shape of the trough (Figure 9).

3.2.3 Inward fault jump

The inward fault jump (e.g., Fig. 8) occurs in all the models when an old fault locks and a new one forms near the ridge axis [e.g., Lavier et al., 2000, Olive et al., 2014]. The new fault usually forms on the high M side first and it connects to the existing detachment developing on the lower M side ($M < 0.5$), producing a curved fault termination. Unlike fault alternation described below, the inward fault jump occurs on the same side of the ridge axis with its along-axis extent corresponds to the $M > 0.5$ region.

3.2.4 Fault alternation

When M is high enough, a normal fault first forms on one side of the ridge axis but another fault forms on the other side when the first one locks (Fig. 13) [Buck et al., 2005, Tucholke et al., 2008]. This behavior is termed as “fault alternation”.

3.2.5 Mass wasting

Mass wasting occurs on the exposed surface of a low-angle normal fault. When a weak fault zone becomes gravitationally unstable and is detached from the underlying material, the detached layer flows towards the ridge axis and the lower M side with a velocity ~ 10 times faster than the half spreading rate (Figure 16; Figure 17.b,e (first row))). The mass wasting produces a continuous fault scarp with a relief of ~ 1 km along the initial break-away.

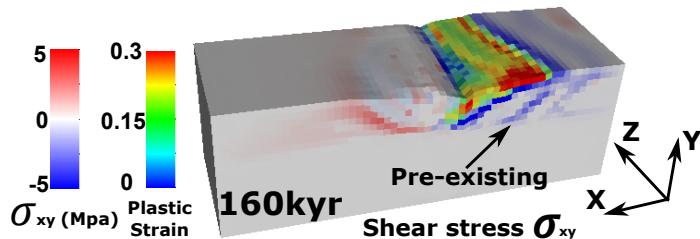


Figure 15: The weak detachment fault tip reaches the pre-existing shear stress. Plastic strain is plotted with opacity linearly proportional to its value.

Immediately after an event of mass wasting, the fault trace jumps towards the ridge axis (Figure 17.d versus f (third row: plastic strain)). Then, σ_{xy} and σ_{xz} soon ^{EC:} fill in the area [What do you mean by “stresses filling an area”?] between cut-back created fault scarp and the new termination (Figure 17.f (fourth and fifth row: σ_{xy} and σ_{xz}). ^{EC:} [don’t understand the rest of this paragraph.] After that, the termination at the higher M side extends faster and further than the lower M side (Figure 20.a~d). This is because during the cut-back, tensional stress due to bending is released at the lower M side but continues to

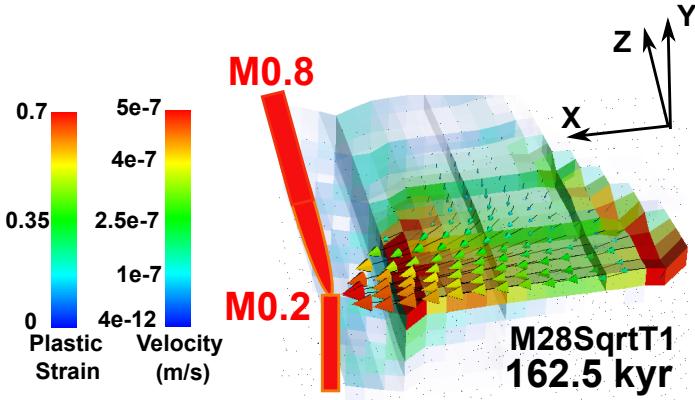


Figure 16: Velocity of the mass wasting hanging wall. Magnitudes of the velocity are shown by the colors of the arrow heads. Plastic strain is plotted with opacity linearly proportional to its value.

accummulate at the higher M side. At the lower M side, it needs time to reach the previous tensional stress state and then starts from there to pull the new termination away from the ridge axis. While at the higher M side the increasing tensional stress directly leads to a fast extending termination.

3.2.6 Hourglass-shaped median valley

All the models develop an hourgalss-shaped median valley although the geometry of the median valley changes with time and with the functional form of M variation. A median valley of the reference model initially has a uniform width along the ridge but is deeper on the lower M side where normal faults first form. By ~ 100 kyr, the fault on the negative x -axis side of the ridge axis does not propagate to the higher M side of the ridge and becomes inactive (Figure 21.b). It produces a depressed topography curve following the inactive fault trace, which is further away from the ridge axis at the lower M side but closer to ridge axis at the higher M side. On the other side of the ridge axis, as the active fault rotates to a lower dip angle, [EC](#):breakaways that are further away from the ridge axis at the lower M side while closer to the ridge axis at the higher M side [So, the same thing is happening on both sides?]. This along-ridge variation in the location of the breakaways act as another boundary of the hourglass. By ~ 170 kyr (Figure 21.c), the hourglass-shaped me-

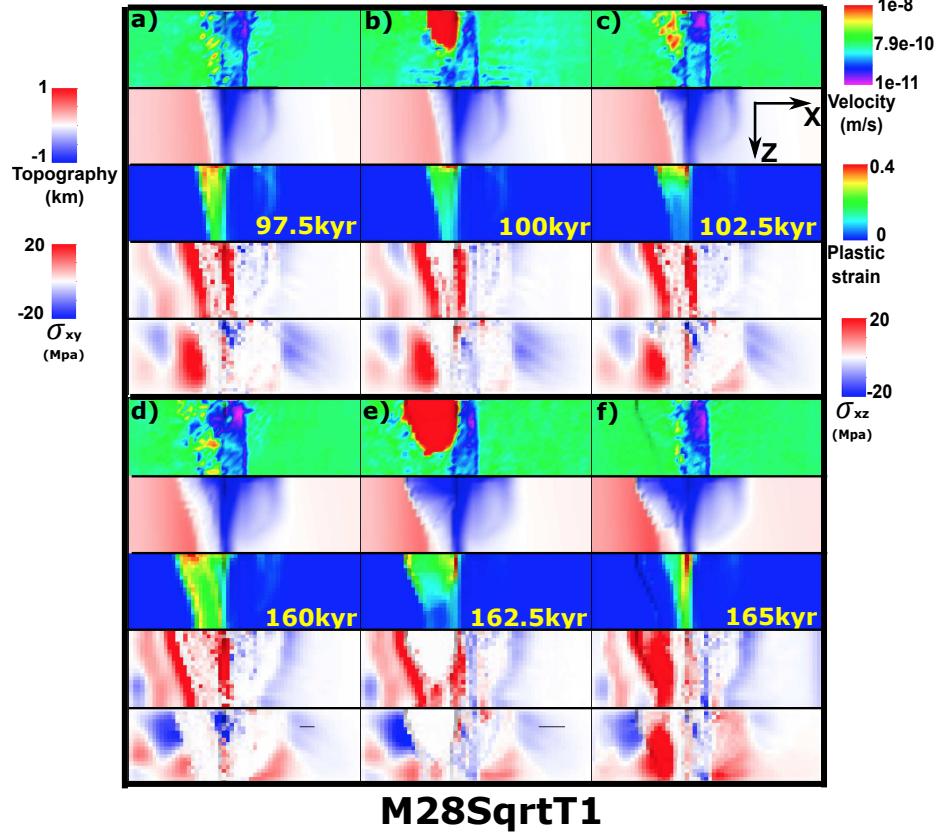


Figure 17: Plastic strain, topography and stresses evolution for M28SqrtT1.

dian valley continues to widen and deepen. Since the area of the cross-section along the ridge inside the hourglass shape median valley is approximately inversely proportional to the local M values, the shape of the hourglass varies with different M variations.

3.2.7 Corrugations and mullion structures

Both corrugations and mullion structures are linear structures parallel to the spreading direction. As shown in Figure 8.f, at the $M < 0.5$ area on top of the OCC surface, corrugations show a uniform wavelength of ~ 2 km with hundreds of meters in amplitude. While at the higher M side of the ridge (Figure 8.g), a mullion structure of a wavelength of ~ 7 km. In spite of morphological similarity, corrugations and mullion structures have different formation mechanisms.

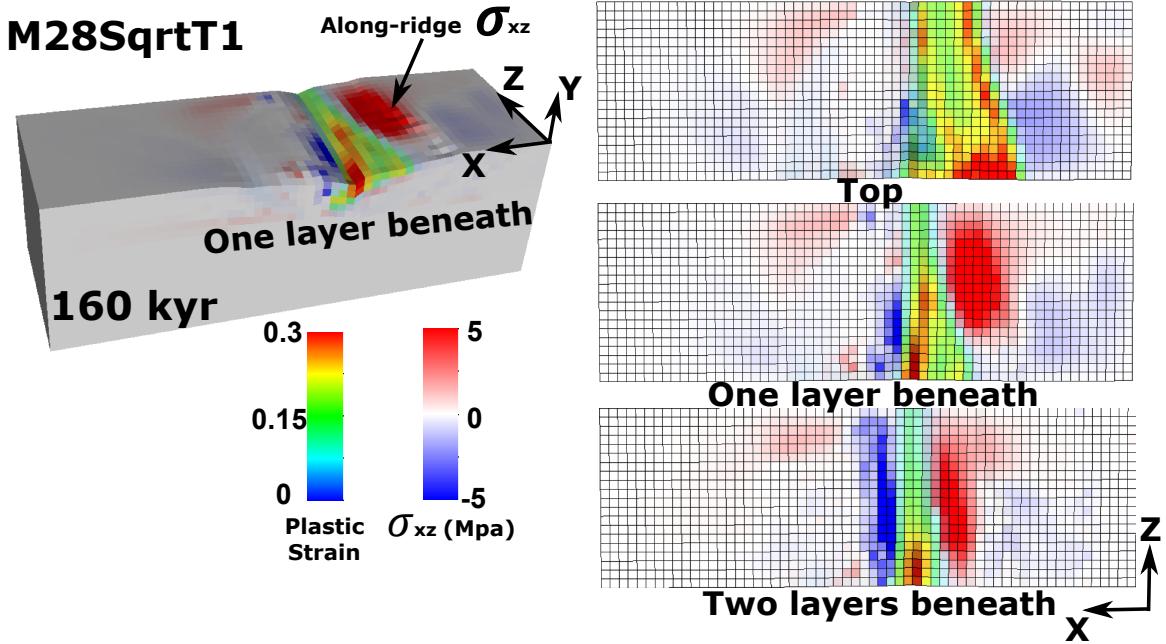


Figure 18: Along ridge axis σ_{xz} (bird's-eye view of three layers of model domain) (positive(red) means clockwise). Plastic strain is plotted with opacity linearly proportional to its value.

Corrugations

The corrugation starts at the breakaway as a response to tensile stress in the z -axis direction. As shown in Figure 23, when the plastic strain reaches or exceeds 0.1 (red color), based on type 1 weakening, the cohesion decreases to 4 Mpa. With a 30° friction angle, tensile failure is declared when the σ_{zz} reaches ~ 7 Mpa (yellow color). The tensile stress is due to the asynchronous faulting along the ridge. Faulting initiates earlier at the lower M side of the ridge. As the fault offsets more on the low M side than on the high M side, the footwall of the fault rotates and gets uplifted more on the low M side. The footwalls along the ridge generate the isochron-parallel tensile stress σ_{zz} . Since σ_{zz} follows along the moving tip of the plastic strain, the plastic strain together with σ_{zz} generate tensile failure that extends away from the ridge axis and thus produces the linear corrugations that are parallel to the spreading velocity. Detail analysis along with simpler model experiments are given in the “Discussion” section.

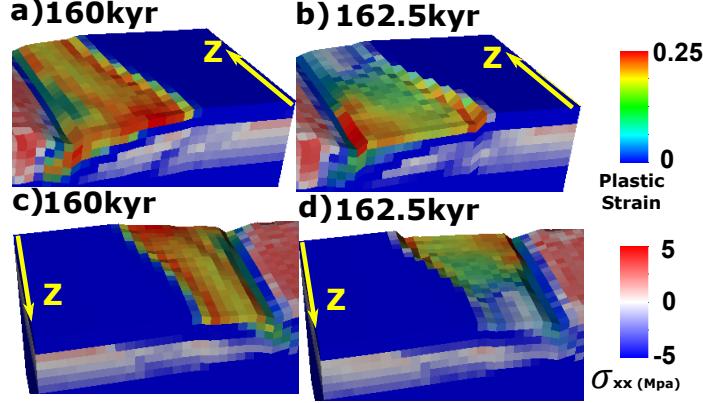


Figure 19: M28SqrtT1 (Table 1). Bending stress drop (a versus b) at the lower M side due to the cut-back. No bending stress drop (c versus d) at the higher M side as a comparison.

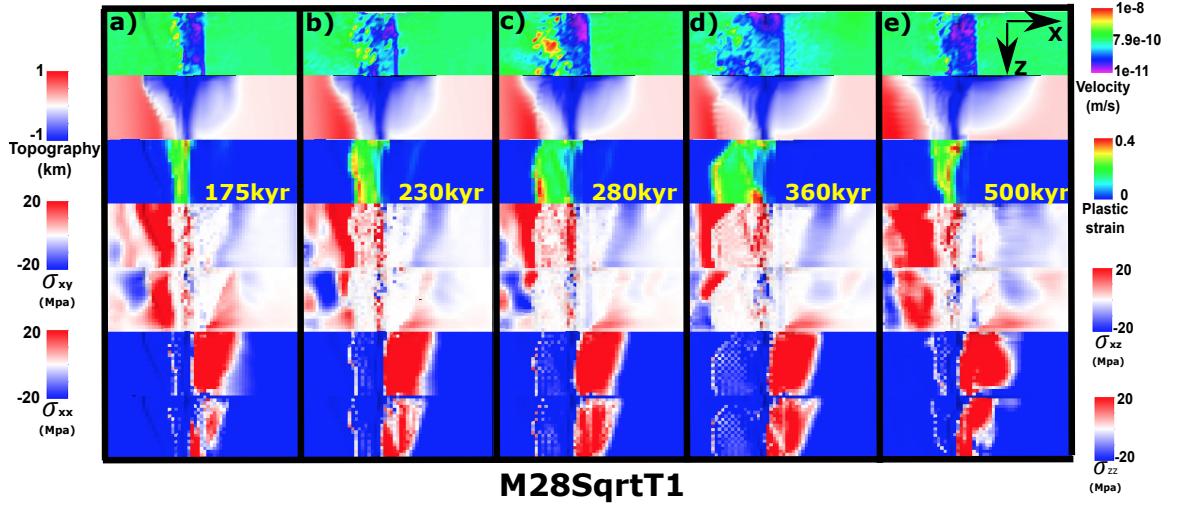


Figure 20: New fault front chase after cut-back.

Mullion structures

Mullion structures observed in the models are formed by the along-ridge variation in the location of the termination due to the evolution of faulting. They usually appear at where the termination is closer to the ridge axis. The shape of the footwall follows the trace of the termination as it is exhumed to the surface. At where the termination is bent inward to the ridge axis, an “initial dome” (Figure 24.a) is produced once the hanging wall is exhumed to the seafloor. The wavelength of the mullion structure is determined by the shape of the termination. If the pattern of the termination lasts for a long time and the

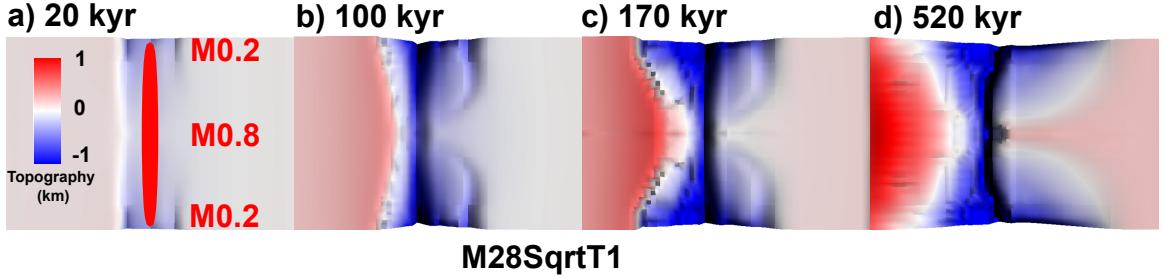


Figure 21: Bird's-eye view of the evolution of the hourglass shape median valley. It is generated by attaching the topography of the M28SqrtT1 model to its mirror reflection by assuming symmetrical M variation (0.2 to 0.8 to 0.2).

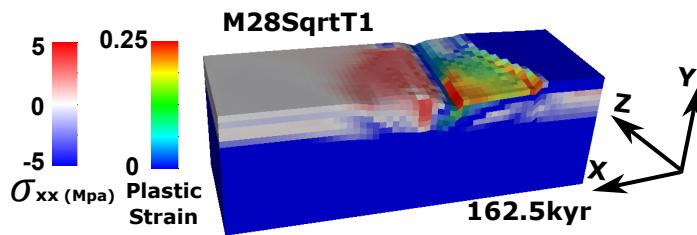


Figure 22: Higher σ_{xx} shows inside the median valley on the negative x -axis direction of the ridge axis.

footwall of the detachment fault keeps being exhumed to the surface following the trace of the detachment fault termination, a mullion structure is produced (Figure 24.b)).

3.3 Effects of the functional forms of M variation

3.3.1 Models with M between 0.2 and 0.8 and Type 1 weakening

Major differences among models with M between 0.2 and 0.8 (M28) and Type 1 (T1) weakening lie in the model behaviors in terms of the inward fault jump, the mass wasting and the hourglass-shaped median valley. None of the three models show fault alternation.

Inward fault jump

Only the models with linear and sinusoidal M variations show inward fault jumps. No inward fault jump occurs when M varies with square root of the along-ridge distance.

EC: [The termination of a detachment fault retreats](#) [How is this different from an in-

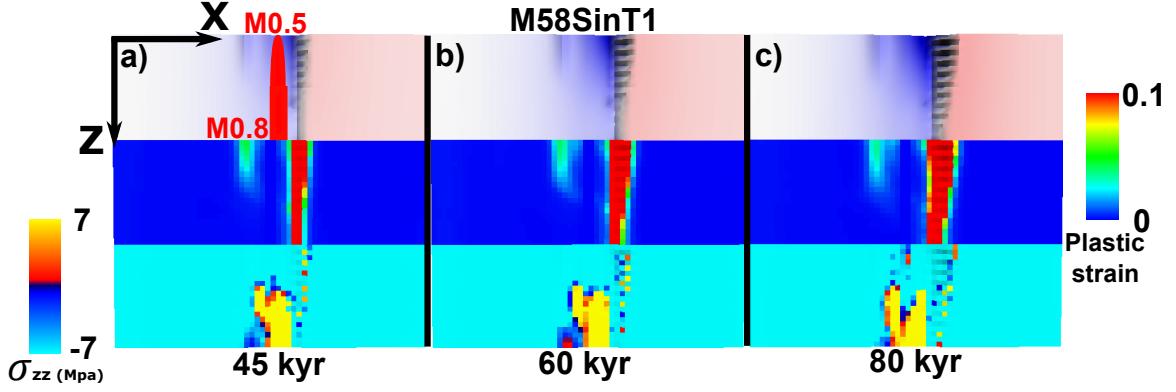


Figure 23: Bird's eye view of the evolution of the corrugations. Color scales for the topography is the same as Figure 8.

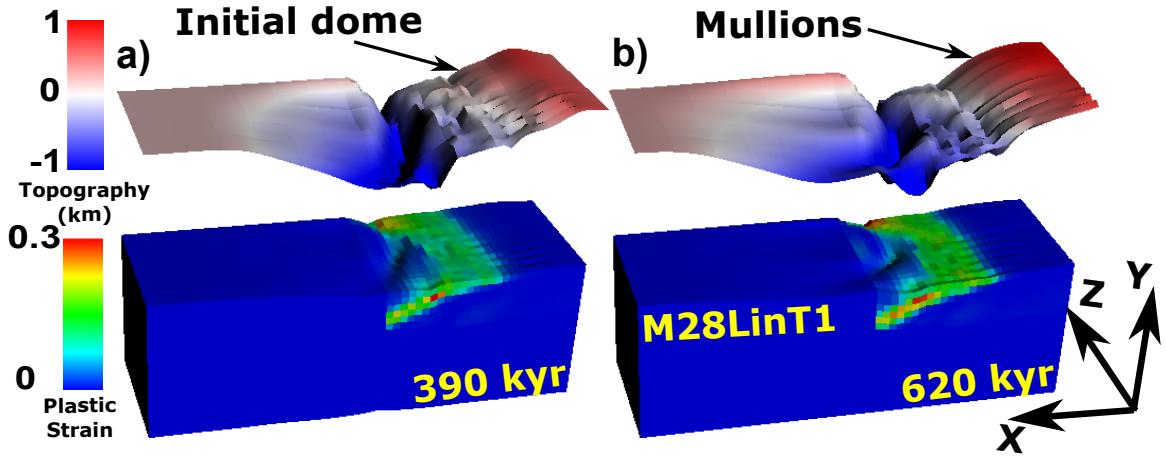


Figure 24: Evolution of mullion structures.

ward fault jump?] towards the ridge axis after a mass wasting and the detachment fault is maintained near the ridge axis ^{EC:} [I don't understand this sentence.]. Timing of inward jumps and the along-strike dimension of the new faults are different between linear and sinusoidal models. The inward jump at the higher M side creates a new fault that accommodates most of the extension at ~ 900 kyr ^{EC:} [in which model? The reference model?](Figure 26, Figure 8.f). The first inward fault jump occurs earlier in a sinusoidal model (e.g., at ~ 550 kyr in MXSinXX) than in the linear reference model. Also, the along-strike length of the new fault in the sinusoidal model is ~ 14 km (Figure 26). The timing difference between the linear and sinusoidal models is because M28SinT1 con-

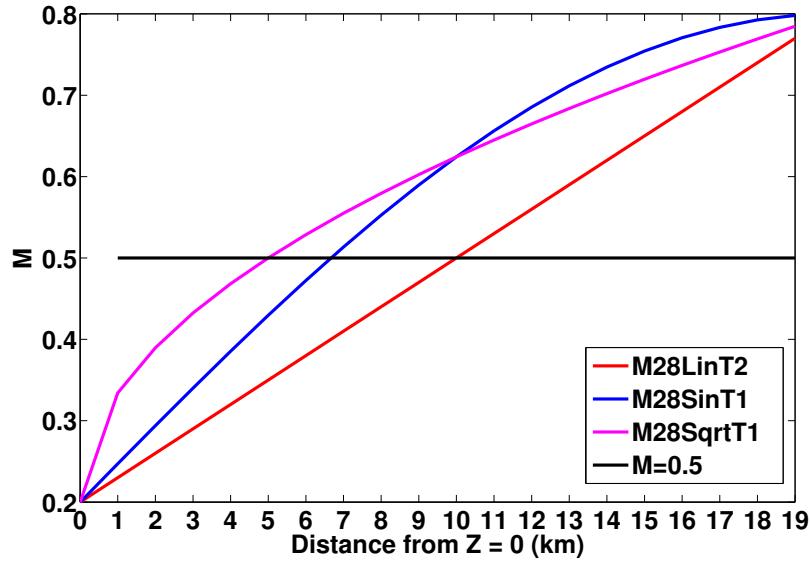


Figure 25: Three functional forms of M variation. M begins to exceed the $M = 0.5$ black line at $Z=10$ km, 7 km, 5 km for M28LinT1, M28SinT1 and M28SqrtT1 respectively.

sistently has a higher M values than the M28LinT1 (Figure 25), which results in that the initial detachment fault at the higher M side ($M > 0.5$) of M28SinT1 is pushed off axis faster than M28LinT1 and thus forming an earlier inward fault jump. The length difference is because M28SinT1 has a greater length along the ridge axis of $M \geq 0.5$ (Figure 25).

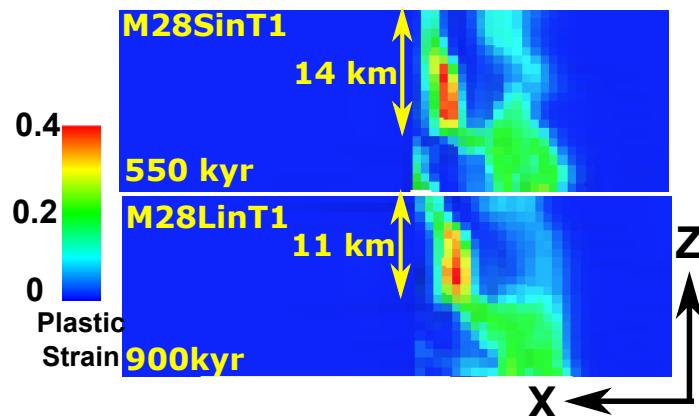


Figure 26: Bird's-eye view for comparing the length and timing of inward fault jump.

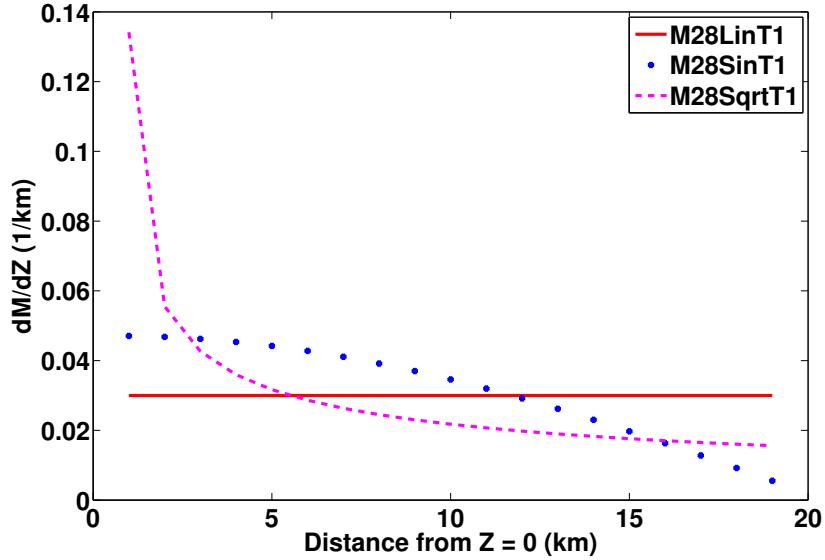


Figure 27: $\partial M / \partial Z$ comparision.

Mass wasting

Mass wasting occurs only in the square root model among thoe group of M28 and T1. Qualitatively, it is because the square root profile of M variation has a much higher value of $\frac{\partial M}{\partial Z}$ at the lower M side (Figure 27), which implies a larger along-ridge shear stress σ_{xz} as well as a larger difference in σ_{xy} along the ridge that result in the decoupling between the higher and lower M sides hanging walls. *EC:* [can't understand the last sentence.]

Hourglass-shaped median valley

As shown in Figure 28, differences among the three models are identified. At 160 kyr, the median valley in the model M28SinT1 has *EC:* the smallest cross-sectional ($x-y$) area [You can't call it a cross-sectional area. Maybe a map-view area? Or just width? And I don't see if it's really the smallest in M28SinT1.] at the higher M side. *EC:* While [You can't use while this way. Please look up the usage of this word.] At the lower M side, the square root model has the smallest area of the cross-section. This is because the *EC:* cross-section area inside width of the median valley is inversely proportional to the local M value along the ridge. Moreover, the *EC:* breakaways [Mark these on the figure.] at the

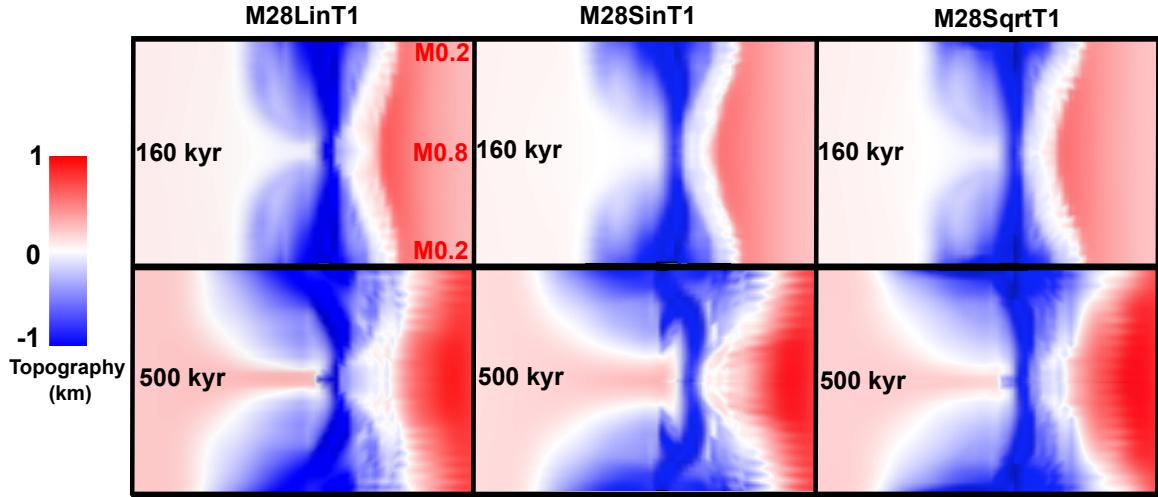


Figure 28: Bird's eye view of the topography. *EC:* (without vertical exaggeration.) [don't need to say. When you did it, you need to say so.] *EC:* [Say somewhere that you put the original and an mirror image together in this figure.]

lowest M value in the linear and sinusoidal models becomes parallel to the ridge axis while the breakaway of the square root model maintains the oblique trend. *EC:* In addition, M28SinT1 has a trough inside the median valley with the highest curvature. [I can't see this. Can you mark this on the figure?] At 500 kyr, M28SinT1 has *EC:* the narrowest median valley [Still can't agree.] at the higher M side and the high topography zone on the left hand side of the ridge axis is the widest. Integrating the topography at the left hand side of the ridge axis of the three models, M28Sqrt has the largest value of integration since it has the largest integration of M along the ridge axis *EC:* [Did you actually do the integration? Integrating over what? Why integration? Why the largest integration of M means the greatest integration of topography?].

3.4 Effects of the weakening rate

Among the eleven 3D models, three pairs of models have the same parameters but differ only in the weakening rate: M57SinT1/T2, M58SinT1/T2 and M58SqrtT1/T2. I compare them pairwise to single out the effects of the weakening rate on the model behaviors.

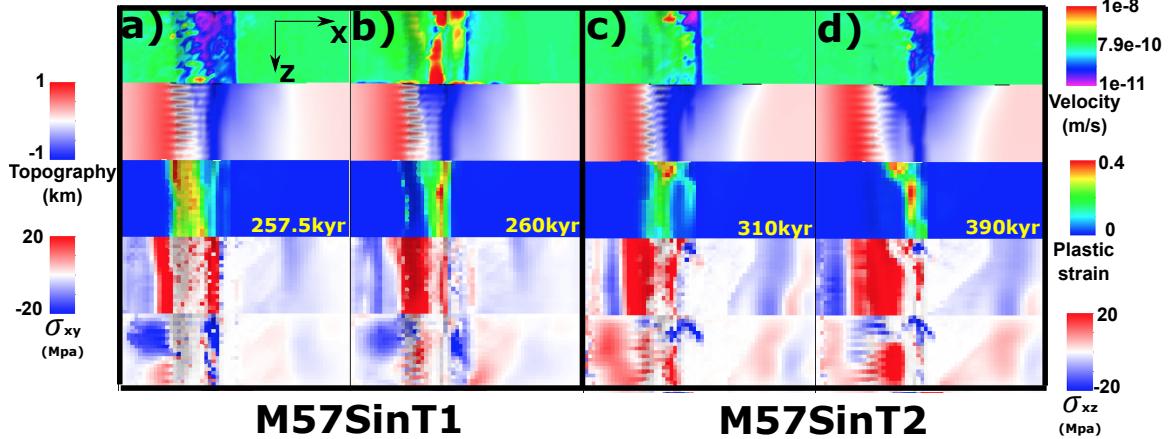


Figure 29: M57SinT2 versus M57SinT1 (Table 1)

Initially, both M57SinT1 and M57SinT2 develop normal faults on both sides of the ridge axis at the lower M side. However, the faults complete propagation towards the higher M side earlier in the fast-weakening model (M57SinT1) than in the slow-weakening model (M57SinT2). In M57SinT1, a mass wasting event occurs at ~ 260 kyr but the initial fault remains active without experiencing an inward fault jump (Figure 29.a and b). In contrast, the slower-weakening model experiences an inward fault jump at the higher M side ($M > \underline{EC:} 0.55$ [Since the range is 0.5-0.7, isn't this value on the lower M side?]) but the initial fault remains active in the $M < 0.55$ region (Figure 29.c and d).^{EC:} [Without further annotations on the figure, referring to the figure wouldn't help much.] The width of the “median valley at the lower M side is wider in M57SinT2 than in M57SinT1 (Figure 29.c, d versus a, b)^{EC:} [Again, the figure doesn't help much because one wouldn't know where to look to visually measure the width.] The amplitude of the corrugations of the faster-weakening model (M57SinT1) is greater because the faster weakening rate allows a faster decrease in the cohesion and the lowered cohesion promotes tensile failure in the isochron-parallel direction, which produces the corrugations.

The pairs, M58SinT1/T2 and M58SqrtT1/T2, show a consistent difference that the fault alternation occurs only in the slow weakening models.

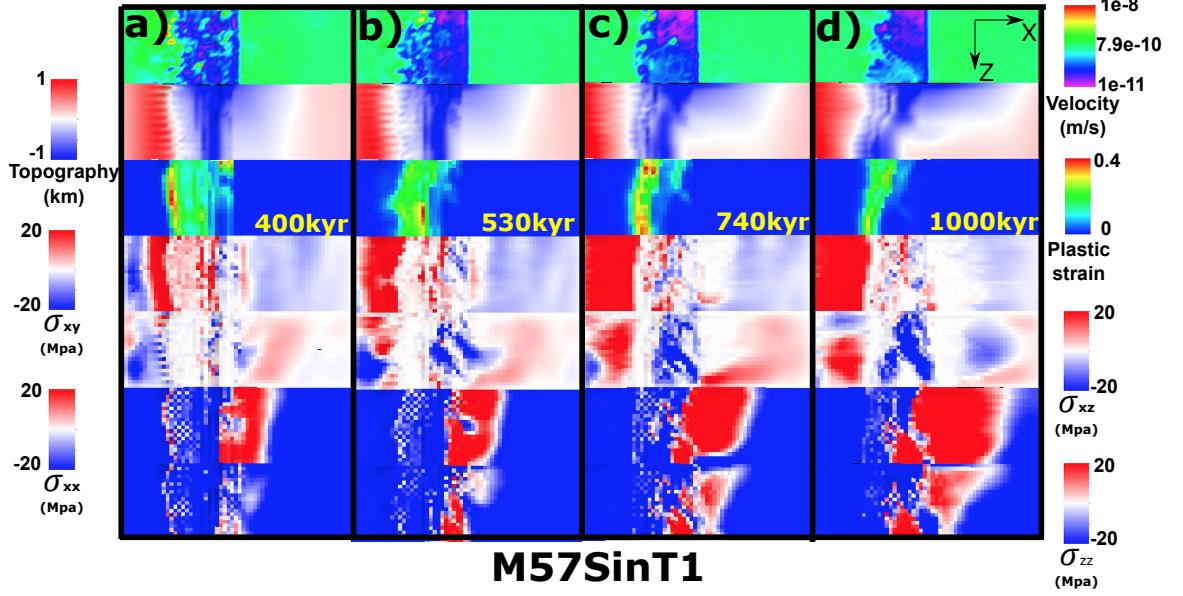


Figure 30: Faulting and stress evolution for M57SinT1.

3.5 Effects of the range of M variation

Generally, M57 and M58 models create a median valley much narrower and shallower than that of M28 models. Although the difference between the M ranges, 0.5-0.7 and 0.5-0.8, is seemingly small, the behaviors of M57 models and M58 ones are noticeably different. A major difference between M57SinT1 and M58SinT1 is that the model with a greater M range (M58) shows a higher frequency of inward fault jumps, mass wasting events and connection of the offset fault zones. Also, the greater M range appears to promote the fault alternation when combined with a nonlinear M profile and a slow weakening (T2): Only the models M58SinT2 and M58SqrtT2, exhibit fault alternation.

4 Discussion

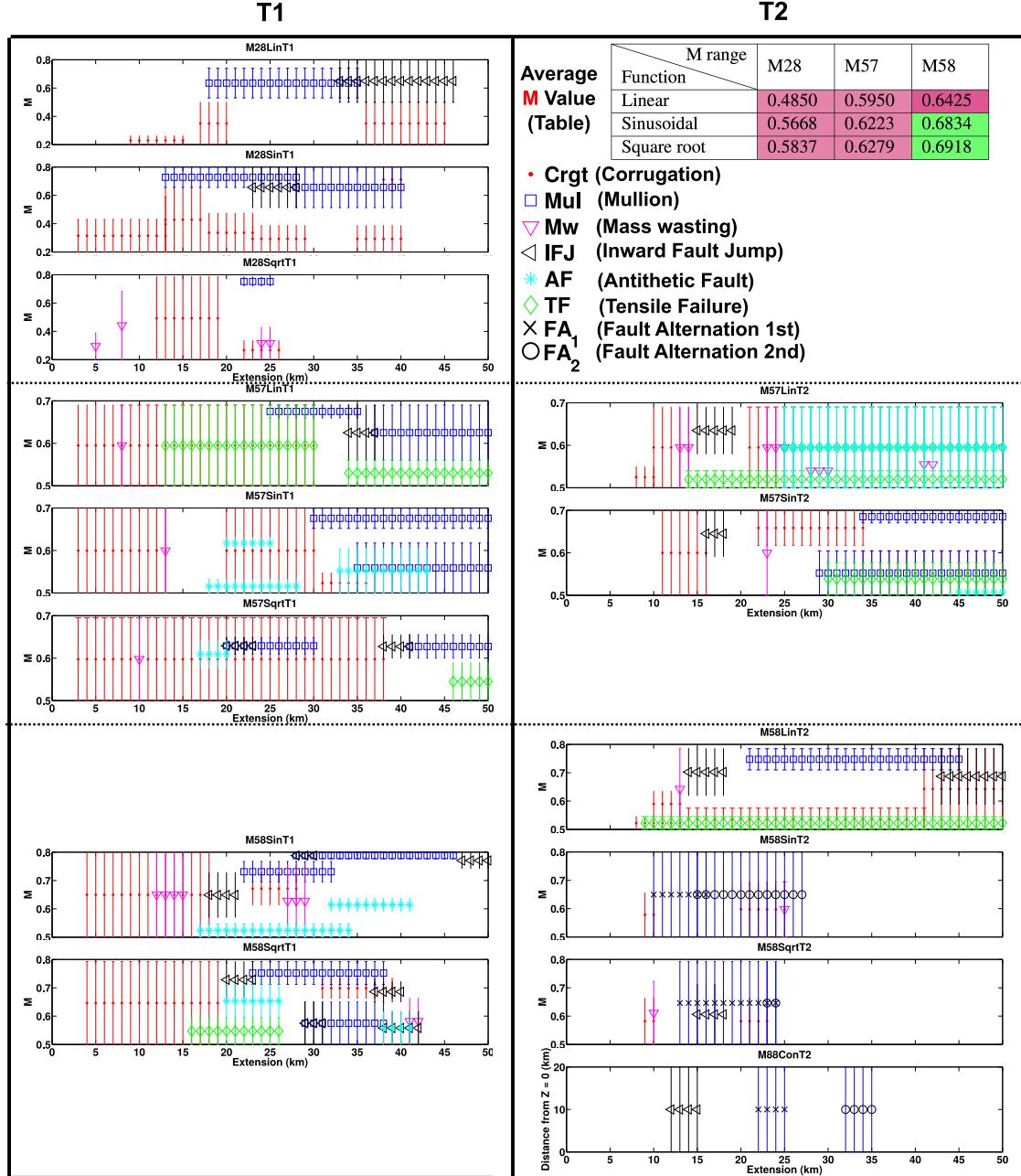


Figure 31: Initiation and duration of main structural features. *EC*: [Labels of an individual plot are too small. Symbols can be bigger, too.]

Model results show systematic changes with the average M value (\bar{M}), which is the integration of M along the ridge divided by the length of the ridge segment. As seen in the table of Figure 31, $\bar{M}_{M58} > \bar{M}_{M57} > \bar{M}_{M28}$ and within each M range, \bar{M} for functional form with square root is higher than that for sinusoidal and linear.

Comparing the models with slower weakening rate (type 1) (left column of Figure 31), when \bar{M} increases, faulting (i.e. inward fault jump, antithetic fault) as well as other main model behaviors (i.e. corrugation, mullion, mass wasting, tensile failure) become more complex and dynamical with higher recurrence frequencies.

Comparing M28SinT1 and M28LinT1 shows that the inward fault jump for M28SinT1 (with higher \bar{M}) happens earlier and lasts a shorter period of time. As \bar{M} increases to 0.5837 for M28SqrtT1, mass wasting first shows up. During the mass wasting, a curved fault scarp with ~ 1 km relief is produced and it has a shape very similar to 13 °N, Mid-Atlantic Ridge (please refer to a detailed discussion in the next section).

When M57 are compared to M28 models, corrugations show up earlier along the whole 20 km ridge segment. All three models have mass wasting. For M57LinT1 ($\bar{M} = 0.5950$), it begins to have tensile failure which releases tensional stress in the hanging wall and results in the delay of inward fault jump. For M57SinT1 ($\bar{M} = 0.6223$), antithetic fault first shows up and together with mass wasting lead to the absence of inward fault jump. The antithetic fault at the lower M side helps to maintain a closer to ridge axis termination which results in a larger mullion structure (Figure 33.a) at the lower M side and thus produces an OCC with a shape similar to the Atlantis Massif at 30 °N Mid-Atlantic Ridge. For M57SqrtT1 ($\bar{M} = 0.6279$), it has twice of inward fault jumps and each of them produces a curved toward ridge axis termination. The curved termination is responsible for the following small scale mullion structure.

Comparing M58 and M57 models shows that the two M58 models have 3 and 4 times of inward fault jump respectively while M57 models have at most twice of that. Also, corrugations of M57 show up earlier and it might imply that corrugation favors specific range

of $\frac{\partial M}{\partial Z}$, neither too high, nor too low. Between M58SinT1 ($\bar{M} = 0.6834$) and M58SqrtT1 ($\bar{M} = 0.6918$), the major difference is M28SqrtT1 has twice inward fault jumps at the lower M side because M values are higher at the lower M side.

When comparing models with different weakening rate (type 1 versus type 2), the most obvious difference is that only two of type 2 models with $\bar{M} > 0.6425$ generates alternating faults, they are M58SinT2 and M58SqrtT2. They don't have mullion structure because the high frequency of fault alternation allows a specific pattern of termination to last only for a shorter period of time. Also, they neither have tensile failure nor antithetic fault. Note that M58LinT2 ($\bar{M} = 0.6425$) does not generate fault alternation and it indicates that not only the range of M (M58) but also the average M value along the ridge \bar{M} are responsible for the fault alternation. This provides an upper limit of $\bar{M} = 0.6425$ in our 3D models for producing long lasting detachment faults that can generate OCCs. Comparing M57T2 and M57T1 models shows that M57T2 models generally have earlier inward fault jumps but later corrugations. For the model with constant $M = 0.8$ along the ridge axis, the inward fault jump and fault alternations have a period of ~ 10 km of extension which is consistent with that of field observations. Corrugation, mullion structures, mass wasting, antithetic fault and tensile failure are not produced in constant M model M88ConT2 and it implies that varying M is necessary for producing those features.

In addition, the upper limit of $\bar{M} = 0.6425$ for allowing a long lasting detachment fault to produce an OCC in our model is consistent to the results from a near-bottom sidescan bathymetric profiler survey and sampling study of the Mid-Atlantic Ridge near 13 °N [MacLeod et al., 2009]. As shown in Figure 32, the average M value $\bar{M} = 0.63 \pm 0.05$ for the pink area where OCCs are present while when there is no OCC, $\bar{M} = 0.73 \pm 0.03$.

Under similar physical conditions of our model setup (e.g. thermal structure, rheology relationship, spreading velocity, weakening rate), the upper limit of $\bar{M} = 0.6425$ predicts a boundary value between two observed morphological end members for slow

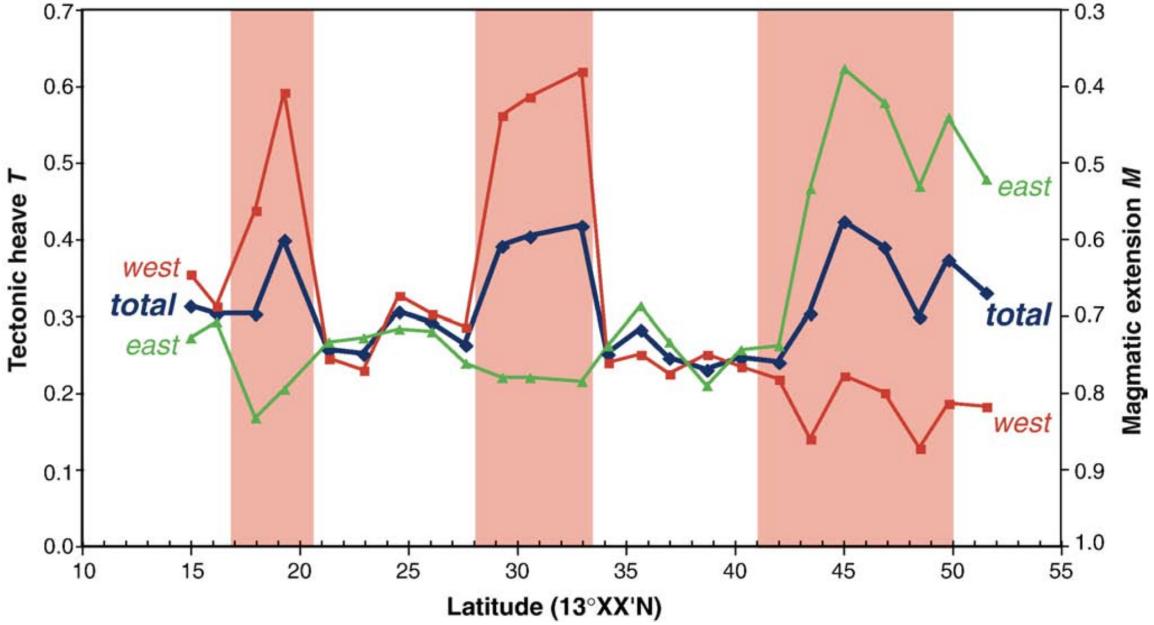


Figure 32: Along-axis variations in total accumulated tectonic heave T in the past 1.86 Ma (since chron C2n), and consequent inferred magmatic component M ($= 1 - T$) as a proportion of total plate separation (blue line). Pink shaded areas delineate loci of the active or recently active OCCs. The relative contributions of tectonic strain from the western and eastern flanks of the axis that give rise to the total heave T are shown by red and green lines respectively. Adapted from [MacLeod et al., 2009].

spreading ridges: 1) long wavelength OCCs generated by asymmetrical spreading on the detachment fault; 2) high frequency abyssal hills results from symmetrical spreading of alternating high angle normal faults. Although the number of $\bar{M} = 0.6425$ is highly consistent with natural observation (e.g. [MacLeod et al., 2009]), we still need more works for a comprehensive result because model parameter such as viscosity of the underlying asthenosphere also contributes to whether fault alternates. For example, [Allken et al., 2012] shows that higher value of viscosity of the ductile asthenosphere leads to better coupling between crust and mantle and promotes distributed multiple faulting rather than a focused long-lived detachment fault.

Previous 2D studies suggest that OCCs are most likely to form when $M = 0.3 \sim 0.5$ [Buck et al., 2005; Tucholke et al., 2008]. However, conflicts exist between model prediction and nature observation in both the upper and lower limits. For the upper limit conflict

that OCCs are observed with $M > 0.5$, Olive et al. [2010] suggests an explanation from a 2D model study that magma supplied in the ductile lower crust and upper mantle will not affect the faulting pattern and thus allows OCCs to be created under excessive diking. However, our 3D model results provides an alternative explanation that due to the along ridge coupling (i.e. torsion and shear), the region ($M = 0.3 \sim 0.5$) along the ridge that promotes stable spreading by detachment faulting helps maintain the normal fault outside the region along the ridge. Once the detachment fault initiates along the whole ridge segment, it is very hard to modify the faulting pattern especially for a faster weakening rate (type 1). Thus, the detachment fault at the higher M side ($M > 0.5$) can still last for more than 20 km of plates separation (Figure 31.M28LinT1) before the fault alternation or inward fault jump ceases the exhuming process of the ultramafic mantle rocks. This along ridge coupling can also explain the conflict at the lower limit end when OCC is produced with observed $M < 0.3$ (e.g., Dick et al., 2008; Grimes et al., 2008; Baines et al., 2008). Our 3D result suggests an extended range of M for OCCs formation than previous 2D studies ($M = 0.3 \sim 0.5$) [Buck et al., 2005; Tucholke et al., 2008].

4.0.1 Models with M between 0.5 and 0.8 and Type 2 weakening^{*EC: M58(Lin,Sin,Sqrt)T2*}

Table 3: Mean M values of the models with ^{*EC: Type 1 (or is it Type 2?)*} weakening. ^{*EC: 20 km segment. (The value is calculated by integrating M along the ridge axis and divided by the length of the model domain in z-axis.)*}

M range Function	M28	M57	M58
Linear	0.4850	0.5950	0.6425
Sinusoidal	0.5668	0.6223	0.6834
Square root	0.5837	0.6279	0.6918

Only the models with M between 0.5 and 0.8 and Type 2 weakening exhibit the fault alternation. However, the model with the linear M profile (M58LinT2) is an exception,

which shows only an inward fault jump until the end of the model time, 1.1 Myr. The behaviors of this group of models indicate that there is an lower limit in the average M values with which fault alternation occurs. According to the tabulated values of M averaged over the along-ridge distance in Table 3, the lower limit is between 0.6425 of M58LinT2 and 0.6834 of M58SinT2. Detailed analysis of the fault alternation is given in “Discussion”. ,

4.1 Comparing model results with nature observation

In this section, I compare the model behaviors with nature observations in terms of the following major model features: inward fault jump, fault alternation, mass wasting, hour-glass shape median valley, mullion structure and corrugation.

4.1.1 Inward fault jump

The term “inward fault jump” is first suggested in a study of geological and geophysical data from the Mid-Atlantic Ridge [Tucholke et al., 1998]. It is the end phase of a general evolution of an OCC as is illustrated in Figure 33. In the begining of a long amagmatic phase (Figure 33.a~c) of seafloor spreading, a high angle normal fault cuts throught the brittle lithosphere and roots in the brittle-ductile transition (BDT) (Figure 33.a). When the fault keeps slipping, the breakaway moves off axis and the fault begin to rotate to a lower dip angle (Figure 33.b). Then, the exposed fault surface roll over and as the detachment fault keeps exhuming lower crust and upper mantle rocks, it generates a dome shape megamullion (OCC). The high angle normal faults cut the detachment fault surface where it is exposed to the seafloor is probably caused by the bending stresses during footwall roll over (Figure 33.c). Then, when magma supply at the ridge center increases and pushes the detachment fault away from the ridge axis, the OCC formation is terminated by the new fault near the ridge axis which is termed as the “inward fault jump”. As shown in the figure, initially, the footwall of this new fault is mostly composed of crust material

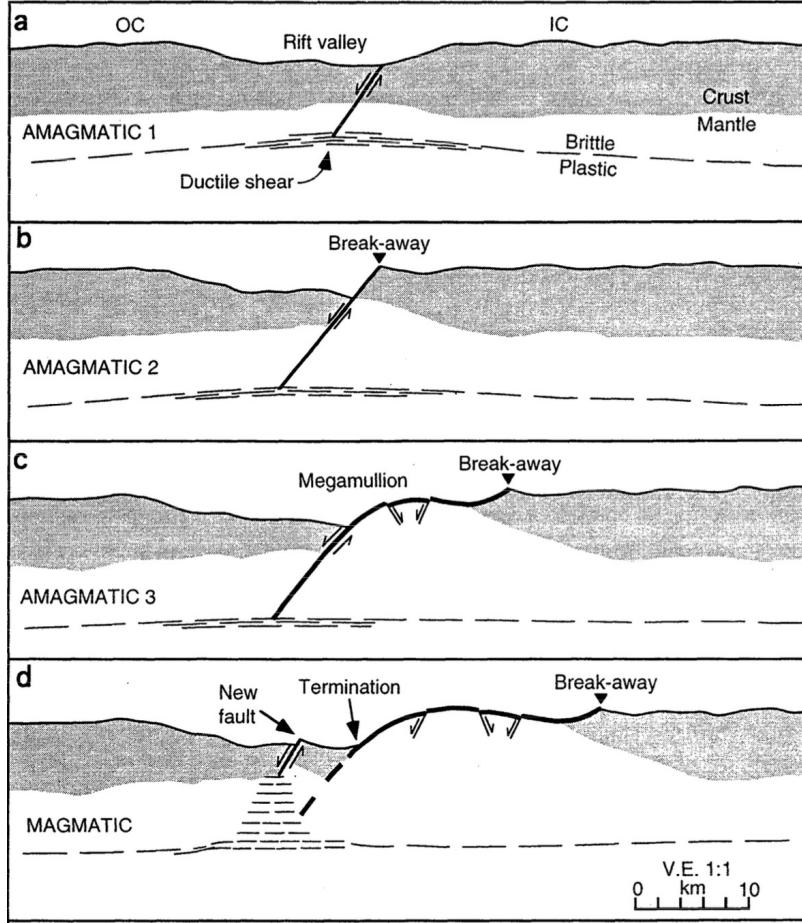


Figure 33: Schematic development of a megamullion. No vertical exaggeration. (a)~(c) shows the detachment fault evolution during amagmatic phase. (d) Increased magma supply pushed the detachment fault away from ridge axis and forms a new normal fault near the ridge axis (“inward fault jump”). Adapted from [Tucholke et al., 1998].

like basalt, however, if this new fault can last a long period of time, it can also exhume lower ultramafic material.

In our model, most of the inward fault jumps last less than 5 km of plates extension before the mantle materials can be exhumed to the seafloor. However, M28LinT1 produces an inward fault jump lasts for \sim 15 km of extension (Figure 31.M28LinT1) and produces a dome shape OCC adjacent to the initial one further way from the ridge axis (Figure 8.h). This behavior might explain the formation mechanism of the brother Abel and Cain domes of the Kane megamullion at 23 °N MAR. As shown in Figure 34, our model behaviors are consistent with the nature observation in terms of the breakaway and the

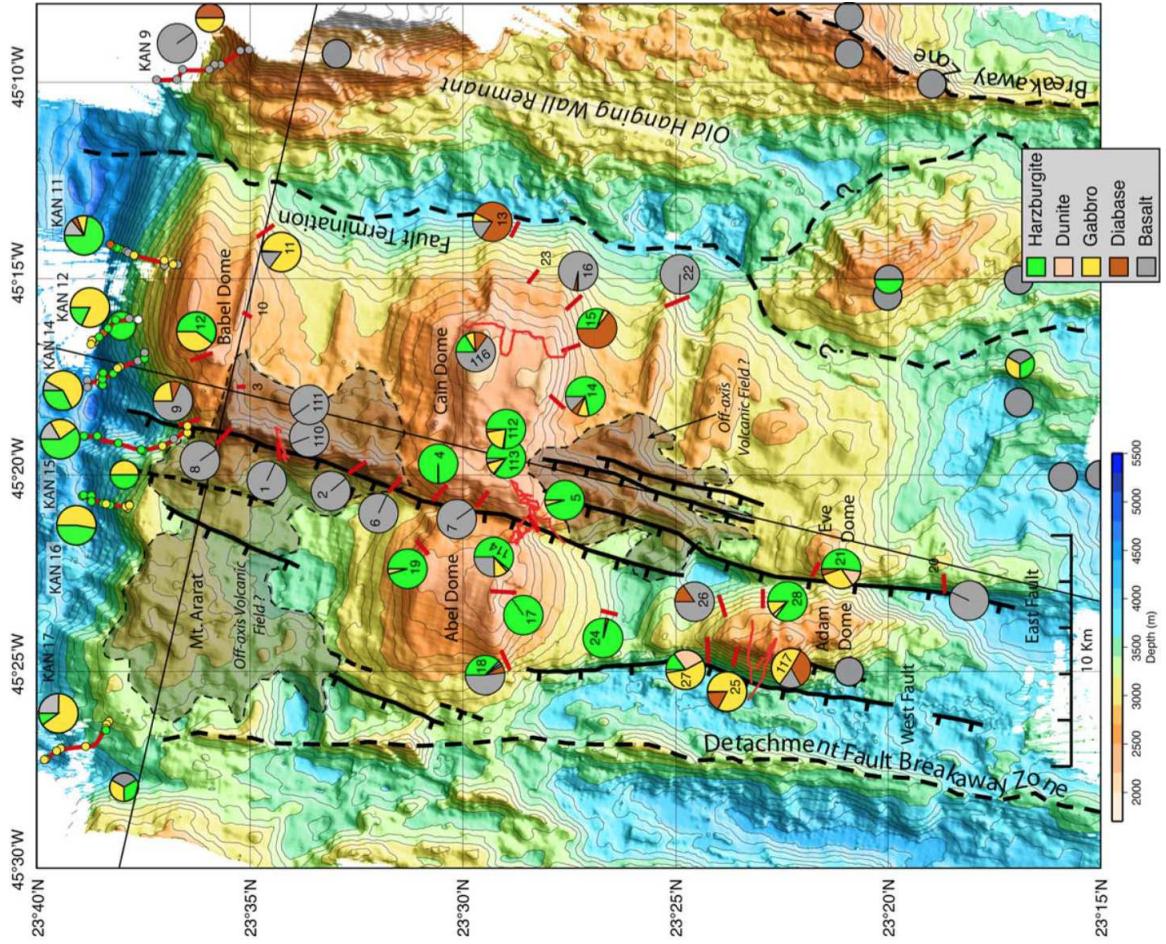


Figure 34: Adapted from [Dick et al., 2008].

wavelength of the domes assuming M decreases from south to north along the ridge axis. First of all, the breakaway of the detachment fault is further away from the ridge axis at the northern than the southern end. Second, the Abel and Cain domes are larger than the Adam and Eve domes because the inward fault jump lasts longer at where M is relatively lower.

In addition, it is mentioned by a seismic study from [Xu et al., 2009] that the Kane OCC is terminated at around 2.1 Myr when an eastward fault jump occurred, i.e. when a new normal fault formed in the rift valley and captured a segment of the basaltic hanging wall. The velocity structure from their P wave tomography study verifies that the hanging wall block eastern to the Cain dome has a lower velocity corresponds to basaltic rocks (Figure 35).

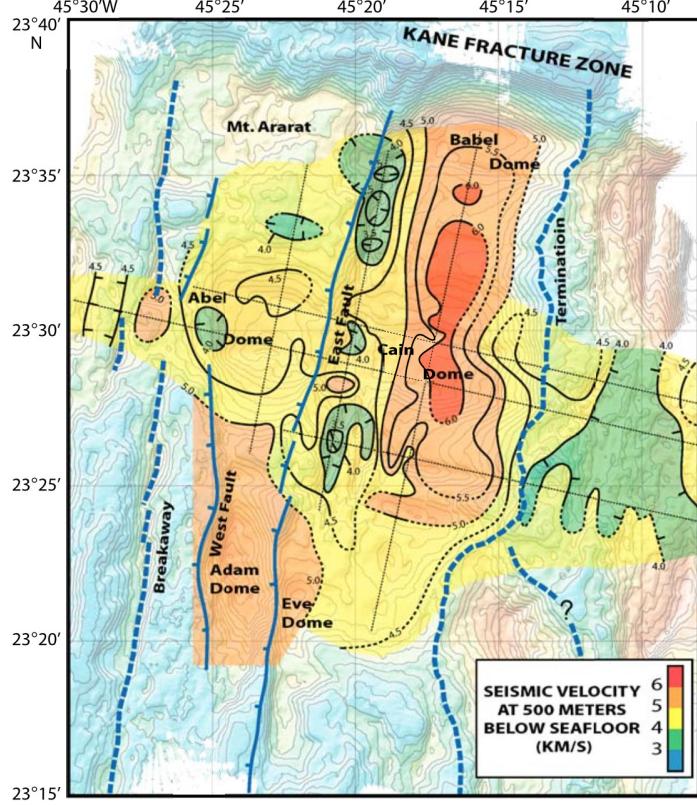


Figure 35: Adapted from [Xu et al., 2009].

4.1.2 Fault alternation

For slow-to-intermediate spreading ridges, two end members govern the off-axis morphologies. One is the higher frequency symmetrically spreading abyssal hills which is usually closer to the ridge segment center (crossection A~A' in Figure 36). The other is longer wavelength asymmetrically spreading OCCs (crossection B~B' in Figure 36). What is the mechanism for this distinct difference along the ridge? The fault alternation behavior in our model provides a 3D perspective for answering the question. When average $M\bar{M}$ is higher than 0.6425 with slower weakening rate (type 2), high frequency abyssal hills are generated. For example, M88ConT2 produces abyssal hills with ~ 10 km in wavelength due to fault alternation (Figure 31.M88ConT2) ^{XT: add the hyperref of the figure of ConM88T2 here}. Note that the wavelength of the abyssal hills in our models is consistent with the nature observation as marked in Figure 36.

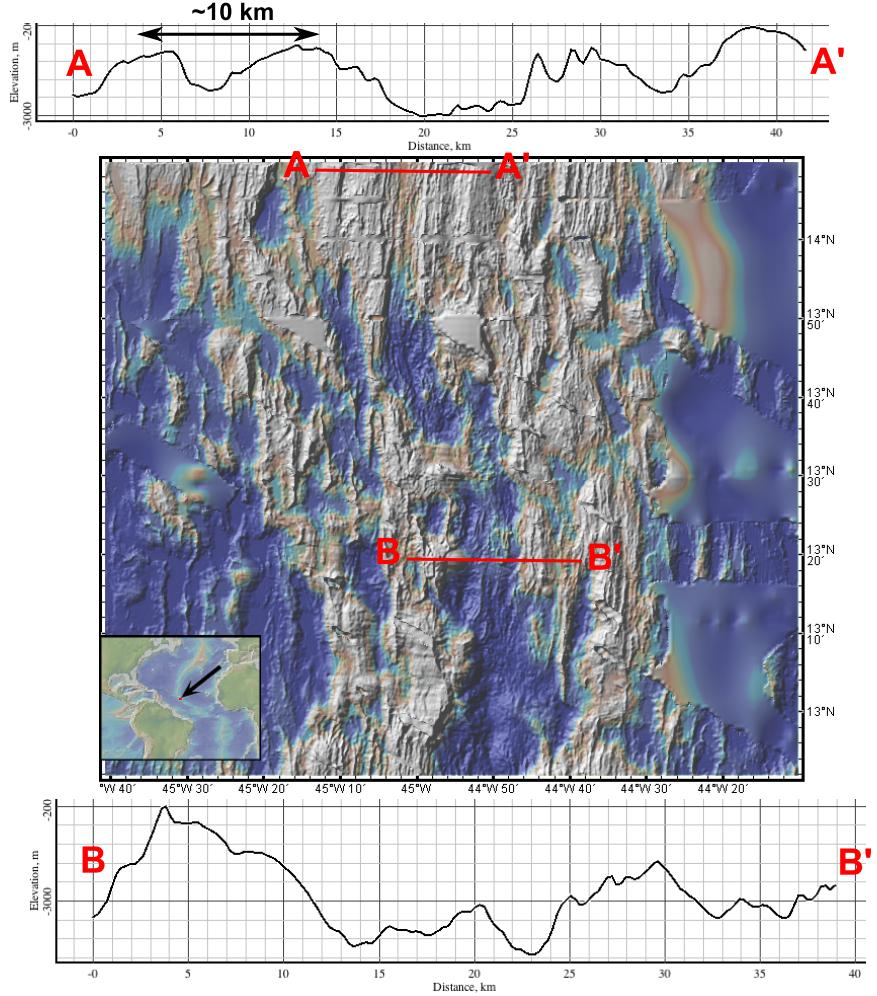


Figure 36: Bathymetry from 12.8~14.2 °N Mid-Atlantic Ridge. Crossection A~A' and B~B' are 5 times vertical exaggerated. From GeoMapApp.

The parameters that controls fault alternation is studied by [Lavier et al., 2000]. There is a trade-off between change in bending force ΔF_b as a function of fault offset ΔX and force change ΔF_w as a function of ΔX due to strain weakening. As described in [Lavier et al., 2000], higher characteristic fault offset (ΔX_c) or slower strain weakening results in multiple faults rather than only one fault lasting. Whether conjugate fault and even multiple faults can be produced depends on the local stress condition. The strength weakening of the existed fault combines with how much bending force resists the fault to keep offsetting play a major role in determining the stress state at the regions other than the active fault. As the oceanic plates keep spreading and ΔX increasing, the change in bending

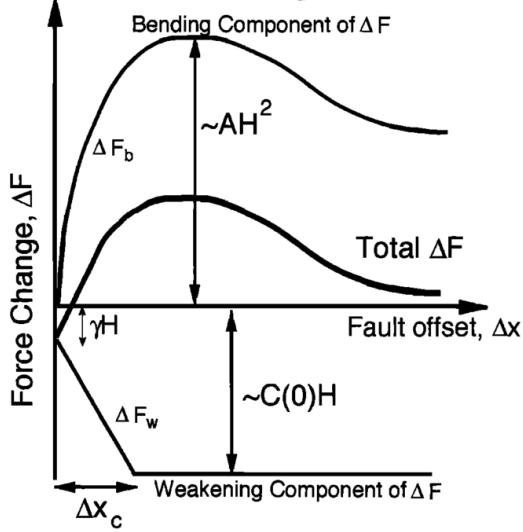


Figure 37: Trade-off between change in bending force ΔF_b and weakening in the fault interface ΔF_w . H is the thickness of the brittle crust and γ is the size of initial weak perturbation and A defines the maximum bending force change. Adapted from [Lavier et al., 2000]. (For more details, please refer to [Lavier et al., 2000])

force ΔF_b increases and the strength at the fault interface decreases due to weakening ΔF_w (Figure 37). If the net force change $\Delta F = \Delta F_b + \Delta F_w$ is positive, it means that it is getting harder and harder to maintain the existing fault and stress will begin to accumulate at the other areas which eventually break another fault. ΔF_b initially increases fast with respect to ΔX and then when the detachment fault surface roll over, ΔF_b reaches its peak value and begins to decrease a little and maintains at a constant value. If the strain weakening is fast enough that the net effect force ΔF is always negative, then most of the stress will be released by the existing fault and thus no conjugate or multiple faults will be generated.

Our model results verify this analysis that only Type 2 weakening (slower weakening with higher ΔX_c) can produce an alternating normal fault on the conjugate plate.

4.1.3 Mass wasting

The mass wasting behavior in M28SqrtT1 model produces a fault scarp of $\sim 1\text{km}$ in relief, 40km in length parallel to the ridge axis. It forms when the detachment fault rolls over

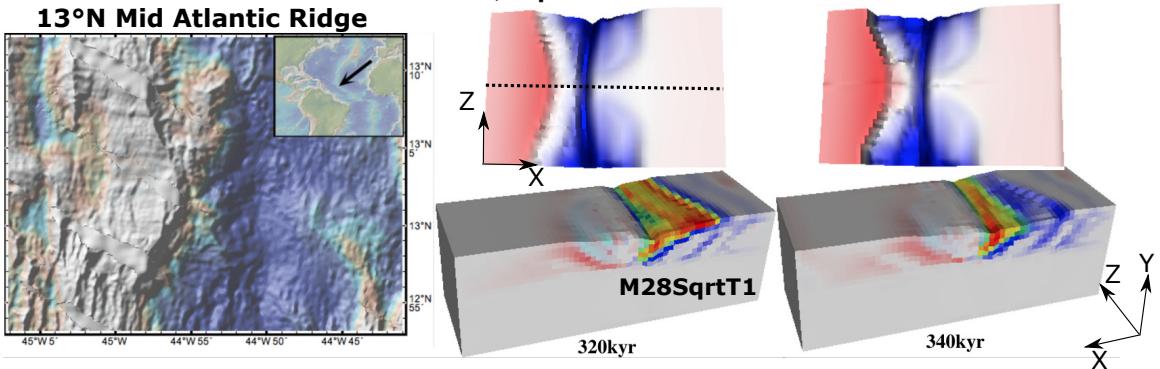


Figure 38: Comparing bathymetry at 13°N Mid-Atlantic Ridge to the mass wasting behavior of M28SqrtT1. The model topography is a mirror symmetric flip according to the dash line, it shows the case of M varies in a square root functional form from 0.2 to 0.8 to 0.2. The bathymetry image is generated by GeoMapApp [Ryan et al., 2009].

and induces bending stresses that generate the high angle fault which cuts the weak detachment fault surface and decouples the spreading breakaway and the highly defomed fault hangingwall. The topography at 13°N Mid-Atlantic Ridge also has a fault scarp with very similar curved geometry with $\sim 1\text{ km}$ in relief. In addition, a magnetic anomaly study of the region from [Smith et al., 2008] reveals a perfect match at the 13° 5' N between the bathymetry and the magnetization (Figure 39). It implies that the block at 13° 5' N adjacent to the curved fault scarp has a relative displacement toward the East and thus provides an evidence for the mass wasting behavior of the block.

Due to the variation in diking along the ridge-axis, an hourglass shape of median valley is also produced in the model where the narrowest center corresponds to the region with higher magma supply ($M=0.8$). This hourglass shape is also frequently observed in the nature along the Mid-Atlantic Ridges [Sempéré et al., 1993].

4.1.4 Hourglass shape median valley

Hourglass shape median valley is frequently observed in the nature along the slow-to-intermediate spreading ridges where the waist of the hourglass is usually narrower and shallower. From an analysis of the sea beam bathymetry along the MAR between 24°

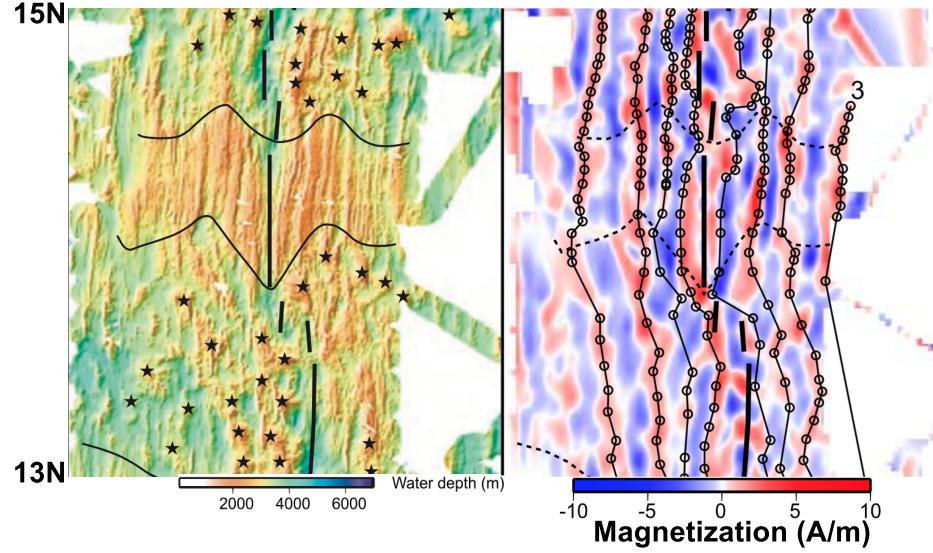


Figure 39: Bathymetry and magnetization of 13~15 °N MAR. Adapted from [Smith et al., 2008].

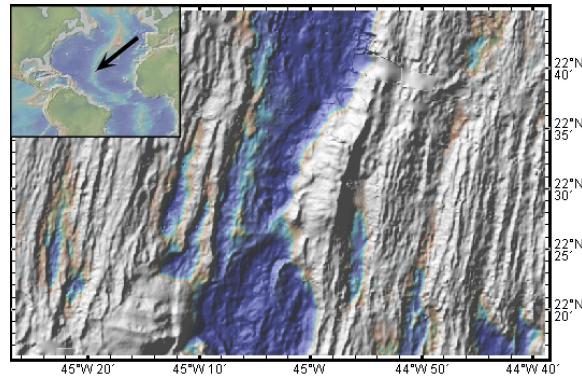


Figure 40: Hourglass median valley at 22 ° 30 'N MAR. Image is generated from GeoMapApp.

00 °N and 30 ° 40 'N [Sempéré et al., 1993], nine hourglass shape valleys are identified. They share similar dimensional scale (~ 40 km \times ~ 40 km) with our model. For example, there is an hourglass at 22 ° 30 ' with along ridge axis length of 45 km and 15 km in width (Figure 40). The observed width and depth of the hourglass can be used for inferring which evolution stage the rift valley is in. For instance, the one shown in Figure 40 corresponds to a early stage of our model because it has a relative symmetrical and narrow shape *^{XT}*[add a hyperref figure of model example for this.](#)

4.1.5 Mullion structure

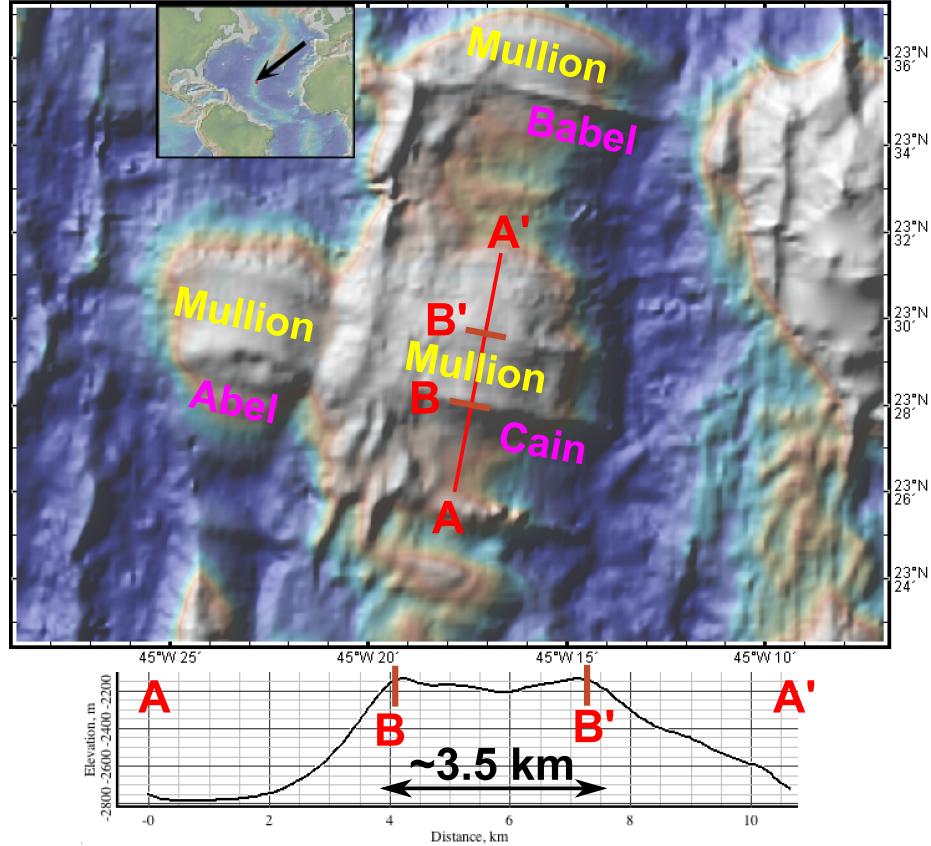


Figure 41: Mullion structures on the surface of Kane OCC at 23 °N MAR. Image is generated from GeoMapApp.

Mullion structure shows up frequently on the surface of OCCs. One of the most characteristic nature observation comes from Kane megamullion. The mullion structure has a wavelength of ~ 3.5 km on the surface of Cain dome as marked in the Figure 41.B~B'. The geometry and length scale of the mullion structure is consistent with our model result (M28LinT1) as shown in Figure 24. Model results indicate that the mullion structure mostly forms when the termination has a curved shape that can last for a long period of time. The spreading parallel mullion structure is produced following the shape of the termination. Where the termination is curved toward the ridge axis corresponds to the larger and higher part of the mullion structure. As shown in Figure 34, the spatial distribution

of the mullion structure of the Cain dome is consistent with the termination that is curved inward to the ridge axis.

4.1.6 Corrugations

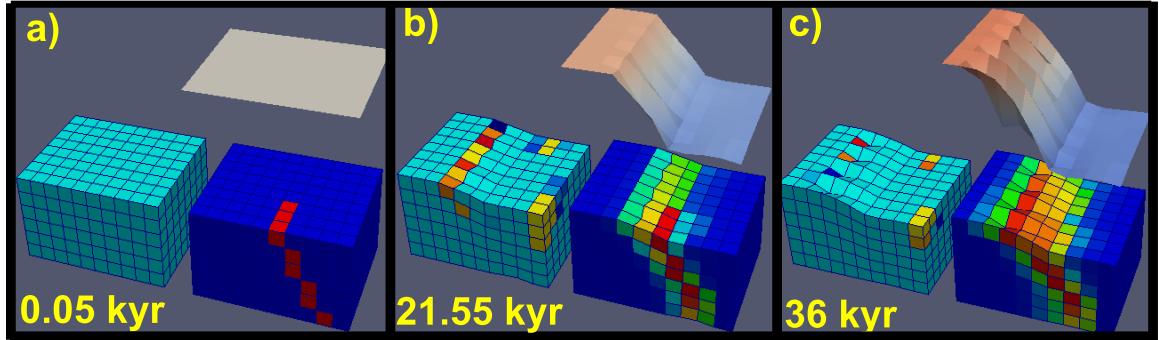


Figure 42: Simpler 3D model with 4 km of seawater pressure on top. Seeds are partially added along the ridge segment. Color scale is the same as Figure 23.

As described in the “Results” section, the corrugations are formed due to the moving off axis tensile failure which results from the asynchronous faulting along the ridge. This mechanism is further verified by simpler 3D models.

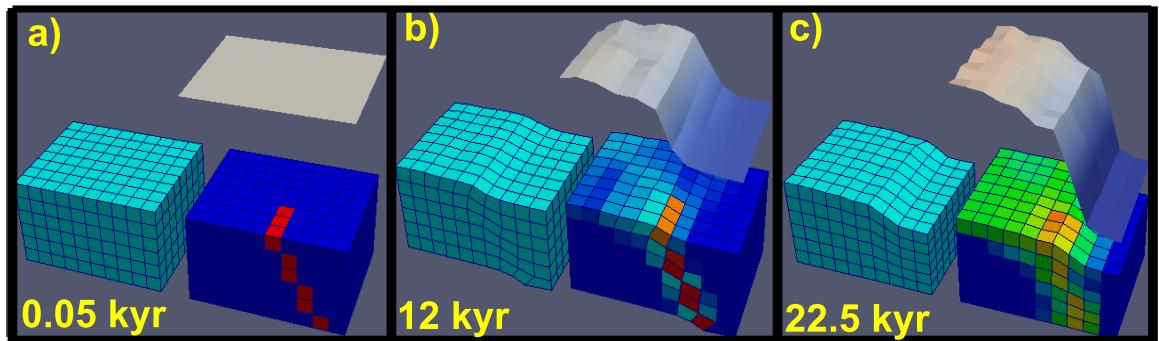


Figure 43: Simpler 3D model with 20 km of seawater pressure on top. Color scale is the same as Figure 23.

When seeds are partially added along the ridge segment for simulating the effects of

asynchronous faulting where normal fault initiates earlier at the region with initial seeds, corrugations are produced by the model (Figure 42). However, when 20 km of sea water pressure is added on top of the seafloor to suppress tensile failure, no corrugation is produced (Figure 43). This verifies that the corrugation is formed by tensile failure.

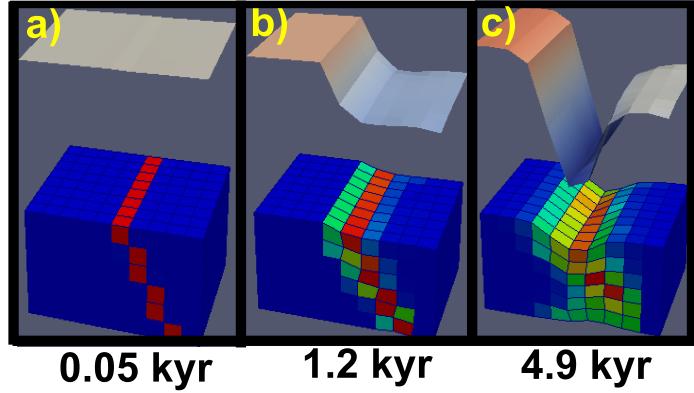


Figure 44: Simpler 3D model with 4 km of seawater pressure on top. Seeds are added along the whole ridge segment. Color scale is the same as Figure 23.

In addition, when seeds are added along the whole ridge segment, which results in synchronous faulting, no corrugation is generated (Figure 44). This verifies that along ridge asynchronous faulting is necessary for creating corrugations.

As mentioned in [Smith et al., 2014], corrugations are observed in many scales from ~ 10 km wavelength megamullion to ~ 1 km wavelength corrugations or even smaller $10\sim 100$ m striations. On most of the OCC surface, corrugations coexist on a wide range of scales (e.g. MacLeod et al., 2009). However, the formation mechanism is still enigmatic [Smith et al., 2006]. Several hypotheses were proposed. Spencer [1999] suggests a casting model that ductile footwall of the detachment fault is continuously exhumed to the surface and is casted by the shape of the cold and strong hanging wall. This hypothesis is consistent with the mullion structure produced in our models. Another hypothesis is given by Smith et al. [2014] that pre-existing offsetted faults or weak zones along the ridge break through and connect to each other forming a curved termination that can produce corrugations.

gations. This anastomosing behavior (e.g. Ferrill et al., 1999; Wong and Gans, 2008) is also produced in our models especially when there is inward fault jumps, antithetic faults or tensile failures happen at part of the ridge segment which perturbs the continuity of the termination along the ridge. In addition, Tucholke et al. [1998] mentioned that with respect to the tectonic configuration at mid-ocean ridges, they expect horizontal, isochron-parallel extension due to contraction of the cooling lithosphere as it move off axis.

Besides all the hypotheses mentioned above, I proposed that the asynchronous faulting due to the varying M along the ridge is possibly responsible for generating those relatively uniform ~ 2 km corrugations. And these corrugations can also be superimposed onto the surfaces of mullion structures or OCCs.

5 Conclusions

To the first time, we model in 3 dimensions on how varying M value (magma supply) along the mid-ocean ridge segment can control the interaction system between tectonics and magmatism and how is it responsible for the major bathymetric features observed on the seafloor.

Six commonly observed MOR features (i.e. inward fault jump, fault alternation, mass wasting, hourglass median valley, corrugation and mullion structure) are produced by our model. By comparing the model results and local field observations, faulting evolution history as well as spatial and temporal variation of magma supply can be inferred.

Three controlling parameters are investigated. They are three ranges of M (i.e. M28, M57 and M58); three type of functional forms of M variation (i.e. linear, sinusoidal and square root) as well as two types of weakening rates (i.e. faster type 1 and slower type 2). As result shows, besides many different structural features generated by different functional forms and ranges of M variations along the ridge, the average M value (\bar{M}) along the whole ridge segment is the major value that is responsible for the two end members of off-axis morphologies, i.e. when $\bar{M} > 0.6425$ with type 2 weakening, the model generates symmetrical spreading, high frequency abyssal hills while when $\bar{M} \leq 0.6425$, fault tends to rotate to a low angle, long lasting detachment fault that can exhume ultramafic rocks to the seafloor producing a domal OCC. Also, our 3D model results resolve the discrepancy between previous 2D model studies and field observations that model studies suggest OCC is produced with M values between 0.3 to 0.5 while field observation reveals cases of OCC forming with observed M beyond both lower and upper limits. Base on this thesis study, we propose a new perspective for explaining cases when OCC forms with local M value is lower than 0.3 or larger than 0.5 by the along ridge coupling argument.

In addition, we propose a new hypothesis of the asynchronous faulting induced tensile failure as one possible formation mechanism for the enigmatic corrugations widely

observed on the surface of OCCs.

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