

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

by

Tian, Xiaochuan

田小川

A Thesis

Submitted in Partial Fulfillment of the
Requirements for the Degree of
Master of Science

Major: Geophysics

The University of Memphis

June, 2015

Copyright © 2015 Xiaochuan Tian

All rights reserved

Dedication

I would like to dedicate this thesis to my mother, Tian, Xia (田霞). 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.

I also want to dedicate this thesis toward my thesis advisor: Professor Eunseo Choi. His mentorship defines what a great advisor is like. Without his guidance, neither this thesis nor my fast personal development during these two years is possible. He has kindled a flame that illuminates the way for my future career as a geodynamic modeler.

Acknowledgements

There is no proper words for me to express my deep gratitude toward my thesis advisor, Professor Eunseo Choi. He has changed my life in a very positive way. From him, I have learned what a great researcher and educator should be like. It is because of his unfailing care and understanding that have given me the courage to persevere, and to carry on during the darkest days as an international student far away from home. It is his vivid and interesting lessons that have inspired me so much that I want to devote my career into geodynamic modeling. It is his numerous inspiration and encouragement that has motivated me to work super hard (enjoy at the same time) on the fascinating research questions. It is his altruistic share that has provided me with great chances to learn from outstanding researchers, references, and to attend meetings. It is his patience to guide me through my endless silly questions that has nurished and cultivated my capabilities of being an independent researcher. Blessings from Eunseo, my role model, are countless. I will keep them in mind and try my best to relay the torch.

I would also like to thank members of my thesis committee, Professor Christine Powell and Professor Jer-Ming Chiu, for their guidance and support. Especially I want to thank Dr. Chiu for his sharing and advice on my course term projects and for his patience to answer all my silly questions. In addition, I want to thank Dr. Powell for your warm-hearted encouragement and benevolence during my studies at the Center for Earthquake Research and Information (CERI).

My sincere thanks must also go to the CERI and the University of Memphis (UM). Thanks a lot for providing me this great chance to study here with full tuition and graduate research assistanship. It is this kind support that has made my two-year studies at this great institution possible.

I am also grateful to the courses instructors for many vivid and interesting courses that have stimulated my interests in Geophysics, helped me build up my foundation and expanded my horizon in the field. They are Professor Robert Smalley for “Data Analysis in

Geophysics”, Professor Charles Langston for “Global Seismology” and “Inverse Methods in Geophysics”, Professor Mitch Withers for “Signal Processing in Earth Science”, Professor Eunseo Choi for “Global Geophysics” and “Geodynamics”, Professor Christine Powell for “Environmental Geophysics”, Professor Randy Cox for “Art of Earth Science”, Professor Jer-Ming Chiu for “Exploration Seismology”. Thank you so much. Especially, I want to thank Professor Withers for your encouragement and your time and efforts for answering all my questions. I would also like to thank Professor Eric Daub for your delicious homemade cakes provided to us graduate students every Wednesday during the “CERI Discussion”. I am probably the person who ate the most of them and enjoyed the most.

Many thanks must also go to my dear classmates and research colleagues, with whom, I have lived, studied and worked during the two years. Especially, I want to thank Yang Yang for your supportive help and share on my term projects. Thank you for being a wonderful roommate. I would also like to thank my best “cubic office mate”, Naeem Khoshnevis, for his best company. We’ve been studied together as the most diligent (long staying) people in CERI (12/7). Thank you for those benevolent and uncountable “how are you”, “good morning”, “good night” as well as many good chats. My sincere thanks must also go to Sabber Ahamed, who has been encouraging me during my darkest days when I lost direction. It is his enthusiastic share and encouragement that have helped me go through the bad days.

Many thanks must also go to CERI staffs. Thank you Chris McGoldrick for many of your kindly help and all the nice chats, from which I have learned many new English words and phrases. Thank you Tanya Broadbent and Kathleen Tucker for sending me birthday gifts which makes me feel so warm as being a member of CERI families. Thank you Christy Chiu and Jer-Ming Chiu for inviting me to your house for Chinese Spring Festival lunch which helped a lot in soothing my nostalgia for hometown. Thank you Gary Patterson for your share and benevolence. Thank you Deshone Marshall for con-

stantly helping me out in using CERI's lab computers. Thank you Josephine Calhoun for providing us tidy and enjoyable environments for studying and doing research everyday.

I also want to sincerely thank Center for Writing and Communication of the UM. Especially, I want to thank Bill Schraufnagel for your very supportive helps on revising this thesis and enhancing the presentation.

I am also thankful to many great researchers that have encouraged and inspired me for this research. Thank you Professor Lin, Jian for your encouragement during 2014 AGU fall meeting. Thank you Jean-Arthur Olive for your feedback and your kind share. Thank you Professor Taras Gerya for your feedback and your great book *Introduction to Numerical Geodynamic Modelling* that has inspired me a lot. Thank you Professor Suzanne Carbotte for your share and encouragement. Thank you Professor Roger Buck for several inspiring talks and sharing as well as feedbacks on this research.

I also want to thank Professor Liu, Jie (刘洁) and Professor Zhang, Ke (张珂) from Sun Yat-sen University, for kindly sharing your computing resources on Tianhe2 to me. Many thanks must also go to the staffs of Penguin, Stampede and Tianhe2 supercomputer centers, thank you so much for all the technical supports. Especially, I would like to thank National Supercomputer Center in Guangzhou, China, for providing me supports with more than 200k of core-hours.

Sincere thanks must also go to the lifeguards and staffs of the swimming pool of the UM Recreation Center. Thank you for your work in providing me such a safe and enjoyable workout environments. Thank you for accompanying me almost everyday and making sure my safty during I am swimming. I could not endure all the stresses from the high pressure school works if without being able to enjoy swimming everyday.

In addition, I also want to express my sincere thank to Professor Cheng, Gu (成谷), my undergraduate thesis advisor. Thank you for providing me a chance to start my career as a geophysicist. Thank you for all your guidance and share.

I should never forget my deep gratitude to my middle school Math teacher Feng, Yan

(冯艳) and high school Math teacher Chen, Shengfang (陈胜方) and Physics teacher Lyu, Lijie (吕黎洁) as well as English teacher Cindy (叶晓靖). It is your great education, inspiration, love and care that have cultivated me into a person, who begins able to appreciate the fascinating science world. Your great efforts and patience in teaching me and answering my silly questions have helped me build up a good foundation for my future studies as a science major student.

Special thanks must also go to a special person, Zhang, Siwan (张思婉). Thank you for planted a seed of love and beauty deeply in my heart that has been encouraging and inspiring me through time.

Finally, I want to thank my families. There simply is no proper word for expressing my gratitude. Thank you so much my dear grandfather Tian, Shengheng (田升恒), and grandmother Wang, Defen (王德芬), who have reared me up with great love and care. Thank you so much my grandgrandfather Tian, Guangzong (田光宗). I will never forget your last words for encouraging me to be a good person and a diligent student.

Thank you my dear mother, Tian, Xia (田霞). You are always a great mother to me. Thank you so much for rearing me up by yourself with your great courage, optimism, love and care. Without you, there is no me.

“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 variations in diking in terms of faulting patterns and the associated structures using a 3D parallel geodynamic modeling code, SNAC. Many observed structural features are reproduced. We also propose 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 ($M = 0.3\sim0.5$) and field observations ($M < 0.3$ or $M > 0.5$) 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	8
2.1 Method of approach	8
2.2 Model Setup	9
3 Results	14
3.1 Reference model	14
3.1.1 Constant M model M88ConT2	18
3.2 Main structural characteristics	19
3.2.1 Location of termination	19
3.2.2 Geometry of trough	20
3.2.3 Inward fault jump	21
3.2.4 Fault alternation	21
3.2.5 Mass wasting	22
3.2.6 Hourglass-shaped median valley	24
3.2.7 Corrugations and mullion structures	26
3.3 Effects of the functional forms of M variation	28
3.3.1 M28T1	28
3.3.2 M58T2	31
3.4 Effects of the weakening rate	32
3.4.1 M57SinT1 versus M57SinT2	32
3.4.2 M58SinT1 versus M58SinT2	35

3.4.3	M58SqrtT1 versus M58SqrtT2	39
3.5	Effects of the range of M variation	42
3.5.1	SinT1	42
3.5.2	M57SinT2 versus M58SinT2	43
3.5.3	M28LinT1 versus M57LinT1	43
4	Discussion	46
4.1	Overview of model behaviors	46
4.2	Comparing model results with natural observations	50
4.2.1	Inward fault jump	50
4.2.2	Fault alternation	53
4.2.3	Mass wasting	56
4.2.4	Hourglass-shaped median valley	57
4.2.5	Mullion structure	58
4.2.6	Corrugations	59
5	Conclusions	63
Bibliography		64

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.	2
3	Relationship between the maximum crustal thickness variations (ΔH_c) along a ridge segment and the segment length (L).	3
4	2D model results adapted from [Buck et al., 2005].	4
5	2D model results adapted from [Tucholke et al., 2008, Whitney et al., 2012].	5
6	Essential components of the numerical method.	8
7	Model setup.	10
8	Reference model M28LinT1.	14
9	Bird's-eye view of a breakaway in the reference model.	15
10	Illustration of bending stresses in lithosphere associated with faulting at the MOR.	16
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)) . . .	17
12	Bird's-eye view of σ_{xz}	17
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.	18
14	The second invariant of strain rates plotted on the reference model's vertical cross-sections along the ridge at 107.5 kyr.	19
15	Velocity of the mass wasting hanging wall.	22
16	The weak detachment fault tip reaches the pre-existing shear stress.	22
17	Plastic strain, topography and stresses evolution for M28SqrtT1.	23
18	Along ridge axis σ_{xz}	24
19	Bird's-eye view of the evolution of the hourglass-shaped median valley. . .	25

20	Higher σ_{xx} shows inside the median valley on the positive x -axis direction of the ridge axis.	25
21	Bird's-eye view of the evolution of the corrugations.	26
22	Evolution of mullion structures.	27
23	Three functional forms of M variation.	28
24	Bird's-eye view for comparing the length and timing of inward fault jump. .	29
25	$\partial M / \partial Z$ comparision.	30
26	Bird's-eye view of the topography.	31
27	Bird's-eye view of M57SinT2 versus M57SinT1 (Table 1).	32
28	Bird's-eye view of faulting and stresses evolution of M57SinT1.	33
29	Bird's-eye view of faulting and stresses evolution of M57SinT2.	34
30	Bird's-eye view of faulting and stresses evolution of M58SinT1.	35
31	Plastic strain for M58SinT1 at 350 kyr.	36
32	Evolution of faulting and morphologies of M58SinT1.	37
33	Bird's-eye view of faulting and stresses evolution of M58SinT2.	38
34	Bird's-eye view of faulting and stresses evolution of M58SqrtT1.	40
35	Faulting and stresses evolution for M58SqrtT2.	41
36	M28SinT1 (Table 1) faulting and stresses evolution with respect to time. .	43
37	M57LinT1 (Table 1) faulting and stresses evolution with respect to time. .	44
38	Initiation and duration of main structural features.	46
39	Field observation results adapted from [MacLeod et al., 2009].	49
40	Schematic development of a megamullion from amagmatic to magmatic, adapted from [Tucholke et al., 1998].	51
41	Bathymetric map and simplified tectonic interpretation of Kane Megamullion. Adapted from [Dick et al., 2008].	52
42	P wave isovelocity contours at 500 m below seafloor superimposed on bathymetry and simplified tectonic interpretation. Adapted from [Xu et al., 2009]. . .	53

43	Bathymetry from 12.8~14.2 °N Mid-Atlantic Ridge.	54
44	Trade-off between change in bending force ΔF_b and weakening in the fault interface ΔF_w . Adapted from [Lavier et al., 2000].	55
45	Comparing bathymetry at 13°N Mid-Atlantic Ridge to the mass wasting behavior of M28SqrtT1.	56
46	Bathymetry and magnetization of 13~15 °N MAR.	57
47	Hourglass median valley at 22 ° 30 'N MAR.	57
48	Bathymetry around the Kane OCC at 23 °N MAR.	58
49	Comparing bathymetry at 16 °N Mid-Atlantic Ridge to the corrugations of M58SinT1 at 550 kyr.	59
50	Simpler 3D models for formation mechanism of corrugation.	60
51	Simpler 3D model with 4 km of seawater pressure on top. Seeds are added along the whole ridge segment.	61

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.

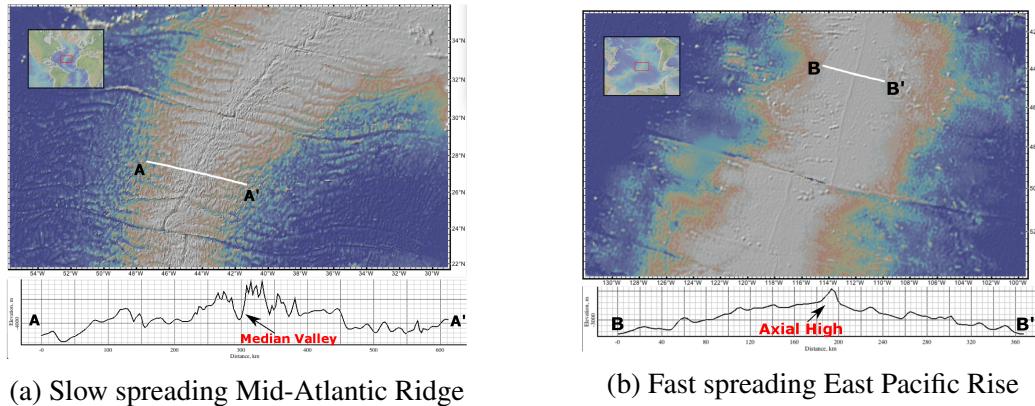


Figure 1: Profiles of bathymetry across MORs.

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).

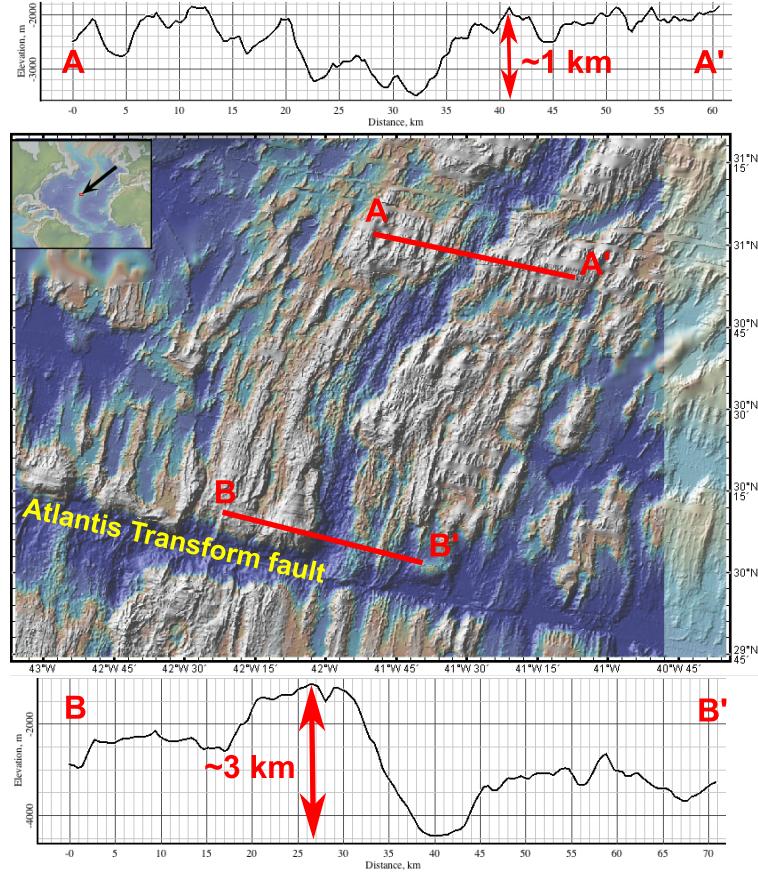


Figure 2: Two bathymetric profiles across the Mid-Atlantic Ridge around 30°N with vertical exaggeration of 10. A-A' is closer to the segment center while B-B' is at the tip of the segment.

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 the center of the ridge segment (A-A') is rather symmetrical and has a shorter wavelength 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 longer wavelength with a median valley wider than 30 km and shows a greater relief (\sim 3 km). Gravity data of different ridge

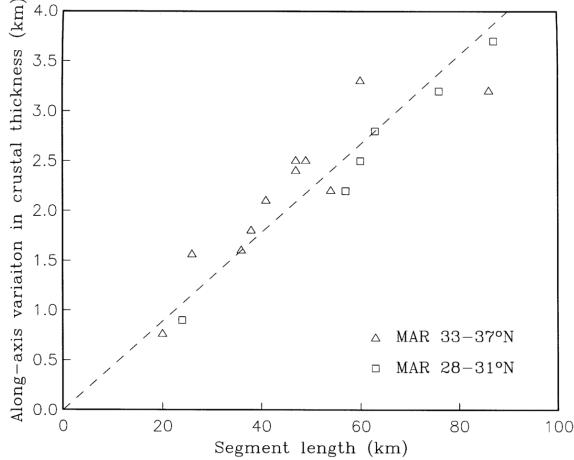


Figure 3: Relationship between the maximum crustal thickness variations (ΔH_c) along a ridge segment and the segment length (L). The dashed line is the best-fit linear regression of the combined data [Chen and Lin, 1999].

segments along MAR (28~31 °N and 33~37 °N) suggests that the maximum along-axis variation in crustal thickness ΔH_c of a single segment increases with the 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 experience more frequent diking, which efficiently releases stresses generated by far-field forces that drives the plate motion. In contrast, diking is less frequent at slow spreading ridges and can only partially release the stresses associated with plate motion. Thus, the brittle lithosphere of slow-spreading ridges experiences internal deformations (e.g. tectonic processes like normal faulting) when the deviatoric stress exceeds the internal strength of the lithosphere.

Buck et al. [2005] defined the ratio between the rates of dike widening and plate sepa-

ration as $M = V_{dx}/2V_x$, where V_{dx} is the rate of horizontally 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.

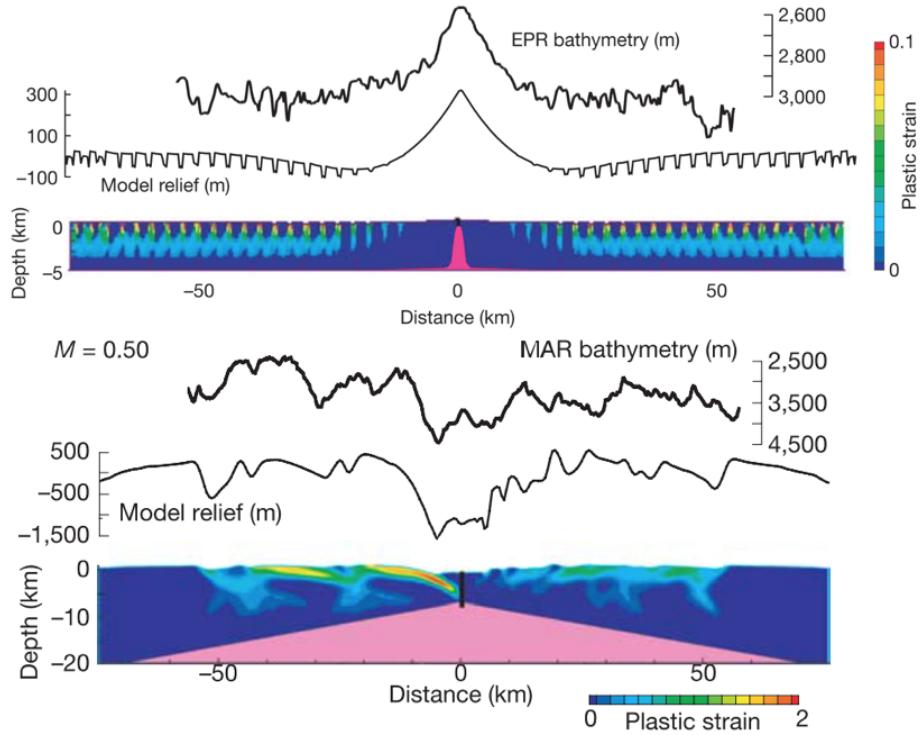


Figure 4: Upper: modeling result for fast spreading agrees well with East Pacific Rise. Lower: modeling result for slow spreading ridges agrees well with the bathymetry of Mid-Atlantic Ridge. Adapted from [Buck et al., 2005].

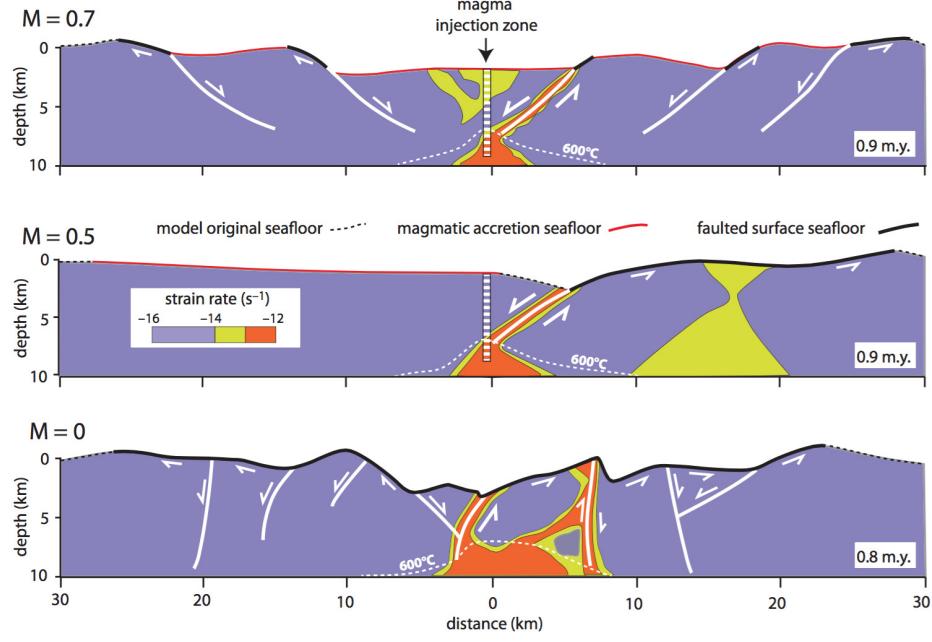


Figure 5: Snapshots of modeled fault behavior and seafloor morphology for $M = 0$, 0.5 , and 0.7 ; model allows thermal evolution. Structural interpretation is superimposed on modeled distribution of strain rate; model time is indicated in panels at lower right; dashed white line at bottom is $600\text{ }^{\circ}\text{C}$ 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. Adapted from [Tucholke et al., 2008, Whitney et al., 2012]

Tucholke et al. [2008] expand the investigation on the role of M in mid-ocean 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 $M = 0.3\sim0.5$, 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 variations 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 variations 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 upwelling magma at the ridge center, 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 vary 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 investigating the factors that controls 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 3D by implementing it into a 3D numerical modeling code SNAC [[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 along-axis variation in diking at slow-to-intermediate mid-ocean ridges is responsible for the observed morpholo-

gies.

2 Methods

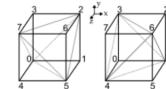
2.1 Method of approach

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.

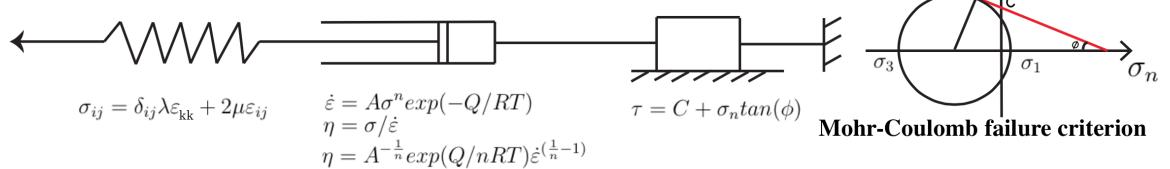
SNAC: a 3D, MPI parallelized, updated Lagrangian explicit finite element code for modeling long-term tectonic evolution of the Earth's elasto-visco-plastic crust and mantle. (Choi et al., 2008)

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: $\mathbf{M} = \mathbf{Vdx} / 2\mathbf{Vx}$ (\mathbf{Vdx} is the dike widening velocity in x direction, $2\mathbf{Vx}$ 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 weakening 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 stress 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 equation $\sigma_{ij} = \lambda\varepsilon_{kk}\delta_{ij} + 2\mu\varepsilon_{ij}$.

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 instantaneous cooling of a semi-infinite half-space model of moving plates [e.g., [Turcotte and Schubert, 2002](#)]. Two

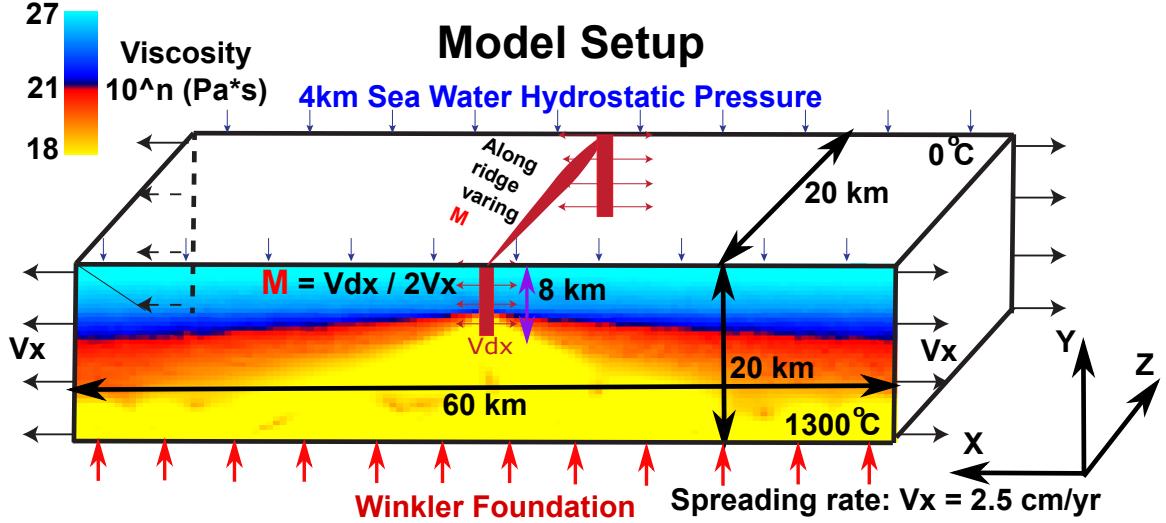


Figure 7: Model setup.

sides perpendicular to the z coordinate axis are free-slip. Hydrostatic sea water pressure is added on top of the seafloor. The heights of the water columns are locally determined as $(4000 - h(x, z))$ m, where $h(x, z)$ is the topography at a location, (x, z) . The bottom boundary is free of shear stresses and is supported by normal stresses that equal to the local lithostatic pressure (Winkler foundation) following [Buck et al. \[2005\]](#). Temperature is fixed at 0 °C on the top surface and at 1300 °C on the bottom surface.

Diking, controlled by the factor M , is assumed to occur in the middle of the domain (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 pseudo-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

(ε_{pe}^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 (ε_{pi}^2) to 0.33 (ε_{pe}^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 $\Delta X_c = 3 \Delta x \varepsilon_{pe}$ 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 supply focuses 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 the 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	60	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 first parameter	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

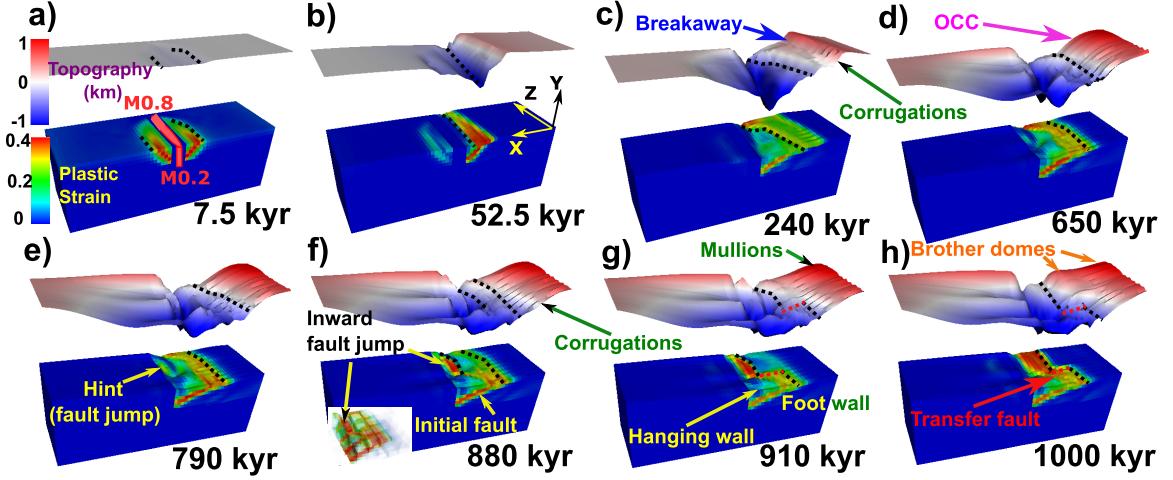


Figure 8: Evolution of plastic strain and surface topography of the reference model M28LinT1. Each snapshot shows plastic strain plotted on the model domain, the five times vertically exaggerated topography and the time at which it is taken. The initial seafloor is at 0 km of elevation. The black dashed 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).

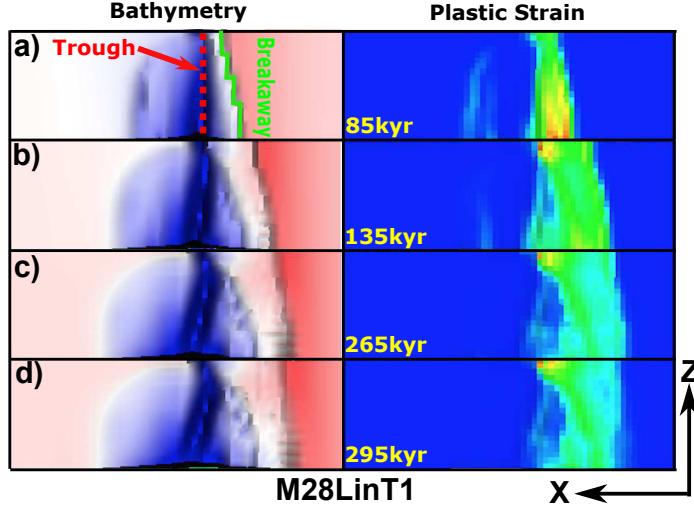


Figure 9: Bird's-eye view of a breakaway in the reference model at a) 85 kyr, b) 135 kyr, c) 265 kyr and d) 295 kyr. The breakaway is marked by green solid line in a). A narrow zone of depressed topography (“trough”) is marked by red dashed line inside the median valley (blue area) in a). Color scales 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. The upper part of the active fault plane (shown as plastic strain in the model) is exposed to the seafloor. As the active fault slips, crust at the footwall bends in a clockwise rotation as is illustrated in Figure 10.

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

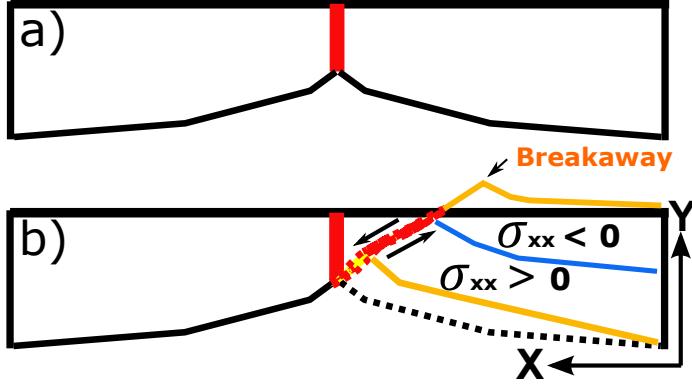


Figure 10: Illustration of the development of bending stress in lithosphere associated with faulting at the MOR. The blue line is the neutral plane where $\sigma_{xx} = 0$. Above the neutral plane is compression ($\sigma_{xx} < 0$) and beneath it is tension ($\sigma_{xx} > 0$).

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 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-

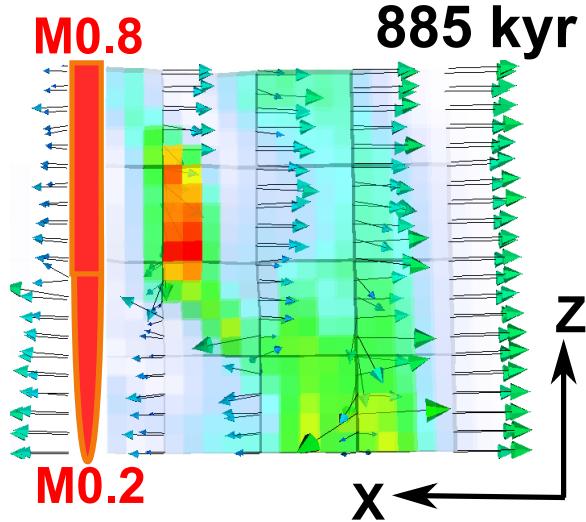


Figure 11: Bird's-eye view of velocity field with plastic strain plotted with opacity linearly proportional to its value. (color scale is the same as Figure 8.a))

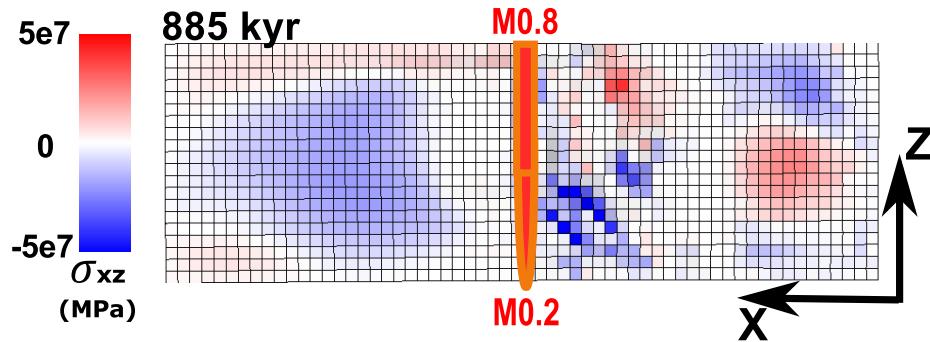


Figure 12: Bird's-eye view of σ_{xz} .

tion. At the lower M side, the detachment 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 adjacent to the initial OCC is produced at the higher M side by 1000 kyr.

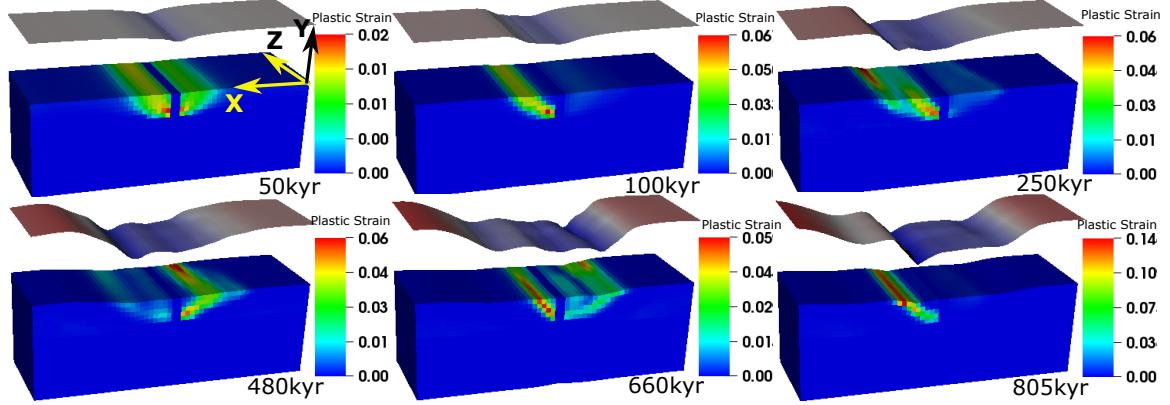


Figure 13: Evolution of plastic strain and surface topography of the model: M88ConT2 (Table 1). Color scale of topography is the same as Figure 8.a.

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 corresponding 2D models 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 observation of the Mid-Atlantic Ridge. 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 of plates extension and produce symmetrical shorter wavelength abyssal hills.

3.2 Main structural characteristics

I describe seven structural characteristics of the models: 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

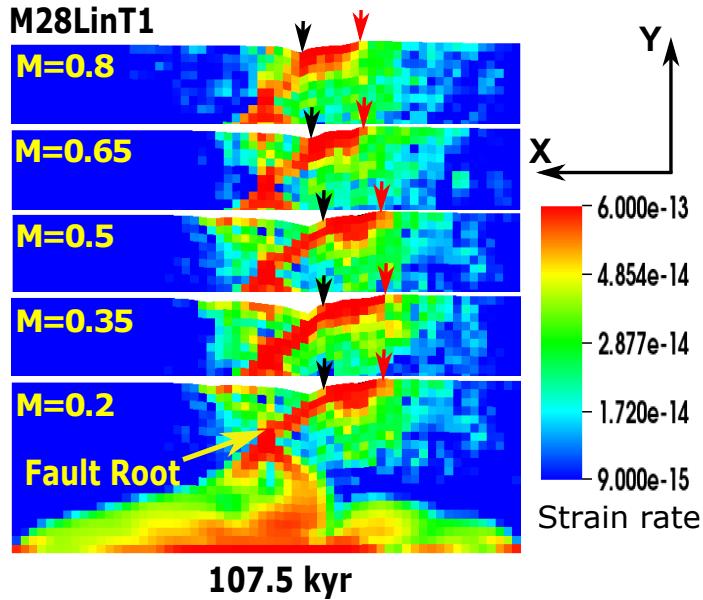


Figure 14: The second invariant of strain rates plotted on the reference model's vertical cross-sections along the ridge at 107.5 kyr. Terminations and breakaways are marked by black and red arrows.

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 highest strain rate regions (red) in Figure 14 can be interpreted as the active detachment fault interfaces. Compared to the two slices with $M > 0.5$, the distances between terminations and the ridge axis at the lower M side ($M \leq 0.5$) is larger and the dip angles of the detachment faults are lower.

For the ridge region with $M > 0.5$, the fault root is pushed away from the ridge axis due to excessive diking while the termination is closer to the ridge axis due to lower slipping rate of the fault and the asynchronous initiation of faulting along the ridge. Thus the distance between the termination and the fault root is smaller at the higher M side and the dip angle of the fault is higher. However, among the three slices of $M \leq 0.5$, the distances and the dip angles are similar. This is because one end of the detachment fault roots at the same places along the ridge axis (intersection between the center dike and the brittle-ductile transition), and for the other end, the moving rate of the termination has a maximum value that is restricted by the far field extension rate V_x . Thus, the bending rates of the detachment faults are similar among the three slices of $M \leq 0.5$.

One thing needs to be noted is that the trough at the higher M side corresponds 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 “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.a-d). 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, 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.d).

3.2.3 Inward fault jump

The inward fault jump (e.g., Fig. 8.f) occurs in most of the models when the energy for maintaining an old fault becomes larger than breaking a new one near the ridge axis. As shown in Figure 14, at the region with $M > 0.5$, the existing normal fault is pushed away from the ridge-axis due to robust diking ($M > 0.5$). As it moves away from the ridge axis, the frictional energy for the fault, the bending energy for the footwall as well as the negative work done by gravity that resists the exhumation of the footwall increase [Lavier et al., 2000, Olive et al., 2014]. The initial detachment fault remains active until the sum of the negative works reaches an upper limit that breaking a new fault near the ridge axis needs less work than to maintain the initial one, the initial detachment fault at the higher M side is then substituted by the inward jumping fault that cuts through the previous hanging wall. As the fault evolves, it connects to the initial detachment at the lower M side ($M < 0.5$) by a transfer fault and generates a curved termination along the ridge. Unlike fault alternation described below, the inward fault jump occurs on the same side of the ridge axis with its along-axis extent corresponding to the $M > 0.5$ region.

3.2.4 Fault alternation

Fault alternation is the behavior that normal faults alternatively show up on each side of the ridge axis when magma supply is robust enough, as shown in M88ConT2 (Figure 13). The normal fault first evolves on the left hand side of the ridge axis and produces an abyssal hill parallel to the ridge axis. By 480 kyr, another normal fault evolves on the other side of the ridge axis and replaces the first one. Note that among the 15 models (Table 1), only three produce fault alternation. They are M88ConT2, M58SinT2 and M58SqrtT2. Fault alternates only when weakening rate is low (type 2 weakening) and the average integration of M along the ridge is larger than 0.65. Analysis on when and why fault alternates is given in the Discussion section.

3.2.5 Mass wasting

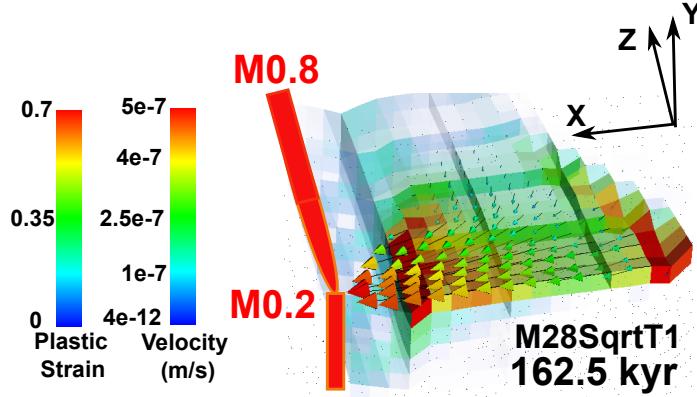


Figure 15: Velocity of the mass wasting hanging wall. Magnitudes of the velocity are shown by the colors of the arrow heads. Plastic strain is plotted with opacity linearly proportional to its value.

Mass wasting occurs on the exposed surface of a low-angle normal fault. When the weak fault zone becomes gravitationally unstable and is decoupled from the spreading plate as well as the underlying material, the detached layer flows towards the lower ridge axis and the lower M side driven by gravity with a velocity hundreds of times faster than the half spreading rate (Figure 15; Figure 17.b,e (first row))). As the top layer of the hanging wall flows down the topography slope, obvious topography drop is observed in 2.5 kyr (Figure 17.b) versus c); d) versus e) (topography)).

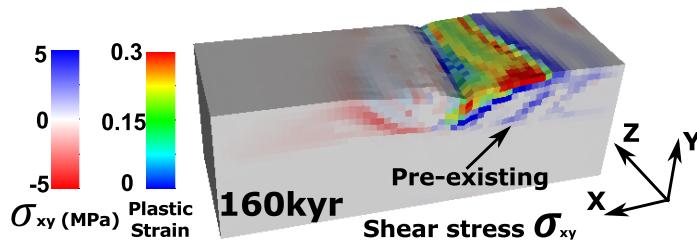


Figure 16: The weak detachment fault tip reaches the pre-existing shear stress. Plastic strain is plotted with opacity linearly proportional to its value.

Mass wasting is triggered by several factors. First, when the tip of the weak fault interface moves away from the ridge axis with the spreading plate and is intersected by a pre-existing shear stress σ_{xy} (Figure 16), the extra shear stress cuts the tip of the weak fault interface and leads to the decoupling between the breakaway and the upper layer of the hanging wall of the detachment fault. Second, during the roll over of the hanging wall at the breakaway, tensional stress ($\sigma_{xx} > 0$) promotes the deviatoric stress for generating small scale high angle normal faults cutting the detachment fault surface [Tucholke et al., 1998] (Figure 40.c,d) and in this case the tip of the weak layer. In addition, along ridge coupling tends to resist the increase in along ridge offset of breakway which is generated by along ridge varying slip rates of the faults. As Figure 18 shows, the dextral sense shear stresses (red σ_{xz}) beneath the top weak layer tends to rotate the upper layer and assists in triggering the decoupling process.

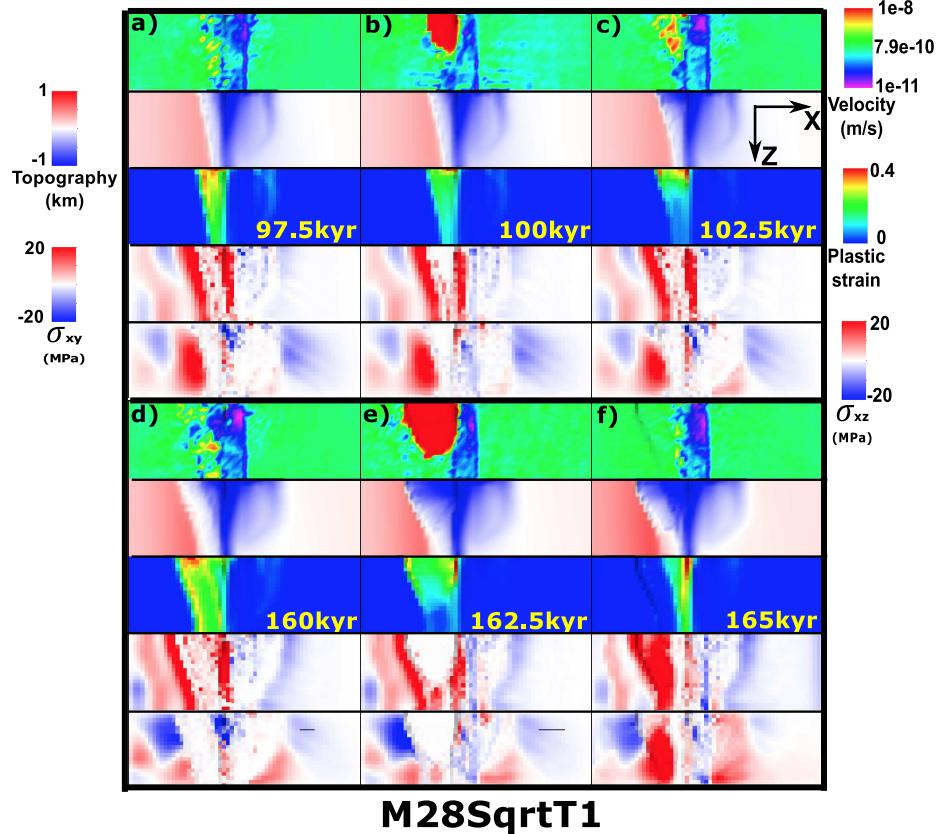


Figure 17: Plastic strain, topography and stresses evolution for M28SqrtT1.

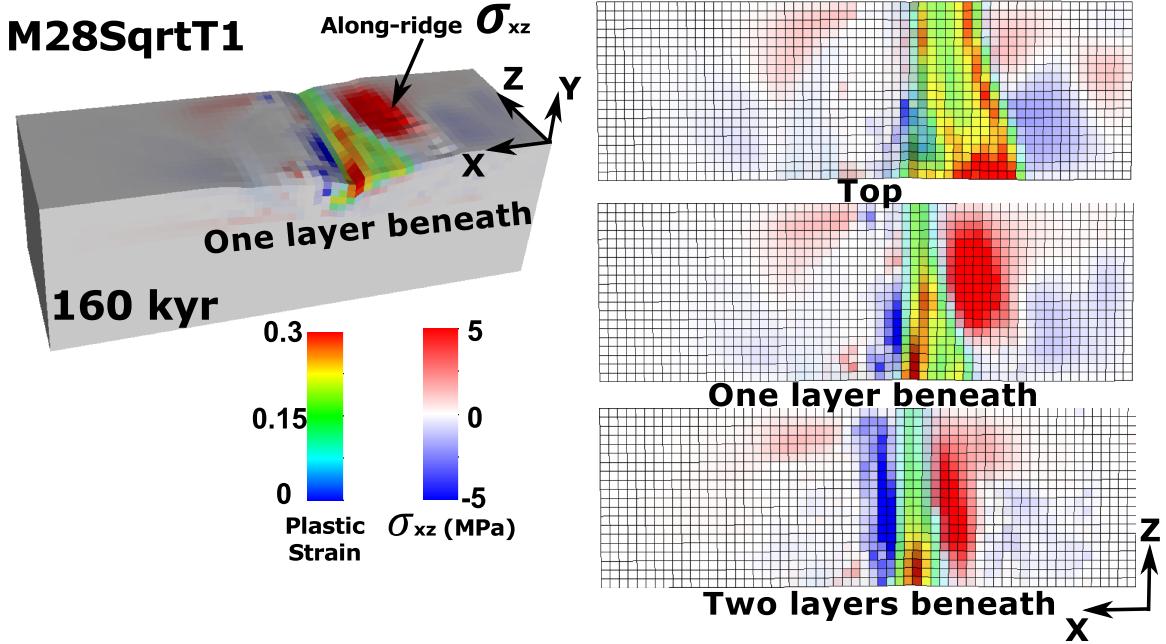


Figure 18: Along ridge axis σ_{xz} (bird's-eye view of three layers of model domain) (positive(red) means clockwise). Plastic strain is plotted with opacity linearly proportional to its value.

3.2.6 Hourglass-shaped median valley

All the models with M variation develop an hourgalss-shaped median valley although the geometry of the median valley changes with time and with the functional forms and ranges of M variation. A median valley of the M28Sqrt model initially has a uniform width along the ridge but is deeper on the lower M side where normal faults first form (Figure 19.a). By ~ 100 kyr, the fault on right hand side of the ridge axis does not propagate to the higher M side of the ridge and becomes inactive (Figure 19.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, breakaways at the lower M side move further away from the ridge axis than the breakaways at the higher M side. This along-ridge variation in the location of the breakaways act as another

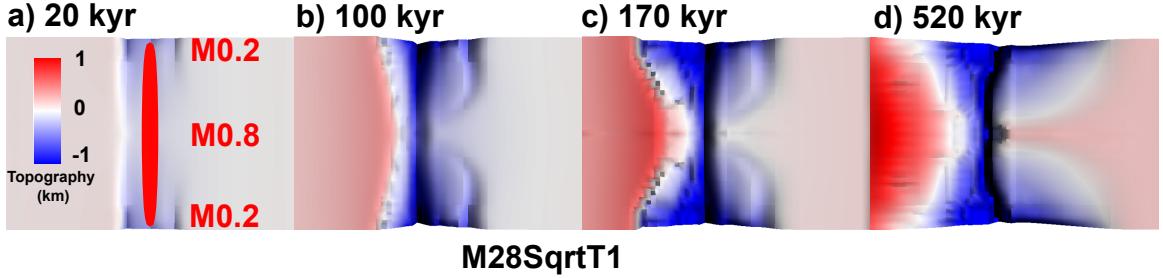


Figure 19: Bird's-eye view of the evolution of the hourglass-shaped median valley. It is generated by attaching the topography of the M28SqrtT1 model to its mirror image by assumming symmetrical M variation (0.2 to 0.8 to 0.2).

boundary of the hourglass on the left. By ~ 170 kyr (Figure 19.c), the hourglass-shaped median valley continues to widen and deepen. Since the area of the cross-section along the ridge inside the hourglass-shaped median valley is approximately inversely proportional to the local M values, the shape of the hourglass varies with different ranges and functional forms of M variations.

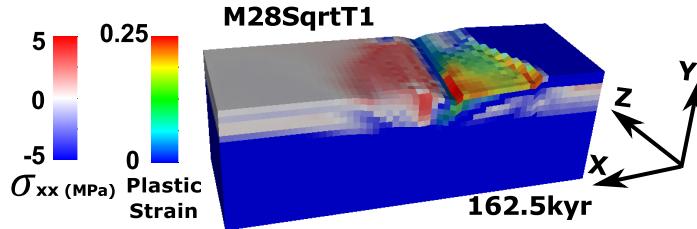


Figure 20: Higher σ_{xx} shows inside the median valley on the positive x -axis direction of the ridge axis.

In addition, the further depression inside the median valley is mostly due to the elastic deformation from crustal extension. As shown in Figure 20, the σ_{xx} in the median valley is higher because the brittle crust is thinner, when same amount of force propagates from far field extension to the median valley, the stress increases. This increased σ_{xx} is responsible for the further depression and extension of the median valley on the side of the ridge axis with no active faulting (Figure 19.d). For the median valley on the other side of the ridge axis, mass wasting between the breakaways and the ridge axis results in further

lowering of topography (Figure 17.d versus e (topography)).

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. At the higher M side of the ridge (Figure 8.g), a mullion structure with a wavelength of ~ 7 km shows up on the surface of the OCC. In spite of morphological similarity, corrugations and mullion structures have different formation mechanisms.

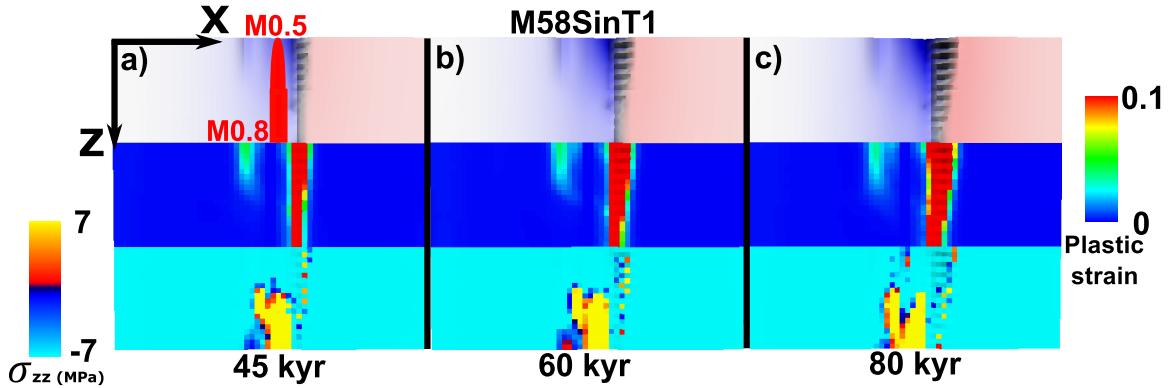


Figure 21: Bird's-eye view of the evolution of the corrugations. Color scales for the topography is the same as Figure 8.

Corrugations

The corrugation starts at the breakaway as a response to tensile stress in the z -axis direction. As shown in Figure 21, 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 generated by the asynchronous faulting along the ridge. Faulting initiates earlier at the lower M side of the ridge. As the fault offsets more on the lower M side than on the

higher M side, the footwall of the fault rotates and get uplifted more on the lower M side. This relative displacement between the footwalls along the ridge generates the isochron-parallel tensional 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 propagates away from the ridge axis and thus produces the linear corrugations that are parallel to the spreading velocity. Detailed analysis along with simpler model experiments are given in the “Discussion” section.

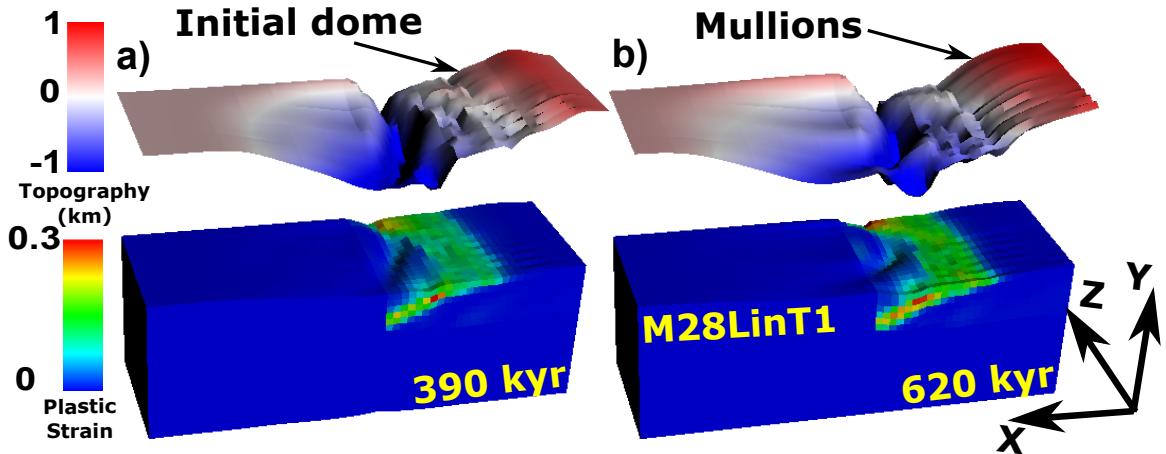


Figure 22: Evolution of mullion structures.

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 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. Where the termination is bent inward to the ridge axis, an “initial dome” (Figure 22.a) is produced once the footwall 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 footwall of the detachment fault keeps being exhumed to the surface following the trace of the detach-

ment fault termination, a mullion structure is produced (Figure 22.b).

3.3 Effects of the functional forms of M variation

3.3.1 M28T1

Major differences among M28LinT1, M28SinT1 and M28SqrtT1 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.

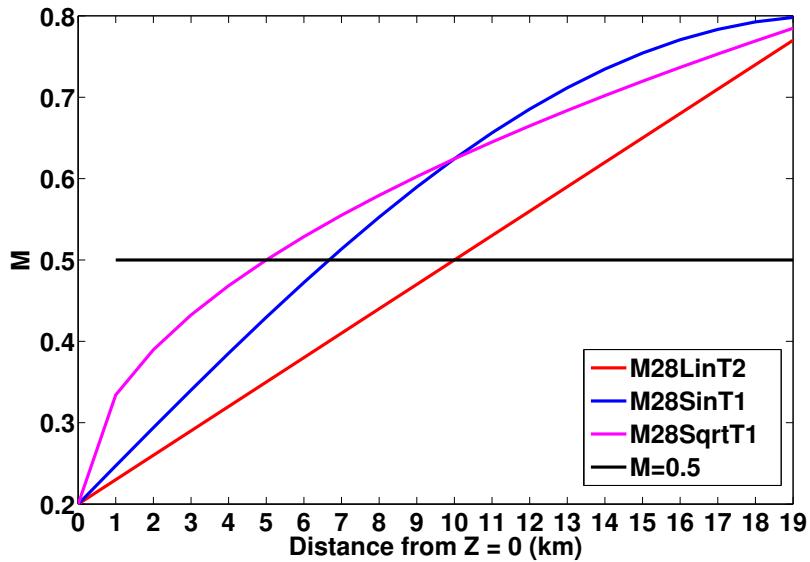


Figure 23: Three functional forms of M variation. M begins to exceed the $M = 0.5$ black line at $Z=10$ km, 7 km, 5 km for M28LinT1, M28SinT1 and M28SqrtT1 respectively.

Inward fault jump

Only the linear (M28LinT1) and sinusoidal (M28SinT1) models have inward fault jump. Sqaure root model (M28Sqrt) does not have inward fault jump because during mass wasting, termination of the detachment fault retreats backward toward the ridge axis and the detachment fault is maintained near the ridge axis.

Comparing linear and sinusoidal models, timing and dimension of the inward jumping faults are different. For the linear functional form, the inward fault jump at the higher M side starts accommodating most of the extension at ~ 900 kyr and replaces the initial detachment fault (Figure 24, Figure 8.f). It nucleates from the ridge center where $M = 0.5$ and then propagates to the $M = 0.8$ end with a length of ~ 11 km. For the sinusoidal functional form, the inward fault jump takes the place of the initial detachment fault earlier at ~ 550 kyr with a length of 14 km (Figure 24).

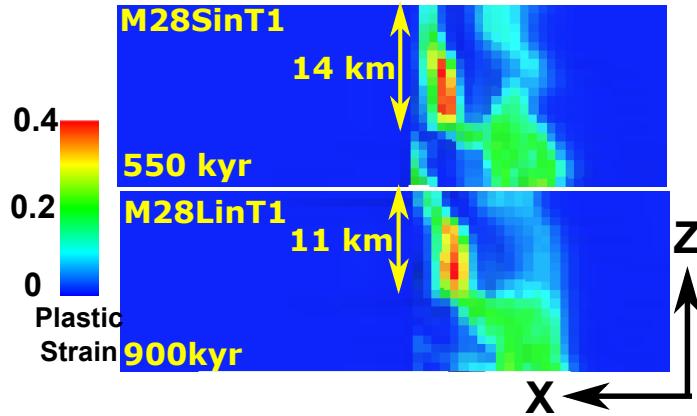


Figure 24: Bird's-eye view for comparing the length and timing of inward fault jump.

The timing difference between the linear and sinusoidal models is because M28SinT1 consistently has a higher M value than the M28LinT1 (Figure 23), which results in the initial detachment fault at the higher M side ($M > 0.5$) of M28SinT1 being 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 23).

Mass wasting

Mass wasting only happens in the M28SqrtT1 model. Qualitatively, it is because M28Sqrt has a much higher value of $\frac{\partial M}{\partial Z}$ at the lower M side (Figure 25), which implies a larger along ridge shear stress σ_{xz} as well as a larger difference in σ_{xy} along the ridge

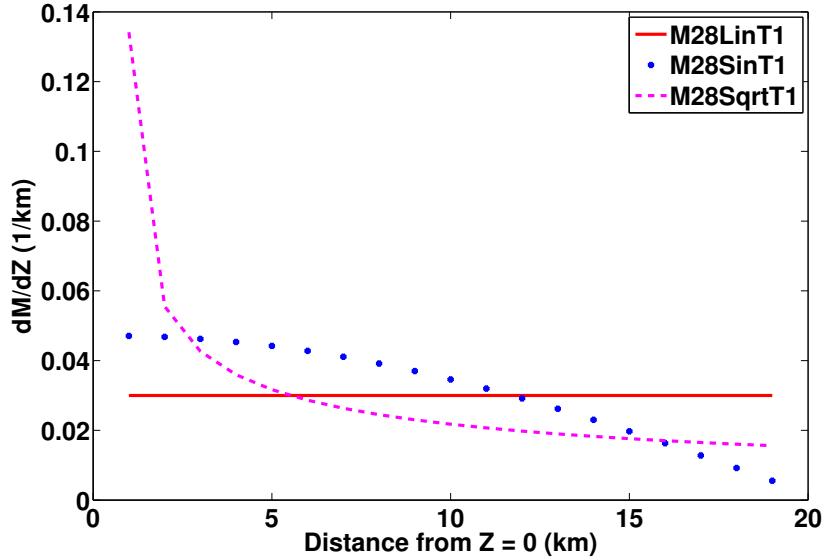


Figure 25: $\partial M / \partial Z$ comparision.

that result in the decoupling between the higher and lower M sides hanging walls.

Hourglass-shaped median valley

As shown in Figure 26, differences among the three models are identified. At 160 kyr, median valley for M28SinT1 has the smallest cross-sectional (x - y) area at the higher M side (center of the ridge segment). Whereas at the lower M side, M28SqrtT1 has the smallest cross-section. This is because the cross-section inside the median valley is inversely proportional to the local M value along the ridge. Moreover, the breakaways at the lower M sides for M28LinT1 and M28SinT1 bend to parallel to the ridge axis while the breakaway for M28SqrtT1 moves further away from the ridge axis. In addition, M28SinT1 has a trough inside the median valley with the highest curvature. At 500 kyr, M28SinT1 has the narrowest median valley 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 below initial elevation (0 km) inside the median valleys of the three models, M28Sin has the lowest value of integration because it has the largest amount of magma supply of $M > 0.5$ (Figure 23). In addition, the termination of the detachment fault of M28SqrtT1 has the highest

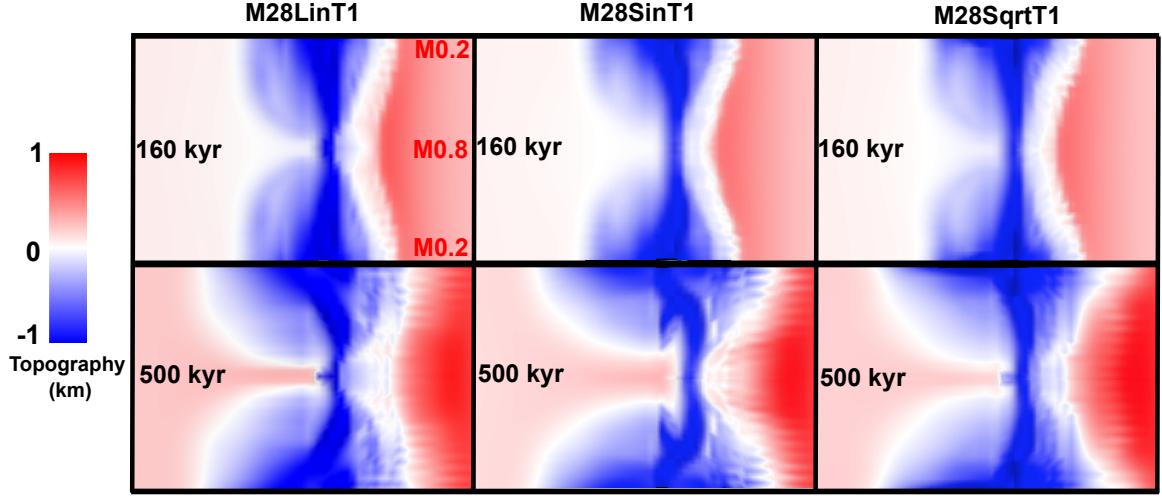


Figure 26: Bird’s-eye view of the topography. (Original and mirror images are put together assuming symmetrical M variation.)

curvature at the lower M sides. All these model behaviors correspond to the M variation (Figure 23).

3.3.2 M58T2

Table 3: Average M values (\bar{M}) of the models. (The value is calculated by integrating M along the ridge axis and divided by the length (20 km) 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

Among M58LinT2, M58SinT2 and M58SqrtT2, the major difference lies in “fault alternation”. Except for the constant M model M88ContT2, among all the models, only the models with type 2 weakening and M ranges from 0.5 to 0.8 (M58) have fault alternation. However, M58LinT2 does not produce alternating fault during the 1.1 Myr model time. Instead, one detachment fault lasts until ~ 300 kyr when the inward fault jump happens at

the higher M side ($0.65 < M < 0.8$) and replaces the initial detachment fault. This provides a lower limit of \bar{M} that prevents fault alternation and allows a long-lived detachment fault to produce an OCC. As shown in Table 3, the lower limit is 0.6425 for M58LinT2. Detail analysis of the fault alternation is given in “Discussion”.

3.4 Effects of the weakening rate

3.4.1 M57SinT1 versus M57SinT2

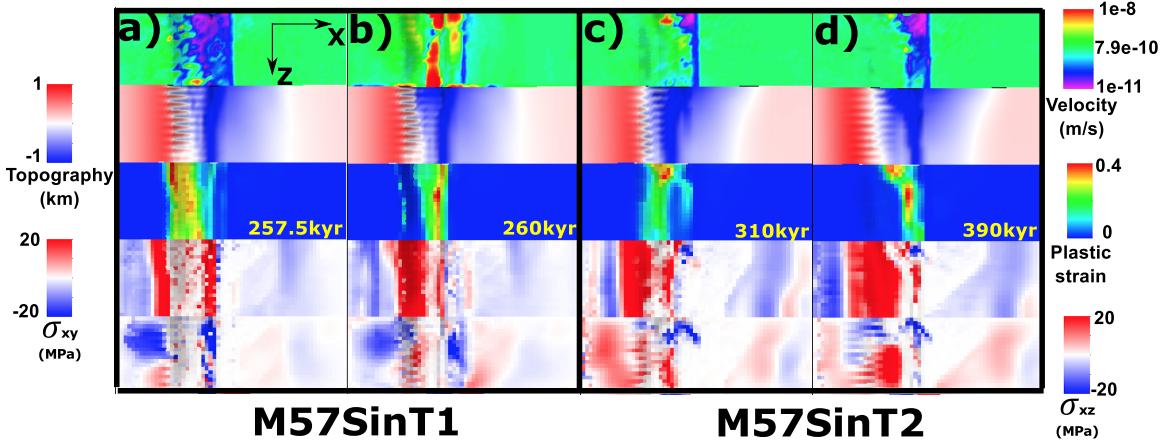


Figure 27: Bird’s-eye view of M57SinT2 versus M57SinT1 (Table 1).

Initially, both models develop normal faults on both sides of the ridge axis at the lower M side. In the model with the faster weakening rate (M57SinT1), faults propagate toward the higher M side and cut through the whole crust by 25 kyr but this process completes 25 kyr later in the model with slower weakening rate (M57SinT2). By ~ 310 kyr, the inward fault jump appears at the higher M side ($M > 0.5908$) of M57SinT2 whereas at $M \leq 0.5908$, the initial fault remains active (Figure 27.c and d). However, when the weakening is fast (M57SinT1), mass wasting happens at ~ 260 kyr and helps to maintain a relatively higher angle fault with a termination closer to the ridge axis. The initial fault remains and

no inward fault jump forms (Figure 27.a and b). In addition, the width of median valley at the lower M side is wider for M57SinT2 than M57SinT1 (Figure 27.c, d versus a, b) because slower weakening (type 2) allows a more distributed tensional stress σ_{xx} rather than fast weakening that once a fault establishes, larger amount of the tensional stress σ_{xx} is released at the fault. The amplitude of the corrugations of M57SinT1 is larger than that of slower weakening M57SinT2. This is because faster weakening rate allows a faster decrease in the cohesion. As the cohesion reaches its minimum of 4 MPa earlier when the plastic strain accumulates to 0.1, tensile failure is easy to happen in the isochron parallel direction and produces the corrugations.

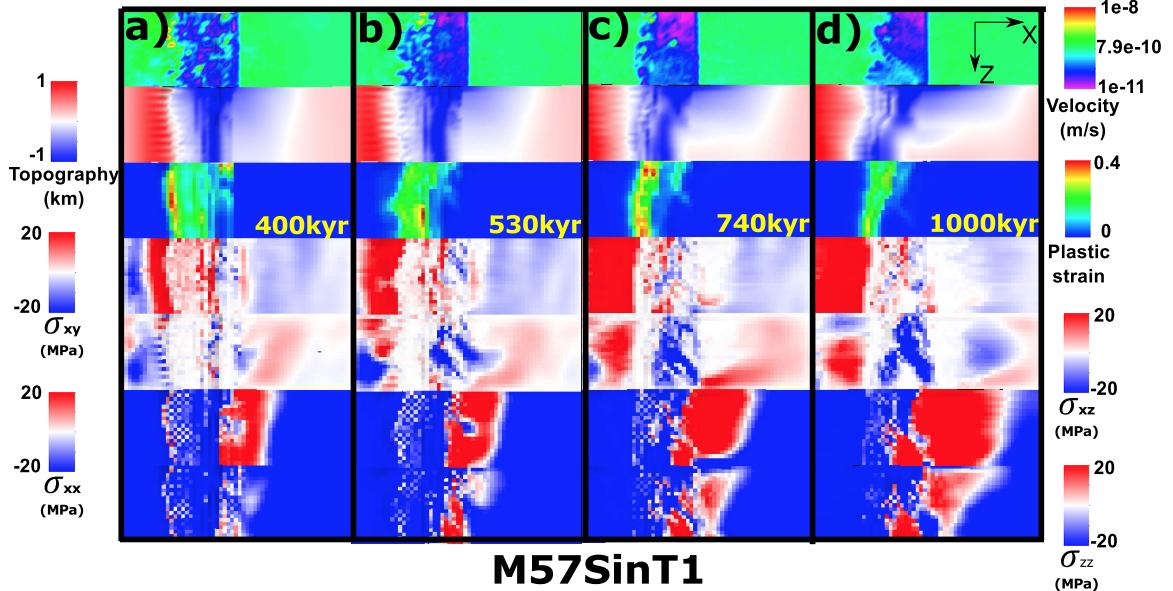


Figure 28: Bird's-eye view of faulting and stresses evolution of M57SinT1.

M57SinT1

For M57SinT1, by 400 kyr (Figure 28.a), two antithetic faults form at the lower M side ($0.5 < M < 0.6521$) accommodating part of the plate extension. This makes the termination at the lower M side retreat backward to the ridge axis. By 530 kyr, the termination at the center of the ridge segment ($M = 0.6299\sim0.6618$) moves further away from

the ridge axis (Figure 28.b). This curved termination leads to a curved topography aligning with it (white curve in the second row). By 740 kyr, another antithetic fault forms at the lower M side (Figure 28.c). It doesn't take the place of initial fault and disappear soon, however, it again releases tensional stress and helps maintain a closer to ridge axis termination at the lower M side. By 1000 kyr (Figure 28.d), an Atlantiss Massif shape OCC is produced (lower M side has a larger dome and higher M side has a smaller dome due to the along ridge termination evolution. Corrugations with wavelength varying from hundreds meters to kilometers are also produced.

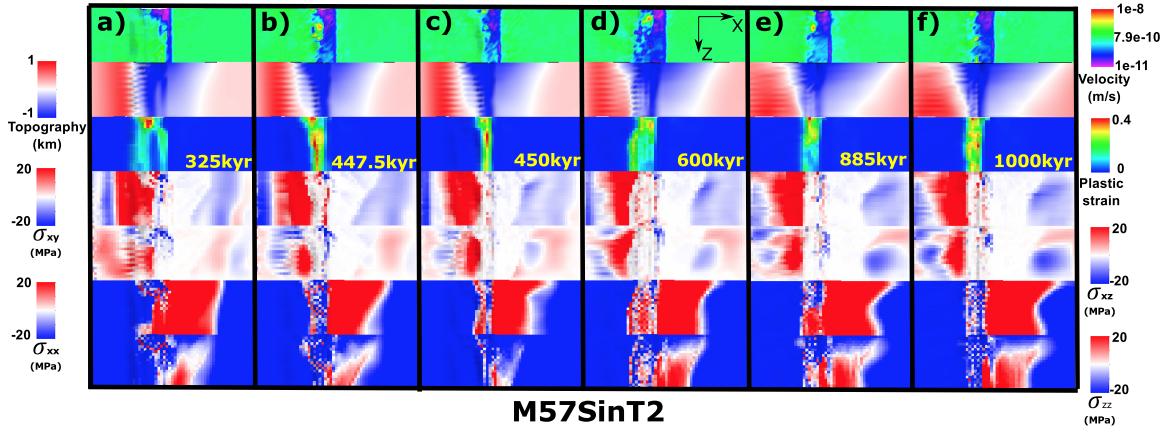


Figure 29: Bird's-eye view of faulting and stresses evolution of M57SinT2.

M57SinT2

For M57SinT2, instead of maintaining a detachment fault like M57SinT1, it produces inward fault jump at the higher M side. By 325 kyr (Figure 29.a), an inward fault jump happens and takes the place of the initial detachment fault at the higher M side. Between 447.5 kyr (Figure 29.b) and 450 kyr (Figure 29.c), a small scale mass wasting happens and the termination recedes backward to the ridge axis. By 600 kyr, the termination at the higher M side extends further (Figure 29.d). By 885 kyr, an inward fault jump happens at the higher M side ($0.62 < M < 0.7$) (Figure 29.e). The width of the median valley at the

lower M side keeps increasing due to the distributed σ_{xx} (Figure 29.a~d).

3.4.2 M58SinT1 versus M58SinT2

A major difference between M58SinT1 and M58SinT2 is that only M58SinT2 has fault alternation.

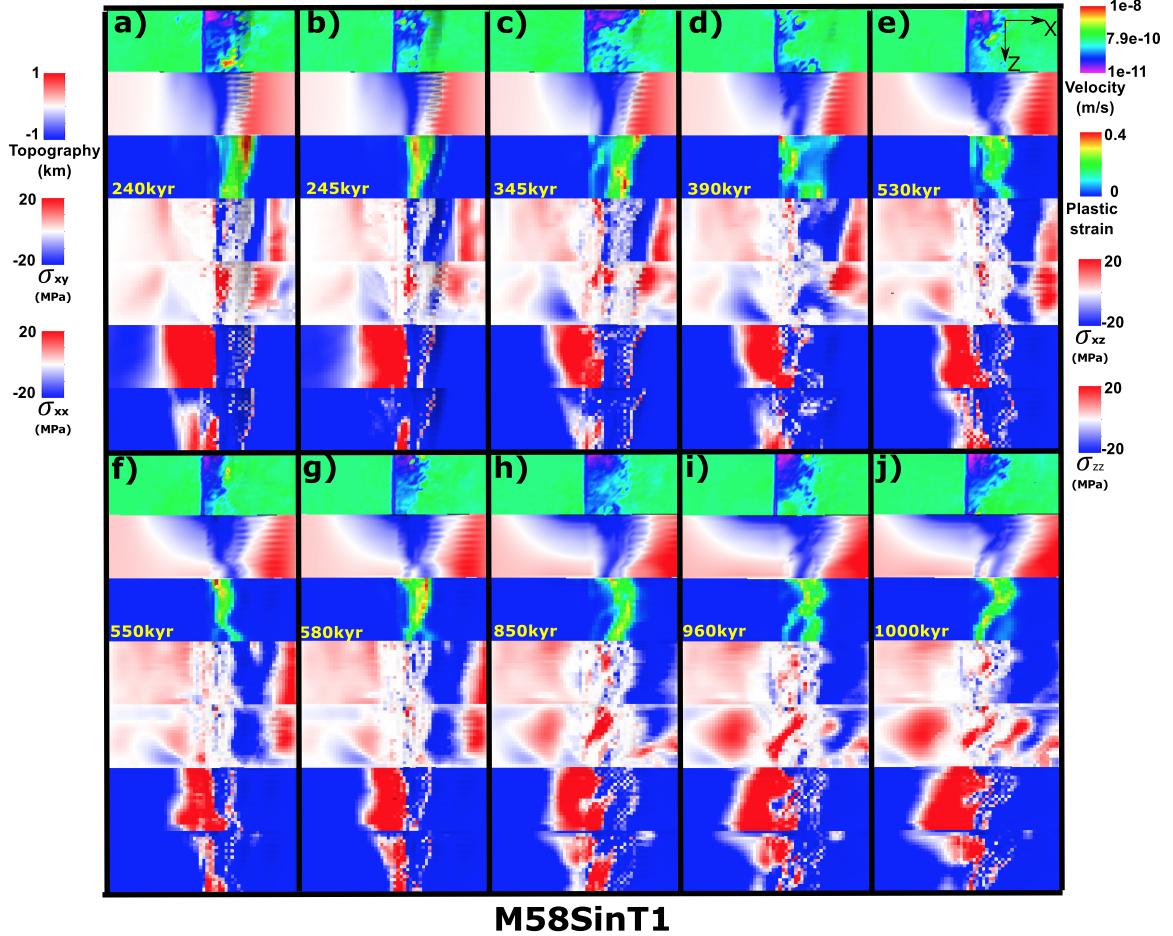


Figure 30: Bird's-eye view of faulting and stresses evolution of M58SinT1.

M58SinT1

During the 1 Myr extension of the model M58SinT1, 10 phases of the evolution of faulting are identified (Figure 30.a~j). Antithetic faults, inward fault jumps, mass wasting

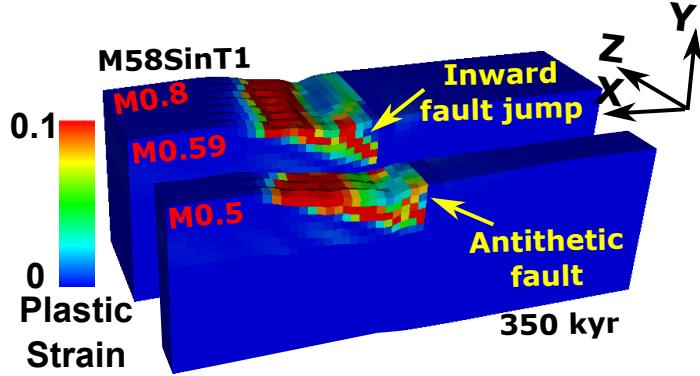


Figure 31: Plastic strain for M58SinT1 at 350 kyr.

happens with a rider block, corrugations and mullion structures produced.

By 240 kyr (Figure 30.a), due to fast weakening (type 1), cohesion along the termination is low. Stress σ_{zz} takes the advantage and generates ~ 2 km wavelength corrugations parallel to the spreading direction as the termination moves further away from the ridge axis. Between 240 kyr (Figure 30.a) and 245 kyr (Figure 30.b), a mass wasting happens at the lower M side. The termination recedes toward the ridge axis during the mass wasting. By 345 kyr (Figure 30.c), an antithetic fault forms at the lower M side ($0.5 < M < 0.5469$) with an inward fault jump happening at ridge segment with $M \in (0.5927, 0.6763)$ (Figure 31). 45 kyr later, the two weak zones break through and connect to each other and take the place of the initial detachment fault at the lower M side (Figure 30.d). Due to this inward fault jump, a sinistral σ_{xz} zone (blue area in the σ_{xz} panel) forms and is bounded by the termination of the inward fault jump near the ridge axis at the lower M side and the termination of the initial detachment fault at the higher M side. By 530 kyr (Figure 30.e), the termination of the inward fault jump at the lower M side evolves to a curve with its center moving further away from the ridge axis because the inward fault jump initiates at the center and the new fault starts slipping earlier. However, the lower M side of the curve remains closer to the ridge axis due to the antithetic fault and the other end of the curve is also closer to the ridge axis because the fault initiates later. This curved termination at the lower M side also connects to the initial detachment fault at the higher M side which

is further away from the ridge axis. Together, the curved termination is like a mirror reflected letter “S”. This flipped “S” shape termination is also shown in topography. As the curved termination at the higher M side lasts for ~ 300 kyr since ~ 390 kyr, following the shape, a ~ 10 km in wavelength and ~ 7 km in along spreading direction mullion structure is formed (Figure 32). By ~ 680 kyr, an inward fault jump happens at the higher M side ($0.7853 < M < 0.7991$). It perturbs the curved shape termination and ceases the further exhumation of the mullion structure. This inward fault jump also produces a rider block that covers the inactive detachment fault and moves off axis following the exhuming footwall of the inward jumped fault.

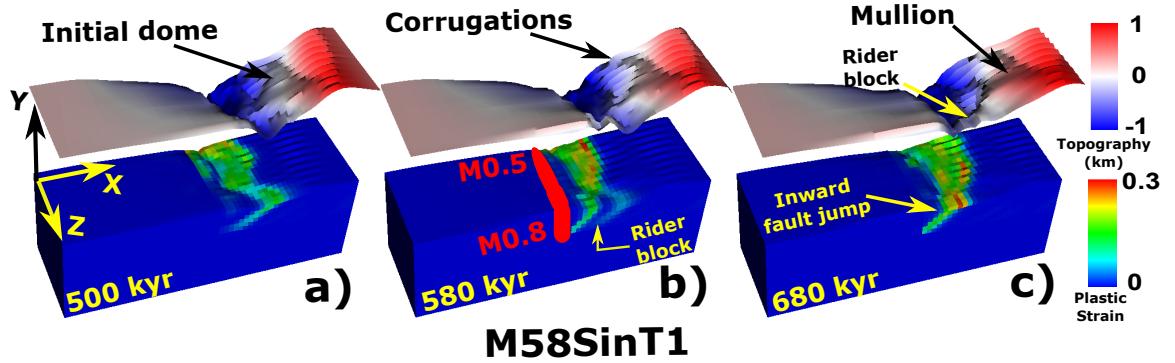


Figure 32: Evolution of faulting and morphologies of M58SinT1.

Between 530 kyr (Figure 30.e) and 550 kyr (Figure 30.f), another mass wasting happens at the lower M side ($0.5 < M < 0.5469$) where a slump block with a surface area of $\sim 9 \text{ km}^2$ flows down the topography slope into the trough. Termination recedes backward to the ridge axis. By 580 kyr, termination at the lower M side moves further away from the ridge axis due to less magma supply. Between 580 kyr and 850 kyr, due to two antithetic faults at the lower M side ($M \in (0.5, 0.5469) \cup (0.5927, 0.6567)$), the termination at the lower M side recedes and the previous mirror reflected “S” shape termination evolves to a half circle curve (Figure 30.h). The shape is also reflected in the topography. By 960 kyr (Figure 30.i)), at the ridge segment with $M \in (0.6763, 0.7121)$, another in-

ward fault jump replaces the detachment fault away from the ridge axis and retreats the termination backward to the ridge axis forming two half circle curves with wavelengths of around half of the model domain in z -axis. A large dextral shear zone (red region $\sim 40^\circ$ oblique to ridge axis) is seen in the σ_{xz} panel. The shear stress σ_{xz} is produced by the inward fault jump at the center of the ridge segment that previous hanging wall changes to the footwall of the inward jumped fault and generates an offset between the old hangingwall at the lower M end and the new footwall of the inward jumped fault. By 1000 kyr, due to the along ridge coupling, the inward fault jump propagates to the end of the high M side (Figure 30.j).

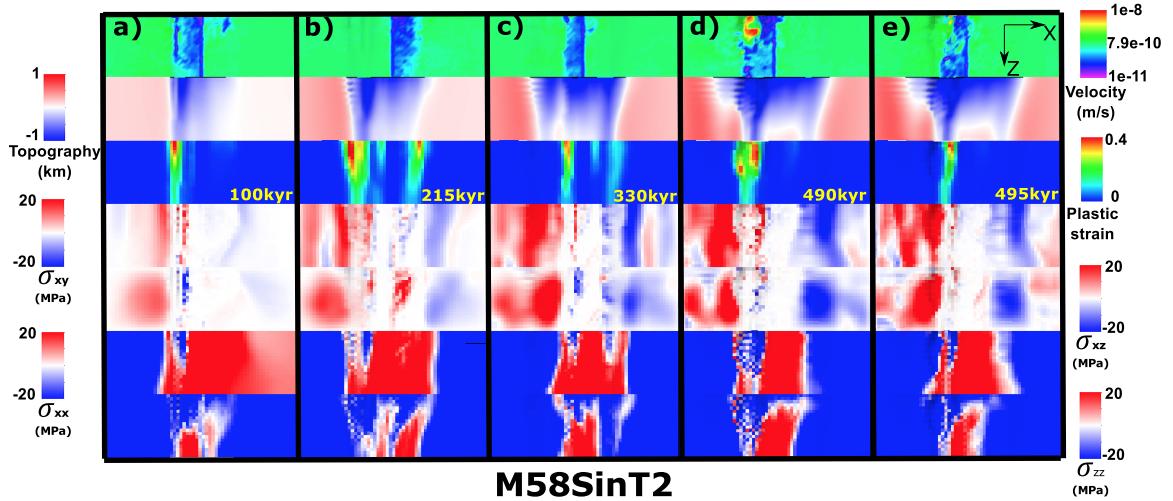


Figure 33: Bird's-eye view of faulting and stresses evolution of M58SinT2.

M58SinT2

As shown in Figure 33, a fault initiates on the left hand side of the ridge axis (Figure 33.a). The breakaway at the lower M side moves away from the ridge axis further than that of the higher M side. It takes longer time of ~ 100 kyr to form a localized fault plane through the whole ridge segment due to the slower rate of weakening (type 2). By 215 kyr (Figure 33.b), fault alternates to the conjugate plate and gradually replaces the initial one.

Corrugations are only produced at the lower M side ($M \in (0.5, 0.6763)$). By 330 kyr, fault alternates again. Between 490 kyr (Figure 33.d) and 495 kyr (Figure 33.e), a mass wasting happens with termination receding. A slump block with an area of $\sim 16 \text{ km}^2$ flows down the topography slope into the trough. With fault alternation, the shape of the median valley is no longer a typical hourglass. However, at the lower M side, the median valley is still wider and deeper. Smaller wavelength abyssal hills are produced at the higher M side.

3.4.3 M58SqrtT1 versus M58SqrtT2

The major difference between M58SqrtT1 and M58SqrtT2 is also whether the normal fault alternates or not.

M58SqrtT1

By 260 kyr, breakaway at $M = 0.5$ moves $\sim 5 \text{ km}$ further away from the ridge axis than that at the higher M end (Figure 34.a). Corrugations with a wavelength of $\sim 2 \text{ km}$ are produced along the ridge. By 370 kyr (Figure 34.b), due to larger value of $\frac{\partial M}{\partial Z}$ at the lower M side, a vertical tensile failure takes place at $M \in (0.5, 0.5949)$. Two parallel sinistral shear stress zones (blue) are seen in the σ_{xz} panel. By 400 kyr (Figure 34.c), an inward fault jump happens where $M \in (0.5949, 0.7121)$ and it propagates to the end of the higher M side and replaces the initial detachment fault at the higher M side by 460 kyr (Figure 34.d). By 590 kyr (Figure 34.e), an inward fault jump happens at the lower M side ($M < 0.5949$) and connect with the normal fault at the higher M side replacing the initial detachment fault. An $\sim 18 \text{ km}^2$ triangular shape (bird's-eye view) rider block is produced at the lower M side. Termination at the center of the ridge segment moves the furthest away from the ridge axis. This is because the previous inward fault jump first initiates there and starts slipping earlier. It is also because the value of M is lower at the segment center than the higher M end. By 660 kyr (Figure 34.f), as the previous inward jumped

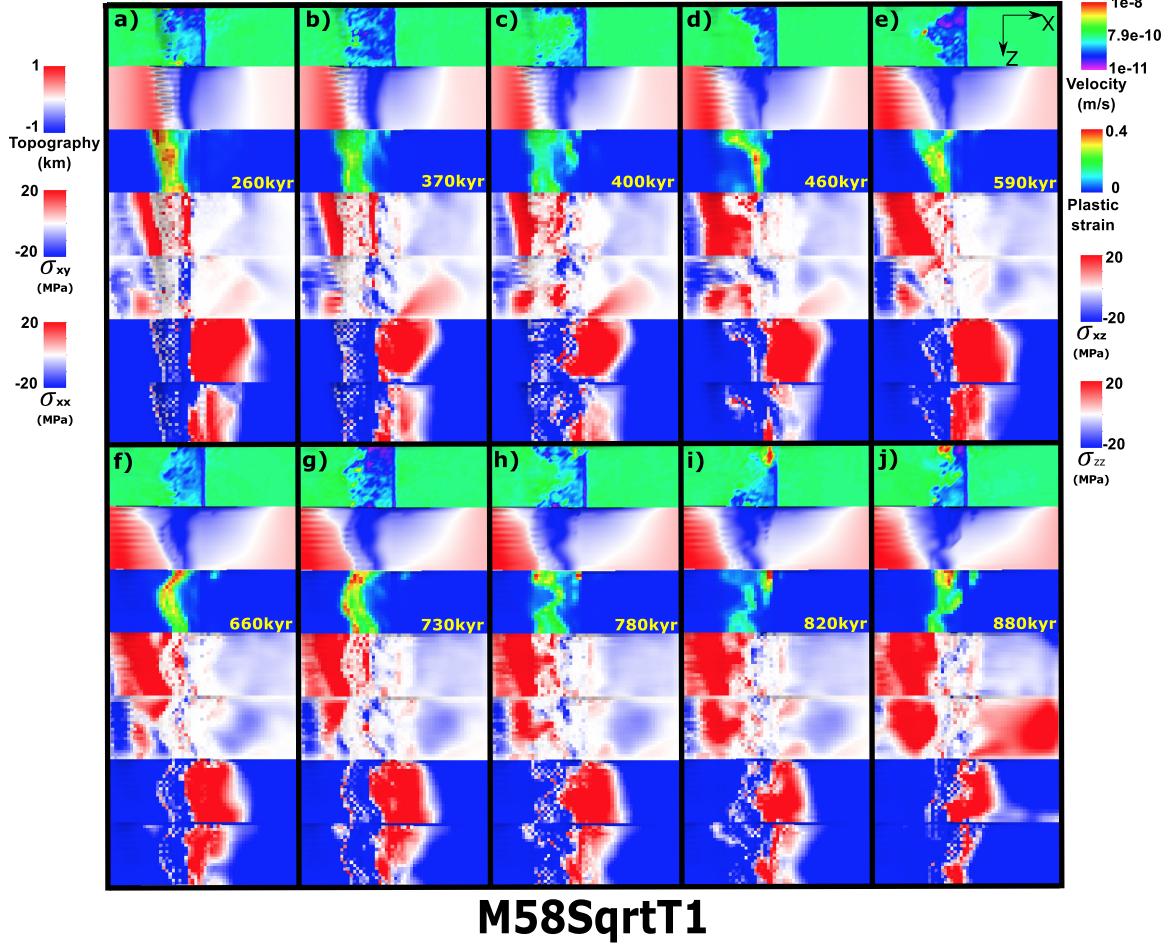


Figure 34: Bird’s-eye view of faulting and stresses evolution of M58SqrtT1.

fault at the higher M side evolves, another dome is produced. There is a hint of high angle normal fault at the region with $M < 0.5949$ on the conjugate plate. But it doesn’t develop. By 730 kyr (Figure 34.g, the termination evolves to a “half circle” and the shape is also seen in the topography. By 780 kyr (Figure 34.h), another inward fault jump appears at the ridge segment with $M \in (0.6342, 0.7121)$ and produced a curved termination with a wavelength of ~ 10 km. It has the potential to create a large wavelength mullion structure. Meanwhile, at the lower M side ($M < 0.5949$) near the ridge axis, an antithetic fault forms and propagates toward the higher M side (Figure 34.i). It triggers another inward fault jump at the lower M side and produces another rider block. The inward jumped fault later connects with the detachment fault at the higher M side (Figure 34.j). In addition, a

tensile failure shows its hint at the lower M side ($M < 0.6342$) of the conjugate plate.

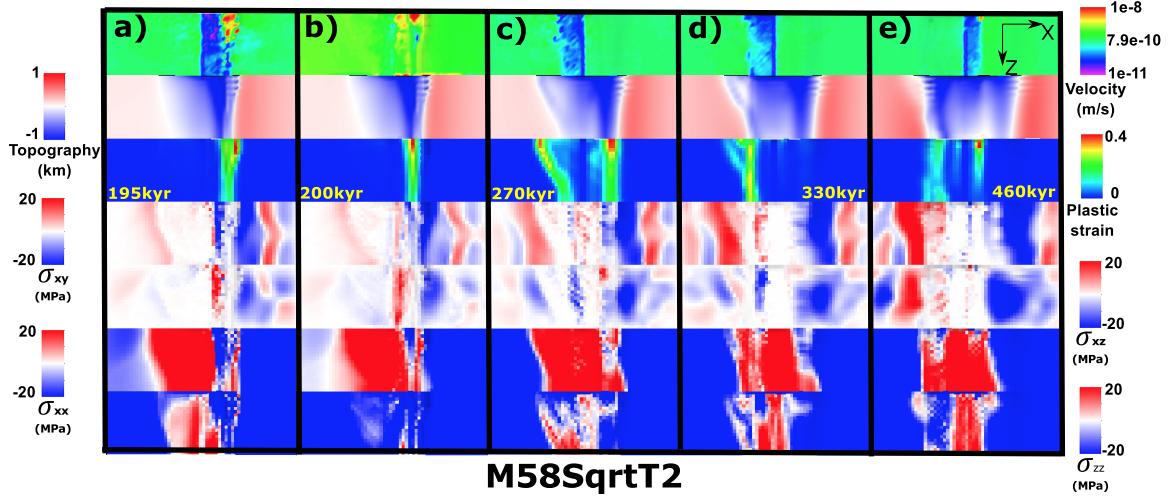


Figure 35: Faulting and stresses evolution for M58SqrtT2.

M58SqrtT2

By 195 kyr (Figure 35.a), breakaway at the lower M side moves further away from the ridge axis. Three corrugations begin to show up due to isochron-parallel tensile failure. Median valley on the conjugate plate is wider at the lower M side because less magma supply results in larger amount of elastic depression. Between 195 kyr (Figure 35.a) and 200 kyr (Figure 35.b), mass wasting happens along the ridge and is followed by the termination retreat. By 270 kyr fault alternates (Figure 35.c). The shape of the alternated fault follows the curved shape shear σ_{xy} zone (red) as seen since ~ 195 kyr on the left hand side of the ridge axis. By 330 kyr (Figure 35.d), at the lower M side, an inward fault jump happens and takes the place of the old fault further away from the ridge axis. By 460 kyr (Figure 35.e), fault alternates again to the right hand side of the ridge axis.

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.

3.5.1 SinT1

M57SinT1 versus M58SinT1

For description of M57SinT1 evolution with respect to time, please refer to Section 3.4.1 and Figure 28. For description of M58SinT1 evolution with respect to time, please refer to Section 3.4.2 and Figure 30. Comparing M57SinT1 and M58SinT1, the major difference is that the faulting pattern evolution for M58SinT1 is much more dynamic with a higher frequency of inward fault jumps, mass wasting and connection of the offsetted fault zones. For M58SinT1, the inward fault jumps and antithetic faults usually replace the old ones. However, for M57SinT1, diking is not robust enough to create big enough stress perturbation along the ridge axis for inward fault jumps or antithetic faults to take the place of the original one. At the lower M side, antithetic faults only help to accommodate tensional stress which assists in maintaining a termination near the ridge axis while the termination at the higher M side gradually moves off axis. This produces an OCC with larger dome at the lower M side than higher M side which is opposite to the shape of the OCC produced by M58SinT1.

M28SinT1

As shown in Figure 36, faulting evolution is much less dynamic than that of M58SinT1. One detachment fault keeps active on the right hand side of the ridge axis. Only inward fault jump happens \sim 540 kyr (Figure 36.d). By \sim 100 kyr (Figure 36.a), breakaway at the lower M side moves \sim 4 km further away from ridge axis than the higher M side. By 250 kyr (Figure 36.b), \sim 4 corrugations begin to evolve at the lower M side ($M \in (0.2, 0.5135)$). By 420 kyr, at the higher M side, a hint of inward fault jump begins to show up.

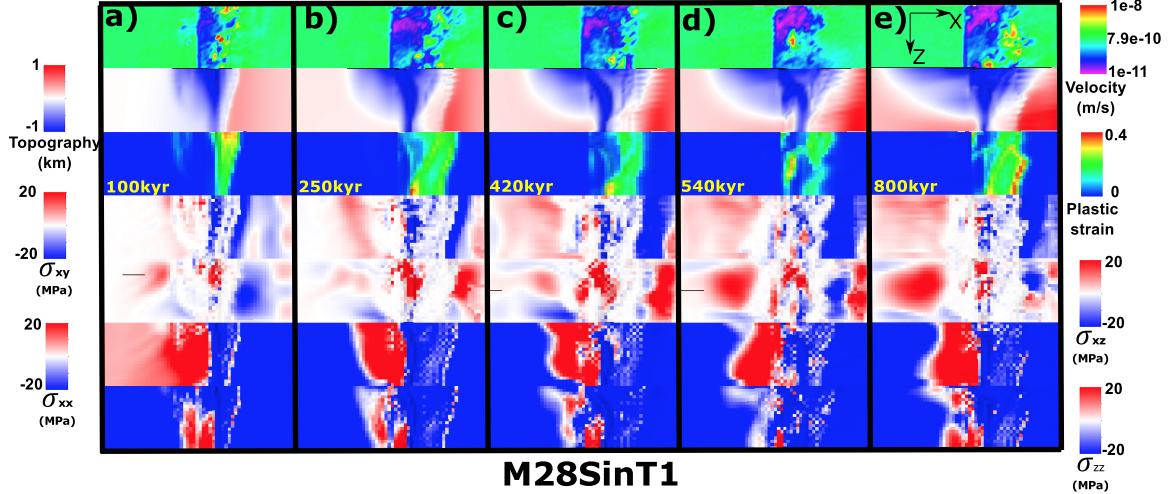


Figure 36: M28SinT1 (Table 1) faulting and stresses evolution with respect to time.

It develops into an inward fault jump by 540 kyr (Figure 36.d) and propagates toward higher M side. At the lower M side ($M \in (0.2, 0.2939)$), a tensile failure takes up part of the plate extension and helps maintain a closer to ridge axis termination than the moving termination at the region with $M \in (0.4724, 0.6562)$ (Figure 36.e). The curved toward ridge axis termination at the lower M side produces a mullion structure.

3.5.2 M57SinT2 versus M58SinT2

For description of M57SinT2 evolution with respect to time, please refer to Section 3.4.1 and Figure 29. For description of M58SinT1 evolution with respect to time, please refer to Section 3.4.2 and Figure 33. A major difference is that M57SinT2 has no fault alternation while M58SinT2 has.

3.5.3 M28LinT1 versus M57LinT1

For description of M28LinT1 evolution with respect to time, please refer to Section 3.1.

M57LinT1

As shown in Figure 37, between 160 kyr (Figure 37.a) and 162.5 kyr (Figure 37.b), a mass wasting happens. The corrugations have a relatively high amplitude. Due to less M

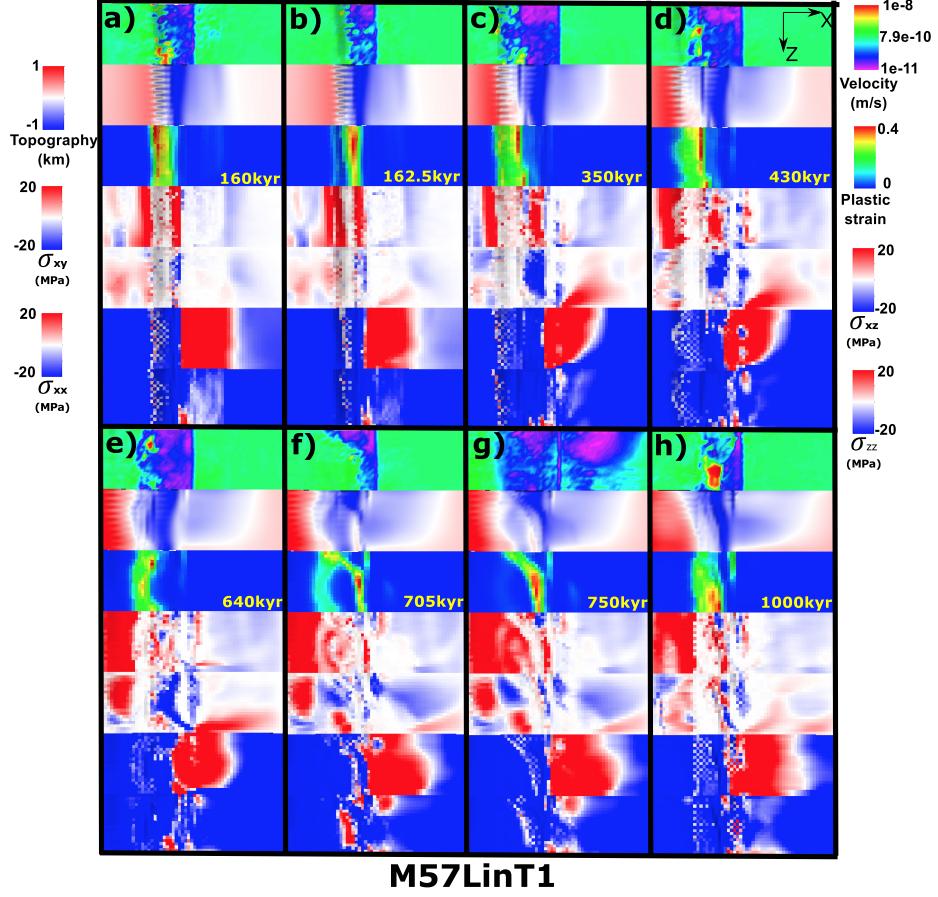


Figure 37: M57LinT1 (Table 1) faulting and stresses evolution with respect to time.

variation (0.2), the along ridge offset in breakaways is smaller. The median valley almost has a constant width along the ridge. By 350 kyr (Figure 37.c), a tensile failure happens at the region with $M \in (0.5, 0.61)$ and generates a linear topography low. Another shorter tensile failure at the higher M side ($M > 0.64$) helps maintain a high angle normal fault near the ridge axis. By 430 kyr (Figure 37.d), the tensile failure at the lower M side ($M < 0.52$) is responsible for the retreat of the termination. Whereas at $M \in (0.54, 0.59)$, the termination moves further away from the ridge axis. This curved termination results in a “dog bone” shape topography as seen at 640 kyr (Figure 37.e) and is also responsible for the large sinistral shear (blue) stress σ_{xz} . By 705 kyr (Figure 37.f), an inward fault jump happens and replaces the original detachment fault at the higher M side with a length of ~ 15 km. This inward jumped fault connects with the detachment fault at the lower M

side. The new “L” shape termination is responsible for the topography seen at 1000 kyr (Figure 37.h). The termination at the higher M side soon catches up the termination at the lower M side due to the presence of a tensile failure.

4 Discussion

4.1 Overview of model behaviors

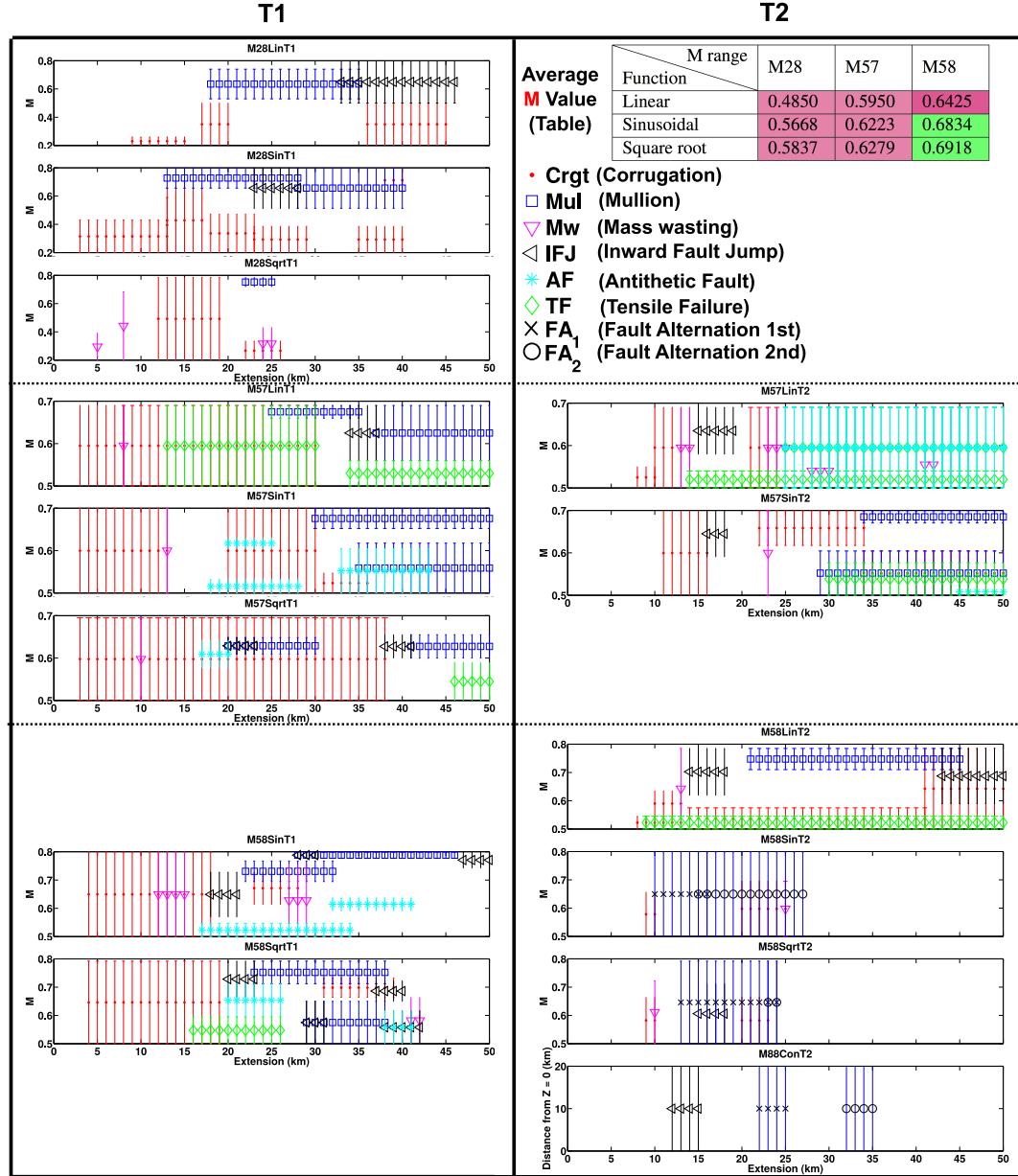


Figure 38: Initiation and duration of main structural features.

Model results show systematic changes with the average M value (\bar{M}), which is the

integration of M along the ridge divided by the ridge length. \bar{M} values of all the models are listed in Table 3. $\bar{M}_{M58} > \bar{M}_{M57} > \bar{M}_{M28}$ and within each M range, \bar{M} with square root functional form is higher than that with sinusoidal form, which in turn is higher than the linear form.

To facilitate the challenging task of describing the complicated behavior of each model as well as comparing different models, I visualize the model behavior as shown in Fig. 38. Each plot has the amount of extension in km as the horizontal axis and M values as the vertical axis. The plot can succinctly show what structure appears, how long it takes to be created or remains active, and where it is created along the ridge. For instance, from the plot for M57SinT1 ($\bar{M} = 0.6223$) in Fig. 38, one can see that an antithetic fault first shows up after about 17 km of extension on the low M end, a mass wasting event occurs after about 13 km of extension and the inward fault jump is absent in this model. Similarly, the plot for the M57SqrtT1 model ($\bar{M} = 0.6279$) clearly shows that this model experiences two inward fault jumps.

Models with type 1 weakening, on the left column of Figure 38, show more complex and dynamic behavior as \bar{M} increases. In other words, more of the main structural characteristics are created and they tend to have a higher recurrence frequency as \bar{M} increases. For instance, the inward fault jump occurs earlier and lasts a shorter period of time in M28SinT1 than in M28LinT1 due to a higher \bar{M} of M28SinT1. Furthermore, among three models with M28T1, mass wasting occurs only in M28SqrtT1, the model with the highest \bar{M} among these M28T1 models. The trend of greater complexity for higher \bar{M} continues in the group of M57T1 models. Corrugations are created earlier along the whole 20 km ridge segment than in the M28-T1 models. Unlike the M28T1 group, all three models in the M57-T1 group have mass wasting. The two M58 models have 3 or 4 inward fault jumps but the M57 models have at most two of them. Also, corrugations are created earlier in the M57 models than in the M58 ones, which suggests that corrugation favors a specific range of $\frac{\partial M}{\partial Z}$, not necessarily higher \bar{M} . Between M58SinT1

($\bar{M} = 0.6834$) and M58SqrtT1 ($\bar{M} = 0.6918$), the major difference is M58SqrtT1 has two times of inward fault jumps at the lower M side because M values are higher at the lower M side.

When comparing models with different weakening rates, the most obvious difference is that only type 2 models generate alternating faults. The models that exhibit the fault alternation do not create mullion structures because the pattern of termination can not last long enough to produce the mullion structure. Also, they have neither tensile failure nor antithetic fault.

M58LinT2 ($\bar{M} = 0.6425$) does not show fault alternation, which suggests that not only the range of M (M58) but also the average M value along the ridge \bar{M} are responsible for 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 show 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 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 with 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 39, the average M value $\bar{M} = 0.63 \pm 0.05$ for the pink area where OCCs are present whereas 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 spreading

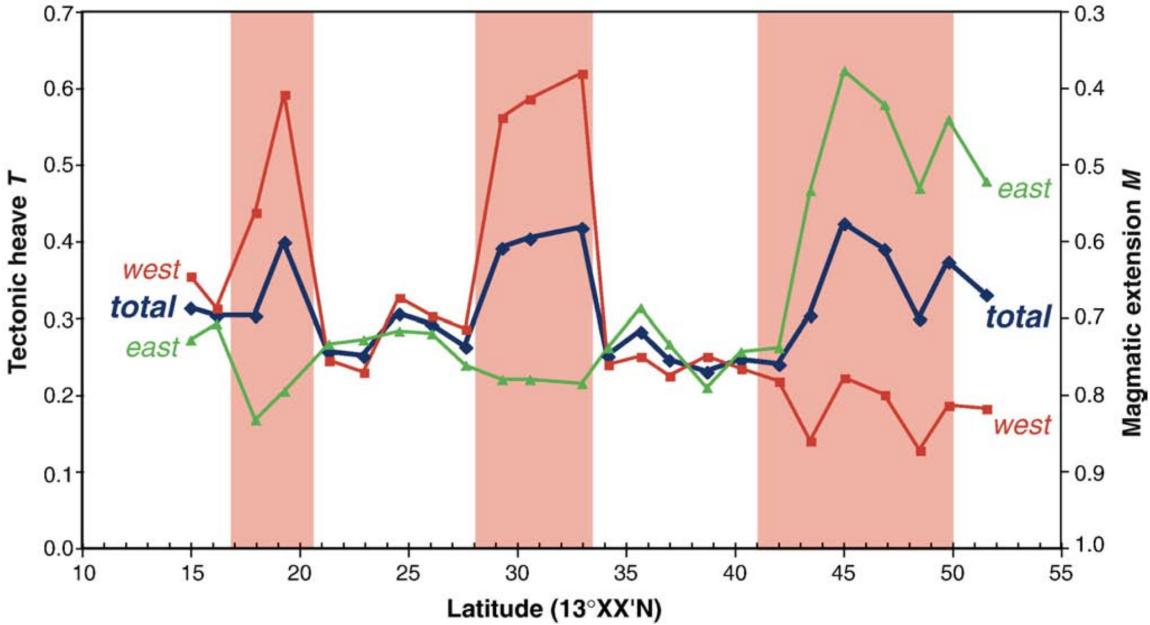


Figure 39: Along-axis variations in total accumulated tectonic heave T in the past 1.86 Ma (since chron C2n), and consequent inferred magmatic component M ($= 1 - T$) as a proportion of total plate separation (blue line). Pink shaded areas delineate loci of the active or recently active OCCs. The relative 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].

ridges: 1) long wavelength OCCs generated by a detachment fault; 2) short wavelength abyssal hills result 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 work 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 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] suggest an explanation from a 2D

model study that magma supplied in the ductile lower crust and upper mantle does not affect the faulting pattern and allows OCCs to be created under excessive diking. However, our 3D model results provide an alternative explanation: 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 38.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].

4.2 Comparing model results with natural observations

4.2.1 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 40. In the beginning of a long amagmatic phase (Figure 40.a~c) of seafloor spreading, a high angle normal fault cuts through the brittle lithosphere and roots in the brittle-ductile transition (BDT) (Figure 40.a). When the fault keeps slipping, the breakaway moves off axis and the fault begins to rotate to a lower dip angle (Figure 40.b). Then, the exposed fault surface rolls 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 cutting the detachment fault surface is probably caused by the bending stresses during footwall roll over (Figure 40.c). Then,

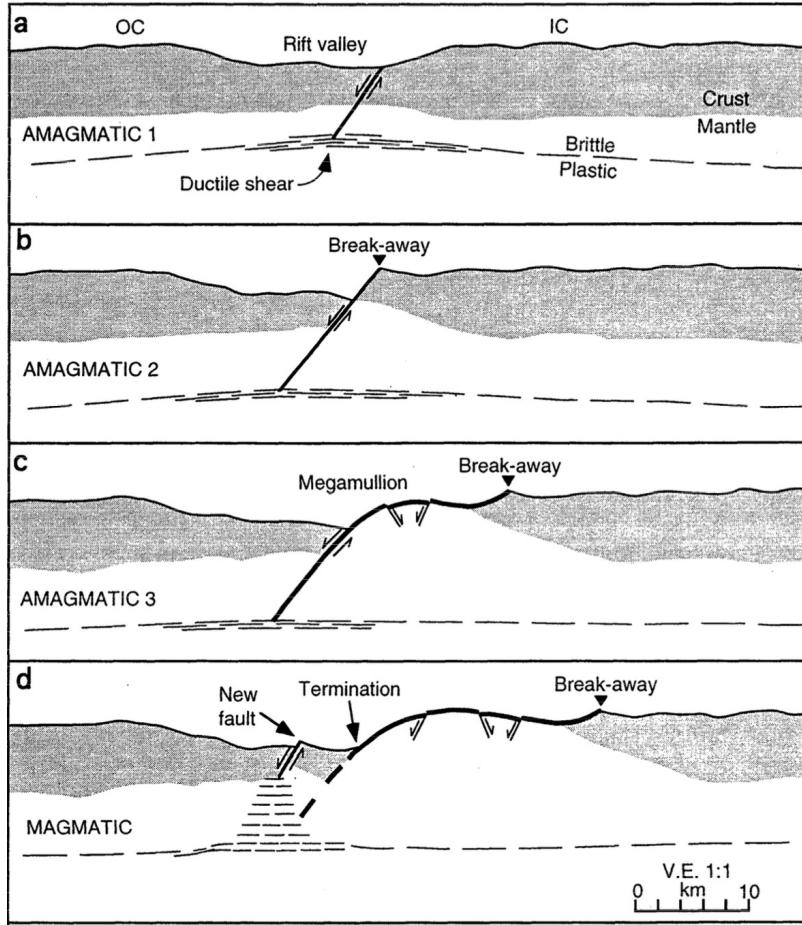


Figure 40: Schematic development of a megamullion. No vertical exaggeration. (a)~(c) shows the detachment fault evolution during amagmatic phase. (d) Increased magma supply pushed the detachment fault away from ridge axis and forms a new normal fault near the ridge axis (“inward fault jump”). Adapted from [Tucholke et al., 1998].

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 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 jumping faults last less than 5 km of plates extension before the mantle material can be exhumed to the seafloor. However, the M28LinT1 model produces an inward fault jump lasts for \sim 15 km of extension (Figure 38.M28LinT1) and produces a dome-shaped OCC adjacent to the initial one further way from the ridge

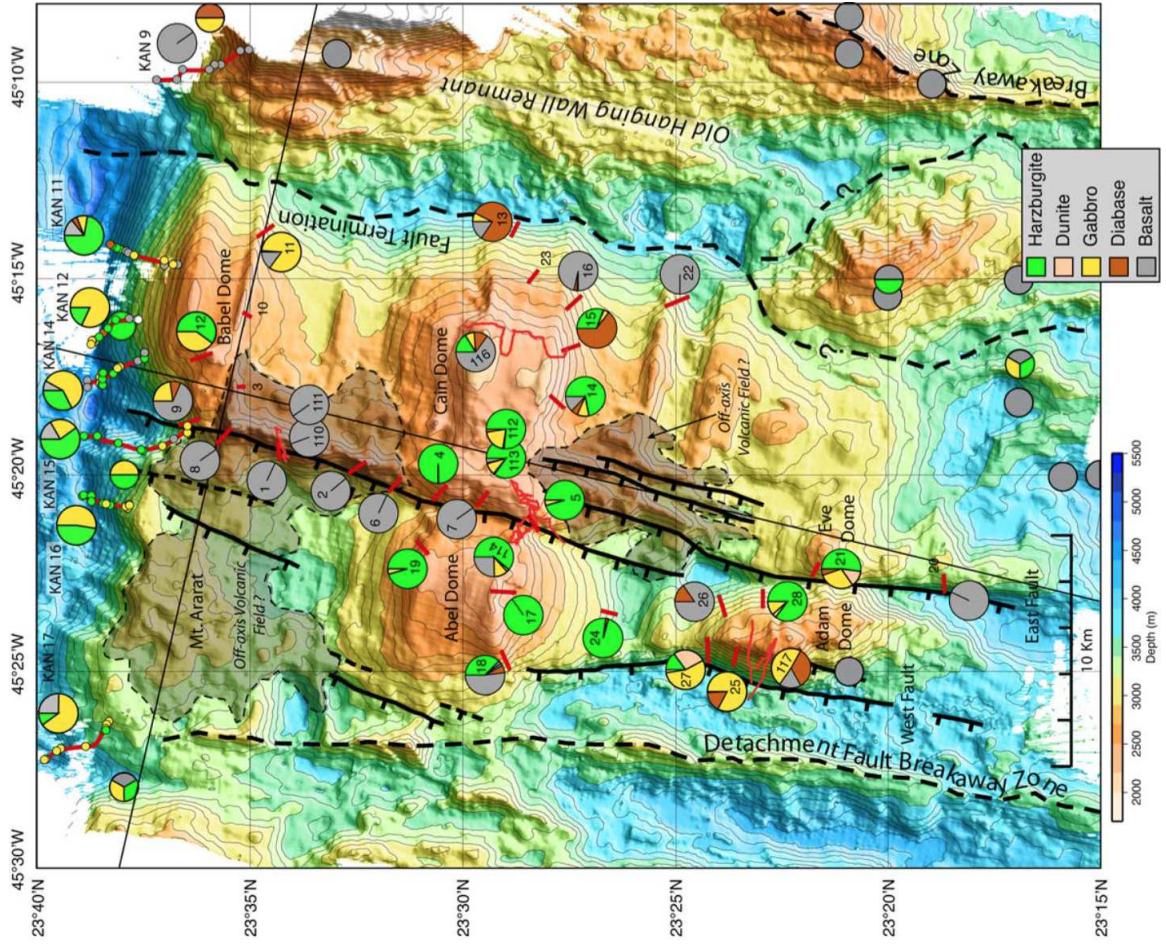


Figure 41: Bathymetric map and simplified tectonic interpretation of Kane Megamullion. Contour interval is 100 m. Pie diagrams show lithologic proportions by weight in dredge and dive collections. Detachemtn fault breakaway and termination are marked as dashed lines. Shaded regions indicate areas of possible off-axis volcanism. Adapted from [Dick et al., 2008].

axis (Figure 8.h). This behavior might explain the brother Abel and Cain domes of the Kane megamullion at 23 °N MAR. As shown in Figure 41, our model behavior is consistent with the observation in terms of the breakaway and the wavelength of the domes assuming M decreases form south to north along the ridge axis. First of all, the breakway of the detachment fault is further away from the ridge axis at the northern than the sourthern end. Second, the Abel and Cain domes are larger than the Adam and Eve domes because the inward fault jump lasts longer where M is relatively lower.

A seismic study [Xu et al., 2009] showed that the Kane OCC is terminated at around

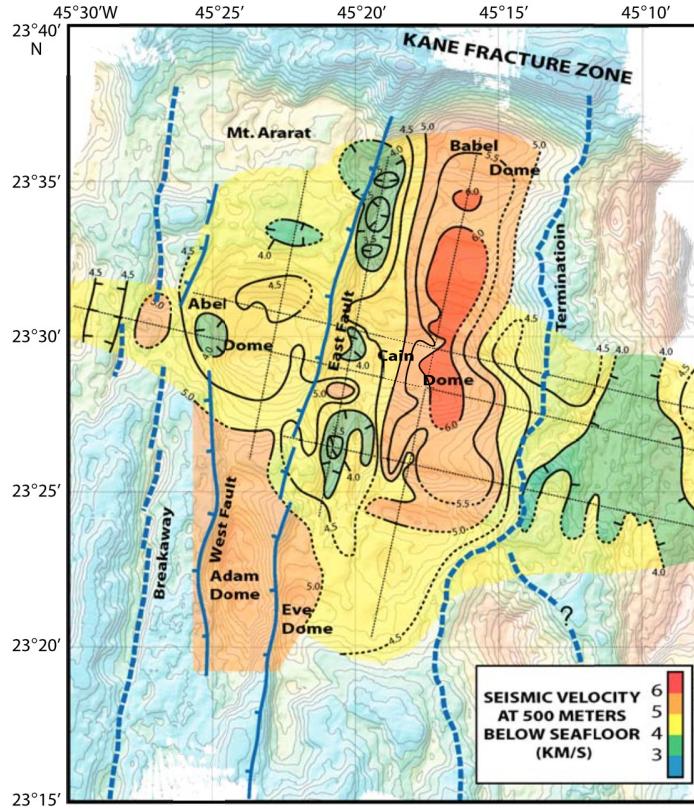


Figure 42: P wave isovelocity contours at 500 m below seafloor superimposed on bathymetry and simplified tectonic interpretation. Adapted from [Xu et al., 2009].

2.1 Myr ago 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 that is eastern to the Cain dome has a lower velocity corresponding to basaltic rocks (Figure 42).

4.2.2 Fault alternation

For slow- to intermediate-spreading ridges, the off-axis morphologies have two end members. One is shorter wavelength abyssal hills that are relatively symmetrical across the ridge axis. They are usually found closer to the ridge segment center (crossection A-A' in Figure 43). The other is the long wavelength asymmetrically spreading OCCs (crossection B-B' in Figure 43). 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

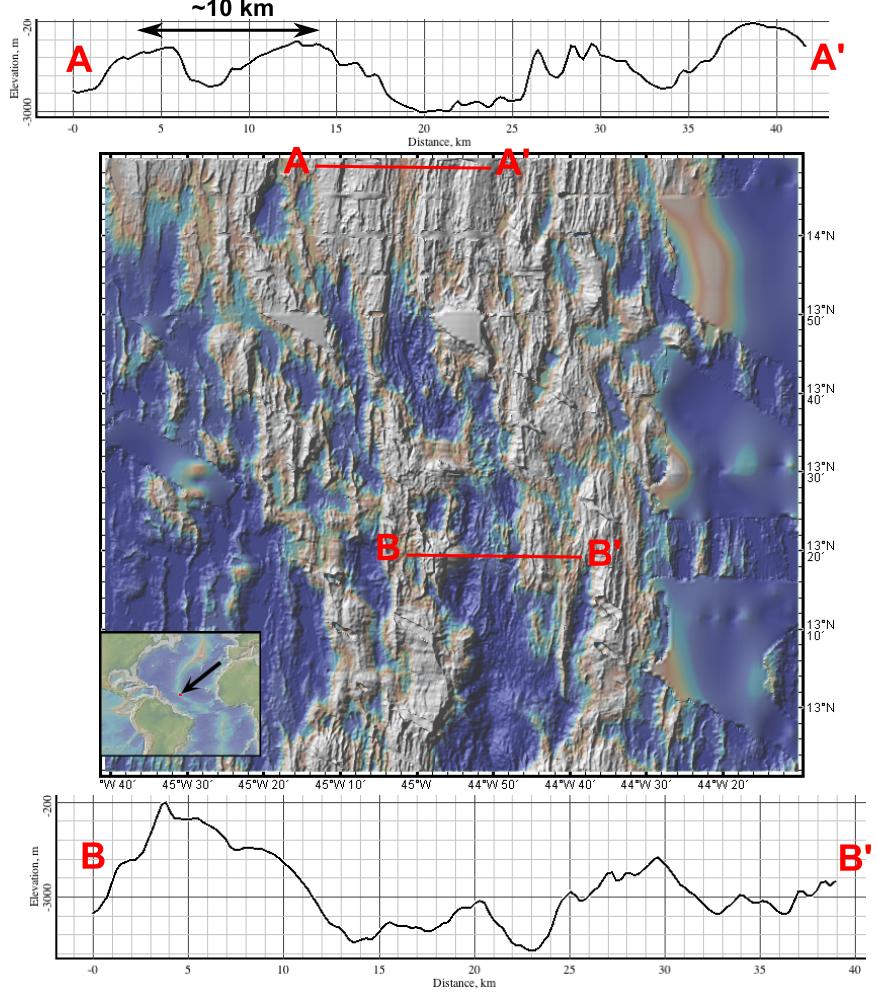


Figure 43: Bathymetry from 12.8~14.2 °N Mid-Atlantic Ridge. Crossection A~A' and B~B' are 5 times vertical exagerated. From GeoMapApp.

question. When \bar{M} is higher than 0.6425 with slower weakening rate (type 2), short wavelength abyssal hills are generated. For example, M88ConT2 produces abyssal hills with a ~ 10 km wavelength due to fault alternation (Figure 38.M88ConT2; Figure 13). Note that the wavelength of the abyssal hills in our models is consistent with the nature observation as marked in Figure 43. While when $\bar{M} \leq 0.6425$, OCC is produced.

The parameters that controls fault alternation have been proposed by Lavier et al. [2000]. Lavier et al. [2000] focused on the trade-off between bending forces (ΔF_b) as a function of fault offset ΔX and force change due to strain weakening (ΔF_w) that is also a function of ΔX . They showed that the strength weakening on the existing fault combined

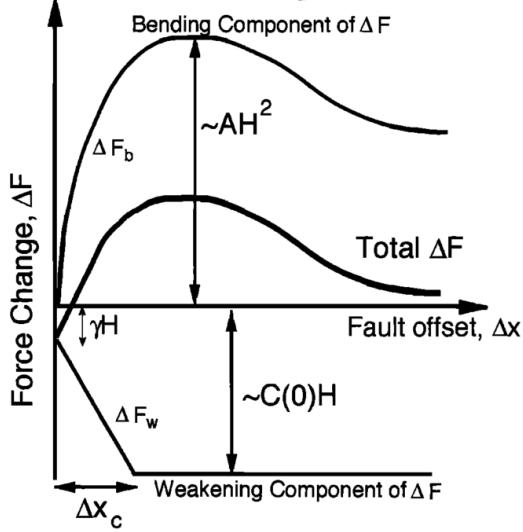


Figure 44: Trade-off between change in bending force ΔF_b and weakening in the fault interface ΔF_w . H is the thickness of the brittle crust and γ is the size of initial weak perturbation and A defines the maximum bending force change. Adapted from [Lavier et al., 2000]. (For more details, please refer to [Lavier et al., 2000])

with bending force resists further offset of the fault plays a major role in determining the stress state at regions other than the active fault. Their analysis explains how higher characteristic fault slip (ΔX_c) or slower strain weakening results in multiple faults rather than only one fault. As ΔX increases on a fault at a spreading center, the change in bending force ΔF_b increases and the strength at the fault interface decreases due to weakening ΔF_w (Figure 44). If the net force change $\Delta F = \Delta F_b + \Delta F_w$ is positive, it means that it becomes harder to maintain the existing fault and stress begins to accumulate at 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 and then maintains 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 are consistent with this analysis in that only weakening (slower weakening with higher ΔX_c) produces an alternating normal fault on the conjugate plate.

4.2.3 Mass wasting

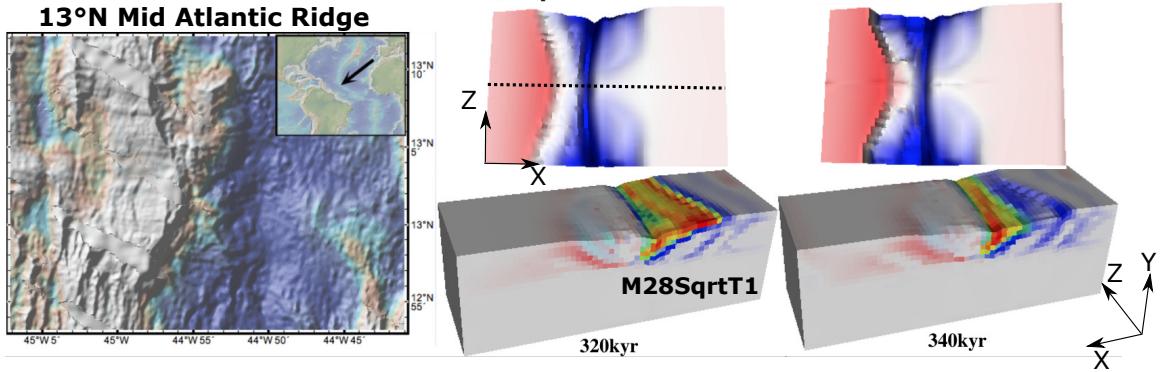


Figure 45: Comparing bathymetry at 13°N Mid-Atlantic Ridge to the mass wasting behavior of M28SqrtT1. The model topography is a mirror symmetric flip according to the dash line, it shows the case of M varies in a square root functional form from 0.2 to 0.8 to 0.2. The bathymetry image is generated by GeoMapApp [Ryan et al., 2009].

The mass wasting behavior in M28SqrtT1 model produces a fault scarp of $\sim 1\text{km}$ in relief, 40km in length along isochron-parallel direction. It forms when the detachment fault rolls over and induces bending stressses that generate the high angle fault which cuts the weak detachment fault surface and decouples the spreading breakaway and the highly deformed fault hanging wall. 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. However, the 1 km relief might be relecting the model resolution of 1 km so further investigation is necessaray. In addition, a magnetic anomaly study of the region by [Smith et al., 2008] reveals a perfect match at 13°5'N between the bathymetry and the magnetization that the suspected eastward moving block corresponds to the positive magnetic anomalies that is detached eastward from the youngest isochron on the American plate side of the ridge axis (Figure 46). It might imply that the block at 13°5'N adjacent to the curved fault scarp has a relative displacement toward the East and thus provides evidence for the mass wasting behavior of the block.

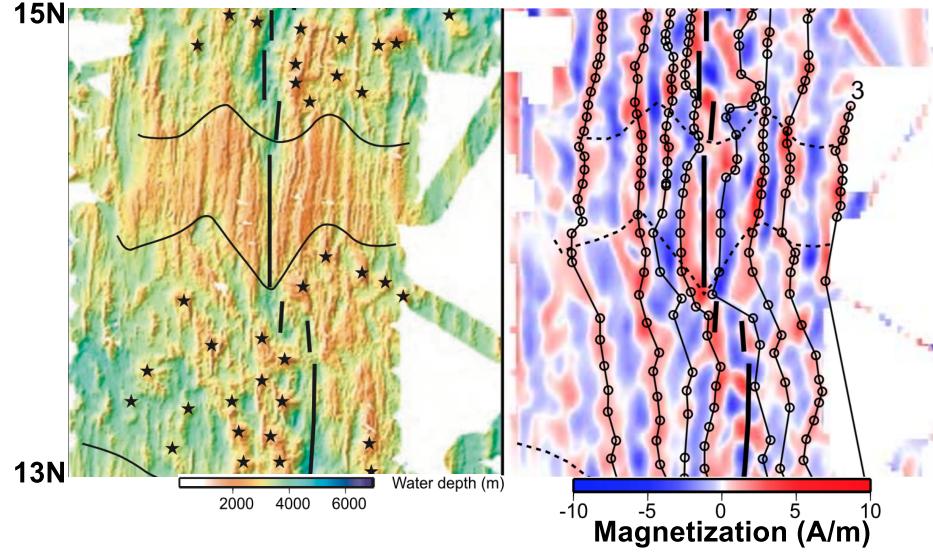


Figure 46: Bathymetry and magnetization of 13~15 °N MAR. Bold lines are ridge axes. Lined up circles are identified isochrons. Adapted from [Smith et al., 2008].

4.2.4 Hourglass-shaped median valley

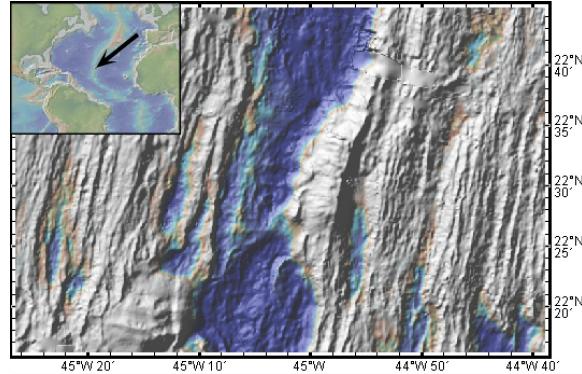


Figure 47: Hourglass median valley at 22 ° 30 'N MAR. Image is generated from GeoMapApp.

Due to the variation in diking along the ridge-axis, an hourglass-shaped median valley is also produced in the model where the narrower center corresponds to the region with higher magma supply ($M=0.8$). Hourglass-shaped median valleys are frequently observed in nature along slow-to-intermediate spreading ridges like the Mid-Atlantic Ridges [Sempéré et al., 1993]. From an analysis of the sea beam bathymetry along the

MAR between $24^{\circ} 00' \text{N}$ and $30^{\circ} 40' \text{N}$ [Sempéré et al., 1993], nine hourglass shape valleys are identified. They share similar dimensional scale ($\sim 40 \text{ km} \times \sim 40 \text{ km}$) with our model. For example, there is an hourglass around $22^{\circ} 30'$ that is 45 km long and 15 km wide (Figure 47). 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 47 corresponds to an early stage of our model because it has a relatively symmetrical and narrow shape (Figure 19.b). In addition, the relatively small variation in the valley's width along the ridge is more corresponding to M57 and M58 models than the M28 models.

4.2.5 Mullion structure

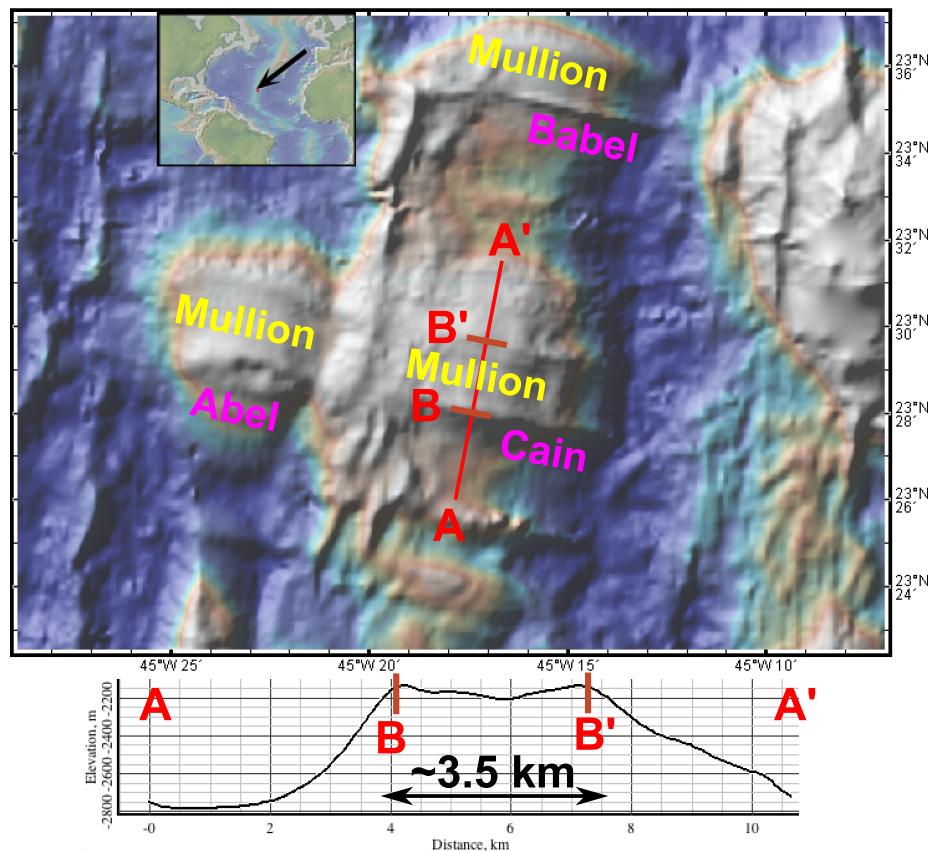


Figure 48: Bathymetry around the Kane OCC at 23°N MAR. Image is generated from GeoMapApp.

Mullion structures are frequently observed on the surface of OCCs. For instance, the mullion structure on the surface of the Cain dome has a wavelength of ~ 3.5 km as marked in Figure 48.B~B'. The geometry and length scale of these mullion structures bear significant visual similarity with those of our reference model, M28LinT1 (Figure 22). Model results indicate that the mullion structure mostly forms when the termination of the detachment fault has a curved shape that can last for a long period of time and 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 41, the spatial distribution of the mullion structure of the Cain dome is consistent with the termination that is curved inward to the ridge axis.

4.2.6 Corrugations

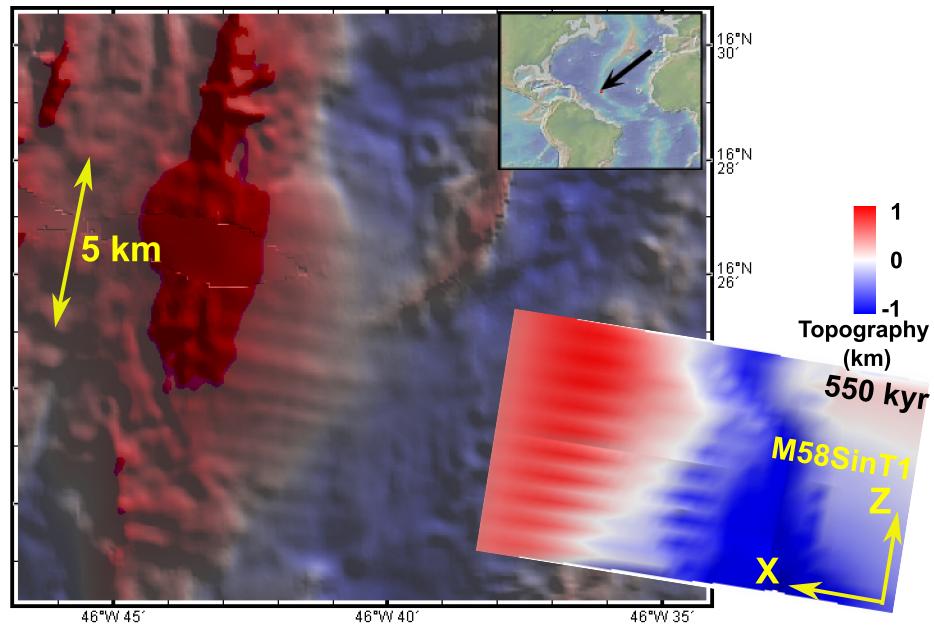


Figure 49: Comparing bathymetry at 16 °N Mid-Atlantic Ridge to the corrugations of M58SinT1 at 550 kyr.

Corrugations are also produced in the models. For instance, the spreading-parallel corrugations with wavelength of several kilometers are produced in the model M58SinT1 (Figure 49).

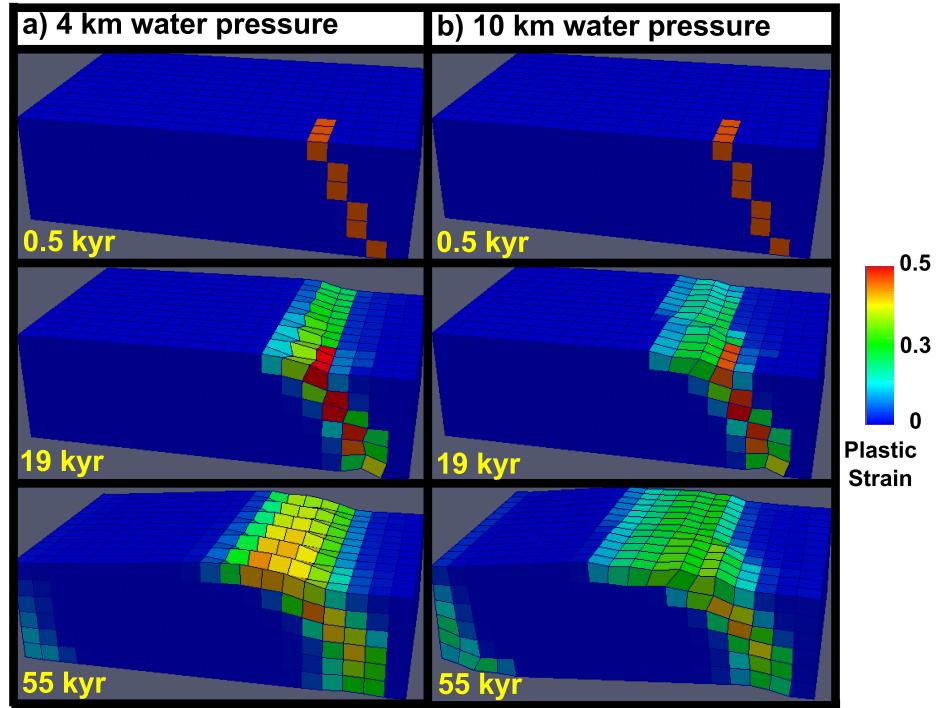


Figure 50: a) Plastic strain evolution of simpler 3D model with 4 km of seawater pressure on top; b) Plastic strain evolution of simpler 3D model with 10 km of seawater pressure on top. Seeds (elements with initial plastic strain) are partially added along the ridge segment.

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

When seeds (elements with prescribed plastic strain of 0.5) 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 50.a). However, when 10 km of sea water pressure is added on top of the seafloor to suppress tensile failure, no corrugation is produced (Figure 50.b). This verifies that the

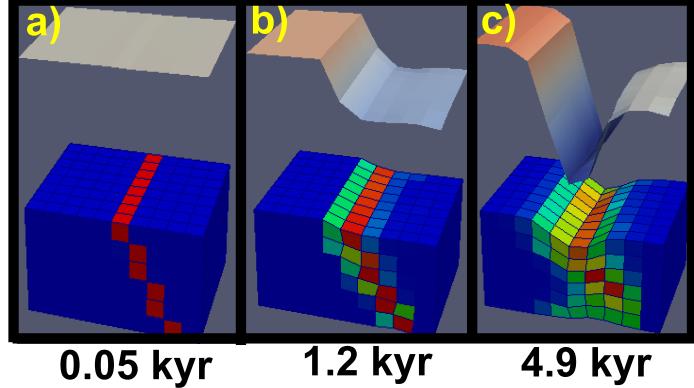


Figure 51: Simpler 3D model with 4 km of seawater pressure on top. Seeds are added along the whole ridge segment. Color scale is the same as Figure 21.

corrugation is related to tensile failure at the exposed fault plane.

In addition, when seeds are added along the whole ridge segment, which results in synchronous faulting, no corrugation is generated (Figure 51). This verifies that along ridge asynchronous faulting induced isochron-parallel tensional stress is necessary for creating corrugations.

As mentioned in [Smith et al. \[2014\]](#), wavelength of corrugations varies from meters to kilometers. Different scales corrugations sometimes coexist on the surface of a single OCC (e.g. [MacLeod et al., 2009](#)). Although corrugation is a widely observed phenomenon, its formation mechanism is still mysterious [[Smith et al., 2006](#)]. Several hypotheses are proposed. [Spencer \[1999\]](#) explains the corrugation by applying the continuous casting model which is analogical to the industrial casting process that the hot and ductile footwall of a detachment fault is continuously exhumed to the surface while being casted by the cold and brittle hanging wall. The domal shape of the corrugation is produced following the shape of the hanging wall. This hypothesis is similar to the formation mechanism of the mullion structure produced in our models. [Smith et al. \[2014\]](#) proposes another hypothesis that pre-existing offsetted faults along the ridge break through and connect to each other and form a curved termination that can produce corrugations. This is called the anastomosing behavior (e.g. [Ferrill et al., 1999](#); [Wong and Gans, 2008](#)) and is also produced in our models especially when inward fault jumps, antithetic faults or ten-

sile failures happen at only part of the ridge segment which perturbs the continuity of the termination along the ridge. In addition, Tucholke et al. [1998] mentioned that as oceanic lithosphere moves off axis, horizontally isochron-parallel extension is possible due to the contraction of the cooling lithosphere.

Besides these hypotheses, I proposed that the asynchronous faulting induced tensile failure due to varying M along the ridge axis as another possible formation mechanism for the relatively uniform ~ 2 km corrugations. These corrugations can also be superimposed onto the surfaces of mullion structures or OCCs.

5 Conclusions

I model in 3D how M varying along a mid-ocean ridge segment can control the interactions between tectonics and magmatism and thereby produce the major bathymetric features observed on the seafloor. Six commonly observed MOR features are produced in our model which are inward fault jump, fault alternation, mass wasting, hourglass median valley, corrugation, and mullion structure. I show that, by comparing these features of the models and field observations, faulting history and spatial and temporal variation of magma supply can be inferred for a ridge segment.

Three controlling parameters are investigated. They are three ranges of M (i.e. M28, M57 and M58); three types 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). Although different structural features are generated by different functional forms and ranges of M variations along the ridge, the average M value (\bar{M}) along the whole ridge segment controls off-axis morphologies. When $\bar{M} > 0.6425$ with type 2 weakening, spreading occurs symmetrically and short-wavelength abyssal hills are produced. In contrast, when $\bar{M} \leq 0.6425$, a long lasting detachment fault forms, which 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 when M varies from 0.3 to 0.5 but field observation reveals cases that OCC forms with observed M beyond both lower and upper limits. We propose a new perspective that along ridge coupling can explain why OCCs can form for M smaller than 0.3 or greater than 0.5. Finally, we propose that the asynchronous faulting can generate along-ridge tensile stresses of a large enough magnitude to cause tensile failure on the surface of the dome of an OCC, producing corrugations widely observed on the OCCs in nature.

Bibliography

- V. Allken, R. S. Huismans, and C. Thieulot. Factors controlling the mode of rift interaction in brittle-ductile coupled systems: A 3D numerical study. *Geochemistry, Geophysics, Geosystems*, 13, 2012. doi: 10.1029/2012GC004077.
- G. Baines, M. J. Cheadle, B. E. John, and J. J. Schwartz. The rate of oceanic detachment faulting at Atlantis Bank, SW Indian Ridge. *Earth and Planetary Science Letters*, 273 (1-2):105–114, 2008. ISSN 0012821X. doi: 10.1016/j.epsl.2008.06.013.
- M. Behn, W. Buck, and I. Sacks. Topographic controls on dike injection in volcanic rift zones. *Earth and Planetary Science Letters*, 246, 2006. doi: 10.1016/j.epsl.2006.04.005.
- M. D. Behn and G. Ito. Magmatic and tectonic extension at mid-ocean ridges: 1. Controls on fault characteristics. *Geochemistry, Geophysics, Geosystems*, 9, 2008. doi: 10.1029/2008GC001965.
- D. Blackman, J. Karson, and D. Kelley. Geology of the Atlantis Massif (Mid-Atlantic Ridge, 30 °N): Implications for the evolution of an ultramafic oceanic core complex. *Marine Geophysical Researches*, 2002.
- R. Buck, L. Lavier, and A. Poliakov. Modes of faulting at mid-ocean ridges. *Nature*, 434 (7034):719–23, 2005. ISSN 1476-4687. doi: 10.1038/nature03358.
- J. P. Canales, B. E. Tucholke, M. Xu, J. a. Collins, and D. L. DuBois. Seismic evidence for large-scale compositional heterogeneity of oceanic core complexes. *Geochemistry, Geophysics, Geosystems*, 9, 2008. doi: 10.1029/2008GC002009.
- S. M. Carbotte, D. K. Smith, M. Cannat, and E. M. Klein. Tectonic and magmatic segmentation of the Global Ocean Ridge System : a synthesis of observations. *Geological Society of London*, 2015.

- Y. J. Chen and J. Lin. Mechanisms for the formation of ridge-axis topography at slow-spreading ridges: A lithospheric-plate flexural model. *Geophysical Journal International*, 136:8–18, 1999. ISSN 0956540X. doi: 10.1046/j.1365-246X.1999.00716.x.
- E. Choi, L. Lavier, and M. Gurnis. Thermomechanics of mid-ocean ridge segmentation. *Physics of the Earth and Planetary Interiors*, 171(1-4):374–386, 2008. ISSN 00319201. doi: 10.1016/j.pepi.2008.08.010.
- A. Dannowski, I. Grevemeyer, C. R. Ranero, G. Ceuleneer, M. Maia, J. P. Morgan, and P. Gente. Seismic structure of an oceanic core complex at the Mid-Atlantic Ridge, 22 ° 19'N. *Journal of Geophysical Research*, 115, 2010. doi: 10.1029/2009JB006943.
- H. J. B. Dick, M. a. Tivey, and B. E. Tucholke. Plutonic foundation of a slow-spreading ridge segment: Oceanic core complex at Kane Megamullion, 23 °30'N, 45 °20'W. *Geochemistry, Geophysics, Geosystems*, 9, 2008. doi: 10.1029/2007GC001645.
- J. Escartín, D. K. Smith, J. Cann, H. Schouten, C. H. Langmuir, and S. Escrig. Central role of detachment faults in accretion of slow-spreading oceanic lithosphere. *Nature*, 455, 2008. doi: 10.1038/nature07333.
- D. A. Ferrill, J. A. Stamatakos, D. Sims, J. P. e. Evans, and S. H. e. Treagus. Normal fault corrugation: implications for growth and seismicity of active normal faults. *Journal of Structural Geology*, 21, 1999.
- C. M. R. Fowler. *The solid earth: an introduction to global geophysics*. Cambridge University Press, 2004.
- C. B. Grimes, B. E. John, M. J. Cheadle, and J. L. Wooden. Protracted construction of gabbroic crust at a slow spreading ridge: Constraints from 206Pb/238U zircon ages from Atlantis Massif and IODP Hole U1309D (30 °N, MAR). *Geochemistry, Geophysics, Geosystems*, 9, 2008. doi: 10.1029/2008GC002063.

G. Ito and M. D. Behn. Magmatic and tectonic extension at mid-ocean ridges: 2. Origin of axial morphology. *Geochemistry, Geophysics, Geosystems*, 9, 2008. doi: 10.1029/2008GC001970.

S. Kirby and A. K. Kronenberg. Rheology of the Lithosphere: Selected Topics. *Reviews of Geophysics*, 25(6):1219–1244, 1987.

L. L. Lavier, W. R. Buck, and A. N. B. Poliakov. Factors controlling normal fault offset in an ideal brittle layer. *Journal of Geophysical Research*, 105, 2000. doi: 10.1029/2000JB900108.

J. Lin, G. M. Purdy, H. Schouten, J.-C. Sempere, and C. Zervas. Evidence from gravity data for focused magmatic accretion along the Mid-Atlantic Ridge. *Nature*, 344:627–632, 1990. ISSN 0028-0836. doi: 10.1038/344627a0.

C. MacLeod, R. Searle, B. Murton, J. Casey, C. Mallows, S. Unsworth, K. Achenbach, and M. Harris. Life cycle of oceanic core complexes. *Earth and Planetary Science Letters*, 287, 2009. doi: 10.1016/j.epsl.2009.08.016.

J.-A. Olive, M. D. Behn, and B. E. Tucholke. The structure of oceanic core complexes controlled by the depth distribution of magma emplacement. *Nature Geoscience*, 3, 2010. ISSN 1752-0894. doi: 10.1038/ngeo888.

J.-A. Olive, M. D. Behn, and L. C. Malatesta. Modes of extensional faulting controlled by surface processes. *Geochemistry Geophysics Geosystems*, 2014. doi: 10.1002/2014GL061507.

T. J. Reston and K. G. McDermott. Successive detachment faults and mantle unroofing at magma-poor rifted margins. *Geology*, 39, 2011. doi: 10.1130/G32428.1.

T. J. Reston and C. R. Ranero. The 3-D geometry of detachment faulting at mid-

ocean ridges. *Geochemistry, Geophysics, Geosystems*, 12, 2011. doi: 10.1029/2011GC003666.

W. B. F. Ryan, S. M. Carbotte, J. O. Coplan, S. O’Hara, A. Melkonian, R. Arko, R. A. Weissel, V. Ferrini, A. Goodwillie, F. Nitsche, J. Bonczkowski, and R. Zemsky. Global multi-resolution topography synthesis. *Geochemistry, Geophysics, Geosystems*, 10, 2009. ISSN 15252027. doi: 10.1029/2008GC002332.

J.-C. Sempéré, J. Lin, H. S. Brown, H. Schouten, and G. M. Purdy. Segmentation and morphotectonic variations along a slow-spreading center : The Mid-Atlantic Ridge (24 °00 ' N ~ 30 ° 40 ' N). *Marine Geophysical Researches*, pages 153–200, 1993.

D. K. Smith, J. R. Cann, and J. Escartín. Widespread active detachment faulting and core complex formation near 13 degrees N on the Mid-Atlantic Ridge. *Nature*, 442, 2006. doi: 10.1038/nature04950.

D. K. Smith, J. Escartín, H. Schouten, and J. R. Cann. Fault rotation and core complex formation: Significant processes in seafloor formation at slow-spreading mid-ocean ridges (Mid-Atlantic Ridge, 13 °-15 °N). *Geochemistry, Geophysics, Geosystems*, 9, 2008. ISSN 15252027. doi: 10.1029/2007GC001699.

D. K. Smith, H. Schouten, H. J. B. Dick, J. R. Cann, V. Salters, H. R. Marschall, F. Ji, D. Yoerger, A. Sanfilippo, R. Parnell-turner, C. Palmiotto, A. Zhelezov, H. Bai, W. Junkin, B. Urann, S. Dick, M. Sulanowska, P. Lemmond, and S. Curry. Development and evolution of detachment faulting along 50 km of the Mid-Atlantic Ridge near 16.5°N. *Geochemistry Geophysics Geosystems*, pages 4692–4711, 2014. doi: 10.1002/2014GC005563.Received.

J. E. Spencer. Geologic continuous casting below continental and deep-sea detachment faults and at the striated extrusion of Sacsayhuaman, Peru. *Geology*, 27, 1999. doi: 10.1130/0091-7613(1999)027<0327:GCCBCA>2.3.CO;2.

- M. Tolstoy, A. J. Harding, and J. A. Orcutt. Crustal Thickness on the Mid-Atlantic Ridge: Bull's-Eye Gravity Anomalies and Focused Accretion. *Science*, 262(October):726–729, 1993. ISSN 0036-8075. doi: 10.1126/science.262.5134.726.
- B. E. Tucholke and J. Lin. A geological model for the structure of ridge segments in slow spreading ocean crust. *Journal of Geophysical Research*, 99, 1994. doi: 10.1029/94JB00338.
- B. E. Tucholke, J. Lin, and M. C. Kleinrock. Megamullions and mullion structure defining oceanic metamorphic core complexes on the Mid-Atlantic Ridge. *Journal of Geophysical Research*, 103, 1998. doi: 10.1029/98JB00167.
- B. E. Tucholke, M. D. Behn, W. R. Buck, and J. Lin. Role of melt supply in oceanic detachment faulting and formation of megamullions. *Geology*, 36(6):455, 2008. ISSN 0091-7613. doi: 10.1130/G24639A.1.
- D. L. Turcotte and G. Schubert. *Geodynamics*. Cambridge, 2002.
- D. L. Whitney, C. Teyssier, P. Rey, and W. R. Buck. Continental and oceanic core complexes. *Geological Society of America Bulletin*, 125, 2012. doi: 10.1130/B30754.1.
- M. S. Wong and P. B. Gans. Geologic, structural, and thermochronologic constraints on the tectonic evolution of the Sierra Mazatán core complex, Sonora, Mexico: New insights into metamorphic core complex formation. *Tectonics*, 27, 2008. doi: 10.1029/2007TC002173.
- M. Xu, J. P. Canales, B. E. Tucholke, and D. L. DuBois. Heterogeneous seismic velocity structure of the upper lithosphere at Kane oceanic core complex, Mid-Atlantic Ridge. *Geochemistry, Geophysics, Geosystems*, 10, 2009. doi: 10.1029/2009GC002586.