

Fatigue, Fracture, and Flow: Modeling the Geologic Resurfacing of Icy Satellites

by

Noah Patrick Hammond

B. A., University of California, Santa Cruz, 2009

Sc. M., Brown University, 2014

A Dissertation submitted in partial fulfillment of the
requirements for the Degree of Doctor of Philosophy
in the Department of Earth, Environmental, and Planetary Science at Brown
University

Providence, Rhode Island

May 2017

© Copyright 2017 by Noah Patrick Hammond

This dissertation by Noah Patrick Hammond is accepted in its present form
by the Department of Earth, Environmental, and Planetary Science as satisfying the
dissertation requirement for the degree of Doctor of Philosophy.

Date _____

E.M. Parmentier, Ph.D., Advisor

Date _____

G. Hirth, Ph.D., Advisor

Recommended to the Graduate Council

Date _____

R.F. Cooper, Ph.D., Reader

Date _____

P. Schultz, Ph.D., Reader

Date _____

G.C. Collins, Ph.D., Reader

Approved by the Graduate Council

Date _____

Andrew G. Cambell, Dean of the Graduate School

Curriculum Vitae

Noah Patrick Hammond

Education

PhD Candidate in Planetary Science, Brown University, 2014 – present

Thesis: *Fatigue, fracture and flow: investigating the deformation and resurfacing of icy bodies and ocean worlds*

Advisors: Greg Hirth and Marc Parmentier (expected defense in Spring 2017).

Master of Science (MSc) in Geological Science, Brown University, 2012 – 2014

Thesis: *Formation of Ganymede's Grooved Terrain by Convection-Driven Resurfacing*

Advisor: Amy C. Barr

Bachelor of Science (BSc) in Earth Science, UC Santa Cruz, 2005 - 2009

Thesis: *Hydrocode Modeling of the South Pole-Aitken Basin Forming Impact on the Moon*

Advisor: Francis Nimmo

Experience

Teaching/Research Assistant, 2012 – present

Brown University, Dept. of Earth, Environmental and Planetary Science, Providence RI

- Conducted research on the evolution of icy satellites and Kuiper belt objects
- Classes Taught: Physical Processes in Geology

Instructor, 2014 - 2016

SPARK Summer Program at Brown University, Providence RI

- Taught summer school course for middle school students on planetary science
- Designed course with emphasis on learning through inquiry

Research Assistant, 2010 – 2012

The SETI Institute, Carl Sagan Center, Mountain View CA

- Created and analyzed digital terrain models of the moons of Saturn.
- Researched the thermal and tectonic evolution of icy satellites.

Research Intern, 2010

The Lunar and Planetary Institute, Houston TX

- Identified scientifically rich landing sites for future human exploration of the Moon.

Park Ranger Intern, 2009 - 2010

US National Park Service, Great Smoky Mountain National Park, Gatlinburg TN

- Designed and delivered curriculum based programs for visiting K-12 students.

Awards and Scholarships

NASA Earth and Space Sciences Fellowship, Brown University, 2013 – 2016

Title: *Investigating the Formation of Ridge and Trough Terrain on Icy Satellites by Convection-Driven Resurfacing*

Referred Journal Publications

Published:

Hammond, N.P., Barr, A. C., & Parmentier, E.M., 2016, Recent tectonic activity on Pluto driven by phase changes in the ice shell, *Geophys. Res. Lett.*, 43 pp. 6775-6782

Hammond, N.P. & Barr, A.C., 2014, Global resurfacing of Uranus's moon Miranda by convection, *Geology*, 42 (11), pp. 931-934.

Barr, A. C., & Hammond, N. P., 2015. A common origin for ridge-and-trough terrain on icy satellites by sluggish lid convection. *Physics of the Earth and Planetary Interiors*, 249, pp.18-27.

Garcia, E. J. L., Rivera-Valentin, E. G., Schenk, P. M., Hammond, N. P., & Barr, A. C., 2014. Topographic constraints on the origin of the equatorial ridge on Iapetus. *Icarus*, 237, 419-421.

Hammond, N.P. and Barr, A.C., 2014. Formation of Ganymede's grooved terrain by convection-driven resurfacing. *Icarus*, 227, pp.206-209.

Hammond, N.H, Phillips C.B., Nimmo F., Kattenhorn S.A., 2013. Flexure on Dione: Investigating subsurface structure and thermal history. *Icarus* 223(1), pp.418-422

In Preparation for Submission

Hammond, N. P., Barr, A. C., Hirth, G., Cooper, R., & Parmentier, E. M., The fatigue of water ice measured through cyclic indirect tensile tests.

Hammond, N. P., Barr, A. C., Hirth, G., Cooper, R., & Parmentier, E. M., The fatigue of icy satellite lithospheres by tidal forces; implications for the tectonic resurfacing and geological evolution.

Conference Abstracts

Hammond, N. P., Barr, A. C., Hirth, G., & Parmentier, E. M., 2015. The Fatigue of Icy Satellite Lithospheres By Diurnal Tidal Forces. *LPSC XLVI*, abs. 1511.

Barr, A.C., Hammond, N.P., 2014, A common origin for icy satellite extensional ridge-and-trough terrain by sluggish lid convection, *LPSC XLV*, abs. 1277

Hammond, N. H., Barr A.C., 2013, Determining ice shell conditions conducive to grooved terrain formation by convection-driven resurfacing, *LPSC XLIV*, abs.1771

Hammond, N.H. and Barr.. A.C., 2013, Convection driven grooved terrain formation on Ganymede: Predictions for JUICE, *EPSC*,2013-863

Phillips, C.B., Hammond, N.P., Robuchon, G., Nimmo, F., Beyer, R.A., and Roberts, J.H. 2012, Stereo imaging, crater relaxation and thermal histories of Rhea and Dione.

Hammond, N.H, Phillips, C.B., Nimmo, F., Kattenhorn, S.A., 2012. Determining elastic thickness on Dione from flexure, *LPSC XLIII* abs.2374

Hammond, N.P., Phillips, C.B., Robuchon, G., Beyer, R., Nimmo, F., Roberts, J., 2011. Crater relaxation and stereo imaging of Rhea, *LPSC XLII*, abs.2633

Phillips, C.B., Hammond, N.P., Nimmo, F., Beyer, R.A., and Roberts, J.H., 2011. Crater Relaxation and Stereo Imaging of Icy Satellites, AGU Fall Meeting, abs. 107

Potter, R. W. K, Donohue, P.H., Gallegos, Z.E., Hammond, N.P., Kring, D.A., 2011. Multi-ring basins: Where and how to best determine their structure, *LPSC XLII* abs.1445

Gallegos, Z.E., Donohue, P.E., Hammond, N.P., Potter, R.W.K., Kring, D.A., 2011. Maander crater: A case study of a landing site designed to fulfill multiple NRC [2007] science objectives, *LPSC XLII* abs. 1958

Hammond, N.P., Nimmo, F., Korycansky, D., 2009. Hydrocode modeling of the South Pole-Aitken Basin-Forming Impact, *LPSC XL*, abs. 1455

Technical Skills

Computer Programs: Experience constructing finite element and finite difference numerical models. Proficient at running and modifying finite element programs such as CITCOM, CitcomS, and Abaqus. Proficient in Latex, Microsoft Office, ArcGIS Software, NASA Ames Stereo Pipeline, and USGS Integrated Software for Images and Spectrometers (ISIS). Familiarity with ENVI/IDL, and Adobe Creative Suite.

Computer Languages: Proficient in Bash Unix shell, MATLAB and FORTRAN. Familiarity with C++, Perl and Mathematica. Experience running program on parallel processors with MPI.

Laboratory Techniques: Experience designing and completing laboratory experiments measuring the mechanical properties of ice. Experience with National Instruments equipment and software including LabView and Signal Express. Operation of servo-hydraulic loading machines. Design and construction of liquid nitrogen cooled cryo-chamber. Measuring acoustic emissions with contact sensors. Calibration of load cell accuracy and actuator PID settings. Preparation of fine-grained water ice samples. Microstructural observations of water ice by making surface replicas of sample.

Media Coverage

“Does Pluto harbor an ocean under all that ice?” by Phil Plait, Slate Magazine, 2016
<http://www.slate.com/blogs/bad_astronomy/2016/06/23/evidence_suggests_pluto_may_have_a_liquid_ocean_under_its_surface.html>

“A year of New Horizons”, by Adam Rutherford, BBC Radio segment, 2016
< <http://www.bbc.co.uk/programmes/bo7jqry1>>

“Checking in on our planetary neighbors”, by Ira Flato, NPR Science Friday, 2016
< <http://www.sciencefriday.com/segments/checking-in-on-our-planetary-neighbors/>>

“The mystery of the solar systems weirdest moon, explained” by Michael D. Lemonick, Time Magazine, 2014, < <http://time.com/3408654/miranda-uranus-solar-system-moon/>>

“Saturn’s moon mysteries multiply”, by Don Vergano, USA Today, 2013
< <http://www.usatoday.com/story/tech/columnist/vergano/2013/01/18/dione-saturn-moons-bigelow/1845419/>>

Acknowledgments

I am extremely grateful I was able to pursue a PhD in planetary sciences at Brown University. I really loved my time here. I found our department to be a very welcoming and exciting environment. I developed a lot of close friendships with both graduate students and faculty that I plan to carry with me for the rest of my life. The open and collaborative environment here also allowed me to follow my scientific curiosity and delve deeply into a wide variety of subjects ranging from rock mechanics, geodynamics, and planet formation. I want to thank everyone in the department for being friendly and helpful and for putting in the work and energy to make our department great.

Of course, I never would have had the chance to attend graduate school without the love and support of my parents. My mother, Joyce Adams, and father, John Hammond, are a huge source of inspiration to me. Joyce Adams is the first woman in her family to become a doctor. Her hard work and long hours as a pediatrician, helping thousands of children, makes me very proud to be her son. John Hammond instilled in me a deep love of music and nature. It was hiking through the mountains as a kid with my father that my fascination with geology took root. Walking through Yosemite Valley you can't help but feel reverence for and curiosity about the great physical forces that shaped the Earth. I thank my parents deeply for raising me with love and joy and teaching me the values of hard work, kindness and respect.

Some other people I should thank for getting me to Brown in the first place are Francis Nimmo and Cynthia Phillips. I took my first planetary science class at UC Santa Cruz from Francis, after which I became somewhat obsessed with icy satellites. He became my undergraduate advisor and gave me an incredibly interesting project to work on, simulating the formation of the largest impact crater on the moon. He also recommended me for a job at the SETI Institute where I worked with Cynthia

Phillips. Cynthia was a very kind and patient research advisor. She helped me build the skills I needed to become a successful planetary scientist.

The person I am most deeply indebted to and whom I am incredibly grateful towards is my advisor Amy Barr. Amy was a truly wonderful advisor to me. She dedicated so much of her time towards teaching me the skills I would need to become an independent and successful academic. She was a fierce advocate for me and was extraordinarily dedicated to teaching students at all levels. In addition to all the academic benefits of working with Amy, she was also a joy just to hang out and talk to. I thank her for everything she was done for me and hope that one day I can be as generous and passionate advisor as she is. It was also extremely considerate of her to give me a yearlong notice that she would be leaving Brown, which helped me smoothly transition to working with Marc Parmentier and Greg Hirth.

Marc and Greg were extremely supportive during my transition. Both Marc and Greg invited me into their research groups made me feel extremely welcome and at home in the solid-Earth group. They were extraordinarily accommodating in allowing me to continue my research on icy satellite evolution, which was far outside their current research interests. Greg introduced me to a huge range of problems in rock mechanics and rheology that have direct applications to the evolution of icy satellites. Marc taught me a great deal about the evolution of magma oceans and melt transport, and really opened my eyes to the geodynamic similarities between icy satellite oceans and magma oceans. Marc was extremely generous with his time, and would work through problems with me for hours on end, helping me build a deeply rooted understanding of geodynamic problems. I am extremely grateful for his guidance and his dedication. Reid Cooper and Pete Schulz were great resources and were always available to me and willing to lend their extensive knowledge towards understanding my ice experiments. Reid was extremely generous in letting me use his lab space and getting me started on performing laboratory experiments.

I also wish to thank a number of graduate students who were extremely helpful.

Tess Caswell taught me the ways of the ice people. She helped me make my first sample, run my first rheology experiment, make my first microstructural observations and put up with me constantly making a mess of things in the lab. She was always willing to help me when something wasn't working in the lab, which was often, and she was always available to chat. Taka Kanaya was also a huge help to me. Taka was constantly lending me his insight, helping me think through difficult problems or doing something as simple as showing me where a tool was in the lab. I relied on Tess and Taka a lot, and my experiments would not have gotten off the ground without their help. I also want to thank all of my friends at Brown for helping make graduate school one most fun and exciting times of my life.

Last I want to thank my girlfriend Hannah Mellon. Hannah is pure magic and joy combined. She makes me smile every single day and reminds me why it is so good to be alive. She is incredibly caring, thoughtful, and kind. Without her love and support, I would not be whole as a person, and how would I write my thesis then!?

Contents

Title	i
Copyright	ii
Signature Page	iii
Vitae	iv
Acknowledgements	vii
Table of Contents	x
List of Figures	xiii
1 Introduction	1
1.1 Preface	2
1.2 Icy Bodies Overview	5
1.3 Endogenic Processes on Icy Bodies	7
1.3.1 Tidal Deformation	9
1.3.2 Solid-State Convection	11
1.4 Understanding Icy Bodies	15
1.5 References	17
1.6 Figures and Tables	26
2 Formation of Ganymede's Grooved Terrain by Convection-Driven Resurfacing	31
2.1 Abstract	32
2.2 Introduction	33
2.3 Methods	34

2.4	Results	36
2.5	Discussion	37
2.6	References	39
2.7	Tables and Figures	44
3	Global Resurfacing of Uranus’s Moon Miranda by Convection	48
3.1	Abstract	49
3.2	Introduction	49
3.3	Formation Conditions	50
3.4	Methods	52
3.5	Results	54
3.6	Discussion	55
3.7	Acknowledgments	56
3.8	Supplemental Material	57
3.8.1	Convection Simulations	57
3.8.2	Convective Yielding	58
3.8.3	Tidal Dissipation on Miranda	59
3.9	References	62
3.10	Tables and Figures	66
4	Recent Tectonic Activity on Pluto Driven by Phase Changes in the Ice Shell	71
4.1	Abstract	72
4.2	Introduction	72
4.3	Methods	74
4.3.1	Initial Conditions	74
4.3.2	Silicate Core	74
4.3.3	Ice Shell	76
4.3.4	Phase Changes in H ₂ O Layer	77

4.3.5	Global Volume Change	79
4.4	Results	79
4.5	Discussion	81
4.6	Conclusion	83
4.7	Acknowledgements	83
4.8	Supplemental Material	84
4.8.1	Stefan Problem Benchmarking	84
4.8.2	Comparison to Robuchon and Nimmo 2011	85
4.8.3	Heat Transfer in the Silicate Core	87
4.8.4	The Effect of Ammonia Concentration	88
4.8.5	Reduced conductivity in the ice shell	89
4.9	References	91
4.10	Tables and Figures	96

5 Experimental Constraints on the Fatigue of Water Ice: Implications for the Strength of Icy Satellite Lithospheres 104

5.1	Abstract	105
5.2	Introduction	105
5.3	Background	108
5.3.1	The Tensile Failure of Water Ice	108
5.3.2	Previous work on the Fatigue of Ice	110
5.4	Methodology	112
5.4.1	Sample Fabrication	112
5.4.2	Four-Point Bending Test	113
5.4.3	Brazil Tests	113
5.4.4	Fatigue Tests	115
5.5	Results	116
5.5.1	Monotonic Failure Results	116
5.5.2	Fatigue Results	116

5.6	Discussion	117
5.6.1	Acoustic Emission Analysis	118
5.6.2	Comparison to Other Fatigue Studies	120
5.6.3	Other Weakening Mechanisms	122
5.6.4	Crack Healing	126
5.7	Conclusion	128
5.8	Supplemental Material	128
5.8.1	Subcritical Crack Growth Mechanisms	128
5.8.2	Numerical Model	130
5.8.3	Crack Healing Comparison	133
5.9	References	134
5.10	Tables and Figures	142
6	Compaction of a Partially Molten Ammonia-Water Ice Shell: Implications for the Geologic Evolution of Triton	159
6.1	Abstract	160
6.2	Introduction	160
6.3	Ice-Ammonia System	163
6.4	Numerical Model	165
6.5	Results for constant $\bar{V}_d$	168
6.6	Icy Satellite Ice Shell Formation	169
6.7	Coupling Thermal Evolution and Melt Migration in the Ice Shell . . .	171
6.7.1	Model	171
6.7.2	Results	172
6.8	Discussion	173
6.8.1	Application to Triton	174
6.8.2	Conditions for Buoyancy	176
6.9	Conclusion	179
6.10	References	180

6.11 Tables and Figures	185
-----------------------------------	-----

List of Figures

Chapter 1

1.1	Image of grooved terrain on Ganymede	26
1.2	Image of coronae on Miranda	27
1.3	Image of possible tectonic features on Pluto	27
1.4	Image of Europa and cycloids	28
1.5	Image of Triton and cantaloupe terrain	29
1.6	Diurnal tidal stress magnitude on icy satellites	30

Chapter 2

2.1	Convection parameters explored for grooved terrain formation on Ganymede	45
2.2	Heat flux and strain rate results for convection models	46
2.3	Temperature field, surface strain rates and heat fluxes during sluggish-lid convection on Ganymede	47

Chapter 3

3.1	Mosaic image of Miranda’s southern hemisphere	66
3.2	Convection parameters and heat flows for models of corona formation on Miranda	67
3.3	Evolution of convection geometry in Miranda’s ice shell	68
3.4	Global heat flow and deformation patterns produced by convection in Miranda’s ice shell	69
3.5	Rayleigh number and viscosity in Miranda’s icy mantle as a function of the temperature	70
3.6	Benchmark comparison of Nusselt number in convection models . . .	70

Chapter 4

4.1	Temperature profiles through Pluto's ice shell	97
4.2	Phase changes, surface stresses and surface strains in Pluto's ice shell through time	98
4.3	Present state of Pluto's ice shell	99
4.4	Benchmark for Stefan problem	100
4.5	Comparison to Robuchon and Nimmo (2011)	100
4.6	The effect of ammonia on ice shell evolution	101
4.7	The effect of reduced thermal conductivity in the ice shell	102
4.8	The effect of different thermal conductivity temperature sensitivities .	103

Chapter 5

5.1	Geologic features on icy satellites that suggest their surfaces are weak	143
5.2	Brazil test geometry and stress in sample	144
5.3	Load and acoustic emissions during monotonic failure test	145
5.4	Results on constant amplitude fatigue test	146
5.5	Failure stress for increasing amplitude tests	147
5.6	Failure stress of ice as a function of previous load cycles	148
5.7	Mechanical data during increasing amplitude test	149
5.8	Cumulative acoustic emission events with time	150
5.9	Acoustic emission events and unloading	151
5.10	Comparison of fatigue data with previous experiments	152
5.11	Magnitude of impact shock waves with depth in the ice shell compared to yield stress	153
5.12	Crack healing rates with depth in the ice shell	154
5.13	Numerical Model of Creep Test	155
5.14	Stress and strain during numerical creep test	156
5.15	Vertical strain measurements of ice samples	157
5.16	Stress and strain during numerical creep test	158

Chapter 6

6.1	Ammonia-water phase diagram and melt fraction	186
6.2	Structure of ice shell with partial melt	186
6.3	Melt fraction with depth after compaction	187
6.4	Compaction zone thickness and internal melt fraction	188
6.5	The effect of different compaction viscosities on compaction	189
6.6	Melt fraction evolution in a 1km ice shell with a temperature dependent compaction viscosity	189
6.7	Early melt fraction evolution in an initially 1 m thick ice layer	190
6.8	Later melt fraction evolution in an initially 1 m thick ice layer	191
6.9	Thermal evolution of Triton with increasing concentration of ammonia in the ocean	192
6.10	Density of the ammonia-rich ocean and ice shell	193

Chapter 1 | Introduction

Noah Hammond¹.

1. Department of Earth, Environmental, and Planetary Science, Brown University,
Providence, RI 02912, USA

1.1 Preface

Our solar system is populated with dozens of icy worlds whose surfaces are geologically diverse and beautifully complex. NASA spacecraft have observed intriguing geologic surface features on the icy satellites of Jupiter, Saturn, Uranus and Neptune, as well as on large Kuiper belt objects. What types of geologic processes operate on these icy bodies to give rise to the diversity of features we observe? In this thesis I test hypotheses about the formation of geologic features on several icy bodies with an emphasis on understanding tectonic deformation, the role of solid-state convection in the ice shell, and the effect of tidal deformation on the lithosphere.

Many icy satellites show a dearth of impact craters on their surfaces, suggesting some kind of past endogenic resurfacing event and in some cases there is evidence of recent or ongoing geologic activity (Zahnle *et al.*, 2003). There is a large debate in the planetary community about how much of this resurfacing is due to tectonic activity, cryovolcanic activity or lithospheric recycling in a process analogous to plate tectonics on Earth (e.g. Kargel 1991; Pappalardo and Greeley 1995; Barr 2008; Kattenhorn and Prockter 2014). Investigating how such processes might operate on an icy body requires a strong understanding of their internal structure, composition, thermal history and orbital history.

Developing models for the geologic resurfacing of icy bodies also requires understanding how water ice deforms in response to loading. At the surface temperatures of icy bodies, a brisk $T < -173^{\circ}\text{C}$, water ice behaves elastically and when stresses are high enough, ice will fracture (behave plastically). Deeper in the ice shell where temperatures approach the melting temperature ice can behave viscously, therefor similar to Earth icy satellites may possess a rigid lithosphere above a viscously deforming asthenosphere (Reynolds and Cassen, 1979). Understanding the brittle failure and viscous deformation of water ice requires careful laboratory experiments that need to be extrapolated to planetary conditions. Numerical models can use rheological data

to better understand geologic processes such as convection, tidal deformation, and tectonic fault motion.

The key scientific questions addressed by this thesis are: 1) did solid-state convection in the ice I shell drive lithospheric deformation and geologic resurfacing on icy bodies, 2) how does the thermal evolution of icy satellites relate to their internal structure and geologic history, 3) how do repeated tidal stresses affect the rheology of the lithosphere, 4) do eutectic melts, such as ammonia-dihydride, play a role in generating near surface liquids on icy satellites? Numerical modeling as well as laboratory experiments are used to investigate these processes. Specific geologic terrains examined in this thesis include grooved terrain on Ganymede, coronae on Miranda, tectonic terrains on Pluto and Europa, and cantaloupe terrain on Triton.

Chapters 2 and 3 focus on understanding the role of convection in resurfacing icy satellites. It is shown that solid-state convection in the ice I shell may drive surface deformation on some icy satellites, leading to the formation of extensional tectonic, ridge and trough terrains. The deformation patterns, strain rates and heat flows generated by this type of convection, called sluggish-lid convection, are consistent with those inferred on the surface of Ganymede and Miranda. Sluggish-lid convection requires the surface to be very weak, with a much lower brittle yield stress than laboratory ice. Chapter 5 explores the hypothesis that the surface of icy satellites may become progressively damaged and fatigued due to repeated stresses from tidal deformation. Fatigue experiments are performed on water ice in the laboratory and the effect of cyclic loading on the tensile failure stress is measured.

Several mechanisms other than tidal deformation or solid state convection may have driven geologic activity at the surface of icy bodies. Chapter 4 explores the hypothesis that the gradual freezing of a subsurface ocean on Pluto generated extensional tectonic features at the surface. Volatiles such as ammonia likely also play an important role in the resurfacing of certain icy satellites. These impurities can have a dramatic effect on the melting temperature of water ice and can alter the density of

the liquids. Chapter 6 introduces models to help constrain the dynamics of partially molten water ice-ammonia systems in order to better understand the role volatiles play during the thermal evolution of icy bodies.

The theme connecting all of these studies is modeling the resurfacing of icy bodies by investigating the physical evolution of water ice systems through numerical models and laboratory experiments. In this chapter, I introduce the concepts relating to the properties of icy worlds, the rheology of water ice, and physics of deformation that are fundamental to subsequent chapters. Specifically I discuss 1) icy satellite composition and structure, 2) the thermo-mechanical properties of water ices, 3) and the mechanics of solid-state convection and tidal deformation.

1.2 Icy Bodies Overview

Icy bodies are one of the most abundant planetary objects in the solar system, in fact icy bodies > 200 km in diameter outnumber rocky bodies of the same size. The bulk density of icy satellites range from $1 - 3$ g/cm³, suggesting that icy bodies are composed of a mixture of silicates and water ice (Burns *et al.*, 1986). Many unmanned NASA spacecraft have flown past these icy worlds and collected detailed information about their surfaces and interiors.

Voyager 2 was the first spacecraft to visit all four giant planets and image their satellites. This seminal mission discovered that many icy satellites are much more geologically active than many scientists expected, despite being incredibly cold and small relative to the Earth (Smith *et al.*, 1979; Smith *et al.*, 1982; Smith, 1986; Smith *et al.*, 1989). Satellites orbiting closer to their parent planet were generally more active than satellites on more distant orbits, very likely due to energy dissipated during tidal deformation (Peale *et al.*, 1979). Like Earth's moon all large icy satellites are tidally locked, showing the same side to their parent planet throughout their orbit (Burns *et al.*, 1986). Satellites on eccentric orbits experience a time varying tidal potential that causes their surfaces to deform, leading to energy dissipation in their interiors. Tidal dissipation can be a major source of energy on icy bodies and influence the amount of endogenic geologic activity (Cassen *et al.*, 1979).

While icy bodies are dominantly composed of water ice and silicates, smaller amounts of volatiles are likely present. Such volatiles include carbonates, salts, sulfides, ammonia, nitrogen, methane and carbon monoxide (Lewis, 1971; McKinnon *et al.*, 2008). Spectroscopic observations show that the icy satellites of Jupiter and Saturn are mostly dominated by water ice (Calvin *et al.*, 1995), although some regions appear to contain hydrated salts (McCord *et al.*, 1998). Some satellites contain darker materials that likely contain organic particles that have been processed by radiation (McCord *et al.*, 1998). Spectroscopic detections of nitrogen, ammonia, and

carbon monoxide are more common in the Kuiper Belt and on Neptune’s moon Triton (McKinnon *et al.*, 2008; Brown, 2012). More volatiles are likely to condense in colder regions of the solar nebula (Lewis, 1971). As discussed in chapters 4 and 6, the volatile concentration in icy satellite ice shells are important in controlling the melting temperature and the dynamics of subsurface oceans.

In addition to observing their surfaces, NASA missions also gathered clues about the internal structure of icy bodies. *Galileo* discovered evidence for subsurface oceans on Europa, Ganymede and Callisto by detecting induced magnetic fields around these bodies, consistent with the presence of a layer of conductive salty water (Khurana *et al.*, 1998; Kivelson *et al.*, 2002). Gravity and topography measurements by the *Cassini* spacecraft helped constrain the moment of inertia of several Saturnian satellites showing that some satellites are differentiated into a silicate core and an H₂O layer (Iess *et al.*, 2010, 2014).

More massive icy bodies are more likely to be differentiated due to the larger amount of heat released during accretion and a higher volume of silicates that produce radiogenic heating (Consolmagno and Lewis, 1978; Reynolds and Cassen, 1979). Differentiated satellites have H₂O layers on the surface, a silicate core, and possibly an iron core. The largest icy satellite, Ganymede, has an intrinsic magnetic field that may be generated in a molten iron core (Schubert *et al.*, 1996). The H₂O layer near the surface can include an ice Ih layer with a density $\rho_{Ih} = 0.92 \text{ g/cm}^3$, above a subsurface ocean $\rho_w = 1 \text{ g/cm}^3$ and possibly higher pressure phases of ice with $\rho_{hp} > 1.2 \text{ g/cm}^3$ underlying the ocean. The depth of the rock core boundary will determine whether or not high pressure ice phases are present. On Europa, a subsurface ocean is likely in direct contact with the silicate core, where as on Ganymede the ocean is likely sandwiched between Ice Ih at the top and high pressure ice phases below. The thermal evolution will strongly control the internal structure. Modeling the thermal evolution of icy satellites can help us understand how the internal structures of icy bodies may have evolved over time.

Energy sources for icy bodies include the gravitational energy released due to accretion and differentiation, the decay of long-lived and short-lived radioactive isotopes and dissipation of energy due to tidal deformation (Schubert *et al.*, 1981; Ellsworth and Schubert, 1983). On icy satellites in orbital resonances, such as Europa, tidal heating can become the dominant source of thermal energy (Ojakangas and Stevenson, 1989b), whereas on satellites without significant tidal deformation, such as Pluto, radiogenic heating controls the long-term thermal evolution (Robuchon and Nimmo, 2011). How heat is transmitted in the interior can make a dramatic difference in the long term thermal evolution of icy bodies. For instance, if the ice and silicate layers on Callisto can transport heat by thermal convection, Callisto could get rid of its accretional heat very efficiently and help prevent it from differentiating (Thurber *et al.*, 1980).

Satellites with strong tidal heating are often more geologically active (c.f. Greenberg *et al.* 1999; Nimmo *et al.* 2014), however satellites without significant tidal heating, such as Pluto, can also be geologically active (e.g. (McKinnon *et al.*, 2016)). Many satellites pass through orbital resonances with neighboring satellites that cause their orbital eccentricity to temporarily increase (Peale, 1976; Peale *et al.*, 1980). Both Ganymede and Miranda are suspected to have passed through resonances that caused them to experience enhanced rates of tidal heating and geologic activity (Showman *et al.*, 1997; Dermott *et al.*, 1988).

Constraints on a satellite’s thermal history can be obtained by analyzing the topography and understanding the influence of the subsurface thermal structure. High surface heat fluxes can cause impact craters to viscously relax and become shallow (Parmentier and Head, 1981; Dombard and McKinnon, 2000; White *et al.*, 2013). Measurements of current crater depth and estimates of initial crater depth (based on depth-diameter scaling laws (Schenk, 1991b)) are used to calculate crater relaxation. Numerical models are then used to estimate how relaxation rates are related to subsurface viscosity. By understanding how ice viscosity and temperature relate,

estimates of the past thermal gradient can be made. Estimates for the thickness of the elastic lithosphere can also be made by measuring the flexural deformation associated with topographic loads (Nimmo and Pappalardo, 2004; Hurford *et al.*, 2005; Giese *et al.*, 2007). These approaches provide clues about the thermal gradient in the ice shell over time and help constrain the thermal environment that geologic features may have formed in.

1.3 Endogenic Processes on Icy Bodies

The diversity in thermal states, compositions and geologic processes operating on icy bodies has generated a wide variety of geologic terrains (see Collins *et al.* 2009 for a recent review). Grooved terrain on Ganymede covers over half the satellite and is composed of a patchwork of smooth, lightly cratered grooves that contain subparallel-ridges and troughs (see figure 1.1). Coronae on Miranda, perhaps the most bizarre looking features on icy satellites, are also composed of sub-parallel ridges and troughs (Pappalardo, 1994) (see figure 1.2). This “ridge and trough terrain” has been observed on several icy satellites and has been described as one of the most common tectonic features on icy satellites (Pappalardo and Greeley, 1995). It is possible, but also uncertain, that ridge and trough terrain may have formed by a common mechanism on icy satellites (Barr and Hammond, 2015). Some geologic features are unique in the solar system and are only found on one body, such as chaos terrains and cycloids on Europa (Greenberg *et al.*, 1999; Hoppa *et al.*, 1999) (see figure 1.5). Cycloids are chains of arcuate lineaments that likely form as fractures in response to tidal stresses. Each arch forms during one phase of the tidal cycle and the fracture curves as it follows the pattern of the tidal stresses (Hoppa *et al.*, 1999).

Extensional tectonic features, such as scarps, graben and tilt block normal faults, have been interpreted on many icy satellites (Nyffenegger *et al.*, 1997; Pappalardo *et al.*, 1997) (see figure 1.3). Voyager 2 images suggested that lightly cratered smooth plains covered several satellites such as Ganymede, Enceladus and Miranda. Some

authors suggested these smooth plains formed by the emplacement of “cryolavas” (Schenk, 1991a; Schenk *et al.*, 2001), where some kind of water-rich liquid or slurry erupted from below and flowed across the surface (Kargel, 1991, 1995) (see figure 1.4). Subsequent missions imaged these terrains in higher resolution and revealed that smooth plains on some satellites were full of scarps, ridges and troughs associated with tectonic deformation (Head *et al.*, 1997; Collins *et al.*, 1998).

Near surface liquids are likely present on several icy satellites (Head and Pappalardo, 1999; Porco *et al.*, 2006), suggesting that melt was either generated at or transported to the near surface (e.g Nimmo *et al.* 2007). The most dramatic evidence for a type of cryovolcanism is the plumes of water vapor erupting from the South Pole of Enceladus (Porco *et al.*, 2006). The plumes may be sourced directly from a subsurface ocean 10 – 30 km below the surface (Postberg *et al.*, 2009; Porco *et al.*, 2014). Geologic evidence for effusive cryovolcanism is less pronounced on icy satellite surfaces, though possible flow features have been identified (Schenk, 1991a; Schenk *et al.*, 2001). Cryovolcanism seems geodynamically difficult, due to the negative buoyancy of liquid water with respect to ice. However, the debate in the planetary community continues about whether resurfacing on some icy bodies was driven more by cryovolcanism or tectonic deformation (c.f. Kargel and Pozio 1996; Figueredo and Greeley 2004; Barr 2008; Neveu and Desch 2015).

Many different driving mechanisms have been suggested for causing tectonic deformation on different icy satellites. Here we focus on stresses generated from diurnal tidal deformation and solid-state convection in the ice shell. However, other important sources of stress may include: 1) global volume changes caused by phase changes in the interior (Squyres, 1980), such as the freezing of a subsurface ocean as discussed in chapter 4, 2) non-synchronous rotation of the ice shell with respect to the core (Wahr *et al.*, 2009), 3) polar wander leading to a readjustment of the tidal bulge (Ojakangas and Stevenson, 1989a), 4) despinning of a satellite leading to a decrease in the equatorial radius (Melosh, 1977).

1.3.1 Tidal Deformation

Tides can generate large stresses at the near surface of icy satellites that may be important in influencing their tectonic evolution. Estimates for the magnitude of the tidal stress on various satellites can be predicted based on Newtonian mechanics as a function of their orbital eccentricity e , semi-major axis a , and tidal response (Murray and Dermott, 1999). The differential gravitational force across the satellite’s surface causes a permanent tidal bulge to develop along the axis aligned with the planet (Takeuchi, 1950). The height of the tidal bulge can be approximated as

$$H = h_2 R_s \left(\frac{M_p}{M_s} \right) \left(\frac{R_s}{a_s} \right)^3 \quad (1.1)$$

where R is radius, M is mass and the subscripts s and P represent the satellite and parent planet, respectively.

The rigidity of the satellite strongly controls how readily it deforms in response to tidal forces. The love number h_2 describes the degree-two radial response of the satellite to tidal forces. For a perfectly deformable body, $h_2 = 2.5$. The tidal love number for a homogeneous elastic spherical body is

$$h_2 = \frac{\frac{5}{2}}{1 + \frac{19\mu}{2\bar{\rho}gR}} \quad (1.2)$$

where μ is the shear modulus of the satellite and $\bar{\rho}$ is the bulk density of the satellite, g is gravity and R is radius. The rigidity can be substantially lower if the satellite deforms viscously in response to the tidal stress. For a satellite with a subsurface ocean, the ice shell becomes decoupled from the interior and deforms more in response to tides. Moore and Schubert (2003) showed that satellites with subsurface oceans have radial love numbers $h_2 \approx 1.5$.

As the satellite orbits, the height of the tidal bulge changes and the axis of the tidal bulge also wobbles back and forth (libration). Tidal stresses vary between com-

pressional, shear and tensional through out one orbital cycle and vary in orientation and magnitude as a function of latitude and longitude (Helfenstein and Parmentier, 1983; Wahr *et al.*, 2009). An estimate for the average magnitude of the stresses is $\sigma_t \approx 3EeH/R_s$, where E is the Young’s modulus of ice (Collins *et al.*, 2009).

Figure 1.6 shows the tidal stress magnitude for a handful of icy satellites. For satellites with evidence of subsurface oceans, we assume $h_2 = 1.5$ (Husmann *et al.*, 2006). The magnitude of tidal stresses on icy satellites ranges from 50 kPa for bodies such as Europa and Enceladus, (Rhoden *et al.*, 2010; Nimmo *et al.*, 2014), to ~ 3 kPa for cold rigid bodies such as Mimas. Satellites on more distant orbits, such as Callisto and Iapetus, have tidal stresses close to zero. Tidal stresses of 10 – 100 kPa appear capable of fracturing the surface of icy bodies. Europa and Enceladus are likely fracturing in response to tides at the present day (Nimmo *et al.*, 2014; Roth *et al.*, 2014), and many satellites display degree-two patterns of tectonic deformation, including Dione and Ganymede, suggesting tidal stresses in the past may have been important on these satellites (Bianchi *et al.*, 1986). The subject of chapter 5 is how tidal forces of this magnitude could fracture icy satellite surfaces despite the tensile yield stress of ice measured in the laboratory being about 1 MPa (Schulson, 2001).

1.3.2 Solid-State Convection

There is abundant evidence that the ice shells of some icy satellites may be convecting. Europa’s surface is pervasively covered with features indicative of convection (Pappalardo *et al.*, 1998). Chaos terrains on Europa may form when warm convective plumes cause partial melting of the lithosphere, leading to the disruption of the surface (Head and Pappalardo, 1999; Collins *et al.*, 2000; Sotin *et al.*, 2002). Pits and domes on Europa may form as the result of compositionally and thermally buoyant diapirs that cause the surface to warp (Pappalardo and Barr, 2004). Bands, also a very common feature on Europa, may be sites of lithospheric spreading where fresh ice is brought up to the surface. Thus bands may form by a similar mechanism to

mid-ocean ridges on Earth (Prockter *et al.*, 2002). Additionally, some authors have suggested that convection may be driving subduction on Europa (Kattenhorn and Prockter, 2014), explaining why the surface of Europa has very few impact craters and an apparent surface age of ~ 50 Myr (Zahnle *et al.*, 2003).

Enceladus also shows strong evidence for ice shell convection (Barr, 2008). There is a large thermal anomaly in the south pole that has been measured by the Cassini spacecraft’s near infrared camera and the heat flow in this region, called the South-Polar terrain (SPT), is estimated to be up to 100 mW/m^2 (Howett *et al.*, 2011). There are no large impact craters in the SPT, suggesting a surface age of less than 0.5 Myr (Porco *et al.*, 2006). The SPT is also the site of one of the most spectacular displays in the solar system; geysers of water vapor periodically erupt out of four long cracks on the surface called the tiger stripes. Convection in the ice shell deforming the surface can help explain the incredibly young surface age, the anomalously high heat flow, and the presence of near surface liquids as a source for the geysers (Barr, 2008). As discussed extensively in chapters 2 and 3, Ganymede and Miranda also show intriguing geologic evidence for convection driven resurfacing.

For convection to occur, the non-dimensional Rayleigh number (described in subsequent chapters), must exceed a critical value. The most uncertain parameter in the Rayleigh number is the viscosity of ice. The viscosity of water ice has been extensively studied in laboratory experiments (Weertman, 1983; Durham and Stern, 2001; Goldsby and Kohlstedt, 2001). The viscosity of ice is strongly temperature dependent and deformation can be limited by different creep mechanisms. If the rheology of water ice can be understood in terms of the temperature, stress and grainsize, then the viscosity in icy satellite ice shells can be approximated.

The steady state strain rate for viscously deforming polycrystalline materials is typically described by the flow law

$$\dot{\epsilon} = Ad^{-p}\sigma^n \exp\left(\frac{-Q}{RT}\right) \quad (1.3)$$

where Q is the activation energy for creep, T is temperature, R is the gas constant, σ is the differential stress, n is the stress exponent, p is the grainsize exponent and A is an empirical constant. When $n = 1$ the strain rate is linearly dependent on the stress and the viscosity $\eta = \sigma/\dot{\epsilon}$ is independent of stress (Newtonian).

Goldsby and Kohlstedt (2001) showed that the viscous deformation of polycrystalline water ice can be described by the composite flow law

$$\dot{\epsilon} = \dot{\epsilon}_{diff} + \left(1/\dot{\epsilon}_{basal} + 1/\dot{\epsilon}_{gsb}\right)^{-1} + \dot{\epsilon}_{disl}. \quad (1.4)$$

The strain rate from diffusion creep, $\dot{\epsilon}_{diff}$ is Newtonian and dependent on the grainsize (d^2 for volume diffusion and d^3 for grain boundary diffusion). Grain boundary sliding can either be rate limited by dislocation slip along the basal plane, $\dot{\epsilon}_{basal}$, or the rate of grain boundary sliding $\dot{\epsilon}_{gsb}$. Both of these are slightly non-Newtonian with a weak grainsize dependence. The strain rate from dislocation creep, $\dot{\epsilon}_{disl}$, has an $n = 4$ and is independent of grainsize.

Dislocation creep is expected to dominate at high stresses and large grainsizes. At smaller grainsizes, grainsize dependent mechanism start to dominate. For very small grainsizes, low stresses and warm temperatures, diffusion creep is expected to be the dominant mechanism, however diffusion creep has not been observed directly in the laboratory (Goldsby and Kohlstedt, 2001).

The deformation mechanism likely to dominate in the ice shells of icy satellites depends strongly on the grainsize. Barr and McKinnon (2007) estimated the grainsize in icy satellites to be between $d = 1 - 80$ mm, assuming that grain boundaries were either pinned by silicate impurities or limited by dynamic recrystallization. However, their model for dynamic recrystallization (which gives the upper bound of $d = 80$ mm) assumed that grain growth could occur in the grain boundary sliding regime, which may not be the case (Caswell, personal communication 2017). For a grainsize of $d < 1$ mm, diffusion creep is likely to dominate (Barr and Pappalardo, 2005), and

grain boundary sliding may dominate in most of the ice shell if grainsizes are larger (Barr and Showman, 2009). Barr *et al.* (2004) modeled the convection of icy satellite ice shells for different ice rheologies, and found that convection with non-Newtonian viscosities behaved similar to convection with Newtonian viscosity, except that the critical Rayleigh number for initiating convection was markedly different. For ice shell convection to occur on many icy satellites the grainsize must be on the order of $d = 1$ mm, or the viscosity of the ice shell becomes too high for convection to start.

The temperature-dependent viscosity of ice strongly affects the dynamics of ice shell convection. At the surface, ice is far too viscous to participate in convection, and behaves essentially like solid rock on the surface of the Earth. Solomatov (1995) modeled the convection of fluids that had a range of temperature dependences in viscosity. He found that convection occurred in three different regimes. Strong temperature dependence led to a stagnant-lid regime where convection is confined below an undeforming conductive lid. When the difference between viscosity between the top and bottom of the layer $\Delta\eta = 10^3 - 10^{4.5}$, convection occurred in the sluggish-lid regime, where the conductive lid slowly deformed and participated in convection. When the temperature dependence was small, convection occurred in the iso-viscous regime, where the surface deformed at the same rate as the convective interior.

Only assuming a temperature-dependent viscosity for the ice shell would suggest convection would occur in the stagnant-lid regime. However, if the surface is fractured the stresses from convection could cause the lid to plastically deform. Plastic yielding in the lid leads to surface strain rates that are similar to those in the sluggish-lid regime (Moresi and Solomatov, 1998). The Earth, which developed plate tectonics, would also be predicted to be in the stagnant-lid regime if only assuming a temperature-dependent viscous rheology (Solomatov, 2004), hence accounting for the effect of brittle failure during convection is essential to understanding convection on icy satellites.

Brittle deformation is often incorporated into numerical convection models by

modeling the viscosity of plastically deforming regions as $\eta_{yield} = \tau_{yield}/\dot{\epsilon}$, where τ_{yield} is the yield stress (Moresi and Solomatov, 1998). The yield stress is typically expressed as a function of depth where $\tau_{yield} = \tau_0 + \mu\rho gz$, where τ_0 is the cohesion, μ is the coefficient of friction, g is gravitational acceleration and z is depth (Byerlee, 1968). Simulations of the initiation of mobile-lid convection on the Earth require the yield stress and coefficient of friction to be low, $\mu \approx 0.05$ (Moresi and Solomatov, 1998). While this is much lower than the coefficient of friction measured on some rocks in the lab (Byerlee, 1968), low frictional resistance has been inferred along large fault zones on Earth (Rice, 1992), suggesting that at large scales, low coefficients of friction are more appropriate for modeling the strength of the crust.

Numerical models of convection in icy satellite ice shells have shown that convection in the sluggish-lid regime can explain many observations on satellites such as Europa and Enceladus (Showman and Han, 2005; Barr, 2008; Han and Showman, 2010; O’Neill and Nimmo, 2010; Han *et al.*, 2012). These models assume that the surface of the icy satellite is weak, either through imposing a low plastic yield criterion of 10 – 80 kPa, or limiting the viscosity contrast across the surface. Each assumption leads to similar surface heat flows and strain rates that match observations of icy satellites, but using a plastic yield criterion can also lead to complex time-dependent behavior, such as the episodic subduction of the surface (Moresi and Solomatov, 1998; O’Neill and Nimmo, 2010). Chapters 2 and 3 explore the hypothesis that sluggish-lid convection may have occurred on Ganymede and Miranda. Chapter 5 explores how the surface of icy satellites may have been weakened by tidal stresses, which may have enabled convective stresses to deform the surface.

1.4 Understanding Icy Bodies

Icy bodies show a mesmerizing array of geologic features on their surfaces. Many different tools can be used to understand how such features form, including geologic mapping from satellite imagery, spectroscopic mapping, and even measurements of

satellite gravity and magnetic field. Building geodynamic models for the formation of surface features helps to connect the surface evolution to the internal structure and thermal evolution. Geodynamic models can help test the physical plausibility of proposed formation mechanism by incorporating our understanding of the rheological behavior of water ice within the context of planetary evolution.

This thesis describes three end-member styles of geologic resurfacing that have been proposed for icy bodies: tectonic deformation, cryovolcanism and convection-driven resurfacing. Is one of these processes responsible for most of the resurfacing on icy bodies, such as tectonic deformation? Which icy bodies have each of these processes operated on? How might these processes act in tandem with one another?

In this thesis, we examine each of these processes independently, applied to specific planetary bodies, in order to better understand the physical parameters that control the behavior in our models. However, in reality, it is more likely that many processes occur simultaneously on icy satellites. For example, we model convection-driven resurfacing on icy satellites and mainly explore the effect of varying the Rayleigh number and viscosity contrast. A more complex treatment for the formation of grooved terrain on Ganymede could include simultaneously modeling convection in the ice shell, the stresses from global surface expansion, the effect of localized tidal heating within the convective plumes, the generation and migration of partial melt within the plumes, the accumulation of damage within the lithosphere from tidal stresses and the localization of brittle failure leading to the formation of linear groove lanes. Including so many processes and parameters in a single model can be cumbersome and impractical, which is why simple models can be beneficial when testing the basic physical plausibility of a geologic process.

The effect of tidal deformation and the issue of strength versus scale are tremendously important to the evolution of icy satellites. The strength of icy satellite lithospheres determines what geologic processes are capable of deforming their surfaces. If icy satellites gradually become weaker due to experiencing billions of cycles of

tidal stress, they could become more geologically active later in their histories. A weak lithosphere would allow convection to deform their surfaces and could lead to lithospheric recycling. Models of convection-driven resurfacing applied to the Earth typically treat the mantle and lithosphere as weak. Micro-mechanical processes have been explored to explain the weak behavior of rocks at large scales, including pore-fluid pressurization, dynamic weakening, and grain size evolution. Advances in our understanding of rock mechanics and mantle deformation could be applied to icy satellites and could help us understand how convection and solid-state deformation take place on icy bodies. Our understanding of terrestrial magma oceans can also be applied to icy satellites. Issues such as the formation of the floatation crust, melt migration and the chemical evolution of the “magma” ocean have direct bearing to the formation and evolution of the ice shell and subsurface ocean. As always, there is more work to be done (modeling and experimental) to understand the rich variety of geologic features observed on icy satellites.

1.5 References

- Barr, A. C. 2008. Mobile lid convection beneath Enceladus' south polar terrain. *Journal of Geophysical Research: Planets* 113(E7).
- Barr, A. C., and N. P. Hammond 2015. A common origin for ridge-and-trough terrain on icy satellites by sluggish lid convection. *Physics of the Earth and Planetary Interiors* **249**, 18–27.
- Barr, A. C., and W. B. McKinnon 2007. Convection in ice I shells and mantles with self-consistent grain size. *J. Geophys. Res. Planets* **112**, 2012.
- Barr, A. C., and R. T. Pappalardo 2005. Onset of convection in the icy Galilean satellites: Influence of rheology. *Journal of Geophysical Research: Planets* 110(E12).
- Barr, A. C., R. T. Pappalardo, and S. Zhong 2004. Convective instability in ice I with non-Newtonian rheology: Application to the icy Galilean satellites. *Journal of Geophysical Research: Planets* 109(E12).
- Barr, A. C., and A. P. Showman 2009. Heat transfer in Europa's icy shell. *Europa*, edited by RT Pappalardo, WB McKinnon, and KK Khurana, University of Arizona Press, Tucson, 405–430.
- Bianchi, R., R. Casacchia, P. Lanciano, S. Pozio, and R. G. Strom 1986. Tectonic framework of grooved terrain on Ganymede. *Icarus* 67(2), 237–250.
- Brown, M. E. 2012. The compositions of Kuiper belt objects. *Annual Review of Earth and Planetary Sciences* **40**, 467–494.
- Burns, J. A., M. S. Matthews, et al. 1986. *Satellites*. Number 77. University of Arizona Press.
- Byerlee, J. D. 1968. Brittle-ductile transition in rocks. *Journal of Geophysical Research* 73(14), 4741–4750.
- Calvin, W. M., R. N. Clark, R. H. Brown, and J. R. Spencer 1995. Spectra of the icy Galilean satellites from 0.2 to 5 μ m: A compilation, new observations, and a recent summary. *Journal of Geophysical Research: Planets* 100(E9), 19041–19048.

- Cassen, P., R. T. Reynolds, and S. Peale 1979. Is there liquid water on Europa? *Geophysical Research Letters* 6(9), 731–734.
- Collins, G., J. Head, R. Pappalardo, and N. Spaun 2000. Evaluation of models for the formation of chaotic terrain on Europa. *Journal of Geophysical Research: Planets* 105(E1), 1709–1716.
- Collins, G. C., J. W. Head III, and R. T. Pappalardo 1998. grooved terrain: Insights from Galileo high-resolution. *Geophysical research letters* 25(3), 233–236.
- Collins, G. C., W. B. McKinnon, J. M. Moore, F. Nimmo, R. T. Pappalardo, L. M. Prockter, and P. M. Schenk 2009. Tectonics of the outer planet satellites. *Planetary Tectonics* 11, 264.
- Consolmagno, G. J., and J. S. Lewis 1978. The evolution of icy satellite interiors and surfaces. *Icarus* 34(2), 280–293.
- Dermott, S. F., R. Malhotra, and C. D. Murray 1988. Dynamics of the Uranian and Saturnian satellite systems: A chaotic route to melting Miranda? *Icarus* 76(2), 295–334.
- Dombard, A. J., and W. B. McKinnon 2000. Long-term retention of impact crater topography on Ganymede. *Geophysical research letters* 27(22), 3663–3666.
- Durham, W., and L. Stern 2001. Rheological Properties of Water Ice-Applications to Satellites of the Outer Planets 1. *Annual Review of Earth and Planetary Sciences* 29(1), 295–330.
- Ellsworth, K., and G. Schubert 1983. Saturn’s icy satellites: Thermal and structural models. *Icarus* 54(3), 490–510.
- Figueredo, P. H., and R. Greeley 2004. Resurfacing history of Europa from pole-to-pole geological mapping. *Icarus* 167(2), 287–312.
- Giese, B., R. Wagner, G. Neukum, P. Helfenstein, and P. C. Thomas 2007. Tethys: Lithospheric thickness and heat flux from flexurally supported topography at Ithaca Chasma. *Geophysical Research Letters* 34(21).
- Goldsby, D. L., and D. L. Kohlstedt 2001. Superplastic deformation of ice: Experi-

- mental observations. *J. Geophys. Res.* **106**, 11017.
- Greenberg, R., G. V. Hoppa, B. R. Tufts, P. Geissler, J. Riley, and S. Kadel 1999. Chaos on Europa. *Icarus* **141**, 263–286.
- Han, L., and A. P. Showman 2010. Coupled convection and tidal dissipation in Europa’s ice shell. *Icarus* **207**, 834–844.
- Han, L., G. Tobie, and A. P. Showman 2012. The impact of a weak south pole on thermal convection in Enceladus’ ice shell. *Icarus* **218**, 320–330.
- Head, J., R. Pappalardo, G. Collins, R. Greeley, G. I. Team, et al. 1997. Tectonic resurfacing on Ganymede and its role in the formation of grooved terrain. In *Lunar and Planetary Science Conference*, Volume 28, pp. 535.
- Head, J. W., and R. T. Pappalardo 1999. Brine mobilization during lithospheric heating on Europa: Implications for formation of chaos terrain, lenticula texture, and color variations. *Journal of Geophysical Research: Planets* **104**(E11), 27143–27155.
- Helfenstein, P., and E. Parmentier 1983. Patterns of fracture and tidal stresses on Europa. *Icarus* **53**(3), 415–430.
- Hoppa, G. V., B. R. Tufts, R. Greenberg, and P. E. Geissler 1999. Formation of cycloidal features on Europa. *Science* **285**, 1899–1902.
- Howett, C. J. A., J. R. Spencer, J. Pearl, and M. Segura 2011. High heat flow from Enceladus’ south polar region measured using 10-600 cm^{-1} Cassini/CIRS data. *Journal of Geophysical Research (Planets)* **116**, 3003.
- Hurford, T., R. Beyer, B. Schmidt, B. Preblich, A. Sarid, and R. Greenberg 2005. Flexure of Europa’s lithosphere due to ridge-loading. *Icarus* **177**(2), 380–396.
- Hussmann, H., F. Sohl, and T. Spohn 2006. Subsurface oceans and deep interiors of medium-sized outer planet satellites and large trans-neptunian objects. *Icarus* **185**(1), 258–273.
- Iess, L., N. J. Rappaport, R. A. Jacobson, P. Racioppa, D. J. Stevenson, P. Tortora, J. W. Armstrong, and S. W. Asmar 2010. Gravity field, shape, and moment of

- inertia of Titan. *Science* 327(5971), 1367–1369.
- Iess, L., D. Stevenson, M. Parisi, D. Hemingway, R. Jacobson, J. Lunine, F. Nimmo, J. Armstrong, S. Asmar, M. Ducci, et al. 2014. The gravity field and interior structure of Enceladus. *Science* 344(6179), 78–80.
- Kargel, J. 1995. Cryovolcanism on the icy satellites. In *Comparative Planetology with an Earth Perspective*, pp. 101–113. Springer.
- Kargel, J. S. 1991. Brine volcanism and the interior structures of asteroids and icy satellites. *Icarus* 94(2), 368–390.
- Kargel, J. S., and S. Pozio 1996. The volcanic and tectonic history of Enceladus. *Icarus* 119(2), 385–404.
- Kattenhorn, S. A., and L. M. Prockter 2014. Evidence for subduction in the ice shell of Europa. *Nature Geoscience* 7(10), 762–767.
- Khurana, K., M. Kivelson, D. Stevenson, G. Schubert, C. Russell, R. Walker, and C. Polanskey 1998. Induced magnetic fields as evidence for subsurface oceans in Europa and Callisto. *Nature* 395(6704), 777–780.
- Kivelson, M., K. Khurana, and M. Volwerk 2002. The permanent and inductive magnetic moments of Ganymede. *Icarus* 157(2), 507–522.
- Lewis, J. S. 1971. Satellites of the outer planets: Their physical and chemical nature. *Icarus* 15(2), 174–185.
- McCord, T., G. Hansen, F. Fanale, R. Carlson, D. Matson, T. Johnson, W. Smythe, J. Crowley, P. Martin, A. Ocampo, et al. 1998. Salts on Europa’s surface detected by Galileo’s near infrared mapping spectrometer. *Science* 280(5367), 1242–1245.
- McCord, T. a., G. Hansen, R. Clark, P. Martin, C. Hibbitts, F. Fanale, J. Granahan, M. Segura, D. Matson, T. Johnson, et al. 1998. Non-water-ice constituents in the surface material of the icy Galilean satellites from the Galileo near-infrared mapping spectrometer investigation. *Journal of Geophysical Research: Planets* 103(E4), 8603–8626.
- McKinnon, W. B., F. Nimmo, T. Wong, P. M. Schenk, O. L. White, J. Roberts,

- J. Moore, J. Spencer, A. Howard, O. Umurhan, et al. 2016. Convection in a volatile nitrogen-ice-rich layer drives Pluto’s geological vigour. *Nature* **534**(7605), 82–85.
- McKinnon, W. B., D. Prialnik, S. A. Stern, and A. Coradini 2008. Structure and evolution of Kuiper belt objects and dwarf planets. *The Solar System Beyond Neptune* **1**, 213–241.
- Melosh, H. J. 1977. Global tectonics of a despun planet. *Icarus* **31**(2), 221–243.
- Moore, W. B., and G. Schubert 2003. The tidal response of Ganymede and Callisto with and without liquid water oceans. *Icarus* **166**(1), 223–226.
- Moresi, L., and V. Solomatov 1998. Mantle convection with a brittle lithosphere: thoughts on the global tectonic styles of the Earth and Venus. *Geophysical Journal International* **133**(3), 669–682.
- Murray, C. D., and S. F. Dermott 1999. *Solar system dynamics*. Cambridge university press.
- Neveu, M., and S. J. Desch 2015. Geochemistry, thermal evolution, and cryovolcanism on Ceres with a muddy ice mantle. *Geophysical Research Letters* **42**(23).
- Nimmo, F., and R. T. Pappalardo 2004. Furrow flexure and ancient heat flux on Ganymede. *Geophys. Res. Lett.* **31**, 19701.
- Nimmo, F., C. Porco, and C. Mitchell 2014. Tidally modulated eruptions on Enceladus: Cassini ISS observations and models. *The Astronomical Journal* **148**(3), 46.
- Nimmo, F., J. R. Spencer, R. T. Pappalardo, and M. E. Mullen 2007. Shear heating as the origin of the plumes and heat flux on Enceladus. *Nature* **447**, 289–291.
- Nyffenegger, P., D. M. Davis, and G. J. Consolmagno 1997. Tectonic lineations and frictional faulting on a relatively simple body (Ariel). *Planetary and space science* **45**(9), 1069–1080.
- Ojakangas, G. W., and D. J. Stevenson 1989a. Polar wander of an ice shell on Europa. *Icarus* **81**(2), 242–270.
- Ojakangas, G. W., and D. J. Stevenson 1989b. Thermal state of an ice shell on

- Europa. *Icarus* 81(2), 220–241.
- O’Neill, C., and F. Nimmo 2010. The role of episodic overturn in generating the surface geology and heat flow on Enceladus. *Nature Geoscience* 3, 88–91.
- Pappalardo, R., J. Head, R. Greeley, R. Sullivan, C. Pilcher, G. Schubert, W. Moore, M. Carr, J. Moore, M. Belton, et al. 1998. Geological evidence for solid-state convection in Europa’s ice shell. *Nature* 391(6665), 365–368.
- Pappalardo, R. T. 1994. *The origin and evolution of ridge and trough terrain and the geological history of Miranda*. Ph. D. thesis, Arizona State University Tempe.
- Pappalardo, R. T., and A. C. Barr 2004. The origin of domes on Europa: The role of thermally induced compositional diapirism. *Geophysical Research Letters* 31(1).
- Pappalardo, R. T., and R. Greeley 1995. A review of the origins of subparallel ridges and troughs: Generalized morphological predictions from terrestrial models. *Journal of Geophysical Research: Planets* 100(E9), 18985–19007.
- Pappalardo, R. T., S. J. Reynolds, and R. Greeley 1997. Extensional tilt blocks on Miranda: Evidence for an upwelling origin of Arden Corona. *Journal of Geophysical Research: Planets* 102(E6), 13369–13379.
- Parmentier, E. M., and J. Head 1981. Viscous relaxation of impact craters on icy planetary surfaces: Determination of viscosity variation with depth. *Icarus* 47(1), 100–111.
- Peale, S. 1976. Orbital resonances in the solar system. *Annual review of astronomy and astrophysics* 14(1), 215–246.
- Peale, S., P. Cassen, and R. Reynolds 1979. Melting of Io by tidal dissipation. *Science* 203, 892–894.
- Peale, S., P. Cassen, and R. Reynolds 1980. Tidal dissipation, orbital evolution, and the nature of Saturn’s inner satellites. *Icarus* 43(1), 65–72.
- Porco, C., D. DiNino, and F. Nimmo 2014. How the geysers, tidal stresses, and thermal emission across the south polar terrain of Enceladus are related. *The Astronomical Journal* 148(3), 45.

- Porco, C. C., P. Helfenstein, P. C. Thomas, A. P. Ingersoll, J. Wisdom, R. West, G. Neukum, T. Denk, R. Wagner, T. Roatsch, S. Kieffer, E. Turtle, A. McEwen, T. V. Johnson, J. Rathbun, J. Veverka, D. Wilson, J. Perry, J. Spitale, A. Brahic, J. A. Burns, A. D. Del Genio, L. Dones, C. D. Murray, and S. Squyres 2006. Cassini Observes the Active South Pole of Enceladus. *Science* **311**, 1393–1401.
- Postberg, F., S. Kempf, J. Schmidt, N. Brilliantov, A. Beinsen, B. Abel, U. Buck, and R. Srama 2009. Sodium salts in E-ring ice grains from an ocean below the surface of Enceladus. *Nature* *459*(7250), 1098–1101.
- Prockter, L. M., J. W. Head, R. T. Pappalardo, R. J. Sullivan, A. E. Clifton, B. Giese, R. Wagner, and G. Neukum 2002. Morphology of European bands at high resolution: A mid-ocean ridge-type rift mechanism. *J. Geophys. Res.* **107**.
- Reynolds, R. T., and P. M. Cassen 1979. On the internal structure of the major satellites of the outer planets. *Geophysical Research Letters* *6*(2), 121–124.
- Rhoden, A. R., B. Militzer, E. M. Huff, T. A. Hurford, M. Manga, and M. A. Richards 2010. Constraints on Europa’s rotational dynamics from modeling of tidally-driven fractures. *Icarus* *210*(2), 770–784.
- Rice, J. R. 1992. Fault stress states, pore pressure distributions, and the weakness of the San Andreas fault. *International Geophysics* **51**, 475–503.
- Robuchon, G., and F. Nimmo 2011. Thermal evolution of Pluto and implications for surface tectonics and a subsurface ocean. *Icarus* *216*(2), 426–439.
- Roth, L., J. Saur, K. D. Retherford, D. F. Strobel, P. D. Feldman, M. A. McGrath, and F. Nimmo 2014. Transient water vapor at Europa’s south pole. *Science* *343*(6167), 171–174.
- Schenk, P. M. 1991a. Fluid volcanism on Miranda and Ariel: Flow morphology and composition. *Journal of Geophysical Research: Solid Earth* *96*(B2), 1887–1906.
- Schenk, P. M. 1991b. Ganymede and Callisto: Complex crater formation and planetary crusts. *Journal of Geophysical Research: Planets* *96*(E1), 15635–15664.
- Schenk, P. M., W. B. McKinnon, D. Gwynn, and J. M. Moore 2001. Flooding of

- Ganymede's bright terrains by low-viscosity water-ice lavas. *Nature* **410**(6824), 57–60.
- Schubert, G., D. Stevenson, and K. Ellsworth 1981. Internal structures of the Galilean satellites. *Icarus* **47**(1), 46–59.
- Schubert, G., K. Zhang, M. G. Kivelson, and J. D. Anderson 1996. The magnetic field and internal structure of Ganymede. *Nature* **384**(6609), 544.
- Schulson, E. M. 2001. Brittle failure of ice. *Engineering Fracture Mechanics* **68**(17), 1839–1887.
- Showman, A. P., and L. Han 2005. Effects of plasticity on convection in an ice shell: Implications for Europa. *Icarus* **177**, 425–437.
- Showman, A. P., D. J. Stevenson, and R. Malhotra 1997. Coupled Orbital and Thermal Evolution of Ganymede. *Icarus* **129**, 367–383.
- Smith, B. A. 1986. Voyager 2 in the Uranian system: imaging science results. *Science* **233**, 43–65.
- Smith, B. A., L. Soderblom, D. Banfield, C. Barnet, A. Basilevsky, et al. 1989. Voyager 2 at Neptune: Imaging science results. *Science* **246**(4936), 1422.
- Smith, B. A., L. Soderblom, R. Batson, P. Bridges, J. Inge, H. Masursky, E. Shoemaker, R. Beebe, J. Boyce, G. Briggs, et al. 1982. A new look at the Saturn system: The Voyager 2 images. *Science* **215**(4532), 504–537.
- Smith, B. A., L. A. Soderblom, T. V. Johnson, A. P. Ingersoll, S. A. Collins, E. M. Shoemaker, G. E. Hunt, H. Masursky, M. H. Carr, M. E. Davies, A. F. Cook, J. M. Boyce, T. Owen, G. E. Danielson, C. Sagan, R. F. Beebe, J. Veverka, J. F. McCauley, R. G. Strom, D. Morrison, G. A. Briggs, and V. E. Suomi 1979. The Jupiter system through the eyes of Voyager 1. *Science* **204**, 951–957.
- Solomatov, V. 2004. Initiation of subduction by small-scale convection. *Journal of Geophysical Research: Solid Earth* **109**(B1).
- Solomatov, V. S. 1995. Scaling of temperature- and stress-dependent viscosity convection. *Physics of Fluids* **7**, 266–274.

- Sotin, C., J. W. Head, and G. Tobie 2002. Europa: Tidal heating of upwelling thermal plumes and the origin of lenticulae and chaos melting. *Geophysical Research Letters* 29(8).
- Squyres, S. W. 1980. Volume changes in Ganymede and Callisto and the origin of grooved terrain. *Geophys. Res. Lett.* 7, 593–596.
- Takeuchi, H. 1950. On the earth tide of the compressible earth of variable density and elasticity. *Eos, Transactions American Geophysical Union* 31(5), 651–689.
- Thurber, C. H., A. T. Hsui, and M. N. Toksoz 1980. Thermal evolution of Ganymede and Callisto-Effects of solid-state convection and constraints from Voyager imagery. In *Lunar and Planetary Science Conference Proceedings*, Volume 11, pp. 1957–1977.
- Wahr, J., Z. A. Selvens, M. E. Mullen, A. C. Barr, G. C. Collins, M. M. Selvens, and R. T. Pappalardo 2009. Modeling stresses on satellites due to nonsynchronous rotation and orbital eccentricity using gravitational potential theory. *Icarus* 200(1), 188–206.
- Weertman, J. 1983. Creep deformation of ice. *Annual Review of Earth and Planetary Sciences* 11(1), 215–240.
- White, O. L., P. M. Schenk, and A. J. Dombard 2013. Impact basin relaxation on Rhea and Iapetus and relation to past heat flow. *Icarus* 223(2), 699–709.
- Zahnle, K., P. Schenk, H. Levison, and L. Dones 2003. Cratering rates in the outer Solar System. *Icarus* 163(2), 263–289.

1.6 Figures and Tables

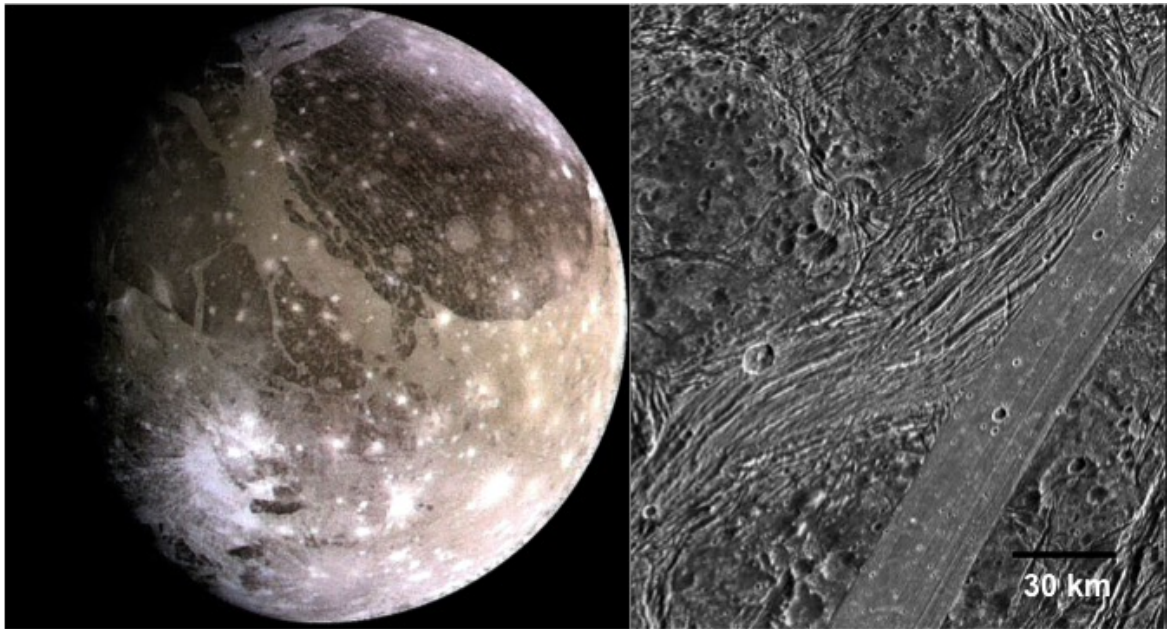


Figure 1.1: (left) Global mosaic of Jupiter's moon Ganymede from images captured by the Galileo spacecraft. (right) Example image of grooved terrain on Ganymede, taken by the Galileo spacecraft. Image credit NASA/JPL photo journal.

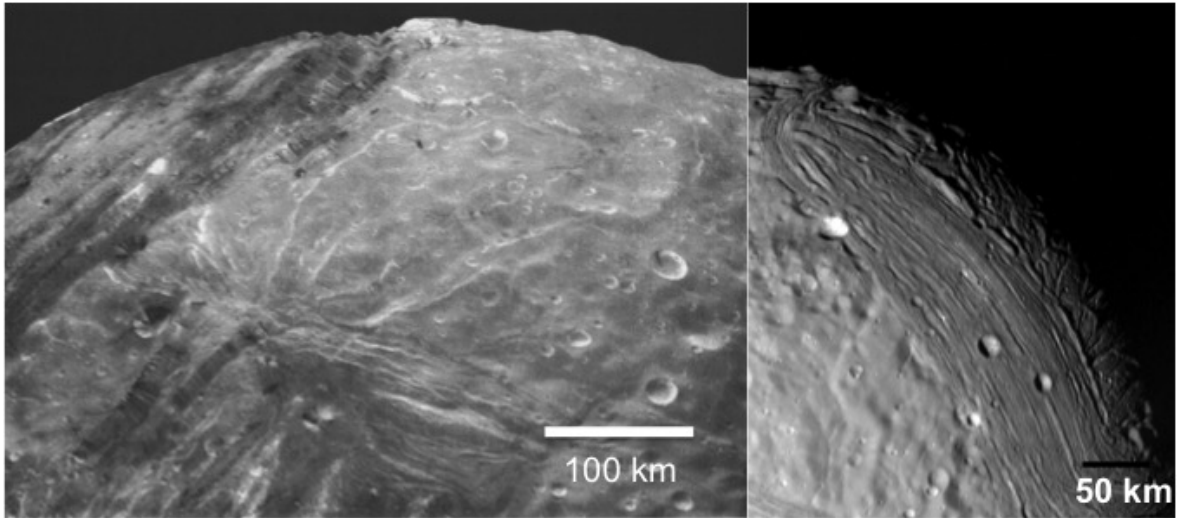


Figure 1.2: (left) Voyager 2 image of Arden coronae on Miranda (right) Voyager 2 image of Elsinore coronae on Miranda. Image credit NASA/JPL.

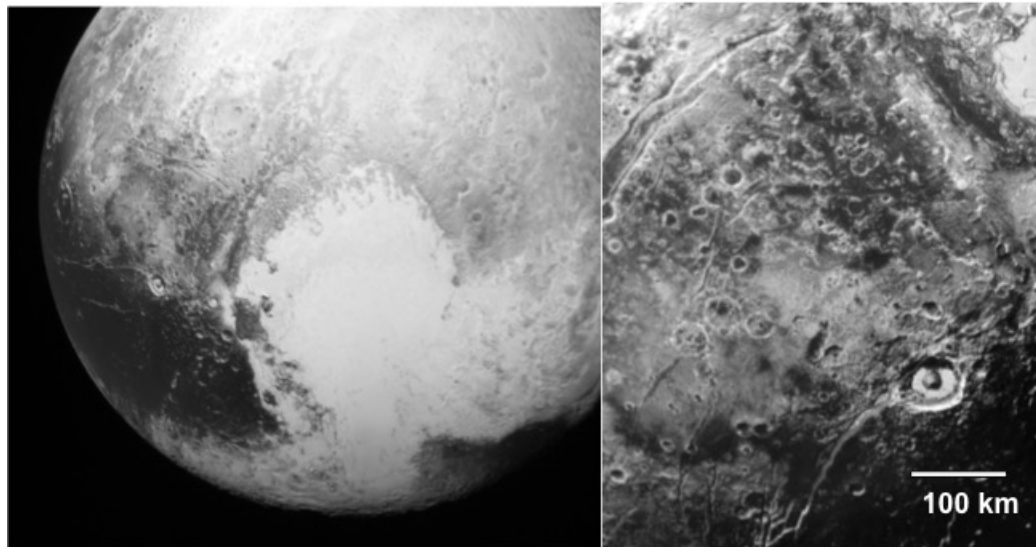


Figure 1.3: (left) Original New Horizons LORRI image of Pluto upon closest approach. Image credit NASA/JPL, image number LORRI_0299148119. (right) Original new horizons image of Pluto focusing on potential extensional tectonic features. Image credit NASA/JPL, image number LORRI_0299168136.

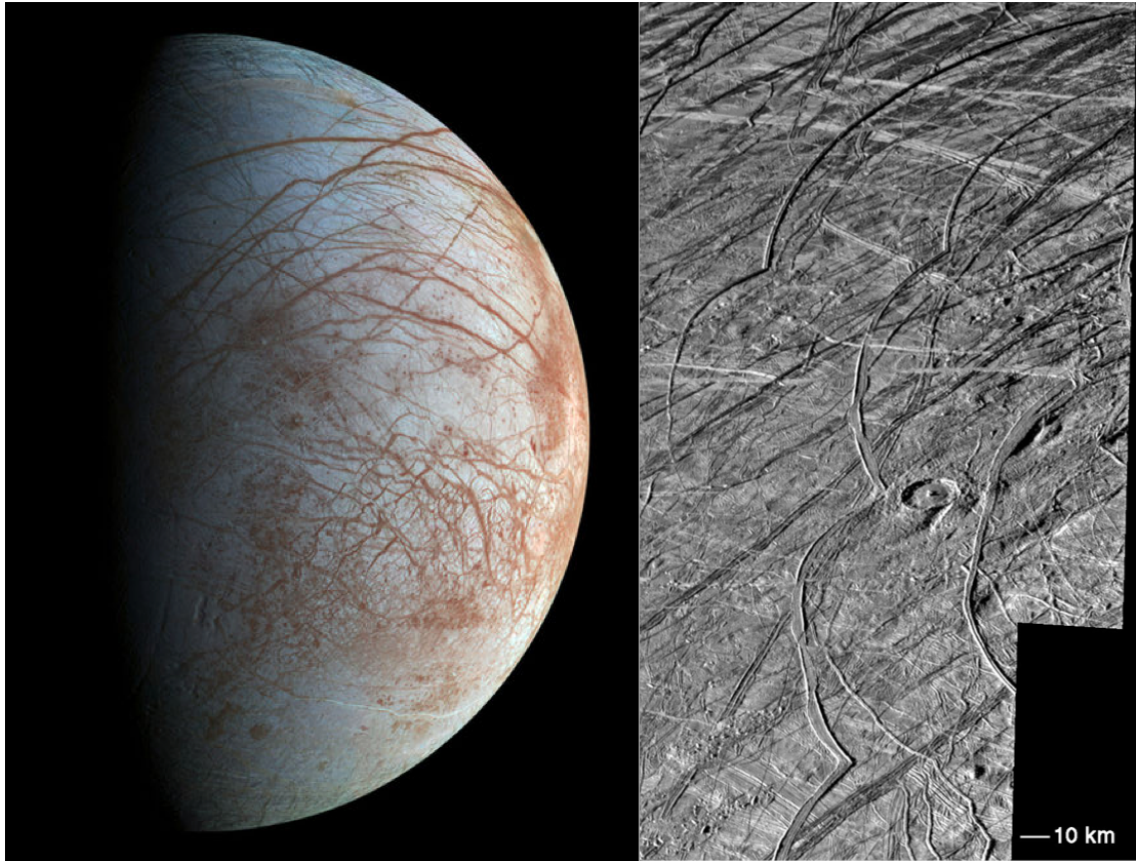


Figure 1.4: (left) Global mosaic of Jupiter's moon Europa made from images captured by the Galileo spacecraft, color enhanced. (right) Close up cycloids on Europa, which may form in response to tidal stress.

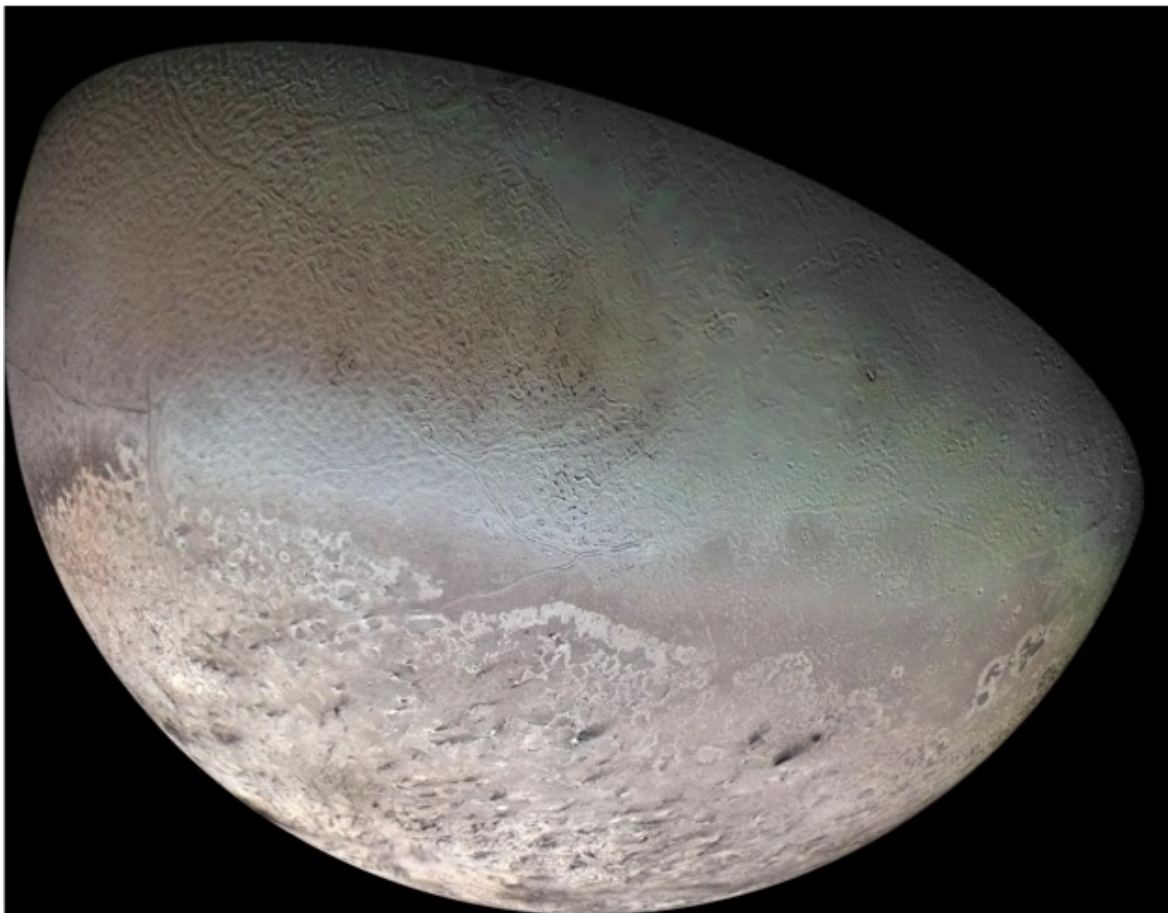


Figure 1.5: (left) Global mosaic of Neptune's moon Triton as captured by NASA's Voyager 2 spacecraft. Image credit NASA/JPL photo journal.

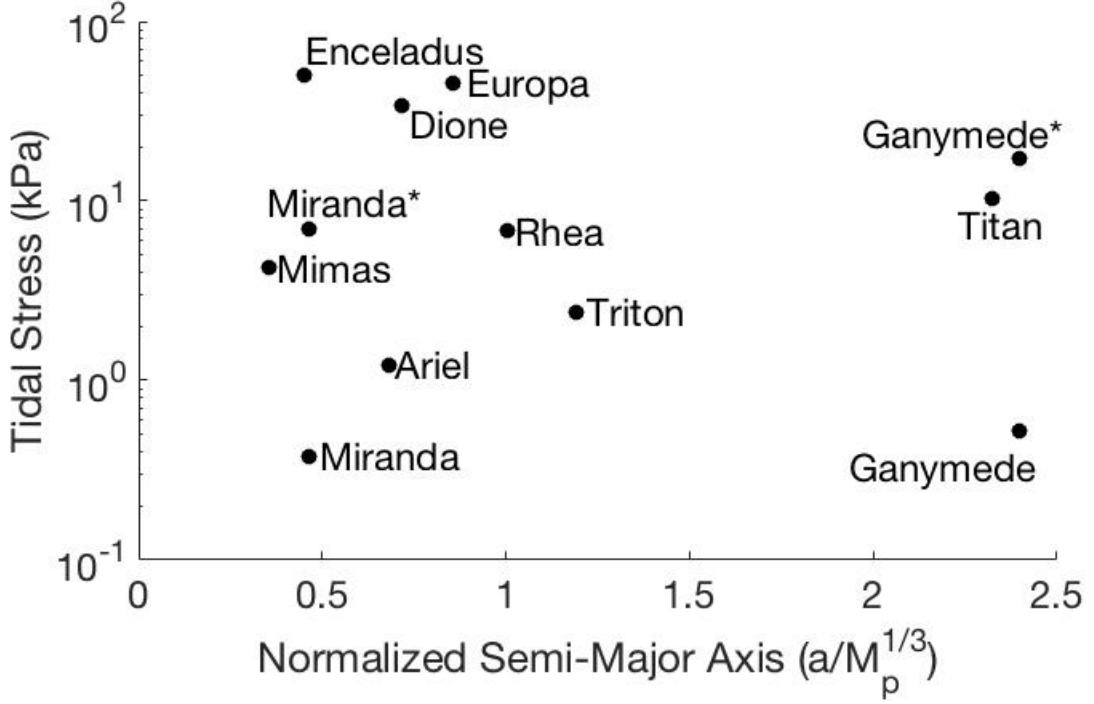


Figure 1.6: The estimated diurnal tidal stress magnitude for various icy satellites (y-axis) versus the normalized orbital distance (x-axis). The normalized distances is the satellites semi-major axis divided by the cube root of the mass of the parent planet. Tidal stress was calculated from equation 1.1 and 1.2 with physical parameters for the satellites and parent planets as described in Collins *et al.* (2009). Satellites we assumed to have a subsurface ocean include Enceladus, Dione, Europa, Rhea, Triton, Ganymede and Titan. The current orbital eccentricity is used to calculate the tidal stress, and for Ganymede and Miranda, an eccentricity of $e = 0.05$ is also used to estimate tidal stresses during past orbital resonances.

Chapter 2 | Formation of Ganymede's Grooved Terrain by Convection-Driven Resurfacing

N.P. Hammond¹ and A.C. Barr².

1. Department of Earth, Environmental and Planetary Science, Brown University,
Providence, RI 02912, USA

2. Planetary Science Institute, 1700 East Fort Lowell, Suite 106, Tucson, AZ
85719-2395, USA

Published in its present form in:

Icarus 227 (2014): 206-209

2.1 Abstract

Grooved terrain on Ganymede is a large geologic province thought to form in an extensional tectonic setting during a time of elevated surface heat flows. Here we use numerical convection models to test the hypothesis that grooved terrain formed in a convecting ice shell with a weak lithosphere. We show that surface heat fluxes and strain rates arising in a convecting ice shell 10 – 100 km thick are consistent with conditions inferred for grooved terrain formation. We suggest that lithospheric stresses and heat flows generated by convection played an important role in the formation of grooved terrain. Smooth linear grooves may have formed primarily by convection-driven lithospheric spreading above convective upwellings. Compressional features, similar to long wavelength folds observed on Europa, would likely form atop convective downwellings, and may be detectable with mapping from ESA’s upcoming Jupiter-Icy Moon Explorer.

2.2 Introduction

Over half the surface of Ganymede, Jupiter’s largest satellite, is covered in grooved terrain, composed of 10 – 100 km wide linear and polygonal swaths (Collins *et al.*, 1998) of sub-parallel ridges and troughs (Murchie *et al.*, 1986; Pappalardo *et al.*, 1998). The ridge and trough spacing is ~ 7 km, though they are often subdivided by narrower grooves 100 m – 1 km wide (Pappalardo *et al.*, 1998). A subset of groove lanes, which we refer to as “subdued grooves”, have straight margins and relatively constant widths over large distances (Head *et al.*, 2002) and are characterized by subdued light material (Patterson *et al.*, 2010).

Grooved terrain is thought to have formed during an era of global surface expansion (Smith *et al.*, 1979) resulting from satellite differentiation (Squyres, 1980) or from the melting of the ice I shell when Ganymede entered a possible Laplace-like resonance with Europa and Io (Showman *et al.*, 1997; Bland *et al.*, 2009). Subparallel grooves may form by tilt-block style normal faulting (Pappalardo *et al.*, 1998) due to tensional stresses in the lithosphere.

Convection in the ice shell has been suggested as a driving mechanism for grooved terrain formation (Lucchitta, 1980; Parmentier *et al.*, 1982), because it can operate globally and generate zones of intense local deformation (Shoemaker *et al.*, 1982). Head *et al.* (2002) argued that subdued grooves form in a similar way to extensional bands on Europa, which likely form by a convection-driven midocean rift-type mechanism (Prockter *et al.*, 2002).

Previous work argues that convective stresses were not strong enough to drive surface deformation on Ganymede (Squyres and Croft, 1986); convective plumes were thought to be confined below a stagnant-lid, a highly viscous layer of ice which inhibits resurfacing (Solomatov, 1995). However, if the near-surface has a yield stress comparable to the thermal buoyancy stresses from convection, plumes can approach the sur-

face, leading to deformation and efficient heat transport (Trompert and Hansen, 1998; Tackley, 2000; Solomatov, 2004; Showman and Han, 2005). This style of “sluggish-lid” convection may be occurring beneath the active South Polar Terrain (SPT) of Enceladus (Barr, 2008). The observed heat flow and surface age of the Enceladus SPT are consistent with heat flows and deformation rates associated with sluggish-lid convection (Barr, 2008; Han *et al.*, 2012).

On Ganymede, surface conditions may also have been consistent with sluggish-lid convection. Models of flexural uplift estimate a heat flux of $100 - 200 \text{ mW m}^{-2}$ for grooved terrain (Nimmo *et al.*, 2002), and a heat flux of $60 - 80 \text{ mW m}^{-2}$ for the adjacent dark terrain (Nimmo and Pappalardo, 2004). While heat flow estimates based on flexure are somewhat uncertain, independent estimates based on models of groove terrain formation by extensional necking predict a similar heat flux of $\sim 50 \text{ mW m}^{-2}$ (Bland *et al.*, 2009). Extensional necking models also predict strain rates between 10^{-16} and 10^{-13} s^{-1} for high-relief grooved material (Dombard and McKinnon, 2001; Bland and Showman, 2007), close to strain rates inferred from folds observed in between the tiger stripes of the SPT (Barr and Preuss, 2010).

Here, we simulate solid-state convection in an ice shell with a weak upper surface to show that the heat flow and deformation rates arising from sluggish-lid convection are consistent with the conditions inferred for grooved terrain formation. Subdued grooves may be areas where convection drove complete lithospheric separation, whereas other grooves may form in response to subsurface flow above a convective upwelling (Lucchitta, 1980).

2.3 Methods

We use the two-dimensional Cartesian finite element model CITCOM (Moresi and Solomatov, 1995) to simulate solid state convection in Ganymede’s ice shell. We explore a wide range of ice shell conditions by varying the basal Rayleigh number,

Ra_1 , which governs the vigor of convection and is related to D , the ice shell thickness,

$$Ra_1 = \frac{\rho g \alpha \Delta T D^3}{\kappa \eta_1}, \quad (2.1)$$

with ice density $\rho = 1000 \text{ kg m}^{-3}$, surface gravity $g = 1.4 \text{ m s}^{-2}$, coefficient of thermal expansion $\alpha = 10^{-4} \text{ K}^{-1}$, basal viscosity η_1 , thermal diffusivity $\kappa = 10^{-6} \text{ m}^2 \text{ s}^{-1}$, and $\Delta T = 150 \text{ K}$ is the difference between the temperature at the base (260 K) and the surface (110 K) of the ice shell. With these parameters, $Ra_1 = 2.1 \times 10^8 (D/100 \text{ km})^3 (10^{14} \text{ Pa s}/\eta_1)$.

We use a simple temperature dependent viscosity (Solomatov and Moresi, 2000), $\eta(T) = \eta_0 \exp(-\gamma T)$, where $\gamma = \theta/\Delta T$, $\theta = \ln(\Delta\eta)$, and $\Delta\eta = \eta_0/\eta_1$ is the viscosity contrast between ice at the surface (η_0), and ice at the base of the shell. On Ganymede, the $\Delta\eta$ predicted for an ice shell deforming by Newtonian volume diffusion is very large (Goldsby and Kohlstedt, 2001), and convection is predicted to occur in the stagnant-lid regime. However, the effective viscosity of the surface can be dramatically reduced if convective stresses exceed the lithospheric yield stress (Solomatov, 2004), leading to so-called “sluggish-lid” convection. Previous work shows that sluggish-lid behavior can occur on Enceladus and Europa for surface yield stress $< 100 \text{ kPa}$ (Showman and Han, 2005; Barr, 2008; O’Neill and Nimmo, 2010). Here, we limit the effective viscosity of surface by imposing a low $\Delta\eta$, which is the simplest way of mimicking the effect of brittle surface ice with a low yield stress (Barr, 2008). We use $\eta_1 = 10^{14} \text{ Pa s}$, for an ice grain size $\sim 0.1 \text{ mm}$ (Barr and McKinnon, 2007).

We simulate convection for $\Delta\eta$ between $10^{2.75} - 10^{4.25}$, close to the boundary between the sluggish-lid regime and stagnant-lid regime, $\Delta\eta \sim 10^4 - 10^5$ (Solomatov, 1995). We model convection in an 8×1 domain, with 512×64 elements, with periodic boundary conditions at the sides to minimize edge effects. We use a basally heated ice shell because the spatial distribution of tidal dissipation in a convecting ice shell is not fully understood (Han and Showman, 2010). Additionally, surface deformation

rates and heat flow are likely insensitive to the precise mode of heating (Solomatov and Moresi, 2000). We allow all our simulations to reach a steady state, then measure how strain rate and heat flux vary at the surface.

The convective heat flux, $F = \frac{k\Delta T}{D}Nu$, where $k = 3.3 \text{ W m}^{-1} \text{ K}^{-1}$ is the thermal conductivity and Nu is the Nusselt number, which describes the efficiency of convective versus conductive heat transport. The strain rate is approximated by $\dot{\epsilon} = (\partial v_{xsf}/\partial x)$, where v_{xsf} is the horizontal velocity at the surface. Simulations successfully match the inferred conditions for grooved terrain formation if regions of surface extension have an average heat flux of $100 - 200 \text{ mW m}^{-2}$, regions of compression have a heat flux of $60 - 80 \text{ mW m}^{-2}$ predicted for dark terrain, and surface strain rates are $10^{-13} - 10^{-16} \text{ s}^{-1}$.

2.4 Results

For $Ra_1 = 10^5 - 10^{8.5}$ and $\Delta\eta = 10^{3.25} - 10^{3.75}$ (see Figure 2.1), convection can match the heat flux inferred for both grooved terrain and dark terrain, at surface strain rates suggested by extensional necking models. Given a basal viscosity of 10^{14} Pa s , ice shell thicknesses of $8 - 110 \text{ km}$ and an effective surface viscosity $\eta_0 < 10^{18} \text{ Pa s}$ are conducive to forming grooved terrain in a convecting ice shell. This corresponds to a surface yield stress $\sigma_{sy} < 60 \text{ kPa}$ for $\eta_0 \sim \sigma_y(z)/2\dot{\epsilon}$, with $\sigma_y(z) = \sigma_{sy} + \mu\rho gz$, coefficient of friction $\mu = 0.1$, $\dot{\epsilon} = 10^{-13} \text{ s}^{-1}$ and depth $z = 1 \text{ km}$, close to elastic thickness estimates for grooved terrain (Nimmo *et al.*, 2002).

Figure 2.2 shows the surface strain rate and heat flow in regions of extension and contraction as a function of $\Delta\eta$. The average heat flux is independent of Ra_1 , however, lower Rayleigh numbers result in larger differences in heat flux over regions of extension and contraction; more vigorously convecting ice shells generate more uniform surface heat flows. For $\Delta\eta \geq 10^4$ the surface heat flux is below that inferred from models of flexure, but may be consistent with estimates based on models of

extensional necking. Surface strain rates are significant even for $\Delta\eta > 10^{3.75}$, suggesting that convection-driven resurfacing can occur on the boundary between the stagnant-lid and sluggish-lid regimes. For $\Delta\eta \leq 10^{3.25}$ the average surface strain rate exceeds the range inferred from extensional necking models, however strain rates as high as 10^{-12} s^{-1} have been inferred for European bands (Stempel *et al.*, 2005), which may be a more appropriate constraint for subdued grooves.

Figure 2.3a shows the temperature field, and the variation of F and $\dot{\epsilon}$ on the surface, of a convection simulation with $Ra_1 = 10^6$ and $\Delta\eta = 10^{3.25}$, corresponding to $D = 17 \text{ km}$. Regions undergoing extension are interpreted as sites of groove formation and are $\sim 100 \text{ km}$ wide, with an average heat flux of 170 mW m^{-2} and strain rates of $10^{-14} - 10^{-13} \text{ s}^{-1}$. Outside this region, the average heat flux is 70 mW m^{-2} . Figure 2.3b shows a case with $Ra_1 = 10^8$ and $\Delta\eta = 10^{3.75}$, with $D \sim 80 \text{ km}$. Here, groove lanes are $20 - 120 \text{ km}$ wide with $F \sim 100 \text{ mW m}^{-2}$ and $F \sim 50 \text{ mW m}^{-2}$ outside this region, just below that inferred for dark terrain. In each successful case, $\sim 50\%$ of the surface is under extension, consistent with proportionality of extensional grooved terrain and dark terrain. Convective downwelling regions remain stationary, allowing for the persistence of relatively undeformed terrain.

2.5 Discussion

Here we use a simple model to show that sluggish-lid convection can provide the heat flux and strain rate inferred for grooved terrain formation. Calculations using more sophisticated plastic rheologies and in a spherical domain are required to test whether convection can create linear groove lanes. Viscosity dependent internal heating modes (e.g. Han and Showman 2010) and the incorporation of damage or strain weakening (e.g. Buck *et al.* 2003; Bland *et al.* 2010) into convection models could be important in localizing surface deformation.

To form grooved terrain by convection, the upper surface must be sufficiently weak

to allow convective plumes to come close to the surface. The effective yield stress of ice in tidally flexed ice shells may be low, as suggested by the presence of cycloids on Europa (Hoppa *et al.*, 1999) and locations of tiger stripe plumes on Enceladus (Hurford *et al.*, 2007), which both seem to be controlled by diurnal tidal stresses $\sim 10 - 100$ kPa. Our results suggest a yield stress of this magnitude may permit convection-driven resurfacing, however this may require a coefficient of friction of ice below the range suggested from laboratory experiments (Schulson and Fortt, 2012). Stresses from global surface expansion, which were likely important (Collins, 2008) may also have increased background extensional stresses and reduced the effective yield stress, allowing convection to drive deformation.

If the total extensional strain on Ganymede exceeds that from global expansion, convection was a likely driving force because passive spreading could not have driven deformation once expansion stresses had been alleviated. Estimates of the strain recorded in grooved terrain depend on the formation mechanism of subdued grooves (Collins, 2006), which could form by lithospheric spreading (Head *et al.*, 2002) or cryo-volcanism (Schenk *et al.*, 2001). Abundant evidence for lithospheric spreading would imply a surface strain beyond what is possible (Collins, 2006) from differentiation and resonance (Squyres, 1980; Bland *et al.*, 2009).

Dark terrain preserved above convective downwelling regions would have experienced significant contraction. Contraction has a variety of surface expressions, but by analogy with Europa, we expect long-wavelength compressional folds to form in the dark terrain (Prockter and Pappalardo, 2000; Pappalardo and Collins, 2005). Bland and McKinnon (2012) show that fold formation and lithospheric thickening on Europa can occur for strain rates of $10^{-12} - 10^{-15} \text{ s}^{-1}$ and $F \sim 20 - 100 \text{ mW m}^{-2}$, close to conditions above convective downwelling regions in our models (see Figure 2.3). No contractional features have been observed in dark terrain (Pappalardo *et al.*, 2004) and deformed craters in dark terrain even show evidence for significant sur-

face extension (Pappalardo and Collins, 2005). However, the presence of pervasive, long-wavelength and low-amplitude folds, would allow dark terrain to be regions of net contraction which accommodate extension in grooved terrain. Identification of long wavelength folds in dark terrain and crustal spreading along subdued grooves by ESA’s upcoming JUICE mission would support convection-driven grooved terrain formation.

Acknowledgements. This work was supported by NASA Planetary Geology & Geophysics grant NNX12AI76G.

2.6 References

- Barr, A. C. 2008. Mobile lid convection beneath Enceladus' south polar terrain. *Journal of Geophysical Research: Planets* **113**(E7).
- Barr, A. C., and W. B. McKinnon 2007. Convection in ice I shells and mantles with self-consistent grain size. *J. Geophys. Res. Planets* **112**, 2012.
- Barr, A. C., and L. J. Preuss 2010. On the origin of south polar folds on Enceladus. *Icarus* **208**, 499–503.
- Bland, M. T., and W. B. McKinnon 2012. Forming Europa's folds: Strain requirements for the production of large-amplitude deformation. *Icarus* **221**, 694–709.
- Bland, M. T., W. B. McKinnon, and A. P. Showman 2010. The effects of strain localization on the formation of Ganymede's grooved terrain. *Icarus* **210**(1), 396–410.
- Bland, M. T., and A. P. Showman 2007. The formation of Ganymede's grooved terrain: Numerical modeling of extensional necking instabilities. *Icarus* **189**, 439–456.
- Bland, M. T., A. P. Showman, and G. Tobie 2009. The orbital thermal evolution and global expansion of Ganymede. *Icarus* **200**, 207–221.
- Buck, W. R., L. L. Lavier, and A. Babeyko 2003. A numerical model of lithospheric extension producing fault-bounded basins and ranges. *International Geology Review* **45**(8), 712–723.
- Collins, G. C. 2006. Global Expansion of Ganymede Derived from Strain Measurements in Grooved Terrain, abstract no. 2077. In *Lunar and Planet. Sci. Conf. Abstracts XXXVII*. Lunar and Planetary Institute, Houston, Tex.
- Collins, G. C. 2008. Driving Mechanisms for Grooved Terrain Formation on Ganymede: Comparison of Theory to Global Groove Database, abstract no. 2254. In *Lunar and Planet. Sci. Conf. Abstracts XXXIX*. Lunar and Planetary Institute,

- Houston, Tex.
- Collins, G. C., J. W. Head, and R. T. Pappalardo 1998. Formation of Ganymede Grooved Terrain by Sequential Extensional Episodes: Implications of Galileo Observations for Regional Stratigraphy. *Icarus* **135**, 345–359.
- Dombard, A. J., and W. B. McKinnon 2001. Formation of Grooved Terrain on Ganymede: Extensional Instability Mediated by Cold, Superplastic Creep. *Icarus* **154**, 321–336.
- Goldsby, D. L., and D. L. Kohlstedt 2001. Superplastic deformation of ice: Experimental observations. *J. Geophys. Res.* **106**, 11017.
- Han, L., and A. P. Showman 2010. Coupled convection and tidal dissipation in Europa’s ice shell. *Icarus* **207**, 834–844.
- Han, L., G. Tobie, and A. P. Showman 2012. The impact of a weak south pole on thermal convection in Enceladus’ ice shell. *Icarus* **218**, 320–330.
- Head, J., R. Pappalardo, G. Collins, M. J. S. Belton, B. Giese, R. Wagner, H. Brene-man, N. Spaun, B. Nixon, G. Neukum, and J. Moore 2002. Evidence for Europa-like tectonic resurfacing styles on Ganymede. *Geophys. Res. Lett.* **29**, 2151.
- Hoppa, G. V., B. R. Tufts, R. Greenberg, and P. E. Geissler 1999. Formation of cycloidal features on Europa. *Science* **285**, 1899–1902.
- Hurford, T. A., P. Helfenstein, G. V. Hoppa, R. Greenberg, and B. G. Bills 2007. Eruptions arising from tidally controlled periodic openings of rifts on Enceladus. **447**, 292–294.
- Lucchitta, B. K. 1980. Grooved terrain on Ganymede. *Icarus* **44**, 481–501.
- Moresi, L.-N., and V. S. Solomatov 1995. Numerical investigation of 2D convection with extremely large viscosity variations. *Physics of Fluids* **7**, 2154–2162.
- Murchie, S. L., J. W. Head, P. Helfenstein, and J. B. Plescia 1986. Terrain Types and Local-Scale Stratigraphy of Grooved Terrain on Ganymede. In *Lunar and Planetary Science Conference Proceedings*, Volume 17 of *Lunar and Planetary Science*

- Conference Proceedings*, pp. 222.
- Nimmo, F., and R. T. Pappalardo 2004. Furrow flexure and ancient heat flux on Ganymede. *Geophys. Res. Lett.* **31**, 19701.
- Nimmo, F., R. T. Pappalardo, and B. Giese 2002. Effective elastic thickness and heat flux estimates on Ganymede. *Geophys. Res. Lett.* **29**, 1158.
- O'Neill, C., and F. Nimmo 2010. The role of episodic overturn in generating the surface geology and heat flow on Enceladus. *Nature Geoscience* **3**, 88–91.
- Pappalardo, R. T., and G. C. Collins 2005. Strained craters on Ganymede. *Journal of structural geology* *27*(5), 827–838.
- Pappalardo, R. T., G. C. Collins, J. Head, P. Helfenstein, T. B. McCord, J. M. Moore, L. M. Prockter, P. M. Schenk, and J. R. Spencer 2004. Geology of Ganymede. *Jupiter: The Planet, Satellites and Magnetosphere*, 363–396.
- Pappalardo, R. T., J. W. Head, G. C. Collins, R. L. Kirk, G. Neukum, J. Oberst, B. Giese, R. Greeley, C. R. Chapman, P. Helfenstein, J. M. Moore, A. McEwen, B. R. Tufts, D. A. Senske, H. H. Breneman, and K. Klaasen 1998. Grooved Terrain on Ganymede: First Results from Galileo High-Resolution Imaging. *Icarus* **135**, 276–302.
- Parmentier, E. M., S. W. Squyres, J. W. Head, and M. L. Allison 1982. The tectonics of Ganymede. *Nature* **295**, 290–293.
- Patterson, G. W., G. C. Collins, J. W. Head, R. T. Pappalardo, L. M. Prockter, B. K. Lucchitta, and J. P. Kay 2010. Global geological mapping of Ganymede. *Icarus* **207**, 845–867.
- Prockter, L. M., J. W. Head, R. T. Pappalardo, R. J. Sullivan, A. E. Clifton, B. Giese, R. Wagner, and G. Neukum 2002. Morphology of European bands at high resolution: A mid-ocean ridge-type rift mechanism. *J. Geophys. Res.* **107**.
- Prockter, L. M., and R. T. Pappalardo 2000. Folds on Europa: Implications for Crustal Cycling and Accommodation of Extension. *Science* **289**, 941–944.

- Schenk, P. M., W. B. McKinnon, D. Gwynn, and J. M. Moore 2001. Flooding of Ganymede's bright terrains by low-viscosity water-ice lavas. **410**, 57–60.
- Schulson, E. M., and A. L. Fortt 2012. Friction of ice on ice. *Journal of Geophysical Research: Solid Earth* **117**(B12).
- Shoemaker, E., B. Lucchitta, D. Wilhelms, J. Plescia, and S. Squyres 1982. The geology of Ganymede. In D. Morrison (Ed.), *Satellites of Jupiter*, pp. 435–520. University of Arizona Press.
- Showman, A. P., and L. Han 2005. Effects of plasticity on convection in an ice shell: Implications for Europa. *Icarus* **177**, 425–437.
- Showman, A. P., D. J. Stevenson, and R. Malhotra 1997. Coupled Orbital and Thermal Evolution of Ganymede. *Icarus* **129**, 367–383.
- Smith, B. A., L. A. Soderblom, T. V. Johnson, A. P. Ingersoll, S. A. Collins, E. M. Shoemaker, G. E. Hunt, H. Masursky, M. H. Carr, M. E. Davies, A. F. Cook, J. M. Boyce, T. Owen, G. E. Danielson, C. Sagan, R. F. Beebe, J. Veverka, J. F. McCauley, R. G. Strom, D. Morrison, G. A. Briggs, and V. E. Suomi 1979. The Jupiter system through the eyes of Voyager 1. *Science* **204**, 951–957.
- Solomatov, V. S. 1995. Scaling of temperature- and stress-dependent viscosity convection. *Physics of Fluids* **7**, 266–274.
- Solomatov, V. S. 2004. Initiation of subduction by small-scale convection. *Journal of Geophysical Research (Solid Earth)* **109**, 1412.
- Solomatov, V. S., and L.-N. Moresi 2000. Scaling of time-dependent stagnant lid convection: Application to small-scale convection on Earth and other terrestrial planets. *J. Geophys. Res.* **105**, 21795–21818.
- Squyres, S. W. 1980. Volume changes in Ganymede and Callisto and the origin of grooved terrain. *Geophys. Res. Lett.* **7**, 593–596.
- Squyres, S. W., and S. K. Croft 1986. The tectonics of icy satellites. In J. A. Burns and M. S. Matthews (Eds.), *Satellites*, Tucson, AZ, pp. 293–341. University of

Arizona Press.

- Stempel, M. M., A. C. Barr, and R. T. Pappalardo 2005. Model constraints on the opening rates of bands on Europa. *Icarus* **177**(2), 297–304.
- Tackley, P. J. 2000. Self-consistent generation of tectonic plates in time-dependent, three-dimensional mantle convection simulations. *Geochemistry, Geophysics, Geosystems* **1**, 1021.
- Trompert, R., and U. Hansen 1998. Mantle convection simulations with rheologies that generate plate-like behaviour. **395**, 686–689.

2.7 Tables and Figures

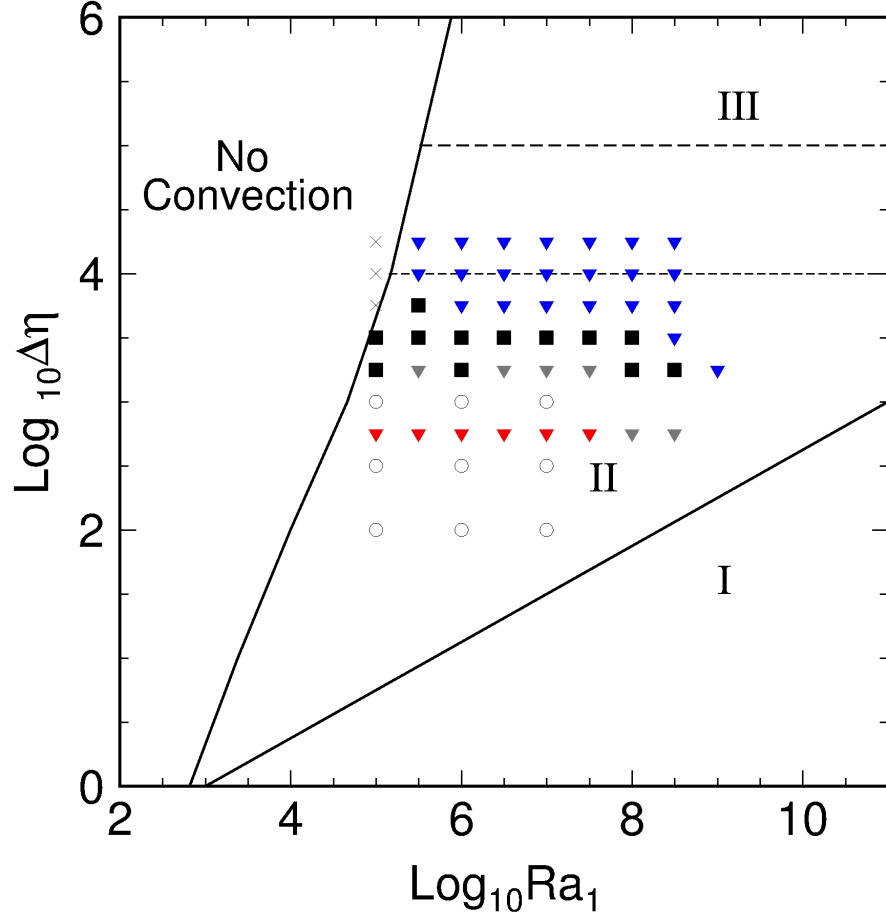


Figure 2.1: Values of Rayleigh number (Ra_1) and viscosity contrast ($\Delta\eta$) explored in this study. Black squares represent simulations which successfully match heat flux estimates based on flexure for grooved terrain and dark terrain at strain rates inferred from models of extensional necking. Inverted triangles show simulations that do not meet these criterion, with blue showing those with a heat flow that is too low, gray showing those with an average strain rate exceeding 10^{-13} s^{-1} , and red showing simulations with an average heat flux exceeding 200 mW m^{-2} in regions of extension. “x” symbols show simulations that do not convect and empty circles represent simulations run for the Enceladus SPT by Barr (2008). Lines indicate the boundaries between convective regimes: isoviscous (I), sluggish-lid (II), and stagnant-lid (III). Dashed lines indicate gradual transition from sluggish-lid regime to stagnant-lid regime.

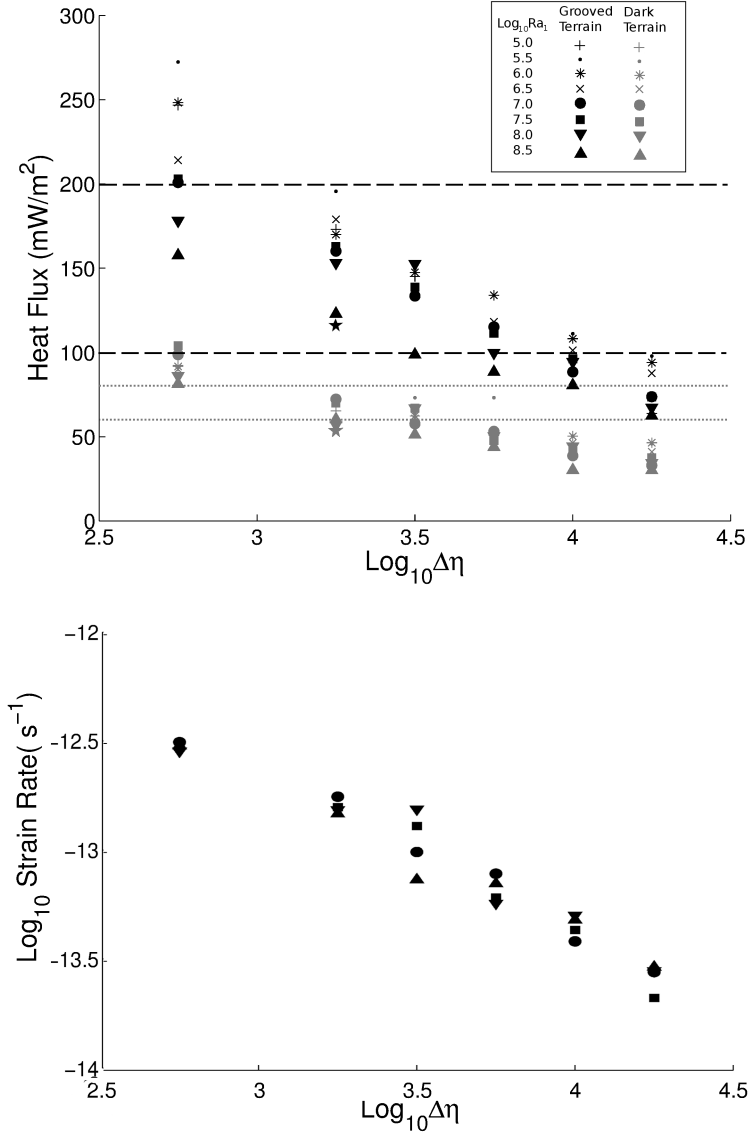


Figure 2.2: (top) Heat flux of convection simulations plotted as a function of viscosity contrast. Basal Rayleigh numbers of $\text{log}_{10}(Ra_1) = (5.0, 5.5, 6.0, 6.5, 7.0, 7.5, 8.0, 8.5)$ are plotted with the symbols, (indicated in legend), plus, dot, asterisk, cross, circle, square, inverted triangle and upright triangle, respectively. Star symbols represent one simulation for $Ra_1 = 10^9$. Black symbols represent average heat flux in regions of extension, and gray symbols show the average heat flux in regions of contraction. Black dashed lines show estimated heat flux for grooved terrain. Gray dotted lines show estimate heat flux for dark terrain. (bottom) Surface strain rate of convection simulations plotted as a function of viscosity contrast. Basal Rayleigh numbers represented by same symbols as above, although many overlap and are not visible.

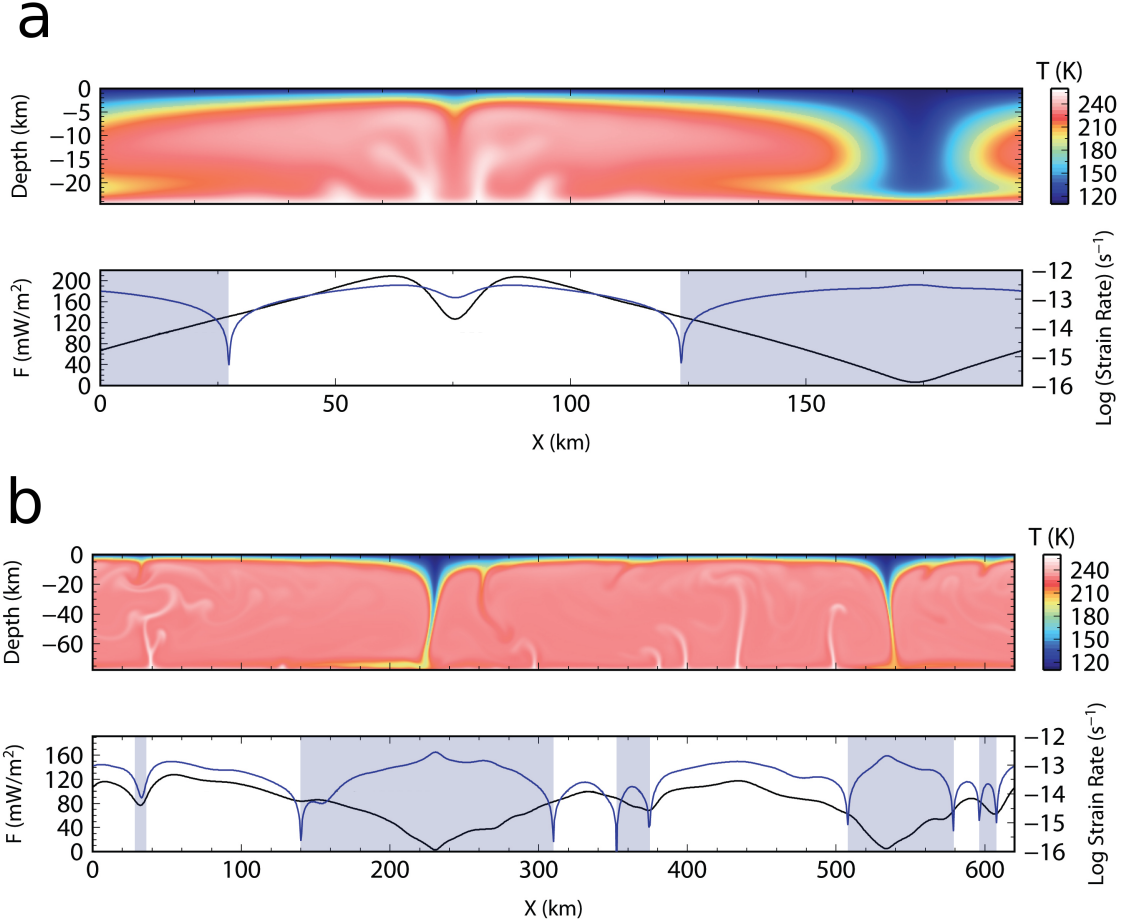


Figure 2.3: Temperature fields of convection simulations with the corresponding surface heat flux (black) and strain rate (blue) plotted below. Shaded areas indicate regions of surface contraction interpreted as dark terrain (a) Simulation with $Ra_1 = 10^6$ and $\Delta\eta = 10^{3.25}$, corresponding to $D = 25$ km. A groove lane 150 km wide forms at $x = 75$ km. (b) Simulation with $Ra_1 = 10^8$ and $\Delta\eta = 10^{3.5}$, corresponding to $D = 80$ km. Groove lanes form near $x = 50$ km (80 km wide) and $x = 450$ km (100 km wide). Narrower groove lanes also form near $x = 10$ km, $x = 330$ km and $x = 590$ km.

Chapter 3 | Global Resurfacing of Uranus's Moon Miranda by Convection

N.P. Hammond¹, A.C. Barr¹.

1. Department of Earth, Environmental and Planetary Science, Brown University,
Providence, RI 02912, USA

Published in it's present form in:

Geology 42.11 (2014): 931-934

3.1 Abstract

Miranda, an icy moon of Uranus, is one of the most visually striking and enigmatic bodies in the solar system. Three polygonal shaped regions of intense deformation, dubbed “coronae”, dominate the surface of Miranda. Here we use numerical methods to show that sluggish-lid convection in Miranda’s ice shell, powered by tidal heating, can simultaneously match the global distribution of coronae, the concentric deformation pattern, and the estimated heat flow during formation. The expected rheological conditions in Miranda’s ice shell lead to the development of low-order convection that produces surface deformation patterns similar to those observed. We find that core size strongly controls convection geometry and that low-order convection patterns are much more stable for core radii less than half the satellite radius.

3.2 Introduction

Miranda, the innermost regular satellite of Uranus, has an average radius, R , of 235.8 km (Thomas, 1988), and a mean density of 1.2 g/cm^3 (Jacobson *et al.*, 1992), suggesting a mixed ice and rock composition. Miranda orbits Uranus at a distance of 5.11 Uranus radii, and its current eccentricity, $e = 0.0013$. Despite its relatively small size, Miranda appears to have experienced an episode of intense endogenic resurfacing (Smith, 1986), resulting in the formation of at least three remarkable and unique surface features called “coronae.” Each corona is at least 200 km across, is ovoidal or polygonal in shape (see Fig. 3.1), and is surrounded by an outer belt of concentric, sub-parallel ridges and troughs (Smith *et al.*, 1986). Arden corona, the largest, has ridges and troughs with up to 2 km of relief, whose slopes and orientations are consistent with tilt-block style normal faults (Pappalardo, 1994). Elsinore corona has an outer belt that is 80 km wide, relatively smooth, and elevated above the surrounding terrain by $\sim 100 \text{ m}$ (Schenk, 1991). Inverness corona has a trapezoidal

shape with a large bright chevron at its center. The northern hemisphere of Miranda was never imaged by the *Voyager 2* spacecraft, therefore it is unknown whether additional coronae exist.

Many hypotheses have been put forth to suggest how the coronae formed. One hypothesis is that Miranda was catastrophically disrupted and reassembled, and that coronae formed as a result of silicate material sinking through an icy mantle causing concentric surface contraction (Janes and Melosh, 1988). In contrast, the majority of researchers suggest that coronae likely formed as the result of buoyant diapirs causing surface extension (Croft and Soderblom, 1991; Greenberg, 1991; Pappalardo *et al.*, 1997). Geologic interpretations that the coronae are composed of extensional tectonic and volcanic landforms support the hypothesis that coronae formed in an extensional environment (Schenk, 1991; Pappalardo *et al.*, 1997). It has also been suggested that heat generated during tidal flexing played an important role in corona formation (Dermott *et al.*, 1988; Pappalardo *et al.*, 1997), but the way in which internal heating leads to the observed surface deformation has not been well described.

Here we simulate convection in Miranda’s ice shell to test the hypothesis that coronae formed by convection-driven resurfacing during an episode of tidal heating. We show that if the surface is weak, sluggish-lid convection can occur, which simultaneously produces the estimated heat flow, the deformation pattern within the coronae and the approximate number of upwellings needed to form the coronae.

3.3 Formation Conditions

Models of lithospheric flexure along the flanks of Arden corona suggest an elastic thickness of 2 km and a thermal gradient of $8 - 20 \text{ K km}^{-1}$ during corona formation (Pappalardo *et al.*, 1997). For a thermal conductivity, k , of $5 \text{ W m}^{-1} \text{ K}^{-1}$ (Petrenko and Whitworth, 1999), this corresponds to a heat flux of $40 - 100 \text{ mW m}^{-2}$. Such a high heat flow implies the past existence of thermal anomalies beneath the coronae,

powered by a heat source other than radiogenic heating, which is not expected to be significant (Greenberg, 1991). One likely heat source is tidal dissipation in the ice shell, which can occur when the time-varying tidal potential during an eccentric orbit causes a satellite to be repeatedly stretched and squeezed (e.g., Ojakangas and Stevenson 1989). Several studies suggest that a past 3 : 1 orbital resonance between Miranda and neighboring satellite Umbriel can explain Miranda’s present orbital inclination of 4.5 deg (Dermott *et al.*, 1988; Tittlemore and Wisdom, 1989; Malhotra and Dermott, 1990; Verheylewegan *et al.*, 2013). During this resonance, Miranda’s eccentricity may have been excited to $e = 0.05$ (Malhotra and Dermott, 1990).

If Miranda were as dissipative as Saturn’s satellite Enceladus, which is similar in size, composition, and orbital frequency to Miranda, we calculate that tidal dissipation could have generated as much as 5 GW of power (cf. Peale and Cassen 1978), for a Miranda with a quality factor $Q = 20$, and Love number $k_2 = 9.3 \times 10^{-4}$ (Supplemental Material S1). Tidal dissipation focused in relatively warm ice at depth could have created conditions more favorable for convection by raising the thermal gradient across the ice shell and lowering the viscosity of ice.

For convection to drive surface deformation, on either terrestrial planets or icy satellites, the surface must behave much weaker than predicted by laboratory experiments (Tackley, 1993; Showman and Han, 2005). Convection with simple temperature-dependent viscosity predicts that vigorous fluid motion would be confined below a thick, undeforming stagnant lid (Solomatov, 1995). For convection to reach the surface, convective stresses must overcome the yield strength in the lid (e.g., Moresi and Solomatov 1998). Once the yield strength is overcome, the surface deforms with strain rates similar to systems in the sluggish-lid regime (Solomatov, 1995), which have moderately low-viscosity contrasts between the top and bottom of the mantle.

Following scaling laws of Solomatov (2004), we calculate that convective stresses

could have deformed the surface of Miranda if its effective surface yield stress is ~ 10 kPa (Supplemental Material S2). Similar yield stress values have been inferred for Enceladus, where near-surface convection is thought to generate the observed thermal anomaly in the geologically active South Polar Terrain (Barr, 2008; O'Neill and Nimmo, 2010). Additionally, the surfaces of Europa and Enceladus fail in response to daily tidal stresses of ~ 20 kPa (Hoppa *et al.*, 1999; Hurford *et al.*, 2007). Laboratory and field experiments, however, suggest the yield stress of ice is an order of magnitude higher, $\sim 0.1 - 1$ MPa (Kehle, 1964), implying that the lithospheres of tidally flexed satellites may be anomalously weak.

3.4 Methods

We simulate convection in Miranda's ice mantle for a wide range of initial conditions using the three-dimensional (3-D) spherical convection model CitcomS (Zhong *et al.*, 2000; Tan *et al.*, 2006). We explore three different sizes for Miranda's rock core: (1) core radius to satellite radius fraction $x = 0.55$, corresponding to a core composed of low-density rock (2.5 g cm^{-3}), (2) $x = 0.4$, appropriate for rock density 3.5 g cm^{-3} , and (3) $x = 0.25$, appropriate for a partially differentiated body with an ice mantle containing 10% rock by volume. We vary the Rayleigh number, which governs the vigor of convection, as

$$Ra = \frac{\rho_i g \alpha \Delta T D^3}{\kappa \eta} \quad (3.1)$$

where $\rho_i = 920 \text{ kg m}^{-3}$ is the density of ice, $g = 0.083 \text{ m s}^{-2}$ is gravity, the thermal expansivity $\alpha = 1.56 \times 10^{-4} (T_i/250 \text{ K}) \text{ K}^{-1}$ (Kirk and Stevenson, 1987), the temperature contrast across the ice shell is $\Delta T = T_b - T_s$ (b is base, s is surface), the ice shell thickness $D = (1 - x) \times R_{sat}$, the thermal diffusivity $\kappa = 1.47 \times 10^{-6} (250 \text{ K}/T_i)^2 \text{ m}^2 \text{ s}^{-1}$ (Kirk and Stevenson, 1987), and η_b is the viscosity of ice at the base. The surface temperature, $T_s = 60 \text{ K}$, T_i is the average internal temperature,

and the temperature at the base of the ice shell, $T_b = 220 - 270K$. Over this range of T_b , the viscosity of ice deforming by volume diffusion, for a grain size $d = 0.2$ mm, $\eta_b = 10^{14} - 10^{17}$ Pa s (Goldsby and Kohlstedt, 2001; Barr and Pappalardo, 2005; Barr and McKinnon, 2007), yielding $Ra = 10^4 - 10^8$ (Supplemental Material S3). Throughout the ice shell, viscosity varies as

$$\eta(T) = \eta_b \exp -\gamma T, \quad (3.2)$$

where $\gamma = \theta/\Delta T$, $\theta = \ln \Delta\eta$, and $\Delta\eta$ is the effective viscosity contrast across the ice shell. We assume sluggish-lid convection, limiting $\Delta\eta = 10^3 - 10^4$ to mimic the effect of surface ice with a low yield stress (King *et al.*, 1992; Trompert and Hansen, 1998; Moresi and Solomatov, 1998). The surface Rayleigh number $R_{as} = Ra/\Delta\eta$. The majority of our simulations are started with an initially conductive temperature profile with a small temperature perturbation near the base, and free slip boundary conditions at the top and bottom of the ice shell. We run simulations until the surface heat flow approaches steady state, and record the velocity and thermal gradient at the surface.

We are interested in whether convection in Miranda’s ice shell leads to surface deformation patterns that resemble the distribution of the coronae, but the number of convective plumes and their arrangement has been shown to be affected by core size (Deschamps *et al.*, 2010), viscosity structure, Rayleigh number (Schubert, 2001) and the distribution of tidal heating (Han and Showman, 2010). It is difficult to determine which of these variables has the largest influence on convection geometry.

In order to independently address each of these effects, we use three different approaches. First we use a simplified approach where we apply an initial temperature perturbation (of spherical harmonic order $l = 3$, $m = 2$) that pushes the system toward low-order tetrahedral convection (e.g. Ratcliff *et al.* 1996). We then vary $Ra, \Delta\eta$ and x to determine the parameter space over which low-order convection

can remain stable. In this suite of simulations, the temperature at the base of the ice shell is held constant, representing tidal energy focused on warm low-viscosity ice at depth. Our second approach is designed to assess the influence of the initial temperature condition. We re-run simulations that led to low-order convection in the first suite of simulations using a random initial temperature perturbation.

Finally, we explore how the distribution of tidal heating might affect convection geometry. For a satellite with a subsurface ocean, a degree-2 pattern is expected to dominate, where heating concentrates near the poles (Tobie *et al.*, 2005). Without an ocean, tidal strain near the base of the ice shell may concentrate near the equator in a cubic type pattern (see Tobie *et al.* 2005; Beuthe 2013). By assuming that tidal heating heterogeneously warmed the ice shell to generate the initial temperature condition, we represent these patterns as the sum of spherical harmonic temperature perturbations, following the equations given by Beuthe (2013).

3.5 Results

We find that rheological conditions in Miranda’s ice shell favor the development of low-order convection patterns. Core size strongly controls convection geometry. Models of Miranda with a core size of $x = 0.55$ generally develop higher-order convective patterns, whereas Miranda with a core size of $x = 0.25$ can sustain tetrahedral convection over the broadest range of rheological constraints (Fig. 3.2a), probably because longer-wavelength plumes are preferred as core fraction decreases (Schubert, 2001; Deschamps *et al.*, 2010). For simulations with low-order convection and with surface Rayleigh numbers between 10^2 and $10^{3.5}$, the average thermal gradient in regions of surface extension is remarkably consistent with estimates suggested from lithospheric flexure (Fig. 3.2b). This shows that sluggish lid convection can simultaneously match the estimated thermal gradient and the approximate number of upwellings necessary to form the coronae. We consider the successful range in plume

number to be 4 – 6, because there are three coronae visible in the southern hemisphere of Miranda. Simulations with random initial temperature perturbations and a fractional core size of $x = 0.25$ also develop low-order convection patterns (Fig. 3.3). This shows that low-order convection easily develops in systems with low Rayleigh numbers, low effective-viscosity contrasts, and small core sizes, regardless of the initial temperature condition. Simulations run with a zero-slip condition at the base of the ice shell had slightly lower Nusselt numbers than those with a free slip condition, but the convection geometry was unaffected.

We also find that convection in Miranda’s ice shell produces the concentric deformation pattern within the coronae. Figure 3.4 shows the surface velocity field and near-surface thermal gradient resulting from two different convection simulations. Above convective upwellings, the thermal gradient is elevated to values predicted from flexure and the surface extends concentrically away from the center, similar to the predicted formation environment of the coronae (Greenberg, 1991). Surface deformation patterns resulting from tetrahedral convection (e.g., Ratcliff *et al.* 1996), shown in Figure 3.4b, can match the locations of two of the coronae, but a third upwelling occurs where no surface deformation is observed.

To match the exact locations of all of the coronae, we must invoke a tidal heating pattern predicted for an icy satellite with no ocean (Tobie *et al.*, 2005; Beuthe, 2013). In this case, four plumes are generated near the equator, separated by ~ 90 deg each. Inverness corona is located near the South Pole, hence in order for this convection pattern to be consistent with the current locations of the coronae, a reorientation is required. Figure 3.4c shows surface deformation patterns produced by this style of convection after a 60 deg rotation from the sub-Uranian point at the equator toward the South Pole. Convective upwellings occur almost precisely beneath the locations of Arden, Inverness, and Elsinore coronae, and the size of extensional regions above the upwellings are in good agreement with the sizes of the coronae. Similar reorientations

of Miranda have been suggested on the basis of the skewed impact crater distribution (Plescia, 1988) and the orientation of scarps that might have been influenced by stresses from reorientation (Pappalardo, 1994). The reorientation of a convective upwelling toward the poles is also suggested to have occurred on Enceladus (e.g., Nimmo and Pappalardo 2006).

3.6 Discussion

We find that convection in Miranda’s ice shell powered by tidal heating can generate the global distribution of coronae, the concentric orientation of sub-parallel ridges and troughs, and the thermal gradient implied by flexure. Models that account for the possible distribution of tidal heating can even match the precise locations of the coronae, after a reorientation of 60 deg. Two aspects of coronae morphology that do not arise naturally in our model are the angular shapes of the coronae, and their differences in morphology. Pappalardo (1994) suggested that the angular appearance may have resulted from radial stresses that interacted with preexisting fractures. We suggest that differences in coronae morphology might be partly explained by variations in extensional strain rates, since spreading rate can control the tectonic expression of mid-ocean ridges on Earth (Morgan and Chen, 1993). Simulations that match the thermal gradient from flexure have total power outputs of close to 10 GW (Fig. 3.2b), somewhat larger than the total power we predict could be generated during orbital resonance. One explanation for this discrepancy is that the thermal gradient inferred at Arden corona may not apply to Inverness or Elsinore corona, therefore convection need not produce the inferred thermal gradient at each upwelling at the same time. Alternatively, it is possible that convection may have episodically deformed the surface, leading to temporary, dramatic increases in surface heat flux (e.g., Moresi and Solomatov 1998; O’Neill and Nimmo 2010). In any case, Miranda must be at least as dissipative as Enceladus for tidal heating to power convection,

and may be required to enter orbital resonance already warm and dissipative, because tidal heating models predict that it is difficult to warm up Miranda from an initially cool state (Verheylewegan *et al.*, 2013). Further studies that incorporate a finite yield stress, and more complex treatments of tidal heating into spherical convection models, could explore the possible time-dependent behavior of corona formation and the feedback between tidal heating and thermal plumes.

3.7 Acknowledgments

This work is supported by NASA Earth and Space Science Fellowship fellowship NNX13AN99H and NASA Planetary Geology and Geophysics Program NNX12AI76G.

3.8 Supplemental Material

3.8.1 Convection Simulations

We use the 3D spherical convection model CitcomS (Zhong *et al.*, 2000; Tan *et al.*, 2006), to simulate convection in Miranda’s ice shell for a wide range of internal structures, ice viscosities and heating rates. We assume Miranda is at least partially differentiated with a fractional core size $X = R_c/R$ of 0.25, 0.4 and 0.7. These core sizes are computed using

$$X = \left(\frac{\bar{\rho} - \rho_m}{\rho_r - \rho_m} \right)^{1/3} \quad (3.3)$$

where $\bar{\rho} = 1.2 \text{ g/cm}^3$ is the bulk density of Miranda (Jacobson *et al.*, 1992), $\rho_r = 2.5$ to 3.5 g/cm^3 is the density of silicate rock and ρ_m is the density the icy mantle. For the fully differentiated case $\rho_m = 0.92$, the density of pure water ice. For the partially differentiated case $\rho_m = 1.178 \text{ g/cm}^3$, corresponding to a 10% volume fraction of silicates.

As discussed in the methods, we calculate the Rayleigh number assuming the

mantle is viscously deforming via diffusion creep. We vary the viscosity of ice at the base of the ice shell between 10^{14} - 10^{17} Pa s. This requires relatively high temperatures at the base of the ice shell and relatively small grain sizes. Figure 3.5 shows how viscosity and Rayleigh number vary as a function of the temperature in the ice shell for a grain size of 0.2 mm. Such a grain size may be plausible if grain growth is restricted by pinning from silicate particles (Barr and McKinnon 2007). These grain size and temperature constraints may be relaxed if ammonia is present in the ice shell. Ammonia hydrate may reduce the viscosity of ice above 173 K. Durham et al. (1993) examined the rheology of the ammonia-water ice system. Melted ammonia-hydrate along ice grains could lower the activation energy of grain boundary diffusion, thereby reducing viscosity. This mechanism may be similar to enhanced deformation rates observed for pure ice above 253 K. Goldsby and Kohlstedt (2001) argue that pre-melting along grains can enhance creep by a factor of 1000. Thus ice with even a small amount of ammonia may have a much lower viscosity above 173 K.

Thermal gradients and surface strain rates are calculated from the model after the convective heat flow and convection patterns reach steady states. The thermal gradient at the surface is calculated as $\frac{dT}{dz} = \frac{\Delta T}{D} Nu$, where Nu is the Nusselt number. Surface deformation rates are calculated as $\dot{\epsilon} = \left| \frac{\delta u_i}{\delta x_j} \right|$, where u is surface velocity, x is direction and $i, j = 1, 2$ are indices representing directions of co-latitude and co-longitude. We run our simulations until the average Nusselt number stabilizes to within 3 significant digits. We compared our results to previous studies by Zhong et al., (2008) and Ratcliff et al., (1996), as well as to scaling relationships for sluggish-lid convection (Barr 2008) derived from convection simulations using the 2D model CITCOM (Moresi and Solomatov 1995). Our benchmarking results, shown in figure 3.6, match the values of previous studies and expected values from scaling laws, to within 5-10%.

3.8.2 Convective Yielding

The critical Rayleigh number for convection in a Newtonian fluid is $Ra_{cr} = 20.9\theta^4$, where $\theta = \frac{Q\Delta T}{RT_i^2}$ (Solomatov 1995). For pure ice deforming by volume diffusion, $Q = 59.6$ kJ/mol. Using the same ΔT and T_i as in the methods section, $Ra_{cr} = 3 \times 10^7$, for pure ice in Miranda's ice shell. However, we imagine that tidal flexing during orbital resonance may have weakened the lithosphere prior to the initiation of convection and that the effective viscosity contrast across the surface was limited by the yield stress of the lithosphere. When convective stresses exceed the yield strength of the lithosphere σ_y , the effective viscosity of the surface η_0 , can be limited to $\eta_0 = \sigma_y/(2\dot{\epsilon})$ (e.g. Tackley 1993, Showman and Han 2005), where $\dot{\epsilon}$ is the surface strain rate. For a yield stress of 10 kPa, similar to the apparent yield stresses of the surfaces of Europa and Enceladus (see Hoppa et al., 1999, Hurdford et al., 2007), and for typical convective strain rates between 10^{-15} to 10^{-13} s $^{-1}$, this yields surface viscosities $\eta_0 = 10^{17}$ to 10^{19} Pa s. This yields a viscosity contrast across the ice shell of $\Delta\eta \approx 10^3$, $\theta \approx 6.908$, and a critical Rayleigh number $Ra_{cr} = 5 \times 10^4$.

Additionally, Solomatov (2004) calculated the critical yield stress for the initiation of convection-driven resurfacing as

$$\sigma_{y,cr} = 13 \frac{\alpha g \rho_m}{\Delta T} \left(\frac{RT_i^2}{E} \right) l_{hor} \quad (3.4)$$

where $l_{hor} = 300$ km is the width of convective upwellings and $E = 59.6$ kJ mol $^{-1}$ K $^{-1}$ is the activation energy of diffusion creep. For an internal temperature $T_i = 200$ K, $\sigma_{y,cr} = 5$ kPa. Tidal flexing during resonance could have fatigued the lithosphere (see Suresh 1999) and lowered the yield stress of the surface to near this value. Additionally, partial melts of ammonia-water ice could generate additional compositional buoyancy stresses, limiting the thermal buoyancy stresses required for convection to deform the surface.

3.8.3 Tidal Dissipation on Miranda

The magnitude of tidal heating on Miranda can be approximated with estimates for Miranda’s past eccentricity during a possible orbital resonance with neighboring satellite Umbriel. The magnitude of tidal heating can be written as

$$\dot{E} = e^2 \frac{21}{2} \frac{k_2}{Q} \frac{n^5 R^5}{G} \quad (3.5)$$

where $R = 238$ km is Miranda’s radius, $n = 5.15 \times 10^{-5}$ 1/s is the orbital frequency, G is the gravitational constant, e is orbital eccentricity, k_2 is the degree two love number and Q is the dissipation factor of the satellite (Peale and Cassen, 1978). The tidal love number for a homogeneous elastic spherical body is

$$k_2 = \frac{3}{2} \left(1 + \frac{19\mu}{2\bar{\rho}gR} \right) \quad (3.6)$$

where μ is the shear modulus of the satellite and $\bar{\rho}$ is the bulk density of the satellite. We assume $\mu = 4$ MPa and $Q = 20$, the same values used by Meyer and Wisdom (2007), and calculate $k_2 = 9.265 \times 10^{-4}$. For an orbital eccentricity of $e = 0.05$, we estimate a maximum tidal heating rate of 5 GW for Miranda.

An important consideration is that the tidal heating and orbital evolution of the satellite is coupled to the evolution of the parent planet. Just as the planet’s gravity causes a tidal bulge to form on the satellite, the satellite’s gravity causes a tidal bulge to form on the planet. Since the planet is rotating faster the satellite is orbiting, the tidal bulge on the planet is constantly being pulled “ahead” of the satellite. The additional gravitation pull of the off-axis tidal bulge acts as a torque on the satellite that causes the satellite to accelerate and move outward and causes the spin of the planet to slow down (see Burns *et al.* 1986). This exchange of angular momentum between the satellite and the planet ultimately control the dynamics of

tidal dissipation on the satellite.

Meyer and Wisdom (2007) showed that the time-averaged rate of energy dissipation in the satellite is limited by the amount of energy that can be extracted out of the parent planet. This rate is governed by the torque, T between the planet and the satellite. The torque is a function of the k_2 and Q of the planet. For two satellites that also exchange angular momentum, the steady state energy dissipation rate for the inner satellite,

$$\dot{E} = \frac{n_m T_m}{\sqrt{1 - e_m^2}} + \frac{n_u T_u}{\sqrt{1 - e_u^2}} \frac{T_m + T_u}{\left(\frac{GMm_m}{a_m} \frac{GMm_u}{a_u} \right)} \quad (3.7)$$

where L is angular momentum, a is the satellite semi major orbital axis, M is the mass of uranus, m is the mass of the satellite, and the subscripts m and u represent Miranda and Umbriel. We use $Q_p = 11000$ (Tittlemore and Wisdom, 1989), $k_{2p} = 0.104$ (Gavrilov and Zharkov, 1977) and calculate an equilibrium tidal heating rate for Miranda of 0.3 GW.

Thus for tides to generate 5 GW of power, the orbital eccentricity must be excited well above equilibrium (e.g. Dermott *et al.* 1988). The eccentricity would quickly dampen toward its equilibrium value and higher tidal dissipation rates would lead to more rapid eccentricity dampening. Dermott *et al.* 1988 calculated the time scale for eccentricity dampening on Miranda to be $t_{damp} = 2Q$ Myr, where Q is the dissipation factor for Miranda. For $Q = 20$ the dampening timescale is 40 Myr.

If we compare this to the timescale for convection driven coronae formation, our models show convection typically begins deforming the surface after less than 10 Myr. This suggests that convection driven coronae formation can occur during a period of increased orbital eccentricity before eccentricities significantly decay.

3.9 References

- Barr, A. C. 2008. Mobile lid convection beneath Enceladus' south polar terrain. *Journal of Geophysical Research: Planets* **113**(E7).
- Barr, A. C., and W. B. McKinnon 2007. Convection in ice I shells and mantles with self-consistent grain size. *J. Geophys. Res. Planets* **112**, 2012.
- Barr, A. C., and R. T. Pappalardo 2005. Onset of convection in the icy Galilean satellites: Influence of rheology. *Journal of Geophysical Research: Planets* **110**(E12).
- Beuthe, M. 2013. Spatial patterns of tidal heating. *Icarus* **223**(1), 308–329.
- Burns, J. A., M. S. Matthews, et al. 1986. *Satellites*. Number 77. University of Arizona Press.
- Croft, S., and L. Soderblom 1991. Geology of the Uranian satellites. *Uranus*, 561–628.
- Dermott, S. F., R. Malhotra, and C. D. Murray 1988. Dynamics of the Uranian and Saturnian satellite systems: A chaotic route to melting Miranda? *Icarus* **76**(2), 295–334.
- Deschamps, F., P. J. Tackley, and T. Nakagawa 2010. Temperature and heat flux scalings for isoviscous thermal convection in spherical geometry. *Geophysical Journal International* **182**(1), 137–154.
- Gavrilov, S., and V. Zharkov 1977. Love numbers of the giant planets. *Icarus* **32**(4), 443–449.
- Goldsby, D. L., and D. L. Kohlstedt 2001. Superplastic deformation of ice: Experimental observations. *J. Geophys. Res.* **106**, 11017.
- Greenberg, R. J. 1991. Uranus. In *University of Arizona Press; Space Science Series*.
- Han, L., and A. P. Showman 2010. Coupled convection and tidal dissipation in Europa's ice shell. *Icarus* **207**, 834–844.
- Hoppa, G. V., B. R. Tufts, R. Greenberg, and P. E. Geissler 1999. Formation of cycloidal features on Europa. *Science* **285**, 1899–1902.

- Hurford, T. A., P. Helfenstein, G. V. Hoppa, R. Greenberg, and B. G. Bills 2007. Eruptions arising from tidally controlled periodic openings of rifts on Enceladus. *447*, 292–294.
- Jacobson, R., J. Campbell, A. Taylor, and S. Synnott 1992. The masses of Uranus and its major satellites from Voyager tracking data and Earth-based Uranian satellite data. *The Astronomical Journal* **103**, 2068–2078.
- Janes, D., and H. Melosh 1988. Sinkers tectonics: An approach to the surface of Miranda. *Journal of Geophysical Research: Solid Earth* *93*(B4), 3127–3143.
- Kehle, R. O. 1964. Deformation of the Ross ice shelf, Antarctica. *Geological Society of America Bulletin* *75*(4), 259–286.
- King, S. D., C. W. Gable, and S. A. Weinstein 1992. Models of convection-driven tectonic plates: a comparison of methods and results. *Geophysical Journal International* *109*(3), 481–487.
- Kirk, R., and D. Stevenson 1987. Thermal evolution of a differentiated Ganymede and implications for surface features. *Icarus* *69*(1), 91–134.
- Malhotra, R., and S. F. Dermott 1990. The role of secondary resonances in the orbital history of Miranda. *Icarus* *85*(2), 444–480.
- Meyer, J., and J. Wisdom 2007. Tidal heating in Enceladus. *Icarus* *188*(2), 535–539.
- Moresi, L., and V. Solomatov 1998. Mantle convection with a brittle lithosphere: thoughts on the global tectonic styles of the Earth and Venus. *Geophysical Journal International* *133*(3), 669–682.
- Morgan, J. P., and Y. Chen 1993. Dependence of ridge-axis morphology on magma supply and spreading rate. *Nature* *364*(6439), 706–708.
- Nimmo, F., and R. T. Pappalardo 2006. Diapir-induced reorientation of Saturn’s moon Enceladus. *Nature* *441*(7093), 614–616.
- Ojakangas, G. W., and D. J. Stevenson 1989. Thermal state of an ice shell on Europa. *Icarus* *81*(2), 220–241.

- O'Neill, C., and F. Nimmo 2010. The role of episodic overturn in generating the surface geology and heat flow on Enceladus. *Nature Geoscience* **3**, 88–91.
- Pappalardo, R. T. 1994. *The origin and evolution of ridge and trough terrain and the geological history of Miranda*. Ph. D. thesis, Arizona State University Tempe.
- Pappalardo, R. T., S. J. Reynolds, and R. Greeley 1997. Extensional tilt blocks on Miranda: Evidence for an upwelling origin of Arden Corona. *Journal of Geophysical Research: Planets* **102**(E6), 13369–13379.
- Peale, S., and P. Cassen 1978. Contribution of tidal dissipation to lunar thermal history. *Icarus* **36**(2), 245–269.
- Petrenko, V. F., and R. W. Whitworth 1999. *Physics of ice*. OUP Oxford.
- Plescia, J. 1988. Cratering history of Miranda: Implications for geologic processes. *Icarus* **73**(3), 442–461.
- Ratcliff, J. T., G. Schubert, and A. Zebib 1996. Effects of temperature-dependent viscosity on thermal convection in a spherical shell. *Physica D: Nonlinear Phenomena* **97**(1), 242–252.
- Schenk, P. M. 1991. Fluid volcanism on Miranda and Ariel: Flow morphology and composition. *Journal of Geophysical Research: Solid Earth* **96**(B2), 1887–1906.
- Schubert, G. 2001. *Mantle convection in the Earth and planets*. Cambridge University Press.
- Showman, A. P., and L. Han 2005. Effects of plasticity on convection in an ice shell: Implications for Europa. *Icarus* **177**, 425–437.
- Smith, B. A. 1986. Voyager 2 in the Uranian system: imaging science results. *Science* **233**, 43–65.
- Solomatov, V. 2004. Initiation of subduction by small-scale convection. *Journal of Geophysical Research: Solid Earth* **109**(B1).
- Solomatov, V. S. 1995. Scaling of temperature- and stress-dependent viscosity convection. *Physics of Fluids* **7**, 266–274.

- Tackley, P. J. 1993. Effects of strongly temperature-dependent viscosity on time-dependent, three-dimensional models of mantle convection. *Geophysical Research Letters* 20(20), 2787–2190.
- Tan, E., E. Choi, P. Thoutireddy, M. Gurnis, and M. Aivazis 2006. GeoFramework: Coupling multiple models of mantle convection within a computational framework. *Geochemistry, Geophysics, Geosystems* 7(6).
- Thomas, P. 1988. Radii, shapes, and topography of the satellites of Uranus from limb coordinates. *Icarus* 73(3), 427–441.
- Tittemore, W. C., and J. Wisdom 1989. Tidal evolution of the Uranian satellites: II. An explanation of the anomalously high orbital inclination of Miranda. *Icarus* 78(1), 63–89.
- Tobie, G., O. Grasset, J. I. Lunine, A. Mocquet, and C. Sotin 2005. Titan’s internal structure inferred from a coupled thermal-orbital model. *Icarus* 175(2), 496–502.
- Trompert, R., and U. Hansen 1998. Mantle convection simulations with rheologies that generate plate-like behaviour. *Nature* 395(6703), 686–689.
- Verheylewegen, E., B. Noyelles, and A. Lemaitre 2013. A numerical exploration of Miranda’s dynamical history. *Monthly Notices of the Royal Astronomical Society* 435(2), 1776–1787.
- Zhong, S., M. T. Zuber, L. Moresi, and M. Gurnis 2000. Role of temperature-dependent viscosity and surface plates in spherical shell models of mantle convection. *Journal of Geophysical Research: Solid Earth* 105(B5), 11063–11082.

3.10 Tables and Figures

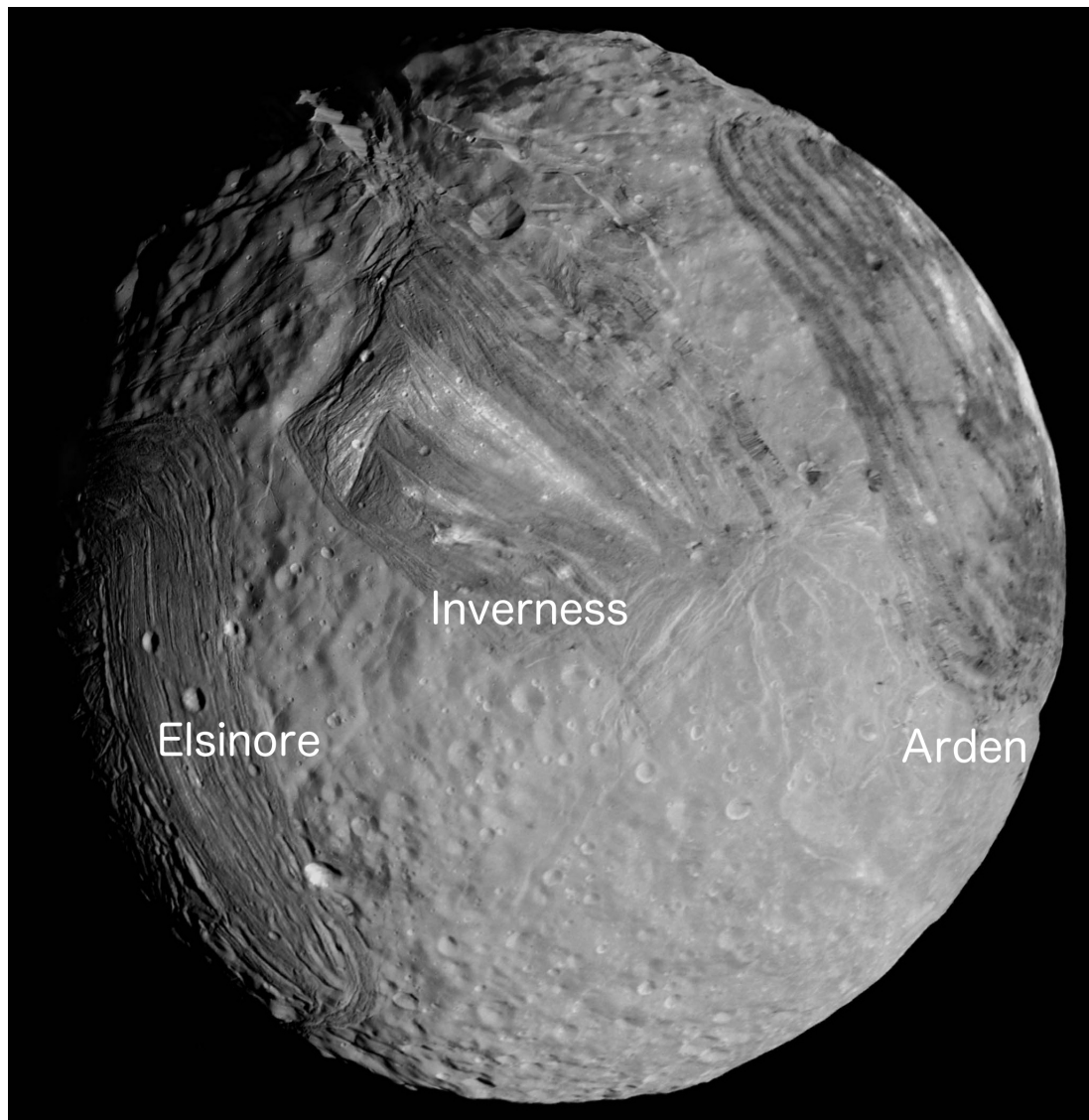


Figure 3.1: Mosaic of southern hemisphere of Miranda, the innermost regular satellite of Uranus, with radius of 236 km. Projection is orthographic, centered on the south pole. Visible from left to right are Elsinore, Inverness, and Arden coronae. Images credit: NASA/Jet Propulsion Laboratory/ Ted Stryk.

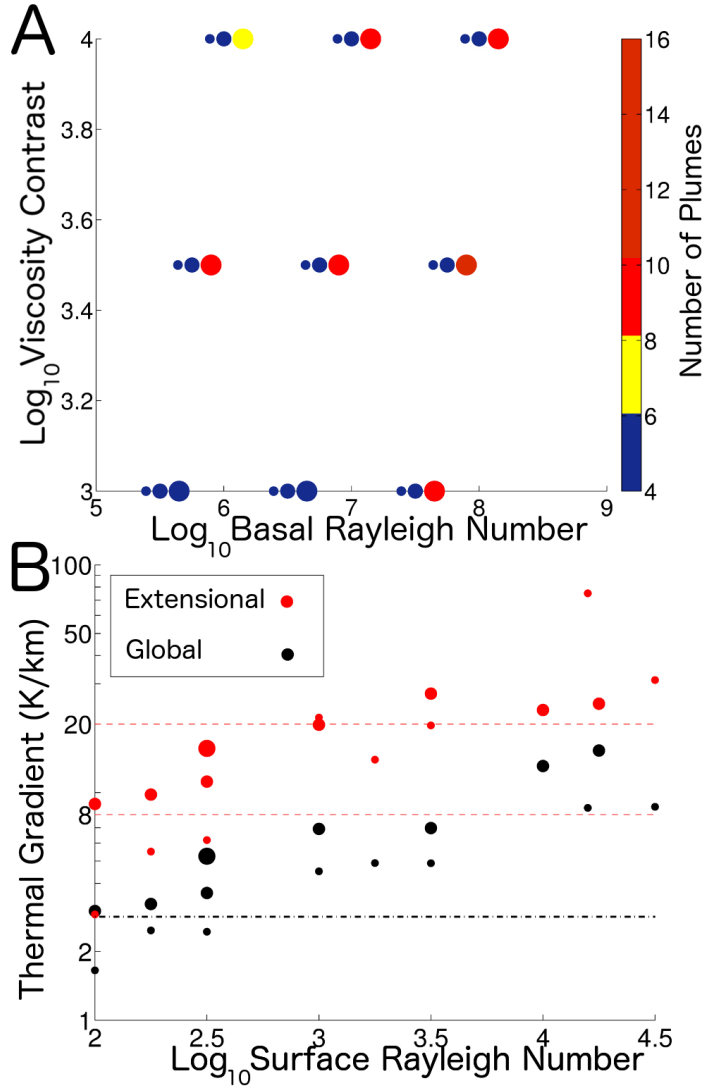


Figure 3.2: A: Convection models from our first set of simulations as a function of basal Rayleigh number and viscosity contrast. Symbol color indicates number of plumes present after convection approaches steady state. Symbol size scales with fractional size of the rocky core, $x = 0.25, 0.4$, and 0.55 . B: Surface Rayleigh number versus thermal gradient for simulations with stable low-order convection. Red points show thermal gradient in regions of extension; black points show global average. Red dashed lines indicate extensional thermal gradient implied by models of lithospheric flexure (Pappalardo et al., 1997). Black dashed line indicates that total surface power of 10 GW is necessary to provide global thermal gradient of $\sim 3 \text{ K km}^{-1}$, given a thermal conductivity of $5 \text{ W m}^{-1} \text{ K}^{-1}$.

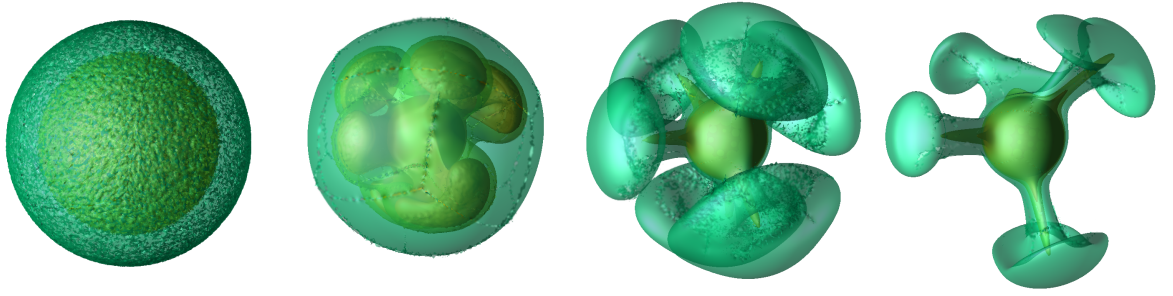


Figure 3.3: Time evolution of convection in Uranus' moon Miranda's ice shell started with random initial temperature perturbations throughout the base of the ice shell, with Rayleigh number $Ra = 10^{6.5}$, $\Delta\eta = 10^3$ (effective viscosity contrast across the ice shell), fractional core size $x = 0.25$, and basal heating. Colors represent temperature isosurfaces, with green and yellow corresponding to 120 K and 170 K, respectively. The convective upwellings merge until reaching a low-order convection pattern with ~ 5 plumes

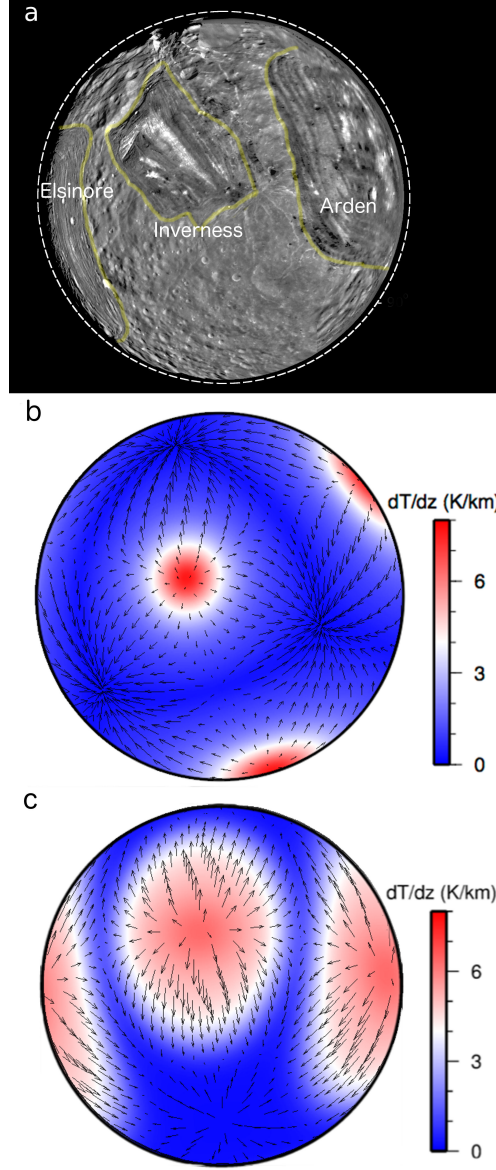


Figure 3.4: A: Mosaic of Uranus' moon Miranda's southern hemisphere with coronae outlined in yellow. Projection is orthographic centered on the south pole. Dashed white line approximates the equator. B, C: Surface conditions of convection simulations, plotted within same projection as A, with near-surface thermal gradient indicated by color, and arrows denoting surface velocity. Above convective upwellings, the thermal gradient is consistent with values predicted from flexure, and the surface is undergoing concentric surface extension. Steady-state tetrahedral convection is shown in B, with Rayleigh number $Ra = 10^{6.75}$, $\Delta\eta = 10^{3.5}$ (effective viscosity contrast across the ice shell), and fractional core size $x = 0.4$. Convection in C results from concentrated tidal heating near the equator, predicted for a satellite with no ocean (Beuthe, 2013), after a 60° rotation from the sub-Uranian point toward the south pole.

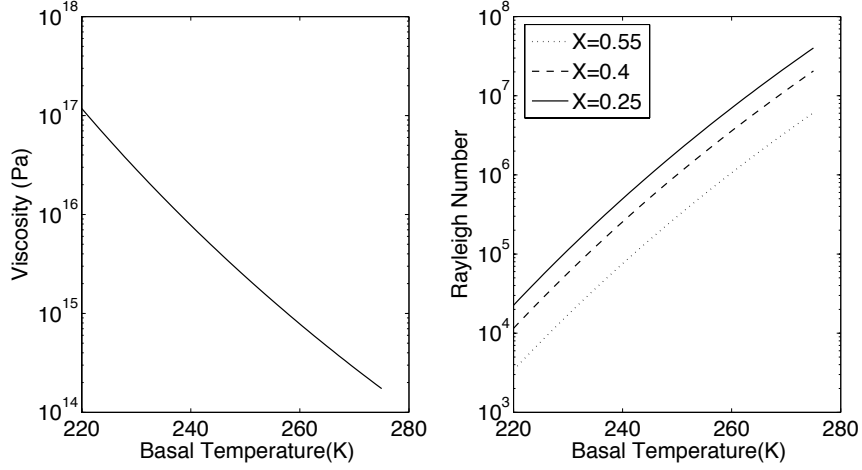


Figure 3.5: (left) Viscosity of ice deforming by Newtonian volume diffusion, for a grain size of $d=0.2$ mm, as a function of temperature. (right) Rayleigh number at the base of the ice shell as a function of basal temperature for different core sizes.

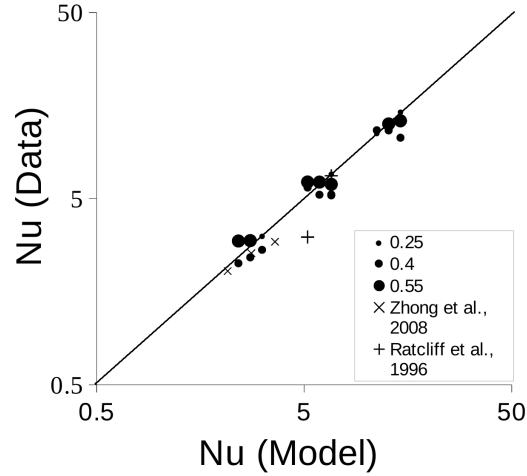


Figure 3.6: : The y-axis shows the average Nusselt numbers of our convection simulations after the Nusselt number approaches a stable value. The x-axis show the expected Nusselt number based on the scaling relationship from Barr (2008), $Nu = 0.33Ra_b^{1/3} \exp(\frac{\theta}{19})$, where $\theta = \ln \Delta\eta$, Ra_b is the Rayleigh number at the base of the ice shell, and $\Delta\eta$ is the viscosity contrast between the top and the bottom of the ice shell. Circles represent the results of this study, where circles size scales with the size of the silicate core, and plus signs and crosses represent the results of Ratcliff et al., (1996) and Zhong et al. (2008), respectively

Chapter 4 | Recent Tectonic Activity on Pluto Driven by Phase Changes in the Ice Shell

N.P. Hammond¹, A.C. Barr², E.M. Parmentier¹.

1. Department of Earth, Environmental and Planetary Science, Brown University,
Providence, RI 02912, USA

2. Planetary Science Institute, 1700 East Fort Lowell, Suite 106, Tucson, AZ
85719-2395, USA

Published in its present form in:

GRL DETAILS

4.1 Abstract

The *New Horizons* spacecraft has found evidence for geologic activity on the surface of Pluto, including extensional tectonic deformation of its water ice bedrock (see Moore *et al.* (2016)). One mechanism that could drive extensional tectonic activity is global surface expansion due to the partial freezing of an ocean. We use updated physical properties for Pluto and simulate its thermal evolution to understand the survival of a possible subsurface ocean. For thermal conductivities of rock less than $3 \text{ W m}^{-1} \text{ K}^{-1}$, an ocean forms and at least partially freezes, leading to recent extensional stresses in the ice shell. In scenarios where the ocean freezes and the ice shell is thicker than 260 km, ice II forms and causes global volume contractions. Since there is no evidence for recent compressional tectonic features, we argue that ice II has not formed and that Pluto’s ocean has likely survived to present day.

4.2 Introduction

Pluto is the largest object in the Kuiper Belt, with a radius $R = 1187 \text{ km}$ (Stern *et al.*, 2015), and a bulk density of $\bar{\rho} = 1.86 \text{ g/cm}^3$ (Stern *et al.*, 2015), suggesting a mixed ice-rock composition (McKinnon *et al.*, 1997, 2008). The *New Horizons* flyby of the Pluto-Charon system recently revealed evidence for diverse geologic activity on the surface of Pluto (Stern *et al.*, 2015). Moore *et al.* (2016) find that while much of the geologic activity seems to involve volatile ices such as nitrogen, carbon monoxide and methane (McKinnon *et al.*, 2016), there is also evidence for geologic activity in Pluto’s water ice bedrock. The many troughs and scarps that cut across Pluto’s surface are often hundreds of kilometers long and up to 4 km deep. These features have been interpreted as extensional faults (Moore *et al.*, 2016). The age of these tectonic features is unknown, though many of the graben do not appear to be heavily degraded, indicating they may be geologically young. No obvious signs of strike-slip

or compressional tectonic features have been found so far (Moore *et al.*, 2016).

What could be driving tectonic activity on Pluto? On many icy moons, tidal forces raised on the satellite by their parent planet can drive recent geologic activity (*e.g.* Hoppa *et al.* (1999); Porco *et al.* (2006)). On Pluto, however, tidal forces are presently insignificant because the Pluto/Charon system has reached the end-point of its tidal evolution, a dual-synchronous state. Another mechanism for driving tectonic activity in icy bodies is global volume change due to phase changes in the ice shell (Mueller and McKinnon, 1988). If Pluto has an ocean that is slowly freezing, it would cause global volume expansion and extensional stresses at the surface (Robuchon and Nimmo, 2011). Conversely, ice Ih converting into a denser phase, such as ice II, would lead to global volume contraction and compressional stresses (McKinnon *et al.*, 1997). The dominance of extensional tectonic features across Pluto’s surface could be explained by the freezing of a subsurface ocean. But would such an ocean have frozen billions of years ago or might it still exist today?

Previous models of Pluto’s thermal evolution by Robuchon and Nimmo (2011) found that if Pluto’s ice shell transmits heat conductively, an ocean forms and likely persists to the present. However, the persistence of a subsurface ocean may be sensitive to the thickness of Pluto’s H₂O layer, the thermal conductivity of the silicate core, and the thermal conductivity of the ice shell. Robuchon and Nimmo (2011) used a constant thermal conductivity for ice of $k_i = 2.26$ W/m/K and did not treat the formation of ice II. However, the thermal conductivity of ice is strongly temperature dependent (Petrenko and Whitworth, 1999). Higher thermal conductivities would allow the ice shell to cool more efficiently, leading to colder final states and possibly, the formation of ice II. Additionally, silicate conductivity is one of the most important parameters influencing the thermal evolution of ice-rock bodies, yet many planetary models do not provide strong justification for their chosen values (Desch *et al.*, 2009). The silicate conductivity in such models vary from 1 W/m/K (Desch *et al.*, 2009), to

4.2 W/m/K (Robuchon and Nimmo, 2011).

We use numerical models to re-examine the thermal evolution of Pluto to test the hypothesis that the partial freezing of an ocean could be driving tectonic activity. We vary the physical properties of both the silicate core and the ice shell to identify the conditions under which a subsurface ocean will survive, and conditions that lead to the formation of ice II.

4.3 Methods

4.3.1 Initial Conditions

We simulate the thermal evolution of Pluto using a one-dimensional, explicit finite-difference model,

$$\rho C_p \frac{dT}{dt} = \frac{H}{\rho} + \frac{1}{r^2} \frac{1}{dr} (r^2 k \frac{dT}{dr}), \quad (4.1)$$

where T is temperature, t is time, r is radial position, C_p is the specific heat, ρ is density, H is the heating per unit mass, and k is thermal conductivity. The value of each of these variables, explained below, depend on temperature, time, and depth. We use a node spacing of 2 km and a time step of less than 1000 years, keeping the Courant number below 0.25.

We use a constant surface temperature of $T_s = 40$ K, appropriate for the long-term average surface temperature of Pluto (Tryka *et al.*, 1994), and an initially homogeneous internal temperature T_0 , which we vary between 150 – 250 K. The initial temperature does not affect the final thermal state because radiogenic heating dominates the thermal evolution (Robuchon and Nimmo, 2011). The Pluto-Charon system likely formed in a giant impact (Canup, 2005, 2010). We assume Pluto becomes tidally locked early in its history (10 – 100 Myr after the impact) (Dobrovolskis *et al.*, 1997; Barr and Collins, 2015). We begin our simulations just after tidal evolution is complete, at $t_0 = 100$ Myr after CAI formation.

4.3.2 Silicate Core

We assume Pluto is fully differentiated into a silicate core and a water ice layer; the energy released by accretion and the decay of radioactive isotopes early in Pluto's history should be more than sufficient to melt water ice and effectively separate icy and rocky components (McKinnon *et al.*, 1997; Hussmann *et al.*, 2006; McKinnon *et al.*, 2008; Desch *et al.*, 2009; Robuchon and Nimmo, 2011). The density of silicates appropriate for Kuiper Belt objects and other ice-rock bodies is uncertain (see e.g., McKinnon *et al.* 2008 for discussion). A density of $\rho_r \approx 2.4 \text{ g cm}^{-3}$ has been inferred for the rocky core of Saturn's satellite Enceladus (Iess *et al.*, 2014), consistent with a core composed of hydrated silicates such as serpentinite. If silicates avoided hydrothermal alteration, or if they were subsequently dehydrated, Pluto's core may be composed primarily of olivine and pyroxene (McKinnon *et al.*, 2008). A reasonable upper bound for core density is $\rho_r \approx 3.5 \text{ g cm}^{-3}$, appropriate for the intrinsic density of carbonaceous chondrites (Yomogida and Matsui, 1983; Macke *et al.*, 2011). The radius of the core,

$$R_c = R \left(\frac{\bar{\rho} - \rho_i}{\rho_r - \rho_i} \right)^{1/3}, \quad (4.2)$$

where $\rho_i = 0.92 \text{ g cm}^{-3}$ is the density of ice Ih. This yields an initial ice shell thickness $D = 180$ to 340 km .

The core is heated by long lived radioisotopes ^{238}U , ^{235}U , ^{232}Th and ^{40}K . We assume an initial abundance of parent isotopes appropriate for CI chondrites (Lodders, 2003). The heating per unit mass of rock is then $H(t) = H_0 \chi \exp(-t/\tau)$, where the values for initial heating H_0 , concentration χ , and half-life τ are taken from Table 3 of Robuchon and Nimmo (2011). Heat transfer in the core occurs by conduction. The core has a specific heat, $C_{p,r} = 800 \text{ J kg}^{-1} \text{ K}^{-1}$. The thermal conductivity of the core depends on its composition, which is uncertain. Ordinary and carbonaceous chondrites have silicate conductivities $k_r = 1 - 2 \text{ W m}^{-1} \text{ K}^{-1}$ (Yomogida and Mat-

sui, 1983). The conductivity of antigorite at low pressure is $k_r = 2.4 \text{ W m}^{-1} \text{ K}^{-1}$ (Osako *et al.*, 2010), the average conductivity of the terrestrial mantle is $k_r = 3 \text{ W m}^{-1} \text{ K}^{-1}$ (Hofmeister, 1999), and olivine has a thermal conductivity of $k_r = 4 \text{ W m}^{-1} \text{ K}^{-1}$ (Osako *et al.*, 2004). Hydrothermal circulation, if present, could increase the effective thermal conductivity of the core by advecting heat toward the surface (*e.g.* Neveu *et al.* (2015)). Because the composition and thermal properties of the core are uncertain, we vary the conductivity from $k_r = 1$ to $4 \text{ W m}^{-1} \text{ K}^{-1}$ to explore how different values affect the thermal evolution.

4.3.3 Ice Shell

It is important to consider whether Pluto’s ice shell is likely to undergo solid state convection. We calculate the Rayleigh Number Ra , a dimensionless parameter that describes the vigor of convection, and compare it to the critical value for convection in the stagnant lid regime (Solomatov, 1995). The critical value for convection is defined as (Solomatov, 1995)

$$Ra_{cr} = 20.9 \left(\frac{Q^*(T_b - T_s)}{RT_i^2} \right)^4 \quad (4.3)$$

where $Q^* = 59.4 \text{ kJ/mol}$ is the activation energy for diffusion in ice (Goldsby and Kohlstedt, 2001), R is the gas constant, T_b is the temperature at the base of the ice shell and T_i is the average temperature of the convecting layer which we approximate as $T_i = T_b$ (McKinnon, 2006). We calculate the critical Rayleigh number for convection in the ice shell to be $Ra_{cr} = 7 \times 10^6$.

The Rayleigh number is defined as

$$Ra = \frac{\rho_i g \alpha (T_b - T_s) D^3}{\kappa \eta_i}, \quad (4.4)$$

where α is the coefficient of thermal expansion, $\kappa = k_i / (\rho_i C_{p,i})$ is the thermal diffu-

sivity of ice and η_i is the viscosity of the ice evaluated at T_i . Under the low stress conditions expected in the ice shell, any plastic deformation will occur via diffusion creep and the viscosity $\eta_i = Ad^2 \exp(Q/RT_i)$, where d is grain size and $A = 8.36 \times 10^9 (T_i/250 \text{ K}) \text{ Pa s/m}^2$ is an experimentally constrained constant (Goldsby and Kohlstedt, 2001; Barr, 2008). The ice grain size in convecting ice shells likely ranges from 1 – 100 mm, depending on whether grain growth is limited by pinning from impurities or by dynamic recrystallization (Barr and McKinnon, 2007). We calculate the Rayleigh number evaluating the physical properties of ice at $T_i = 250 \text{ K}$, and the effects of spherical geometry and variable conductivity on the Rayleigh number are discussed in the supplemental material (S2).

We find that the ice shell is unlikely to convect and that the Rayleigh number only exceeds the critical value for the thickest possible ice shells, $D = 340 \text{ km}$, with the smallest possible grain sizes, $d < 2 \text{ mm}$, when the viscosity of the ice $\eta_i < 6 \times 10^{16} \text{ Pa s}$. Our findings are consistent with Robuchon and Nimmo (2011) who find that the ice shell will convect only if $\eta \leq 10^{16} \text{ Pa s}$. We therefore assume heat transport in the ice shell occurs by thermal conduction and do not model convection. We use a temperature dependent specific heat $C_{p,i} = 8.8T \text{ J kg}^{-1} \text{ K}^{-2}$ (Petrenko and Whitworth, 1999) and temperature dependent thermal conductivity of $k_i = 488/T + 0.48 \text{ W/m/K}$ (Hobbs, 1974). The conductivity near the surface is likely reduced due to porosity and the presence of volatile ices such as nitrogen, therefore, in the top 10 km we use a constant thermal conductivity of $k_i = 1 \text{ W m}^{-1} \text{ K}^{-1}$ (Shoshany *et al.*, 2002).

4.3.4 Phase Changes in H₂O Layer

If the heat flow out of the core is larger than the maximum heat flow capable of moving through the ice shell, melting at the base of the ice shell can occur. The heat flow into the ocean boundary, if the boundary is at element i , is $F_{in} = F_{core}(R_c/r)^2$,

where F_{core} is the heat flow from the core and r is the radial position of the ocean boundary. The heat flow away from the boundary is $F_{out} = \left(\frac{k_i + k_{i-1}}{2}\right) \frac{T_i - T_{i-1}}{\Delta r}$. Over a time step Δt , the boundary moves a distance $\Delta z = \Delta t \frac{F_{in} - F_{out}}{\rho L}$, where L is the latent heat. When $\Delta z > \Delta r$, we consider the phase transition in that element complete. We treat the transformation between ice Ih to liquid and ice Ih to ice II in this way. We benchmarked our code against an exact solution to the classical Stefan problem (see Supplemental Figure 4.4). We have also benchmarked our code against the results of (Robuchon and Nimmo, 2011) for the case of a conductive ice shell (S2).

We assume the ocean maintains a constant temperature, T_m , with depth, because the adiabatic temperature gradient is small (Melosh *et al.*, 2004). The melting temperature of ice depends on pressure, in addition to the presence of volatiles such as salts and ammonia. A small amount of ammonia can significantly reduce the liquidus of the water-ammonia system (Kargel, 1992). Some Kuiper Belt accretion models suggest that Pluto’s ice shell might be 5 percent ammonia by mass (McKinnon *et al.*, 2008). This would reduce the liquidus to $T_m = 250$ K. We use a melting temperature that depends on pressure and ammonia concentration following Leliwa-Kopystyński *et al.* (2002) and assume a constant mass fraction of ammonia in the H_2O layer, X , between 0, 5 and 10%. As the ice-ocean boundary changes depth, the melting temperature changes due to changes in lithostatic pressure. For example, if the ocean is growing, the melting temperature increases as pressures at the ocean boundary decreases. The ocean must therefore warm up in order to reach this new melting temperature. This energy is accounted for in our model.

We also treat the phase transition between ice Ih and ice II. While ice Ih has a hexagonal crystal structure, ice II is a dense polymorph of ice with a rhombohedral structure and a density $\rho = 1.18 \text{ g cm}^{-3}$ (Hobbs, 1974). Pre-encounter interior models by McKinnon *et al.* (1997) suggest that ice II is likely only present in Pluto’s interior if Pluto is undifferentiated, because for a differentiated body, the pressure at the base

of the ice shell may not be high enough for ice II to form. However, Pluto’s more ice-rich composition revealed by *New Horizons* data suggests that ice II could form in a differentiated Pluto. We account for the temperature dependence of the ice Ih to ice II transition. The equilibrium temperature, T_{eq} , between these phases is,

$$T_{eq} = \frac{P - 14}{0.827}, \quad (4.5)$$

where P is pressure in MPa, and temperature is in units of Kelvin (Bridgman, 1912; Whalley, 1984; Leon *et al.*, 2002). As temperature drops, ice II becomes the stable phase in the ice shell. For likely pressures at the base of Pluto’s shell, ice II begins to form when temperatures drop below ~ 230 K. The conversion of ice Ih to ice II liberates ~ 1 kJ mol $^{-1}$ (Handa *et al.*, 1988). Pressure is approximated as $P = \rho_i g_z z$, where z is depth, $g_z = (GM_r)/r^2$ is the depth-dependent gravity, G is the gravitational constant and M_r is the mass below the local radius r . Gravity increases slightly from $g = 0.67$ m s $^{-2}$ at the surface to $g \approx 0.75$ m s $^{-2}$ at the base of the ice shell. McKinnon and Hofmeister (2005) noted that ice XI may be present on the surface of Pluto, but we do not treat this phase because its formation is not associated with any significant volume change (Hobbs, 1974) .

4.3.5 Global Volume Change

Phase changes in the ice shell and thermal expansion can both contribute to global volume change. At each time step, we measure the volume change in each element due to these effects and calculate the change in radius, ΔR , and the surface strain, $\epsilon = \Delta R/R$. We use a constant coefficient of thermal expansion of $\alpha = 10^{-4}$ K $^{-1}$ in the ice shell and $\alpha = 10^{-5}$ K $^{-1}$ in the core. The strain from thermal expansion is over an order of magnitude less than from H $_2$ O phase changes. The stress resulting

from global volume change is (Melosh, 1977),

$$\sigma = 2\mu\left(\frac{1+\nu}{1-\nu}\right)\epsilon, \quad (4.6)$$

where $\mu = 3.5$ GPa is the shear modulus of ice and $\nu = 0.33$ is Poisson’s ratio (Gammon *et al.*, 1983). To account for brittle failure at the surface, we use a plastic failure strength of -10 MPa in compression and 1 MPa in tension (Schulson, 1999), and the stress in the ice shell remains confined between these two values.

4.4 Results

We simulate Pluto’s thermal evolution for a wide range of silicate core properties and ice shell properties. We find that Pluto either reaches a “warm” final state or a “cold” final state depending on the thermal properties of the silicate core. Figure 4.1a shows an example of a warm final state, where an ocean forms and survives to the present day. Here, $k_r = 2.5$ W m⁻¹ K⁻¹ and $\rho_r = 3.05$ g cm⁻³. The ocean would currently be 50 km thick, beneath 250 km of ice. The simulations shown here have a nominal ammonia concentration of 5%. We find that varying the ammonia concentration in the ice shell changes the ocean thickness by 10 – 20 km (see S3).

Figure 4.1b shows the results of a simulation with $k_r = 3.5$ W m⁻¹ K⁻¹ and $\rho_r = 3.05$ g cm⁻³, where Pluto reaches a cold final state. An ocean forms early in Pluto’s history but freezes completely by $t = 4.5$ Gyr. As the last few kilometers of the ocean start to freeze, ice II begins to form. The curved temperature profile through the ice shell, which results from the temperature dependent thermal conductivity of ice, causes more of the ice shell to be at colder temperatures and therefore increases the likelihood that ice II will form.

Figure 4.2 illustrates the phase, strain, and stress evolution of Pluto’s H₂O layer for both the “warm history” and “cold history” depicted in Figure 4.1. In both cases, a

layer of ice II forms early on due to our assumed initial temperature in Pluto’s interior, $T_0 = 200$ K, but the ice II soon disappears after Pluto warms up. The growth of the ocean leads to compressional stresses, which reverse as the ocean begins to freeze after $t \sim 2$ Gyr. In the warm scenario, (Figure 4.2b), the slow freezing of the ocean from a thickness of 100 to 50 km causes an extensional surface strain of 0.3% and generates extensional surface stresses exceeding the tensile failure stress of ice. Contraction due to the cooling of the ice shell is small compared to the expansion from ocean freezing. Extensional failure occurs into the present day. In the cold scenario, however, the ocean freezes and ice II begins to form in the last 300 Myr. The formation of ice II results in a surface contraction of about 1.5% (Figure 4.2b), during which surface stresses reach the compressive failure stress.

Our results illustrate that the evolution of the H_2O layer is sensitive to the thermal properties of the core. Figure 4.3 shows how the present state of the ice shell depends on the density and conductivity of the silicate core. Silicate conductivities less than $k_r \sim 3 \text{ W m}^{-1} \text{ K}^{-1}$ allow Pluto to remain warm enough to sustain a subsurface ocean and preclude the formation of ice II. Higher silicate conductivities allow the core to lose heat more efficiently, leading to scenarios where the ocean freezes and ice II forms. However for ice shells thinner than 260 km, (corresponding to silicate densities less than 2.9 g/cm^3), the pressures at the base of the ice shell are not high enough to form ice II, therefor the ocean may freeze without forming ice II. In the case of high core density and very high conductivity, the ice II layer can reach a maximum thickness of 200 km.

4.5 Discussion

Our results show that if the ocean freezes, and if the ice shell is thicker than 260 km, then ice II will form. The recent formation of ice II would lead to global volume contraction and compressional tectonic activity at the surface. Since there is no strong

evidence of compressional tectonic activity, (Moore *et al.*, 2016), we conclude that ice II has not formed.

Ice II formation can be prevented either by keeping Pluto warm enough to sustain a subsurface ocean, or if the silicate core density is less than 2.9 g/cm^3 and the ice shell is less than 260 km thick. There are several reasons that a thicker ice shell with a subsurface ocean may be more likely than a thin ice shell. We find that the temperatures in the inner 750 km of the silicate core exceed 850 K, the temperature where low density hydrous silicates destabilize (Ulmer and Trommsdorff, 1995). Therefore at least 60% of the core by volume is likely dehydrated into olivine and pyroxene with $\rho_r = 3.5 \text{ g/cm}^3$, giving a minimum average core density of 3.05 g/cm^3 . In this case, the minimum H_2O layer thickness is $\sim 300 \text{ km}$. Additionally, the influence of volatile ices such as nitrogen and methane may be more effective at insulating the ocean than shown in our model. We assume volatiles are concentrated in the top 10 km of Pluto’s ice shell, but if methane clathrates are abundant in the entire ice shell, its thermal conductivity may be significantly reduced, (McKinnon *et al.*, 1997; Waite *et al.*, 2002), increasing the likelihood that the ocean will survive. We find that if the ice shell has a constant thermal conductivity of 3 W/m/K , a subsurface ocean survives even if the thermal conductivity of the core is high (S4). The likelihood of ocean survival further increases when considering that as the ocean begins to freeze, impurities are excluded from the ice shell and ammonia and salt concentrations in the ocean will increase, further reducing the melting temperature.

If Pluto still has an ocean, then global volume expansion would still be occurring, suggesting extensional tectonic activity could be ongoing. This is consistent with the observations of Moore *et al.* (2016) who suggest that the relative youth of tectonic features could be consistent with the partial freezing of an ocean. Additionally, tensional stresses at the surface and ocean pressurization, due to the ice shell thickening, could make cryovolcanic resurfacing more likely (Robuchon and Nimmo, 2011).

The results of our model differ from that of Robuchon and Nimmo (2011) in several important ways. First, we treat the formation of ice II, which is important because we find that if the ocean freezes then ice II is likely to form and generate large compressional strains at the surface. Secondly, we use a temperature dependent thermal conductivity in the ice shell, which causes the thermal gradient to increase with depth. We find that this increases the rate at which the ocean freezes, because heat can be transported more efficiently away from the ice-ocean boundary. Lastly, we limit the stress at Pluto’s surface to account for brittle failure. This affects the timing of when the stress state switches between compression and tension. We find that in cases where the ocean survives, ocean freezing starts generating extensional failure by 1Gyr ago and that extensional tectonics should dominate, whereas Robuchon and Nimmo (2011) find that cases with an ocean generate mostly compressional features, with minor recent extensional features.

The partial freezing of an ocean would produce globally distributed extensional faults with no preferred orientation, although heterogeneities in the ice shell or sets of pre-existing fractures could strongly influence fault orientation. Other potential mechanisms for generating tectonic features on Pluto include tidal de-spinning and polar wander (Keane and Matsuyama, 2016; Nimmo *et al.*, 2016). Tidal stresses from despinning will no longer occur after Pluto and Charon reach a dual synchronous state, which likely occurred less than 100 Myr after the Charon forming impact (Barr and Collins, 2015). Stresses from polar wander would likely create a distribution of compressional, strike-slip and extensional tectonic features with orientations that could be predicted based on the reorientation angle (Leith and McKinnon, 1996). Global surface expansion may be the best explanation for the dominance of extensional tectonic features observed on Pluto, although all of these processes could be important in determining Pluto’s complete tectonic evolution.

4.6 Conclusion

We have shown that while the survival of the ocean depends strongly on the thermal properties of the silicate core, the lack of obvious compressional tectonic features on Pluto’s surface suggests that ice II has not formed. Therefore Pluto either has an ocean today or the ice shell is thinner than 260 km. New Horizons data can be used to distinguish between these hypotheses by characterizing the global geologic stratigraphy, assessing the timing of tectonic deformation events, and searching for evidence of present day extensional tectonic activity.

4.7 Acknowledgements

We thank Francis Nimmo and an anonymous reviewer for their helpful comments which have greatly improved the paper. This work is supported by NESSF NNX13AN99H. Author Barr acknowledges support from NASA PG&G NNX15AN79G. Supporting data and information on our model are available in an SI file.

4.8 Supplemental Material

4.8.1 Stefan Problem Benchmarking

The following summarizes a solution the Stefan problem, which describes the behavior of a moving phase boundary. We have used this solution to benchmark our finite difference model. Consider a moving solid-liquid boundary at the location $z = D(t)$, where $D(t)$ is an unknown function of time, in our case describing the ice shell thickness. Boundary condition 1 is that the temperature at $z = D(t)$ is always at the melting temperature T_m . Boundary condition 2 is that the temperature at the surface $T(z = 0) = 0$. As the boundary moves, the latent heat from freezing or melting must be conducted away from the boundary at the rate it is released, as described by

the equation

$$\rho L \frac{dD}{dt} = k \frac{dT}{dz} - F_c(t). \quad (4.7)$$

Here k is conductivity in the ice shell, $\frac{dT}{dz}$ is the thermal gradient in the ice shell evaluated just above the boundary and $F_c(t)$ is the time dependent heat flow into the base of the ice shell coming from the silicate core.

If each term in equation 1 has the same dependence on time, a similarity solution can be found. To accomplish this we assume ice shell thickness

$$D(t) = 2\alpha\sqrt{\kappa t}, \quad (4.8)$$

where κ is the thermal diffusivity of ice and α is an unknown constant. We set the heat flux

$$F_c = \frac{\beta}{\sqrt{\kappa t}}, \quad (4.9)$$

where β is a constant we select to match the heat flux of Pluto. We note that this function of the core heat flux is a fair approximation for the last few billion years. To solve for $\frac{dT}{dz}$, we differentiate the general solution to the dynamic heat equation

$$T = A + B \operatorname{erf}\left(\frac{z}{2\sqrt{\kappa t}}\right) \quad (4.10)$$

and evaluate it at $z = D(t) = 2\alpha\sqrt{\kappa t}$.

Plugging all these terms into equation 1, time cancels and we can directly relate the unknown constant α to our chosen constant β ,

$$\beta = \frac{T_m k \exp(-\alpha^2)}{\sqrt{\pi} \operatorname{erf}(\alpha)} - \alpha \rho L \kappa \quad (4.11)$$

. Selecting β gives an exact solution for α and for $D(t)$. The figure below shows a comparison between our finite difference model and the analytical solution described

above. The constant β is selected such that the core heat flux at present is 2 mW m^{-2} .

4.8.2 Comparison to Robuchon and Nimmo 2011

In this section we compare our results to that of Robuchon and Nimmo (2011) (referred to as RN) who examined the thermal evolution of Pluto and the survival of the subsurface ocean. For the case of a conductive ice shell, RN found that an ocean would survive to present day and would be 165 km thick. We are able to reproduce this result when we use the same physical properties for the ice shell and the silicate core as mentioned in table 3 of their paper. Figure S2 shows the phase of the ice shell through time where the final ocean thickness is 160 km. We are thus able to reproduce their result within 5%.

There are several important differences between our model and the work of RN. Some of the choices for the physical parameters discussed below lead to surprising differences in the resulting thermal/tectonic evolution:

(1) We treat the formation of ice II and RN do not. This turns out to be very important because we find that if the ocean freezes then ice II is likely to form and generate large compressional strains at the surface. The lack of compressional features observed on Pluto therefor provide an new line of evidence that Pluto has an ocean at present day.

(2) We use a temperature dependent thermal conductivity in the ice shell $k_i = 488/T + 0.48 \text{ W/m/K}$, where as RN use a constant thermal conductivity of ice $k_i = 2.26 \text{ W/m/K}$, because they evaluate the physical properties of ice at the melting temperature. The temperature dependent conductivity causes the thermal gradient to increase with depth. This causes more of the ice shell to be at colder temperatures and actually increases the rate at which the ocean loses heat.

(3), We use a melting temperature that depends on pressure and temperature,

whereas RN use a constant melting temperature $T_m = 273$ K. We account for the energy required to change the temperature of the ocean; As the ocean grows and the pressure on the ice ocean interface decreases, the melting temperature increases and the ocean needs to warm up to continue growing. Neglecting this effect would make a 20% difference in the thickness of the ocean.

(4) We limit the stress at the surface of the ice shell to between -1 to 10 MPa to account for brittle failure and RN do not. This changes the timing of when the stress state switches between compression and tension. We find that in cases where the ocean survives, ocean freezing to starts generating extensional failure by 1Gyr ago and that extensional tectonics should dominate. In RN’s model, cases with the ocean generate mostly compressional features, with minor recent extensional features (compare their figure 2g to our figure 2b).

(5) When calculating the critical Rayleigh number, we do not account for the effect of spherical geometry. RN account for this effect and show that the critical Rayleigh number is about three times smaller. However, it is also important to account for temperature dependent thermal conductivity as the Rayleigh number is typically evaluated at $k(T_i)$, where k is conductivity and T_i is the temperature of the convecting layer. We think these are important effects to consider and have conducted an analysis that shows that, in the case of Pluto, these two effects nearly balance one another out.

As discussed by McKinnon (2006), an effective ice shell thickness should be used when calculating the Rayleigh number to account for the effect of spherical geometry and variable conductivity. The effective ice shell thickness, D' can be found by equating the conductive heat flow through a spherical ice shell with variable conductivity with the heat flow through a Cartesian ice shell with constant conductivity

(McKinnon 2006),

$$q = \left(\frac{488}{T_i} + 0.48 \right) \frac{T_b - T_s}{D'} = \frac{488}{D} \ln(T_b/T_s) (1 - D/R), \quad (4.12)$$

where T_b is the temperature at the base, T_s is surface temperature, D is the real ice shell thickness and R is the radius of the body. The effective ice shell thickness is then

$$D' = D \frac{(488/T_i + 0.5)(T_b - T_s)}{488 \ln(T_b/T_s)} \frac{1}{1 - D/R}. \quad (4.13)$$

The first fraction in this equation accounts for the effects of variable conductivity, which reduces the effective ice shell thickness by 60%, and the second fraction accounts for the effects of spherical geometry, which increases the effective thickness by 30%. Overall the effective ice shell thickness is actually 10% smaller, yielding slightly smaller Rayleigh numbers, making convection less likely.

4.8.3 Heat Transfer in the Silicate Core

We assume that heat transfer in the core occurs solely by thermal conduction, however it is important to consider whether convection is possible. To understand whether or not convection in the silicate core is likely, we calculate the Rayleigh number and critical Rayleigh number for convection. The Rayleigh number for a radiometrically heated sphere is (Weber and Machetel 1992),

$$Ra = \frac{\alpha \rho g H_0 e^{-\lambda t} r^5}{6 C_p \kappa^2 \eta} \quad (4.14)$$

where H_0 is the initial heating rate, λ is the decay rate, $r = 950$ km is the radius of the silicate core and other parameters are as previously defined. We calculate the viscosity of the silicates using a flow law for the diffusion creep of wet-olivine given by Hirth and Kohlstedt (2003), $\eta = A^{-1} f_{H_2O}^{-1} d^3 \exp(Q/RT_i)$. We use a water

fugacity of 100 MPa, an activation energy $Q = 375$ kJ/mol and a grain size of 5 mm, similar to the expected grain size in upwelling mantle plumes on Earth (Karato 1984). The viscosity is evaluated at T_i , the internal temperature of the convecting layer. For spherical convection $T_i < 0.77T_b$ where T_b is the temperature at the base (see Deschamps, Tackley and Nakagawa 2010). In our simulations, T_i reaches a maximum temperature of 1180 K after 2.5 Gyr. Under these conditions the viscosity $\eta = 1.5 \times 10^{18}$ Pa s and the Rayleigh number reaches a maximum of $Ra = 3.5 \times 10^7$.

This is just under the critical Rayleigh number for convection, $Ra_{cr} = 20.9 \frac{Q(T_b - T_s)}{RT_i^2} = 10^8$, for $T_b = 1300$ K, $T_s = 250$ K and $T_i = 1000$ K. We therefore argue that solid-state convection in Pluto's silicate core is not likely. Convection may be possible for smaller grain sizes, or if other deformation mechanisms, such as dislocation creep, accommodate the strain. We expect that even under such conditions, the convection in the silicate core would be very sluggish, would only persist for a fraction of Pluto's geologic history, and would not strongly influence the thermal evolution of the ice shell.

4.8.4 The Effect of Ammonia Concentration

It is important to consider that ammonia may be present in the ice shell since ammonia can significantly reduce the melting temperature of water ice. We treat the melting temperature as the liquidus of the H₂O-ammonia system, which is a function of ammonia concentration by mass X_{Am} and pressure P in pascals, where

$$T_m = 273.1 - bP - aP^2 - AX_{Am} - BX_{Am}^2 - CPX \quad (4.15)$$

from Leliwa-Kopystynski *et al.*, (2002) where $a = 9.6 \times 10^{-17}$ K Pa⁻², $b = 7.95 \times 10^{-8}$ K Pa⁻¹, $A = 53.8$ K, $B = 650$ K and $C = 4 \times 10^{-8}$ K Pa⁻¹.

Figure (S3) shows how the final ocean thickness and Ice II thickness changes as ammonia concentrations increases from 0% to 5% to 10%. The ocean thickness only

changes by 10 – 20 km.

We do not treat the effect of partial melt, even though the ice shell may be partially molten if it is above the solidus of the ammonia-water system is 176 K (Leliwa-Kopystynski *et al.*, 2002). Water ice with a mass concentration of 5% ammonia at $T = 200$ K would be 10% liquid by volume. Partial melt could reduce the viscosity of the ice shell, increasing the likelihood that the ice shell will convect. However, this melt might migrate out of the system. If the melt is positively buoyant (which would require the ice shell have $\sim 10\%$ rock by mass), the partial melt would migrate toward the surface where it would freeze. If it is negatively buoyant, it would migrate toward the base of the ice shell, where ammonia would concentrate and create a low temperature ocean. The relative buoyancy of the ammonia-water ocean could even change overtime as the concentration of impurities change. Therefore the effect partial melt on the thermal and tectonic evolution of Pluto are unclear.

4.8.5 Reduced conductivity in the ice shell

If the thermal conductivity of the ice shell is significantly reduced, it makes the persistence of an ocean to present day much more likely. Some models predict that high porosity in the ice shell due to impact cratering could persist up to depths of 50 km (Bierson et al. LPSC 2016 #2176). There may also be methane clathrates in the ice shell that have a thermal conductivity as low as $k = 0.5$ W/m/K. It is unclear how significant these effects may be because the structure and composition of the ice shell is uncertain. Figure S4 show the final ocean thickness if the thermal conductivity of the ice shell is a constant $k = 3$ W/m/K. An ocean survives to present day regardless of the silicate core properties. We find it unlikely that porosity and methane clathrates could reduce the thermal conductivity by this magnitude, and this calculation serves as an upper bound on how drastically these factors could influence the thermal evolution.

Using a thermal conductivity of $k_i = 651/T$ W/m/K versus $k_i = 488/T + 0.47$ W/m/K does not make a significant difference in the thermal evolution of the ice shell. Figure S5 illustrates this point.

4.9 References

- Barr, A. C. 2008. Mobile lid convection beneath Enceladus' south polar terrain. *Journal of Geophysical Research: Planets* **113**(E7).
- Barr, A. C., and G. C. Collins 2015. Tectonic activity on Pluto after the Charon-forming impact. *Icarus* **246**, 146–155.
- Barr, A. C., and W. B. McKinnon 2007. Convection in ice I shells and mantles with self-consistent grain size. *J. Geophys. Res. Planets* **112**, 2012.
- Bridgman, P. W. 1912. Water, in the liquid and five solid forms, under pressure. In *Proceedings of the American Academy of Arts and Sciences*, pp. 441–558. JSTOR.
- Canup, R. M. 2005. A giant impact origin of Pluto-Charon. *Science* **307**(5709), 546–550.
- Canup, R. M. 2010. On a giant impact origin of Charon, Nix, and Hydra. *The Astronomical Journal* **141**(2), 35.
- Desch, S. J., J. C. Cook, T. Doggett, and S. B. Porter 2009. Thermal evolution of Kuiper belt objects, with implications for cryovolcanism. *Icarus* **202**(2), 694–714.
- Dobrovolskis, A. R., S. J. Peale, and A. W. Harris 1997. Dynamics of the Pluto–Charon binary. *Pluto and Charon* **159**.
- Gammon, P., H. Kiefte, and M. Clouter 1983. Elastic constants of ice samples by Brillouin spectroscopy. *The Journal of Physical Chemistry* **87**(21), 4025–4029.
- Goldsby, D. L., and D. L. Kohlstedt 2001. Superplastic deformation of ice: Experimental observations. *J. Geophys. Res.* **106**, 11017.
- Handa, Y. P., D. Klug, and E. Whalley 1988. Energies of the phases of ice at low temperature and pressure relative to ice Ih. *Canadian journal of chemistry* **66**(4), 919–924.
- Hobbs, P. V. 1974. *Ice physics*. Oxford university press.
- Hofmeister, A. 1999. Mantle values of thermal conductivity and the geotherm from

- phonon lifetimes. *Science* 283(5408), 1699–1706.
- Hoppa, G. V., B. R. Tufts, R. Greenberg, and P. E. Geissler 1999. Formation of cycloidal features on Europa. *Science* **285**, 1899–1902.
- Hussmann, H., F. Sohl, and T. Spohn 2006. Subsurface oceans and deep interiors of medium-sized outer planet satellites and large trans-neptunian objects. *Icarus* 185(1), 258–273.
- Iess, L., D. Stevenson, M. Parisi, D. Hemingway, R. Jacobson, J. Lunine, F. Nimmo, J. Armstrong, S. Asmar, M. Ducci, et al. 2014. The gravity field and interior structure of Enceladus. *Science* 344(6179), 78–80.
- Kargel, J. S. 1992. Ammonia-water volcanism on icy satellites: Phase relations at 1 atmosphere. *Icarus* 100(2), 556–574.
- Keane, J., and I. Matsuyama 2016. Pluto Followed Its Heart: True Polar Wander of Pluto Due to the Formation and Evolution of Sputnik Planum. In *Lunar and Planetary Science Conference*, Volume 47, pp. 2348.
- Leith, A. C., and W. B. McKinnon 1996. Is there evidence for polar wander on Europa? *Icarus* 120(2), 387–398.
- Leliwa-Kopystyński, J., M. Maruyama, and T. Nakajima 2002. The water–ammonia phase diagram up to 300 MPa: Application to icy satellites. *Icarus* 159(2), 518–528.
- Leon, G. C., S. R. Romo, and V. Tchijov 2002. Thermodynamics of high-pressure ice polymorphs: ice II. *Journal of Physics and Chemistry of Solids* 63(5), 843–851.
- Lodders, K. 2003. Solar system abundances and condensation temperatures of the elements. *The Astrophysical Journal* 591(2), 1220.
- Macke, R. J., G. J. Consolmagno, and D. T. Britt 2011. Density, porosity, and magnetic susceptibility of carbonaceous chondrites. *Meteoritics & Planetary Science* 46(12), 1842–1862.
- McKinnon, W., and A. Hofmeister 2005. Ice XI on Pluto and Charon? In *Bulletin of the American Astronomical Society*, Volume 37, pp. 732.

- McKinnon, W., D. Simonelli, and G. Schubert 1997. Composition, internal structure, and thermal evolution of Pluto and Charon. *Pluto and Charon* **1**, 295.
- McKinnon, W. B. 2006. On convection in ice I shells of outer Solar System bodies, with detailed application to Callisto. *Icarus* *183*(2), 435–450.
- McKinnon, W. B., F. Nimmo, T. Wong, P. M. Schenk, O. L. White, J. Roberts, J. Moore, J. Spencer, A. Howard, O. Umurhan, et al. 2016. Convection in a volatile nitrogen-ice-rich layer drives Pluto’s geological vigour. *Nature* *534*(7605), 82–85.
- McKinnon, W. B., D. Prialnik, S. A. Stern, and A. Coradini 2008. Structure and evolution of Kuiper belt objects and dwarf planets. *The Solar System Beyond Neptune* **1**, 213–241.
- Melosh, H., A. Ekholm, A. Showman, and R. Lorenz 2004. The temperature of Europa’s subsurface water ocean. *Icarus* *168*(2), 498–502.
- Melosh, H. J. 1977. Global tectonics of a despun planet. *Icarus* *31*(2), 221–243.
- Moore, J. M., W. B. McKinnon, J. R. Spencer, A. D. Howard, P. M. Schenk, R. A. Beyer, F. Nimmo, K. N. Singer, O. M. Umurhan, O. L. White, et al. 2016. The geology of Pluto and Charon through the eyes of New Horizons. *Science* *351*(6279), 1284–1293.
- Mueller, S., and W. B. McKinnon 1988. Three-layered models of Ganymede and Callisto - Compositions, structures, and aspects of evolution. *Icarus* **76**, 437–464.
- Neveu, M., S. J. Desch, and J. C. Castillo-Rogez 2015. Core cracking and hydrothermal circulation can profoundly affect Ceres’ geophysical evolution. *Journal of Geophysical Research: Planets* *120*(2), 123–154.
- Nimmo, F., C. Bierson, D. Hamilton, J. Moore, W. McKinnon, S. Stern, L. Young, H. Weaver, C. Olkin, K. Ennico, et al. 2016. Loading, Relaxation, and Tidal Wander at Sputnik Planum, Pluto. In *Lunar and Planetary Science Conference*, Volume 47, pp. 2207.
- Osako, M., E. Ito, and A. Yoneda 2004. Simultaneous measurements of thermal

- conductivity and thermal diffusivity for garnet and olivine under high pressure. *Physics of the Earth and Planetary Interiors* **143**, 311–320.
- Osako, M., A. Yoneda, and E. Ito 2010. Thermal diffusivity, thermal conductivity and heat capacity of serpentine (antigorite) under high pressure. *Physics of the Earth and Planetary Interiors* **183**(1), 229–233.
- Petrenko, V. F., and R. W. Whitworth 1999. *Physics of ice*. OUP Oxford.
- Porco, C., P. Helfenstein, P. Thomas, A. Ingersoll, J. Wisdom, R. West, G. Neukum, T. Denk, R. Wagner, T. Roatsch, et al. 2006. Cassini observes the active south pole of Enceladus. *science* **311**(5766), 1393–1401.
- Robuchon, G., and F. Nimmo 2011. Thermal evolution of Pluto and implications for surface tectonics and a subsurface ocean. *Icarus* **216**(2), 426–439.
- Schulson, E. M. 1999. The structure and mechanical behavior of ice. *JOM journal of the Minerals, Metals and Materials Society* **51**(2), 21–27.
- Shoshany, Y., D. Prialnik, and M. Podolak 2002. Monte Carlo modeling of the thermal conductivity of porous cometary ice. *Icarus* **157**(1), 219–227.
- Solomatov, V. S. 1995. Scaling of temperature- and stress-dependent viscosity convection. *Physics of Fluids* **7**, 266–274.
- Stern, S., F. Bagenal, K. Ennico, G. Gladstone, W. Grundy, W. McKinnon, J. Moore, C. Olkin, J. Spencer, H. Weaver, et al. 2015. The Pluto system: Initial results from its exploration by New Horizons. *Science* **350**(6258), aad1815.
- Tryka, K. A., R. H. Brown, D. P. Cruikshank, T. C. Owen, T. R. Geballe, and C. Debergh 1994. Temperature of nitrogen ice on Pluto and its implications for flux measurements. *Icarus* **112**(2), 513–527.
- Ulmer, P., and V. Trommsdorff 1995. Serpentine stability to mantle depths and subduction-related magmatism. *Science* **268**(5212), 858–861.
- Waite, W., B. deMartin, S. Kirby, J. Pinkston, and C. Ruppel 2002. Thermal conductivity measurements in porous mixtures of methane hydrate and quartz sand.

Geophysical Research Letters 29(24).

Whalley, E. 1984. Energies of the phases of ice at zero temperature and pressure.

The Journal of chemical physics 81(9), 4087–4092.

Yomogida, K., and T. Matsui 1983. Physical properties of ordinary chondrites. *Journal of Geophysical Research: Solid Earth* 88(B11), 9513–9533.

4.10 Tables and Figures

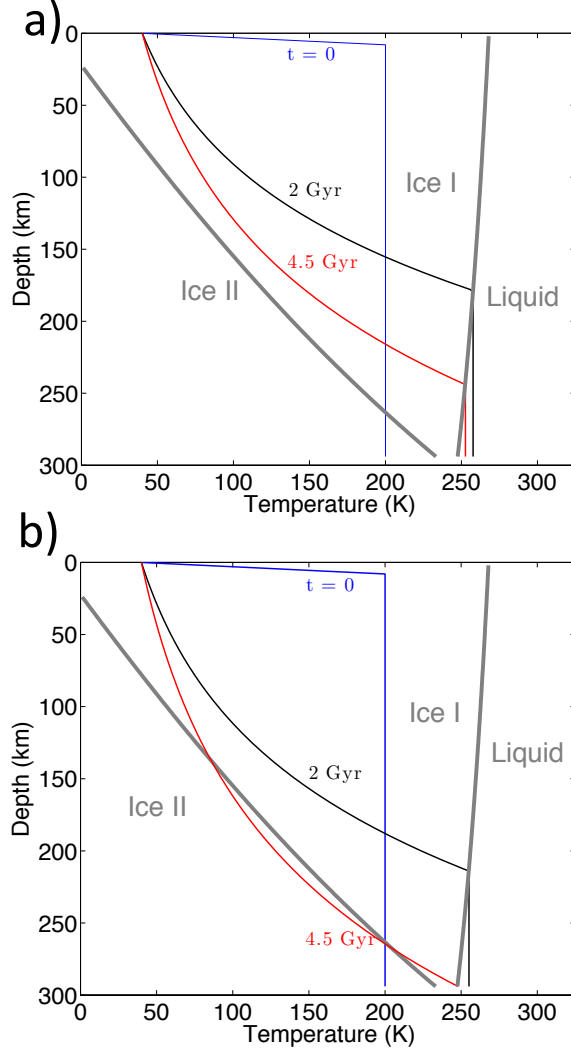


Figure 4.1: Temperature profile in the ice shell of Pluto. Blue lines show the initial temperature condition. Black lines and red lines show the temperature at 2 Gyr and 4.5 Gyr after CAI formation. Thick gray lines show the phase boundary between ice Ih and ice II, and between ice Ih and liquid. The simulations shown have a mass fraction of 5% ammonia in the ice shell. Panel (a) shows a simulation with core density $\rho_r = 3.05 \text{ g cm}^{-3}$, ice shell thickness $D = 285 \text{ km}$, and silicate thermal conductivity $k_r = 2.5 \text{ W m}^{-1} \text{ K}^{-1}$. In this case, the ocean persists to present day and no ice II forms. Panel (b) shows a simulation with $\rho_r = 3.05 \text{ g cm}^{-3}$, $D = 285 \text{ km}$ and $k_r = 3.5 \text{ W m}^{-1} \text{ K}^{-1}$. In this simulation, the ocean freezes and the base of the ice shell cools such that 100 km of ice II forms.

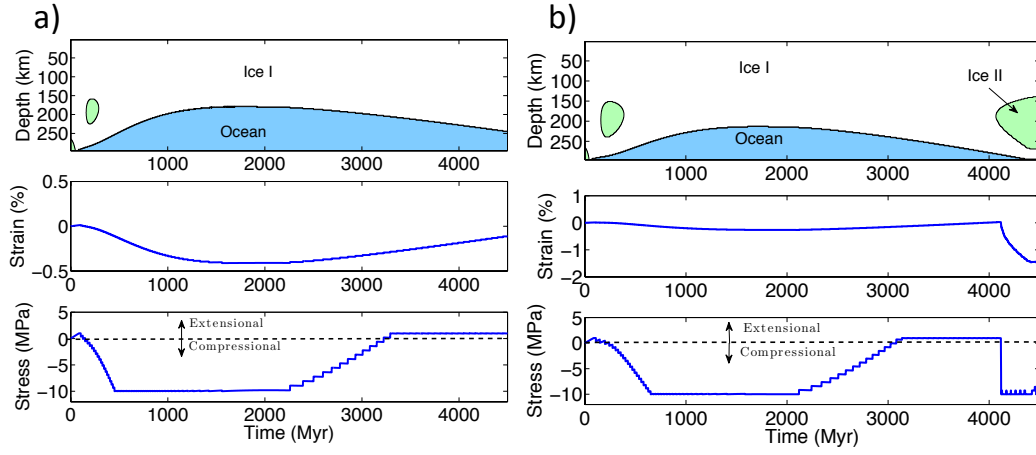


Figure 4.2: The evolution of the ice shell through time, with corresponding stress and strain at the surface, for the two cases shown in Figure 4.1, (left panels correspond to Figure 4.1a, right panels correspond to Figure 4.1b). The top panel shows a cross section of the ice shell through time, with depth on the y -axis and time after CAI formation on the x -axis. Color corresponds to H_2O phase, with white, green and blue representing ice Ih, ice II, and liquid, respectively. The second panel from the top shows the heat flow out of the core (red line) and the surface heat flux (black line). The bottom center panel shows the strain at the surface as the result of phase changes in the ice shell and thermal expansion. The bottom panel shows stress at the surface. Compressive stresses are negative and extensional are positive.

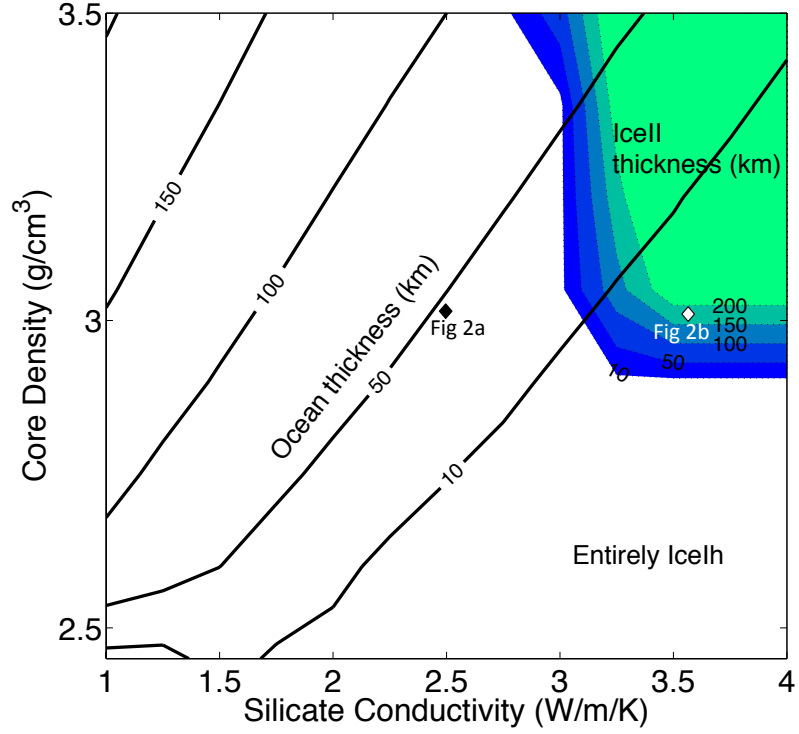


Figure 4.3: Present state of Pluto's H_2O layer, which depends on the silicate core conductivity (x -axis) and density (y -axis). We simulate Pluto's thermal evolution for more than 100 pairs of conductivity and density values. Black lines show contours for final ocean thickness. Blue and green shaded regions show the final ice II thickness. This suite of simulations used an ammonia concentration of $X = 5\%$ by mass. White and black diamonds indicate where the final state of the ice shell plots for the simulations discussed in figures 1 and 2.

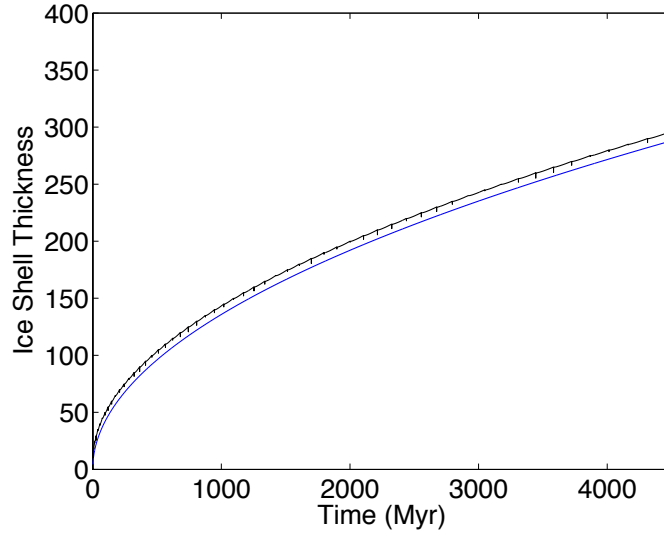


Figure 4.4: Ice Shell thickness (km) as a function of time. Black line shows the numerical solution and blue line shows the analytical solution described above.

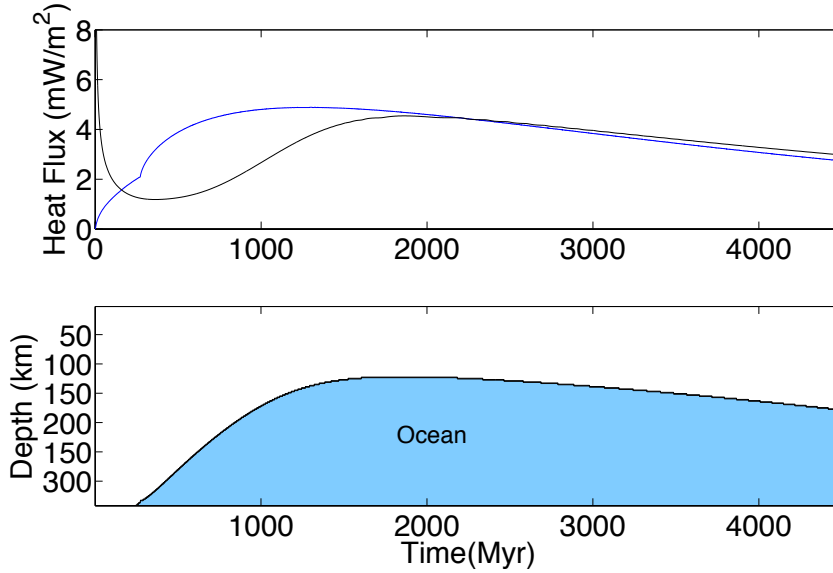


Figure 4.5: This simulations reproduces the results of Robuchon and Nimmo (2011) and uses a silicate conductivity of $k_r = 4.2$ W/m/K and a ice thermal conductivity of $k_i = 2.26$ W/m/K. The top panel shows the core heat flux (blue) and surface heat flux (black) through time. The bottom panel shows the phase of the ice shell through time.

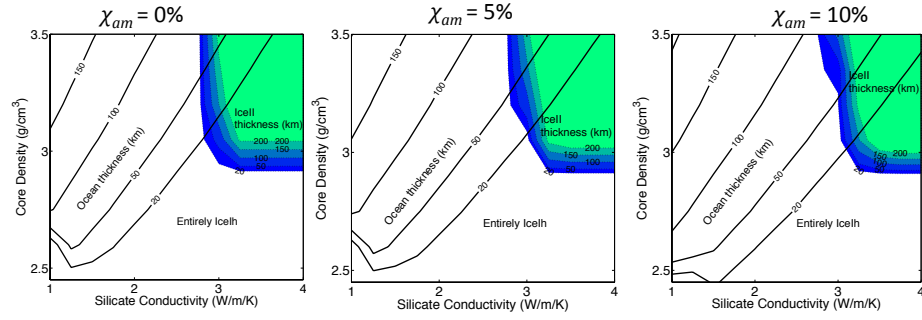


Figure 4.6: The final ocean thickness and ice II thickness as a function of silicate thermal conductivity, silicate core density and ammonia concentration in the ice shell. Black lines show ocean thickness contours and green to blue contours show ice II thickness.

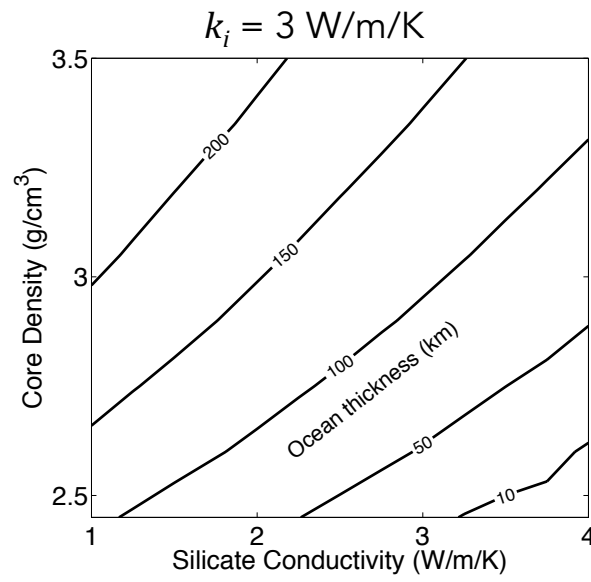


Figure 4.7: The final ocean thickness on Pluto as a function of silicate thermal conductivity and silicate core density for a constant ice shell conductivity of $k_i = 3 \text{ W/m/K}$. Black lines show ocean thickness contours.

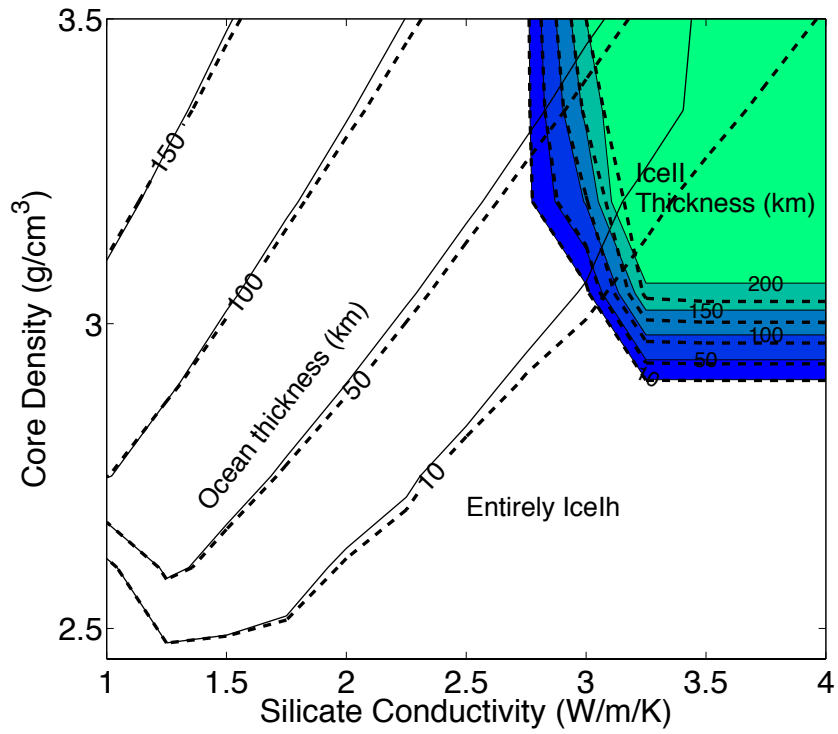


Figure 4.8: The final ocean thickness on Pluto as a function of silicate thermal conductivity and silicate core density. Black dashed contours show the results for models that used $k_i = 651/T$ W/m/K. Solid dashed lines show the results for models that used $k_i = 488/T + 0.47$ W/m/K.

Chapter 5 | Experimental Constraints on the Fatigue of Water Ice: Implications for the Strength of Icy Satellite Lithospheres

N.P. Hammond¹, A.C. Barr², R.F. Cooper¹ G. Hirth¹.

1. Department of Earth, Environmental and Planetary Science, Brown University,
Providence, RI 02912, USA

2. Planetary Science Institute, 1700 East Fort Lowell, Suite 106, Tucson, AZ
85719-2395, USA

5.1 Abstract

The surfaces of many tidally deformed icy satellites appear to be weak; tidal and convective stresses of only $\sim 10\text{--}100$ kPa in magnitude are capable of deforming their surfaces though the strength of water ice when measured in laboratory experiments is typically on the order of 1 MPa. One possible explanation for their weak behavior is that their lithospheres have become fatigued by repeated tidal stresses. Fatigue can cause materials that undergo cyclic loading to experience brittle failure at much lower stresses than for monotonic loading. To test this hypothesis, we perform a series of laboratory experiments to measure the fatigue of water ice at temperatures of $T = 198 - 233$ K and a frequency of 1 Hz. We run cyclic-indirect tensile tests on polycrystalline fresh water ice, as well as ice doped with 50 ppm H_2SO_4 , and measure the effect of cyclic loading on the brittle failure stress. We find that ice is *not* susceptible to fatigue at our experimental conditions. Samples that experienced up to 10^5 cycles at stress amplitudes of 80% the static failure stress ultimately failed at similar stresses to samples that experienced no cyclic loading. We suggest that mechanisms other than fatigue could be responsible for the apparent weak behavior of some icy satellites. However, colder temperatures and impurities in the ice shells of icy satellites may increase the likelihood of fatigue crack growth.

5.2 Introduction

Can repeated stresses from tidal deformation cause the surface of an icy satellite to become fatigued? This is the major question our paper seeks to address. We test the hypothesis that billions of cycles of tidal stress may have caused the lithospheres of some icy satellites to become progressively damaged and fatigued, allowing their surfaces to deform in response to much lower stresses. We begin by discussing the mechanics of tides, geologic evidence that icy satellite surfaces are perhaps much

weaker than expected, and previous work on the fatigue of water ice. We then describe laboratory experiments we performed to measure how cyclic loading affects the brittle yield stress of ice under various stress and temperature conditions. We apply our results to investigate how tidal deformation affects the lithospheric strength of icy satellites.

Tides play a major role in controlling the orbital and geologic evolution of icy satellites. The major icy moons of Jupiter, Saturn, Uranus and Neptune are all in slightly eccentric orbits about their parent planets and are tidally locked, meaning they constantly show the same face toward their parent planet. Their orbital periods range from $\sim 1 - 5$ days with eccentricities ranging from near zero to $e = 0.03$ (Burns *et al.*, 1986). Throughout their orbits, as they get closer to and further from their parent planet, they experience a time-varying tidal potential field that causes them to deform.

Stresses from diurnal tides are an important driver of tectonic activity on icy satellites despite being relatively weak. Laboratory experiments suggest the brittle yield stress of ice in tension is about 1 MPa (Hawkes and Mellor, 1972; Schulson, 2001). Although they are typically much lower in magnitude, tidal stresses are thought to have caused fracturing on the surface of several icy satellites. On Europa, where likely tidal stress magnitudes $\sigma_t \approx 50$ kPa (Wahr *et al.*, 2009), diurnal tidal stresses cause long, arch shaped fractures to form called cycloids (Hoppa *et al.*, 1999). Tidal models can closely match the location and morphology of cycloids but require the brittle failure stress of Europa to be 20 – 60 kPa (Hoppa *et al.*, 1999; Hurford *et al.*, 2007; Rhoden *et al.*, 2010). On Saturn’s satellite Enceladus, tidal stresses of ~ 100 kPa in magnitude regulate the timing of water vapor plumes in the geologically active South Polar Terrain (Fig 1b). This suggests that tidal stresses are initiating surface fractures that propagate down to expose a subsurface reservoir of liquid water (Nimmo *et al.*, 2014; Porco *et al.*, 2014).

Additional evidence that the surfaces of icy satellites are weak comes from models of convection-driven resurfacing (c.f Showman and Han 2005; Barr 2008; O'Neill and Nimmo 2010). Prominent geologic features on icy satellites such as Europa, Ganymede, Enceladus and Miranda are suggested to form as a result of solid-state convection in the ice I shell driving surface deformation (see fig5.1),(e.g. Pappalardo *et al.* 1997; Head *et al.* 2002; Prockter *et al.* 2002; Pappalardo and Barr 2004; Barr 2008). Numerical convection models show that solid-state convection in the ice shell cannot drive surface deformation unless that surface yield stress is less than 10 – 80 kPa (Showman and Han, 2005; Barr, 2008; O'Neill and Nimmo, 2010; Hammond and Barr, 2014a,b).

There are several hypotheses for why icy satellites surfaces appear to deform in response to such weak stresses:

- 1.) The brittle yield stress of icy satellites surfaces may have been reduced due to fatigue from repeated tidal stresses; tidal stresses would act upon populations of microcracks and cause them to slowly propagate over billions of cycles. The growth of microcracks throughout the satellite's lithosphere would progressively reduce the brittle yield strength of the surface by creating large flaws that could be exploited by weak tidal stresses to cause large-scale fractures. The critical question with fatigue is the rate of microcrack growth under various stress and temperature conditions and the rate of weakening.

- 2.) Alternatively, the brittle yield stress may remain close to 1 MPa, but stresses from convection or diurnal tides may couple with other sources of stress to overcome the lithospheric yield stress. For example, global volume expansion due to ice shell thickening could create large tensile stresses that tidal or convective stresses are superimposed on.

- 3.) Finally, other mechanisms for damaging the surface and reducing the brittle yield stress are possible. Damage from impact cratering could create large flaws near

the surface that do not heal due to the incredibly cold temperatures in the outer solar system. Additionally, large flaws may exist in the lithosphere due to the presence of frozen brines or other inhomogeneities.

Here we focus on the hypothesis that repeated loading from tidal stresses can reduce the yield stress through fatigue. Before describing our laboratory experiments designed to measure the rate of weakening in water ice, we first discuss the mechanics of fracture and fatigue and previous studies investigating the fatigue of ice.

5.3 Background

5.3.1 The Tensile Failure of Water Ice

The tensile strength of pure, polycrystalline water ice is influenced by temperature, porosity and grain size (Hawkes and Mellor, 1972; Schulson *et al.*, 1984; Schulson, 1999). For temperatures and grain sizes expected on icy satellites, likely $d = 0.1 - 10\text{mm}$ (Barr and McKinnon, 2007), the predicted brittle yield stress of ice in tension is $1 - 2\text{ MPa}$ (Litwin *et al.*, 2012). Tensional failure occurs by transgranular cleavage, and can be rate limited by either crack propagation or crack nucleation. Schulson (2001) found that for grainsizes of $d = 1 - 2\text{ mm}$ or less, fracture is rate-limited by crack propagation and that nucleated cracks can remain stable if stress is not increased. In this regime the tensional failure stress is given by

$$\sigma_{fail} \approx Kd^{-0.5} \quad (5.1)$$

where d is grain size in meters and $K = 0.044\text{ MPa}\sqrt{\text{m}}$ (Schulson, 2001).

The tensile failure of water ice can also be described in terms of the Griffith failure criterion and the mode I stress intensity K_I (c.f. Irwin 1957). The stress intensity is a parameter used to estimate the stress environment near a crack tip. The mode I

stress intensity for a flaw of length $2a$ under a far field tensile stress σ_{11} is

$$K_I = \sigma_{11}\sqrt{\pi a}. \quad (5.2)$$

Unstable fracture occurs when the stress intensity reaches a critical value called the fracture toughness K_{Ic} (Irwin, 1957). The average fracture toughness of water ice is $K_{Ic} \approx 0.14 \text{ MPa } \sqrt{\text{m}}$ (Nixon and Schulson, 1987; Rist *et al.*, 1999; Litwin *et al.*, 2012) with most experiments finding a range of values between $0.08 - 0.2 \text{ MPa}\sqrt{\text{m}}$.

Materials containing larger flaws fail at lower stresses, and for a given stress magnitude, the critical flaw size for brittle failure is

$$a_{cr} = \frac{K_{Ic}^2}{\sigma^2 \pi}. \quad (5.3)$$

Glaciers and ice sheets tend to fracture at much lower stresses than laboratory ice. Surface crevasses initiate at tensile stresses of $\sigma_t \approx 0.1 - 0.4 \text{ MPa}$ (Kehle, 1964; Vaughan, 1993; Weiss, 2004). Such low fracture strengths require near surface flaws $5 - 50 \text{ cm}$ in length to localize stress, but it is unclear how such large flaws form, since microcracks generated during plastic flow are typically no larger than a few times the grain size of a few millimeters (Nath and Vaughan, 2003). Similarly, if Europa's surface was populated with sharp flaws $\sim 5 \text{ m}$ long, it could reduce the surface yield stress to 50 kPa . How would flaws of this size be generated in glaciers and icy satellite surfaces?

Weiss (2004) suggested subcritical crack growth as the primary mechanism for the initiation surface flaws in glaciers. During subcritical crack growth, microcracks slowly propagate at stress intensities below the fracture toughness. Subcritical crack growth has been extensively studied in engineering and geologic materials (c.f. Atkinson 1984; Johnson and Paris 1968), and can occur under either static loading or cyclic loading, when it is usually referred to as fatigue crack growth. Experimental obser-

vations of crack growth velocity during fatigue are often fit to the empirical formula Paris' Law

$$\frac{da}{dt} = C \left(\Delta K_I \right)^m, \quad (5.4)$$

where t is time, ΔK_I is the change in stress intensity at the crack tip over one loading cycle and C and m are empirical constants. Although the stress intensity strongly controls fatigue crack growth rates, small changes in microstructure, temperature, and chemical environment can also have a dramatic influence (Atkinson and Meredith, 1981). For example, subcritical crack growth rates in wet quartz increase an order of magnitude in response to an increase of temperature by only 40 K (Atkinson, 1979).

Different micro-mechanical mechanisms can contribute to fatigue crack growth, including fatigue corrosion and reverse plastic slip. Fatigue corrosion (also known as stress corrosion) involves the weakening of molecular bonds near the crack tip by reactive impurities (Atkinson and Meredith, 1981). Stress corrosion is the dominant mechanism for subcritical crack growth in most geologic materials and typically involves the presence of a reactive fluid in the crack. Reverse plastic slip promotes crack growth through the interaction of dislocation with the crack tip (Davidson and Lankford, 1992).

5.3.2 Previous work on the Fatigue of Ice

Relatively few studies have examined the effect of cyclic loading on the brittle failure of ice. In general, two opposing conclusions have been found: 1) cyclic loading causes brittle failure at much lower stresses 2) cyclic loading does not cause weakening and may even increase the failure strength of ice. Previous studies used a variety of in situ and laboratory experiments with differing loading geometries, ice properties, temperatures and loading frequencies, making their results difficult to reconcile.

Several groups have conducted in situ cyclic loading experiments on sea ice (Haynes *et al.*, 1993; Kerr, 1996; Haskell *et al.*, 1996; Bond and Langhorne, 1997;

Langhorne *et al.*, 1998). These authors generated S-N failure curves which represent the results of fatigue experiments by indicating the number of cycles to failure (N) under a cyclic load of constant magnitude (S) (c.f. Suresh 1998). These studies find that the brittle yield stress of sea ice decreases with increasing number of loading cycles and may decrease by as much as 50% (Haynes *et al.*, 1993; Haskell *et al.*, 1996; Bond and Langhorne, 1997). Haskell *et al.* (1996) suggested that sea ice may exhibit a threshold stress for fatigue, where for cyclic load amplitudes below 50% of the static failure stress, sea ice can endure a practically unlimited number of loading cycles without failure. Nixon and Smith (1987) performed cyclic reversed bending tests on polycrystalline ice samples in the laboratory and also found that ice weakened with an increasing number of loading cycles. Although there is considerable scatter in their data, Nixon and Smith (1987) found that ice eventually fractured at stresses as low as 30% the static failure stress after experiencing over $> 10^6$ loading cycles.

Other fatigue studies performed on fresh water ice in the laboratory found that cyclic loading increases the brittle failure stress. Illiescu and Schulson (2002) examined the brittle compressive failure of S2 columnar ice. They found that samples that experienced a cyclic load with a progressively increasing amplitude failed at stresses 1.5 times higher than those that experienced monotonic loading. They interpreted that cyclic loads relieved stress concentrations at grain irregularities where cracks initiate through the motion of dislocations. The hypothesis that cyclic loading may suppress fracture by increasing dislocation activity has been suggested by several authors (Cole, 1990; Nixon and Smith, 1987). Cole (1990) performed fully reversible tension-compression cyclic loading on ice, and found that samples failed in tension at 1.6 MPa, significantly higher than for monotonic loading at similar conditions. Cole and Durell (1995) performed a similar experiment on saline ice and found no difference between the failure stress for monotonic and increasing cyclic loading.

Weber and Nixon (1997) also found that dislocation activity could relieve stresses

at crack tips and prevent fatigue. They performed cyclic four point bending tests on polycrystalline ice with a 5mm long pre-cut flaw at the region of maximum tensile stress. The flaw was subjected to stress intensities ranging from 50 – 90% the fracture toughness of ice and crack growth was monitored during the experiment using a traveling microscope with a laser illuminating the crack tip. They found that cracks initially grew but then stopped after growing less than 1mm. They observed striations emanating from the crack tip that they interpreted to be dislocation slip bands. They suggested that the motion of the dislocations relieved stress ahead of the crack and halted crack growth. Weber and Nixon (1997) concluded that water ice may not be susceptible to fatigue, however all of their experiments were performed at $T = 268K$, and therefore crack healing and blunting could have strongly inhibited fatigue crack growth.

5.4 Methodology

To help resolve the discrepancy between various ice fatigue studies, we performed several types of experiments to understand the fatigue of water ice at conditions relevant to the near surface of icy satellites. Our aim was to construct SN-failure curves for water ice at various temperatures and with varying impurity content. Because direct tensile loading of ice is difficult to achieve in the lab (see Cole 1990), we perform two types of indirect tensile loading tests: four point bending of rectangular bars and diametrical compression of circular discs (Brazil tests).

5.4.1 Sample Fabrication

Polycrystalline water ice samples were prepared via the “standard ice” method (Stern *et al.*, 1997). Ice made from deionized water was crushed, sieved to the desired grain size and packed into a mold. The mold was then subjected to vacuum to remove pore gases and subsequently flooded with 0°C, degassed-deionized water. The flooded mold

was then allowed to freeze slowly, from the bottom-up, over 24 hours. This method produced fully dense samples (porosity $< 5\%$). Samples for both types of mechanical tests were prepared in this manner by using different mold geometries. Cylindrical samples 2.54 cm in diameter were cut into cylindrical pucks of thickness $t = 1.1 \pm 0.1$ cm using a band saw. Rectangular samples had a grain size of $d = 1 - 1.4$ mm and cylindrical samples had a grain size of $d = 0.5 - 0.8$ mm. Grain size was measured by making surface replicas of ice samples using a liquid Formvar solution, which were photographed with a reflected light microscope (c.f. Klein-Paste and Sinha 2010).

To investigate the potential for stress corrosion cracking we performed a suite of fatigue tests on ice samples doped with 50 ppm H_2SO_4 . These samples were prepared in a manner identical to pure ice samples, except that sulfuric acid was added into the degassed-deionized water that was used to flood the mold. The sulfuric acid naturally concentrates along grain boundaries and triple junctions during freezing (c.f. Baker *et al.* 2003).

5.4.2 Four-Point Bending Test

4-point bending tests were performed to measure the fracture toughness of water ice. Samples used in these tests were $L = 20$ cm long, $w = 4$ cm wide and $h = 2$ cm height. The horizontal distances between the top and bottom loading pins was 8 cm and 16 cm, respectively. A 7.5 mm-long flaw was cut along the bottom center of the sample, parallel to the short-axis, using a band saw and sharpened with a razor blade.

Experiments were conducted in a liquid-nitrogen-cooled, circulating air cryo-chamber. All tests were conducted at -40°C . Load was applied by an Instron-1350 servo-hydraulic testing apparatus, with a fixed upper piston and load applied from below. Position and load were measured using the Instron’s internal DCDT and a 500-lb load cell.

After the sample was brought into contact with the 4-point bending pieces, posi-

tion was increased at 0.15 mm per minute, equivalent to increasing the stress intensity $5 \times 10^4 \text{ Pa } \sqrt{\text{m}} \text{ s}^{-1}$. Samples were loaded until failure and the failure load was recorded. The mode-I stress intensity on the pre-cut flaw was calculated with the formula,

$$K_I = \frac{3PL}{4wh^2} \left(C_0 + C_1X + C_2X^2 \right) \quad (5.5)$$

where P is the load X is the flaw length over the sample thickness and C_0 , C_1 and C_2 are constants derived by (Rooke and Cartwright, 1976).

5.4.3 Brazil Tests

To measure the fatigue of water ice, we perform cyclic indirect tensile tests using the Brazil test geometry. Brazil tests are most often used as a simple way to quantify the ultimate tensile strength of materials (Hawkes and Mellor, 1972). Our experimental approach is similar to Erarslan and Williams (2012) who performed cyclic Brazil tests to investigate the fatigue of Brisbane tuff.

In the Brazil test, circular disks are compressed along their diameter between curved steel plates (Fig 5.2a) (astm3967). This loading geometry generates tensile stresses in the center of the sample. Sufficient loading causes fractures to initiate in the center of the sample and propagate parallel to the loading direction. The magnitude of the tensile stresses in the sample center is

$$\sigma_t = \beta \frac{2P}{\pi Dt} \quad (5.6)$$

where D and t are sample diameter and thickness, and β is a constant of order ~ 1 (ASTM). We also calculated stresses in our sample using the finite element package Abaqus. In our numerical model we replicated our sample loading geometry in two-dimensions, with a refined mesh in areas of high stress (see supplemental material).

Brazil tests were conducted in a liquid nitrogen-cooled, ethanol-bath cryostat

capable of maintaining sample temperature within $\pm 0.5^\circ\text{C}$ (Caswell *et al.*, 2015). The cryostat is installed on a servo-mechanical Instron 1361 materials testing apparatus. This system is capable of higher-precision displacement control required to observe fatigue in our experiments.

In all tests, prior to applying cyclic loading, the load was increased to 20 lbs, (200 kPa) at a rate of 0.5 lbs/s and then held constant for 100 s. This loading procedure allowed a smooth contact to form between the sample and the steel plates and reduced stress concentrations at the contact points. This prevented fractures from initiating near the edge of the sample and promoted fracture initiation from the center.

Temperatures in icy satellite lithosphere $T_i < 150\text{K}$ and the tidal stress frequency $\omega_{\text{tidal}} \sim 10^{-5}$ are difficult to replicate in the lab. One solution to this dilemma is to trade-off frequency and temperature to create a similar amount of viscous dissipation per cycle in the experiments as occurs in the natural system. This can be approximated

$$\frac{\omega_n \eta_n}{\mu} = \frac{\omega_e \eta_e}{\mu} \quad (5.7)$$

where η is the effective viscosity, μ is the shear modulus, and the subscripts n and e represent the natural and experimental systems. We run experiments at $\omega_e = 1$ Hz and $T = 198 - 233\text{K}$, such that viscous deformation is very limited, similar to the conditions expected near the surface of an icy satellite.

5.4.4 Fatigue Tests

Two types of cyclic loading experiments were conducted using the Brazil test geometry, constant amplitude tests and increasing amplitude tests. Constant amplitude tests used a constant-amplitude sinusoidal load with a constant mean load. Each of these experiments used a different peak load, varying from 50 – 90% the load required for monotonic failure, and the minimum load was held at $L_{\text{min}} = 44.5$ N. Cyclic loading was applied until the sample fractured or until more than 10^4 cycles

were completed without failure. Some samples that did not fail in cyclic loading were subsequently deformed with a linear load ramp (44.5 N/s) to determine the failure stress. For comparison, some linear load tests were also conducted without any cyclic loading.

In increasing amplitude tests, a cyclic load ΔL was applied for N cycles. If failure did not occur, then the average load and ΔL were increased (keeping L_{min} constant) and N loading cycles were applied again. This process was repeated until the sample failed. At a given temperature, three sets of experiments were performed in which $N = 10, 100$, or 1000 cycles were applied before the load was increased.

Sample failure (crack propagation through the middle of the sample) coincides with a sharp reduction in load and a jump in vertical displacement, both of which were readily detectable in our measurements. To further characterize crack nucleation and growth, we measured acoustic emissions (AEs) and used a small snake-scope camera to monitor the sample surface during the experiments. AEs were measured throughout the experiment (at 10 kHz) with a contact ultrasonic sensor. The sensor was attached a macor rod which was in direct contact with the upper steel platen. Figure 5.3 shows the coincidence between AEs and mechanical data for a monotonic failure test, illustrating that we were able to successfully detect failure. Visual monitoring of the crack also coincided with our measured time of failure, though the camera was not operational for all experiments.

5.5 Results

5.5.1 Monotonic Failure Results

Six fracture toughness values were measured from four-point bending experiments performed at $T = 233$ K and crack loading rates of 5×10^4 Pa $\sqrt{\text{m}}$ s⁻¹. The mean value of fracture toughness measured is $K_{Ic} = 0.136$ MPa $\sqrt{\text{m}}$, with a range of

$K_{Ic} = 0.09 - 0.16$ MPa. We also measured the monotonic yield stress of water ice during Brazil tests at a loading rate of 10 lbs/s. We found the range of tensile failure stress of water ice to be $\sigma_{fail} = 0.92 - 1.7$ MPa with an average failure stress of 1.27 MPa.

5.5.2 Fatigue Results

Data from constant-amplitude tests are shown in Fig. 5.4. Most samples either fractured during the first cycle or they did not fail at all, suggesting the ice samples were not especially susceptible to fatigue. One experiment did fail after 100 cycles, but the cyclic load amplitude was quite high (1.6 MPa). Several other samples experienced cyclic stress magnitudes greater than 1 MPa for over 50000 cycles without experiencing brittle failure. AE events were detected in the first few thousand cycles but AE activity decreased with increasing loading cycles (see section 5.61).

Figure 5.5 show the stresses during three suites of increasing amplitude tests at temperatures of $T = 193$ K, and $T = 233$ K. A suite of increasing amplitude tests were also performed on ice doped with sulfuric acid. The failure stress did not decrease with increasing loading cycles in any of these experimental sets.

In figure 5.6 the failure stress of water ice is shown as a function of the number of loading cycles the sample experienced. Both the failure stress measured during increasing amplitude tests as well as the monotonic failure stress measured after constant amplitude tests are shown. Experiments on fresh water ice at 233 K and 198 K, as well as experiments run on ice doped with sulfuric acid at 233 K, all show that cyclic loading does not reduce the tensile failure stress and that cyclic loading might slightly increase the failure stress. Experiments at 233 K generally fail at slightly higher stresses than samples at 198 K, suggesting that increased plastic dissipation at warmer temperatures might increase the strength of the ice through rounding out flaws or increasing the fracture toughness of the ice. Experiments doped with sulfuric

acid fail at slightly lower stresses than pure water ice at the same temperature.

5.6 Discussion

Our fracture toughness and tensile yield stress measurements for water ice are consistent with previous work. The largest grainsize in our Brazil tests ($d = 0.8$ mm) gives a predicted tensile yield stress (from equation 1 Schulson (2001)) of 1.46 MPa, closely matching our average failure stress during monotonic loading of 1.27 MPa. The average fracture toughness value we measure $K_{Ic} = 0.136$ MPa $\sqrt{\text{m}}$, and the range is of fracture toughness values we measure, are both consistent with previous studies. (Nixon and Schulson, 1987; Rist *et al.*, 1999; Litwin *et al.*, 2012).

To compare our fracture toughness measurements with our tensile yield stress measurements during Brazil tests, we calculate the stress intensity on the starter flaws in our ice samples. At high strain rates grain boundaries can act as stress concentrators where fractures initiate (Gold, 1972), therefore we assume the flaw size and grainsize are related. Calculating the stress intensity on a grain face at our measured failure stress only gives a value $K_1 = \sigma_{fail} \sqrt{\pi d/2} = 0.04$ MPa $\sqrt{\text{m}}$, three times lower than the fracture toughness. This suggests that either the stress intensities at grain boundaries are higher than predicted from equation (2), which is appropriate for isolated flaws, or microcracks or pores larger than the grainsize act as the initial flaw where fracture initiates.

We find that ice does not weaken when cyclically loaded at $T = 193 - 233$ K with frequency of 1 Hz. In fact, cyclic loading under these conditions may have a slight strengthening effect. We analyze our acoustic emission (AE) data to understand the extent and distribution of microcracking and damage accumulation during our experiments.

5.6.1 Acoustic Emission Analysis

The mechanical data from an increasing amplitude tests is shown in figure 5.7. The stress shown is σ_{11} in the center of the sample. AE activity intensified when the maximum tensile stress reached $\sigma = 0.78$ MPa and AE events continued to increase in occurrence and amplitude as the cyclic load was increased. The sample failed once the tensile stress was increased to 1.24 MPa. Macroscopic crack formation coincided with a load drop, a large AE event and a tertiary creep signal evident in the displacement data.

The sharp increase in AE events when the tensile stress reached ~ 0.8 MPa could indicate the onset of crack nucleation in the sample. In water ice under tension with a grain size less than 1 – 2 mm, crack may nucleate at lower stresses before unstably propagating at higher stresses (Schulson, 2001). The mechanism for crack nucleation in ice is likely dislocation pile-ups at grain boundaries (Currier and Schulson, 1982; Schulson *et al.*, 1984; Cole, 1986, 1988), although other mechanisms, such as elastic anisotropy between adjacent grains, may also lead to crack nucleation (Sunder and Wu, 1990; Weiss and Schulson, 2000). For dislocations to pile up, dislocations must be generated within the grain (Frank-Read sources), before they move to the grain boundary. The critical shear stress for initiating crack nucleation by dislocation pile-ups (assuming the slip plane and crack are ideally oriented) is (Smith and Barnby, 1967; Cole, 1988),

$$\tau_{cr} = \sqrt{\frac{\pi\gamma_s G}{(2(1-\nu)l)}} \quad (5.8)$$

where $G = 2.3$ GPa is the shear modulus (Gammon *et al.*, 1983), $\gamma_s = 0.1$ J/m² is the surface energy of ice (Hobbs, 1974), and l is the characteristic distance that dislocations pile up. For $l \approx d$, and a grainsize of $d = 0.8$ mm, we estimate a critical shear stress of 0.8 MPa for the onset of crack nucleation by dislocation pile-ups, consistent with previous estimates (Schulson *et al.*, 1984; Cole, 1988).

Our theoretical calculations of the crack nucleation stress precisely matches the tensile stress in the center of the sample when AE activity increases. However, mysteriously, we do not observe microcracks in our un-failed samples, and we did not observe weakening with cyclic loading. It may be that microcracks formed by this process are similar in length to the grain size and their formation did not increase the largest flaw size available for brittle failure (Schulson *et al.*, 1984; Cole, 1986). Our lack of observations of microcracks could also be explained by the difficulty of identifying inter-granular fractures and our limited number of microstructure observations.

Alternatively, dislocation emission could be responsible for the increased AE activity above 0.8 MPa, due to reaching the critical stress for generating dislocations. Acoustic emissions associated with the motion of clusters of dislocations have been detected in ice signal crystals (Weiss *et al.*, 2001; Weiss and Marsan, 2003). Dislocations generated within the grain may not necessarily lead to crack nucleation by dislocation pile-ups if dislocations do not have enough time to accumulate at grain boundaries. We performed experiments at 1 Hz, equivalent to a cyclic strain rate of $\dot{\epsilon} \sim 10^{-4} \text{ s}^{-1}$. Cole (1988) and weiss2000grain argued that at high strain rates above 10^{-4} s^{-1} , the kinematics of dislocation pile-ups are too slow to generate sufficient stresses to nucleate cracks.

Figure 5.8 shows the cumulative number of AE events that occurred during a constant amplitude test out to 6×10^4 cycles that did not lead to failure. AE activity peaks early and continues at a slower rate after 5000 cycles. The cumulative AE rate matches nicely with the permanent vertical deformation of the sample. Our measured vertical displacement correlates well with our numerical model in which the ice is deforming by dislocation creep (see supplemental material). The correlation between AE and sample deformation supports the hypothesis that AE are associated with dislocation motions. Periodic burst of AE activity that occur later during steady state creep show a similar pattern to AE activity associated with dislocation avalanches

(Weiss *et al.*, 2001). A surprising observation is that AE activity appears to peak during the unloading phase. Figure 5.9 shows a histogram of AE events that occurred during for the same constant amplitude test is show in figure 5.8. The majority of AE events occur during the unloading phase when the load returns from the peak to the mean load. A smaller peak in AE activity correlates with peak load. For nearly all our experiments we observe that macro-fracture formation still occurs at peak load (see Figure 5.7).

5.6.2 Comparison to Other Fatigue Studies

We do not observe fatigue for our experimental conditions but we can shed light on how fracture and damage accumulation in ice vary under different conditions by comparing our results of previous ice fatigue studies. Figure 5.10 shows our S-N failure data compared to previous studies. We normalize the failure stress measured in each study by the monotonic failure stress measured in each study. Similar to our results, Iliescu and Schulson (2002) observed that cyclic loading of columnar ice led to an increase in the compressive failure stress. Although we perform indirect tensile failure tests, it is interesting to note that we do apply a compressive load to our samples and we also find a slight increase in failure stress with the number of loading cycles. In contrast, cantilever beam tests on sea ice show a weakening effect, with the failure stress being reduced by as much as 50% (Haynes *et al.*, 1993; Haskell *et al.*, 1996; Bond and Langhorne, 1997).

The presence of salts and brines in the sea ice may promote fatigue and explain the difference in behavior compared to fresh water ice. Stress corrosion requires the presence of a corrosive fluid in the crack (c.f. Atkinson and Meredith 1981; Atkinson 1984). Sodium chloride and other salts may be more reactive at crack tips than sulfuric acid, which had little effect in our experiments, making stress corrosion more active in sea ice. McCarthy *et al.* (2011) performed creep tests on ice-magnesium

sulfate mixtures at temperatures of 230 – 240 K. They observed that magnesium sulfates increased the viscosity of the samples by an order of magnitude and that semi-brittle flow occurred in magnesium-hydrate crystals. Increasing the viscosity of the ice reduces the amount of dissipation near crack tips, and semi-brittle flow involves microcrack formation. Both of these effects are likely to promote the accumulation of damage during cyclic loading. Several types of salts, including magnesium sulfate, may be present on the surface of Europa (McCord *et al.*, 1998), therefore repeating fatigue experiments on ice doped with various salts is important for understanding how fatigue might operate on the surface on Europa.

Loading geometry is important in promoting or suppressing fatigue. Our Brazil testing geometry was optimal for running many tensile loading tests, but as discussed above, the stress gradients in the sample created a distribution of damage that did not reduce the failure stress. The only experiment to show significant fatigue of fresh water ice is Nixon and Smith (1987). Their experiments used a fully reversible twisting load, high frequencies of 2.5 Hz, and large grain sizes of $d = 5 - 8$ mm. Reversible loading, (transitioning between tension and compression) could influence processes occurring near the crack tip. Fully reversible loading also prevents permanent deformation of the sample, which can affect the stress distribution even if the load is constant (see Supplemental material). The fatigue behavior of ice may change under tidal loading conditions, which alternates between tensile, shear and compressive (Helfenstein and Parmentier, 1983) and replicating these conditions would require a bi-axial apparatus.

Changes in temperature can dramatically change the fatigue behavior of rocks and minerals (Atkinson, 1984). In our experiments, ice at 233 K failed at slightly higher stresses than ice at 198 K (see figure 5.6), suggesting that viscous dissipation played a role in strengthening the ice. Higher temperatures could lead to faster rates of crack healing, crack tip blunting and dissipation near the crack tip which

effectively increases the fracture toughness (Nixon and Schulson, 1987; Greil, 2012). Weber and Nixon (1997) suggested that plastic dissipation ahead of the crack tip prevented fatigue crack growth, and observed dislocation slip lines ahead of the crack tip. To reduce the amount of plastic dissipation per cycle, we ran experiments at lower temperatures than Weber and Nixon (1997), but the temperatures are still much colder on icy satellite surfaces. To further reduce viscous dissipation per cycle, we increased the frequency relative to the tidal frequencies, but our fast cyclic strain rates of 10^{-4} may have prevented dislocation pile-ups from initiating crack nucleation (Weiss and Schulson, 2000).

Although we do not observe fatigue in our experiments, fatigue may still occur on icy satellites. Fatigue experiments on metals and ceramics show that slight changes in material properties or test conditions can produce dramatic differences in fatigue behavior (Suresh, 1998). We have explored a limited parameter space and found that ice does not fatigue at our experimental conditions. Our micromechanical understanding of fatigue crack growth in ice suggests that fatigue may be more likely on ice satellites due to the presence of corrosive salts and colder temperatures. However, it is possible that the conditions that promote fatigue in ice are not present on icy satellites, and that there is another explanation for the apparent weak behavior of icy satellites. Other potential weakening mechanisms include: 1) the generation of flaws in the ice shell from impact cratering, 2) the superposition of tidal stresses with other larger sources of tensile stress, 3) the inherent weakness of larger scale structures due to inhomogeneities. Below we discuss these possibilities and we also discuss crack healing in the ice shell.

5.6.3 Other Weakening Mechanisms

Superposing Stress Fields

Several other mechanisms may be capable of generating large stresses on icy satel-

lites that convective and diurnal tidal stresses could combine with to fracture the ice shell. These processes include global surface expansion, polar wander, obliquity tides and non-synchronous rotation of the ice shell with respect to the core (Collins *et al.*, 2009).

There is evidence that different stress mechanisms may combine to fracture the surfaces of various icy satellites. On Ganymede, weak convective stresses were likely important in the formation of grooved terrain (Hammond and Barr, 2014a), but it is possible that convective stresses combined with stresses from global surface expansion. On Europa, stresses from diurnal tides are thought to be the main driver of cycloid formation, however stresses from non-synchronous rotation could combine with diurnal tides and increase the magnitude of the cyclic stresses (Wahr *et al.*, 2009).

It remains a strong possibility that weak tidal and convective stresses are only capable of deforming the surface because they are combining with other larger sources of stress. In this case, fatigue crack growth from tidal stresses would not be necessary to explain how tides could fracture the surface. To understand how different stress mechanisms combine, more studies are required to examine how different sources of stress affect fracture patterns on icy satellites (Rhoden and Hurford, 2013).

Large-Scale Inhomogeneities

On earth, large scale features naturally form fractures at much lower stresses. For instance, Dempsey *et al.* (1999) studied the failure strength of sea ice beams over a large size range and found that 80 m long beams failed at tensile stresses as low as 40 kPa. One argument to explain this behavior is that larger scale features contain larger flaws that localize stress. Scale dependent failure models often use a Weibull flaw distribution, where the failure strength of a volume of material V goes as

$$\sigma_{fail} \sim V^{-1/12}. \quad (5.9)$$

This relation assumes that failure is controlled by the “weakest-link”, where fractures initiate from the largest available flaw (Parsons *et al.*, 1992). Over very large areas of ice the mean failure stress would approach zero, but it is unclear that this relationship is applicable at the global scale.

On Europa, diurnal tidal stresses 50 kPa in magnitude appear to cause wide spread tensile failure. If Europa’s lithosphere was full of sharp flaws 2.5 m in length, then Europa’s low brittle yield stress could be explained. Based on the size effect discussed above, one could argue that it is likely that flaws of this size would exist at the global scale. Possible mechanisms for generating such flaws would be the freezing of brine pockets, thermal expansion stresses, or damage from impact cratering.

Damage from Impact Cratering

Impact cratering can generate extensive fracturing on planetary surfaces. Most of the damage from impact cratering is localized near the surface in a thin layer of highly porous debris called regolith. On Europa, the estimated impact regolith thickness is 0.1 – 1 m (Moore *et al.*, 2009). However, impact cratering may cause fracturing at much greater depths due to impact shock waves propagating into the subsurface. Viscous relaxation or thermal contraction after the impact can also cause stress and damage in the lithosphere (Bratt *et al.*, 1985). A deeper fractured zone called the mega-regolith layer can form as a result of impact cratering. The mega-regolith on the moon is 5 – 15 km thick with 10 – 20% porosity (Besserer *et al.*, 2014; Soderblom *et al.*, 2015).

High velocity impact laboratory experiments show that a highly fractured zone can exist directly beneath impact craters and radial and concentric fractures can be found further away (Polanskey and Ahrens, 1990). Impacts into ice show that shear fractures form in response to the passage of impact shock waves (Arakawa *et al.*, 1995), and the depth of fracturing can be predicted by comparing the shock wave pressure to the shear brittle failure stress (Ahrens *et al.*, 2002). The formation of

damage during impact crater formation in silicates was studied by Ai and Ahrens (2004). They found that impact shock waves generated compressive radial stresses that resulted in relatively tensional concentric stresses, leading to the formation of radial fractures deep within the target (Ai and Ahrens, 2004).

Shock wave pressure as a function of radial distance r from the impact point can be approximated as

$$P(r) = P_0(r/r_i)^{-n} \quad (5.10)$$

where r_i is the radius of the impactor, P_0 is the peak pressure near the impact point, and n is the decay exponent. The peak contact pressure during a high velocity impact can be approximated as $P_0 \approx \rho_i v^2$ where ρ_i is the density of the impactor and v is the impact velocity (Melosh and Ivanov, 1999). For ice at $T = 150K$, the shock wave decay exponent $n = 1.5$ for low pressures and $n \approx 4$ for pressures above 300 MPa, due to energy absorbed by high pressure phase transitions in ice (Kraus *et al.*, 2011). Further away from the crater, the shock pressures decay elastically, with a decay exponent of $n = 2$.

Figure 5.11 shows how the maximum shock pressure changed with depth generated from a comet impacting at a velocity of $v = 5$ km/s with an impactor diameter of $r_i = 300$ m. On Europa, such an impact would result in a crater with a diameter $D \approx 10$ km. Shock would be capable of causing shear failure to a depth of 3 km and tensile failure to a depth of 4 km. In contrast, the formation of a 1 km diameter crater could cause fracturing to a depth of 300 m, only a ~ 50 m below the crater floor.

Our simplified analysis suggest that large impact craters could generate a significant amount of subsurface damage and that smaller impactors are unlikely to generate much damage other than directly beneath the crater. Could damage from impact cratering explain why icy satellites such as Europa appear to fracture in response to weak tidal stresses? Although stresses from 10 km diameter impacts could cause signifi-

cant damage, not many larger craters exist on Europa's surface. Therefore recent impact cratering is unlikely to generate widespread and deep lithospheric damage on Europa. However, its possible that damage from an earlier more intense period of impact cratering could have remained in the lithosphere. As discussed below, cracks and damage may not heal at the cold surface temperatures of icy satellites.

5.6.4 Crack Healing

How quickly will microcracks and macrocracks heal in an ice shell? Several processes could contribute to the removal of cracks in an ice shell including elastic compaction, bulk viscous strains from convective flow, the refreezing of melt in the crack and surface energy reduction accommodated by diffusion. Elastic compaction is not very efficient at closing pore space on icy satellites due to low gravity and low pressure gradients. Durham *et al.* (2005) showed that ice could sustain a porosity of over 10% at pressures up to 150 MPa. Viscous strains from convection would likely efficiently remove cracks. The question then becomes, how deep in the lithosphere of the ice shell could cracks be present and how quickly would they heal due to diffusion.

There are surprisingly few studies of crack healing in ice. Barrette and Jordaan (2002) studied crack healing in icebergs, but these cracks began healing by being refilled with melt that refroze. This type of process may only occur at the base of the ice shell, but could also occur below Chaos terrain on Europa where near surface melt is expected (Greenberg *et al.*, 1999). Colbeck (1986) discussed a theory for microcrack healing in ice driven by vapor diffusion but we focus on crack healing driven by crack tip radius curvature limited by volume diffusion. We know of one study that observed the healing rates of microcracks in ice (Cole, 1986) and although their data is limited our crack healing model is consistent with their observations (see Figure 5.16).

Observations of crack healing in quartz, calcite and polycrystalline ceramics show that as cracks heal, the crack tip retreats and the crack may be replaced by a series of

aligned spherical pores (Hickman and Evans, 1987; Brantley *et al.*, 1990; Greil, 2012). Models of crack tip retreat driven by surface energy reduction can closely match the rates of crack healing observed in laboratory experiments (Greil, 2012).

For simplicity, imagine a straight crack with an initial length a , and a constant crack tip radius r . The driving force for crack retreat is proportional to the curvature of the crack tip radius, $F_b = \gamma/r$, where $\gamma = 0.1 \text{ J/m}^2$ is the surface energy of ice (Hobbs, 1974). The velocity of the crack boundary is $V_b = F_b M_b$ (e.g. Evans *et al.*, 2001), where M_b is the mobility of the crack tip. In the case where crack healing is accommodated by volume diffusion, the mobility

$$M_b = \frac{\Omega D_g}{\delta k_{bz} T} \quad (5.11)$$

where Ω is molar volume, k_{bz} is Boltzmann's constant, δ is the width of the crack boundary and D_g is the diffusion rate through the ice. From Ramzier (1967), $D_g = D_o \exp((-Q)/RT)$ and D_o is constrained from experiments by Goldsby and Kohlstedt (2001). Since we are assuming that crack tip retreat is controlled by the same physical process that controls grain growth (driven by surface curvature and accommodated by diffusion), we can verify our estimate of M_b by comparing to measured values of the parabolic grain growth constant k_{gr} (see Evans *et al.* 2001). We calibrated our estimate of M_b as a function of temperature to various values of k_{gr} estimated for glaciers at different temperatures (see Table 3.1 in (Cuffey and Paterson, 2010)).

Assuming the crack tip radius remains constant with time, the crack healing rate can be written as:

$$\frac{da}{dt} = \frac{\gamma}{r} \frac{\Omega D_o}{\delta k_{bz} T} \exp(-Q/RT) \quad (5.12)$$

Figure 5.12 shows how long it would take to heal a crack $a = 1 \text{ m}$ long for a given temperature. Our simple analysis suggests that in order to completely heal a 1 m long crack after 1 Myr, for a crack tip radius of $r = 0.01 - 0.1 \text{ mm}$, temperatures

of $T > 170 - 200$ K would be required. This suggests that in the near surface of icy satellites where $T < 170$ K, cracks could remain present for long time periods without healing due to diffusion. Below this depth, there could still be fractures and microcracks but they could potentially heal over geologic timescales.

5.7 Conclusion

Although cyclic loading from tidal stresses could increase damage and reduce the yield stress of icy satellites, our experiments show that ice is not susceptible to fatigue at our experimental conditions. More experiments are required to better understand subcritical crack growth and damage accumulation at temperatures, frequencies and chemical environments that more closely approach the conditions in icy satellite lithospheres. There are several other explanations for why some icy satellite surfaces appear weak, including damage from impact cratering, naturally large flaws due to the large scale of icy satellite surfaces and the coupling of weak stresses with other sources of stress. Each of these mechanisms remains equally likely for various satellites and it is difficult to determine which might be more important. Our calculated crack healing rates in the ice shell show that in the lithosphere where $T < 180$ K, large cracks could remain present for millions to billions of years. This suggests that the lithosphere of Europa could support a high density of fractures down to depths of 1 – 5 km, depending on the thermal gradient. Fatigue in water ice remains a poorly understood yet highly important subject, since cyclic loading of ice occurs not only on icy moons but on ice shelves on Earth as well. Future experiments on ice fatigue with more detailed observations of microcrack behavior will enable us to better understand and predict icy satellite tectonics and ice rift propagation.

5.8 Supplemental Material

5.8.1 Subcritical Crack Growth Mechanisms

Fatigue crack propagation can occur by a number of mechanisms that operate near the crack tip. The rate of crack propagation is a function of stress intensity and often varies with many material parameters including temperature.

Stress Corrosion Cracking

Stress corrosion is defined as environment-induced subcritical crack growth under sustained stress (Anderson and Grew 1977). It is thought to occur when a chemical agent, such as water, reacts with strained atomic bonds near a crack tip, thereby weakening the bonds and allowing the crack to propagate. When loading is periodic, the process is called corrosion fatigue. Stress corrosion is extremely well studied in metals and ceramics and has also been observed in rocks and minerals such as quartz (Atkinson 1984). Crack growth during stress corrosion can be rate limited by chemical reaction rates near the crack tip (referred to as reaction limited) or by the diffusion rate of chemicals toward the crack tip (diffusion limited). Stress corrosion is the dominant mechanism of subcritical crack growth in geologic materials (Atkinson 1984).

Thermal Barrier Propagation

Lawn (1975) proposed that in the absence of chemical agents to weaken bonds, cracks may grow by thermally jumping through energy barriers in the lattice. His theory proposes that for a crack under sufficient tension, the energy barrier for crack growth is lower than for crack healing, thus thermal perturbations will lengthen the crack over time. This mechanism has proven difficult to validate with experiments, however subcritical crack growth has been observed in glass under vacuum, and thermal barrier propagation has been proposed to explain this behavior (Atkinson 1984).

Plastic Slip

Fatigue crack growth in ductile solids may occur by a duplex slip mechanism, where flow occurs along two perpendicular slip planes which intersect the crack tip at 45° to the direction of crack growth (Forsyth 1962). During the tensional phase of the load cycle, the crack opens and motion along the slip plane causes the crack tip to advance. During the compressional phase, the crack closes but motion along the slip plane does not entirely reverse due to the irreversibility of plastic slip (Laird 1967, Suresh 1999). Thus crack may grow slightly each cycle due to cyclic motion along the slip bands and a crack can continue to grow over subsequent cycles by activating more slip planes. Fatigue crack growth by plastic slip commonly causes fatigue striations, or ductile striations, to form. Fatigue striations typically occur 45° to the direction of crack growth and are thought to be dislocation slip bands. Fatigue striations are very commonly observed steel, and other aluminum alloys.

In many materials, growth by plastic slip is thought to be the dominant mechanism limiting fatigue crack growth and the spacing of fatigue striations is thought to correlate well with fatigue crack growth velocity (each striation may form over one loading cycle). However in many ceramic materials fatigue striations are not observed. In ice, fatigue striations have been observed (Nixon and Smith 1987), but the relationship between crack growth and ductile striations in ice is not well established. For instance, Weber and Nixon (1998) suggested that plastic slip near the crack tip served to retard fatigue crack growth in ice, rather than facilitate it. Indeed, when plastic deformation near the crack tip becomes too large, cracks are unable to grow because stress near their tips are relieved. Schulson (2001) calculated a brittle-ductile transition in ice based on the relative size of the plastic deformation zone near a crack tip and the size of the initial crack.

5.8.2 Numerical Model

We use the numerical model ABAQUS to simulate the loading conditions during our Brazil tests to understand the stress state in our samples. ABAQUS is a finite element software package used for many engineering applications (Hibbett *et al.*, 1998). We replicate our loading conditions in 2-D simulating the contact between the curved steel loading plates and the circular ice sample. The contact between the ice and the steel is initially only one element wide (less than 1 degree) but the contact area is allowed to increase as loading progresses. We kept the bottom steel plate fixed and applied a constant load to the top steel piece. We increased the element density near the contact points and near the center of the ice sample where the stresses are high.

We assume the steel is elastic, with a Young's Modulus $E_{st} = 210$ GPa and we model the ice as viscoelastic with $E = 9$ GPa and $\nu = 0.3$ (Gammon *et al.*, 1983). For the stresses, temperatures and grainsizes in our ice samples, the dominant creep mechanism is expected to be grain boundary sliding or dislocation creep (Goldsby and Kohlstedt, 2001; Barr and Showman, 2009). The dominant creep mechanism also may vary in different parts of the sample experiencing different stresses. We tested both grain boundary sliding and dislocation creep independently, using the parameters given by Goldsby and Kohlstedt (2001) with a grainsize of 0.7 mm and a temperature of $T = 233$ K.

Figure 5.9 shows a comparison between the vertical strain (defined as the vertical displacement at the top contact point divided by the sample diameter) for our experiments and our numerical model at a load of $L=255$ N. We used both grain boundary sliding and dislocation creep and found a good match between the model and data, although primary creep is not captured in our model since we treat the ice as a Maxwell viscoelastic solid. When we use a purely elastic rheology for the ice we get an excellent match to analytical solutions for the stress distribution in the sample (see figure 5.2) (Hawkes and Mellor, 1972).

One important effect we investigate with our model is that the relationship between load and tensile stress in the sample may change as vertical displacement increases and the contact area between the ice and steel gets wider. This may be significant for our experiments because as the ice sample permanently deforms during cyclic loading the max cyclic stress in the center of the sample may reduce with time. Wang and Cao (2015) found that the tensile stress in the sample is a function of the contact angle α as well as Poisson's ratio such that

$$\sigma_t = (1.06 + 0.21\nu - 0.01\alpha) \frac{2P}{\pi Dt} \quad (5.13)$$

where α is in degrees, hence increasing the contact area from 10 to 30 degrees reduces the tensile stress magnitude by about 20%.

We ran creep tests in our numerical model to see how the contract area and stress changed with time. We ran tests for 10^5 seconds at loads of 450N, equivalent to a tensile stress of ≈ 1.1 MPa in the elastic case. The total amount of vertical strain that occurred during the simulation was 5% and the contact area increased from 2 to 15 degrees. Figure 5.13 shows the tensile stress in the sample after $\epsilon_v = 0.1, 1.0, 3.0$ and 5.0% and figure 5.14 shows the tensile stress as a function of time. In this case we assumed grain boundary sliding. The elastic stress of 1.1 MPa is achieved after only $\epsilon_v = 0.1$. At a strain of 1.0% the stress has actually increased to 1.4 MPa. This may be because the increased deformation leads to greater horizontal strain before stress can be relaxed by grain boundary sliding. However, after 3 and 5% strain, the tensile stress decreases down to 1.2 MPa because the contact area begins getting wider. This latter effect makes sense but it seems odd that the stress initially increases above the elastic solution. That behavior may disappear if we used a more realistic viscoelastic material and a finer mesh near the contact area, but if vertical strain increases with out increasing the contact area, then the tensile stress may indeed increase.

What do these results imply about our fatigue experiments? The vertical strain is less than 5% for all our experiments. We calculated vertical strain by measuring vertical displacement during the experiment and also by imaging the samples after the experiment. A circle and ellipsoid were manually fit to the image and the difference in radius at the contact point was measured (see figure 5.15). Since the vertical strain remains low, over all we expect that the stress in the samples is not less than $\sim 10\%$ of elastic solution. A small reduction in the stress with increased strain could help explain why we see a slight apparent increase in tensile yield stress with previous cyclic loading. However, that does not change the fact that no fatigue weakening occurred during our experiments.

5.8.3 Crack Healing Comparison

(Cole, 1986) performed the only observations of ice microcrack healing rates that we are aware of. He ran uniaxial creep tests on polycrystalline ice with a grain size of 1.4 – 6 mm at stresses of ~ 2 MPa and at temperatures of 268 K. He observed that microcracks were initially about half the grain diameter on average and made observations of the crack lengths over an 8-week period. Microcracks became shorter and more rounded with time until they eventually evolved into spherical pores (Cole, 1986). A comparison between his data and our crack healing model (figure 5.16) shows they are fairly consistent, although it is clear that a linear trend is not a good match for their data. We suggest that if the crack tip radius is getting larger with time it could explain why the crack healing rates decrease with time.

5.9 References

- Ahrens, T. J., K. Xia, D. Coker, M. D. Furnish, N. N. Thadhani, and Y. Horie 2002. Depth of cracking beneath impact craters: New constraint for impact velocity. In *AIP Conference Proceedings*, Volume 620, pp. 1393–1396. AIP.
- Ai, H.-A., and T. J. Ahrens 2004. Dynamic tensile strength of terrestrial rocks and application to impact cratering. *Meteoritics & Planetary Science* 39(2), 233–246.
- Arakawa, M., N. Maeno, and M. Higa 1995. Direct observations of growing cracks in ice. *Journal of Geophysical Research: Planets* 100(E4), 7539–7547.
- ASTM, D. 3967, Standard Test Method for Splitting Tensile Strength of Intact Rock Core Specimens. *ASTM International, West Conshohocken, PA*.
- Atkinson, B. K. 1979. A fracture mechanics study of subcritical tensile cracking of quartz in wet environments. *Pure and Applied Geophysics* 117(5), 1011–1024.
- Atkinson, B. K. 1984. Subcritical crack growth in geological materials. *Journal of Geophysical Research: Solid Earth* 89(B6), 4077–4114.
- Atkinson, B. K., and P. G. Meredith 1981. Stress corrosion cracking of quartz: a note on the influence of chemical environment. *Tectonophysics* 77(1-2), T1–T11.
- Baker, I., D. Cullen, and D. Iliescu 2003. The microstructural location of impurities in ice. *Canadian Journal of Physics* 81(1-2), 1–9.
- Barr, A. C. 2008. Mobile lid convection beneath Enceladus’ south polar terrain. *Journal of Geophysical Research: Planets* 113(E7).
- Barr, A. C., and W. B. McKinnon 2007. Convection in ice I shells and mantles with self-consistent grain size. *J. Geophys. Res. Planets* 112, 2012.
- Barr, A. C., and A. P. Showman 2009. Heat transfer in Europa’s icy shell. *Europa*, edited by RT Pappalardo, WB McKinnon, and KK Khurana, University of Arizona Press, Tucson, 405–430.
- Barrette, P. D., and I. J. Jordaan 2002. Healed cracks in iceberg ice. *Journal of*

- Glaciology* 48(163), 587–591.
- Besserer, J., F. Nimmo, M. A. Wieczorek, R. C. Weber, W. S. Kiefer, P. J. McGovern, J. C. Andrews-Hanna, D. E. Smith, and M. T. Zuber 2014. GRAIL gravity constraints on the vertical and lateral density structure of the lunar crust. *Geophysical Research Letters* 41(16), 5771–5777.
- Bond, P. E., and P. J. Langhorne 1997. Fatigue behavior of cantilever beams of saline ice. *Journal of cold regions engineering* 11(2), 99–112.
- Brantley, S. L., B. Evans, S. H. Hickman, and D. A. Crerar 1990. Healing of microcracks in quartz: Implications for fluid flow. *Geology* 18(2), 136–139.
- Bratt, S. R., S. C. Solomon, and J. W. Head 1985. The evolution of impact basins: Cooling, subsidence, and thermal stress. *Journal of Geophysical Research: Solid Earth* 90(B14), 12415–12433.
- Burns, J. A., M. S. Matthews, et al. 1986. *Satellites*. Number 77. University of Arizona Press.
- Caswell, T. E., R. F. Cooper, and D. L. Goldsby 2015. The constant-hardness creep compliance of polycrystalline ice. *Geophysical Research Letters* 42(15), 6261–6268.
- Colbeck, S. 1986. Theory of microfracture healing in ice. *Acta metallurgica* 34(1), 89–95.
- Cole, D. 1988. Crack nucleation in polycrystalline ice. *Cold Regions Science and Technology* 15(1), 79–87.
- Cole, D. M. 1986. Effect of grain size on the internal fracturing of polycrystalline ice.
- Cole, D. M. 1990. Reversed direct-stress testing of ice: Initial experimental results and analysis. *Cold regions science and technology* 18(3), 303–321.
- Cole, D. M., and G. D. Durell 1995. The cyclic loading of saline ice. *Philosophical Magazine A* 72(1), 209–229.
- Collins, G. C., W. B. McKinnon, J. M. Moore, F. Nimmo, R. T. Pappalardo, L. M. Prockter, and P. M. Schenk 2009. Tectonics of the outer planet satellites. *Planetary*

- Tectonics* **11**, 264.
- Cuffey, K. M., and W. S. B. Paterson 2010. *The physics of glaciers*. Academic Press.
- Currier, J., and E. Schulson 1982. The tensile strength of ice as a function of grain size. *Acta Metallurgica* **30**(8), 1511–1514.
- Davidson, D., and J. Lankford 1992. Fatigue crack growth in metals and alloys: mechanisms and micromechanics. *International Materials Reviews* **37**(1), 45–76.
- Dempsey, J., R. Adamson, and S. Mulmule 1999. Scale effects on the in-situ tensile strength and fracture of ice. Part II: First-year sea ice at Resolute, NWT. *International journal of fracture* **95**(1-4), 347.
- Durham, W. B., W. B. McKinnon, and L. A. Stern 2005. Cold compaction of water ice. *Geophysical research letters* **32**(18).
- Erarslan, N., and D. Williams 2012. Investigating the effect of cyclic loading on the indirect tensile strength of rocks. *Rock mechanics and rock engineering* **45**(3), 327–340.
- Evans, B., J. Renner, and G. Hirth 2001. A few remarks on the kinetics of static grain growth in rocks. *International Journal of Earth Sciences* **90**(1), 88–103.
- Gammon, P., H. Kiefte, and M. Clouter 1983. Elastic constants of ice samples by Brillouin spectroscopy. *The Journal of Physical Chemistry* **87**(21), 4025–4029.
- Gold, L. W. 1972. The process of failure of columnar-grained ice. *Philosophical Magazine* **26**(2), 311–328.
- Goldsby, D. L., and D. L. Kohlstedt 2001. Superplastic deformation of ice: Experimental observations. *J. Geophys. Res.* **106**, 11017.
- Greenberg, R., G. V. Hoppa, B. R. Tufts, P. Geissler, J. Riley, and S. Kadel 1999. Chaos on Europa. *Icarus* **141**, 263–286.
- Greil, P. 2012. Generic principles of crack-healing ceramics. *Journal of Advanced Ceramics* **1**(4), 249–267.
- Hammond, N. P., and A. C. Barr 2014a. Formation of Ganymede’s grooved terrain

- by convection-driven resurfacing. *Icarus* **227**, 206–209.
- Hammond, N. P., and A. C. Barr 2014b. Global resurfacing of Uranus’s moon Miranda by convection. *Geology* *42*(11), 931–934.
- Haskell, T., W. Robinson, and P. Langhorne 1996. Preliminary results from fatigue tests on in situ sea ice beams. *Cold regions science and technology* *24*(2), 167–176.
- Hawkes, I., and M. Mellor 1972. Deformation and fracture of ice under uniaxial stress. *Journal of Glaciology* *11*(61), 103–131.
- Haynes, F. D., A. D. Kerr, and C. R. Martinson 1993. Effect of fatigue on the bearing capacity of floating ice sheets. *Cold regions science and technology* *21*(3), 257–263.
- Head, J., R. Pappalardo, G. Collins, M. J. S. Belton, B. Giese, R. Wagner, H. Brene-man, N. Spaun, B. Nixon, G. Neukum, and J. Moore 2002. Evidence for Europa-like tectonic resurfacing styles on Ganymede. *Geophys. Res. Lett.* **29**, 2151.
- Helfenstein, P., and E. Parmentier 1983. Patterns of fracture and tidal stresses on Europa. *Icarus* *53*(3), 415–430.
- Hibbitt, Karlsson, and Sorensen 1998. *ABAQUS/standard: User’s Manual*, Volume 1. Hibbitt, Karlsson & Sorensen.
- Hickman, S. H., and B. Evans 1987. Influence of geometry upon crack healing rate in calcite. *Physics and Chemistry of Minerals* *15*(1), 91–102.
- Hobbs, P. V. 1974. *Ice physics*. Oxford university press.
- Hoppa, G. V., B. R. Tufts, R. Greenberg, and P. E. Geissler 1999. Formation of cycloidal features on Europa. *Science* **285**, 1899–1902.
- Hurford, T., A. Sarid, and R. Greenberg 2007. Cycloidal cracks on Europa: Improved modeling and non-synchronous rotation implications. *Icarus* *186*(1), 218–233.
- Iliescu, D., and E. Schulson 2002. Brittle compressive failure of ice: monotonic versus cyclic loading. *Acta materialia* *50*(8), 2163–2172.
- Irwin, G. 1957. Analysis of Stresses and Strains Near the End of a Crack Traversing a Plate. *J. Appl. Mech.*.

- Johnson, H. H., and P. C. Paris 1968. Sub-critical flaw growth. *Engineering Fracture Mechanics* 1(1), 3–45.
- Kehle, R. O. 1964. Deformation of the Ross ice shelf, Antarctica. *Geological Society of America Bulletin* 75(4), 259–286.
- Kerr, A. D. 1996. Bearing capacity of floating ice covers subjected to static, moving, and oscillatory loads. *Applied Mechanics Reviews* 49, 463–476.
- Klein-Paste, A., and N. K. Sinha 2010. Microstructural investigation of ice surfaces after rubber–ice and sand–ice sliding friction tests. *Tribology International* 43(5), 1151–1157.
- Kraus, R. G., L. E. Senft, and S. T. Stewart 2011. Impacts onto H₂O ice: Scaling laws for melting, vaporization, excavation, and final crater size. *Icarus* 214(2), 724–738.
- Langhorne, P. J., V. A. Squire, C. Fox, and T. G. Haskell 1998. Break-up of sea ice by ocean waves. *Annals of Glaciology* 27(1), 438–442.
- Litwin, K. L., B. R. Zygielbaum, P. J. Polito, L. S. Sklar, and G. C. Collins 2012. Influence of temperature, composition, and grain size on the tensile failure of water ice: Implications for erosion on Titan. *Journal of Geophysical Research: Planets* 117(E8).
- McCarthy, C., R. F. Cooper, D. L. cooGoldsby, W. B. Durham, and S. H. Kirby 2011. Transient and steady state creep response of ice I and magnesium sulfate hydrate eutectic aggregates. *Journal of Geophysical Research: Planets* 116(E4).
- McCord, T. a., G. Hansen, R. Clark, P. Martin, C. Hibbitts, F. Fanale, J. Granahan, M. Segura, D. Matson, T. Johnson, et al. 1998. Non-water-ice constituents in the surface material of the icy Galilean satellites from the Galileo near-infrared mapping spectrometer investigation. *Journal of Geophysical Research: Planets* 103(E4), 8603–8626.
- Melosh, H., and B. Ivanov 1999. Impact crater collapse. *Annual Review of Earth and*

- Planetary Sciences* 27(1), 385–415.
- Moore, J. M., G. Black, B. Buratti, C. B. Phillips, J. Spencer, and R. Sullivan 2009. Surface properties, regolith, and landscape degradation. *Europa, edited by RT Pappalardo, WB McKinnon, and K. Khurana, University of Arizona Press, Tucson*, 329–349.
- Nath, P., and D. Vaughan 2003. Subsurface crevasse formation in glaciers and ice sheets. *Journal of Geophysical Research: Solid Earth* 108(B1).
- Nimmo, F., C. Porco, and C. Mitchell 2014. Tidally modulated eruptions on Enceladus: Cassini ISS observations and models. *The Astronomical Journal* 148(3), 46.
- Nixon, W., and E. Schulson 1987. A micromechanical view of the fracture toughness of ice. *Le Journal de Physique Colloques* 48(C1), C1–313.
- Nixon, W., and R. Smith 1987. The fatigue behavior of freshwater ice. *Le Journal de Physique Colloques* 48(C1), C1–329.
- O'Neill, C., and F. Nimmo 2010. The role of episodic overturn in generating the surface geology and heat flow on Enceladus. *Nature Geoscience* 3, 88–91.
- Pappalardo, R. T., and A. C. Barr 2004. The origin of domes on Europa: The role of thermally induced compositional diapirism. *Geophysical Research Letters* 31(1).
- Pappalardo, R. T., S. J. Reynolds, and R. Greeley 1997. Extensional tilt blocks on Miranda: Evidence for an upwelling origin of Arden Corona. *Journal of Geophysical Research: Planets* 102(E6), 13369–13379.
- Parsons, B., M. Lal, F. Williams, J. Dempsey, J. Snellen, J. Everard, T. Slade, and J. Williams 1992. The influence of beam size on the flexural strength of sea ice, freshwater ice and iceberg ice. *Philosophical Magazine A* 66(6), 1017–1036.
- Polanskey, C. A., and T. J. Ahrens 1990. Impact spallation experiments: Fracture patterns and spall velocities. *Icarus* 87(1), 140–155.
- Porco, C., D. DiNino, and F. Nimmo 2014. How the geysers, tidal stresses, and

- thermal emission across the south polar terrain of Enceladus are related. *The Astronomical Journal* 148(3), 45.
- Prockter, L. M., J. W. Head, R. T. Pappalardo, R. J. Sullivan, A. E. Clifton, B. Giese, R. Wagner, and G. Neukum 2002. Morphology of European bands at high resolution: A mid-ocean ridge-type rift mechanism. *J. Geophys. Res.* **107**.
- Rhoden, A. R., and T. A. Hurford 2013. Lineament azimuths on Europa: Implications for obliquity and non-synchronous rotation. *Icarus* 226(1), 841–859.
- Rhoden, A. R., B. Militzer, E. M. Huff, T. A. Hurford, M. Manga, and M. A. Richards 2010. Constraints on Europa’s rotational dynamics from modeling of tidally-driven fractures. *Icarus* 210(2), 770–784.
- Rist, M., P. Sammonds, S. Murrell, P. Meredith, C. Doake, H. Oerter, and K. Matsuki 1999. Experimental and theoretical fracture mechanics applied to Antarctic ice fracture and surface crevassing. *Journal of Geophysical Research: Solid Earth* 104(B2), 2973–2987.
- Rooke, D. P., and D. J. Cartwright 1976. Compendium of stress intensity factors. *Procurement Executive, Ministry of Defence. H. M. S. O. 1976, 330 p(Book)*.
- Schulson, E. M. 1999. The structure and mechanical behavior of ice. *JOM journal of the Minerals, Metals and Materials Society* 51(2), 21–27.
- Schulson, E. M. 2001. Brittle failure of ice. *Engineering Fracture Mechanics* 68(17), 1839–1887.
- Schulson, E. M., P. Lim, and R. Lee 1984. A brittle to ductile transition in ice under tension. *Philosophical Magazine A* 49(3), 353–363.
- Showman, A. P., and L. Han 2005. Effects of plasticity on convection in an ice shell: Implications for Europa. *Icarus* 177(2), 425–437.
- Smith, E., and J. Barnby 1967. Crack nucleation in crystalline solids. *Metal Science Journal* 1(1), 56–64.
- Soderblom, J. M., A. J. Evans, B. C. Johnson, H. J. Melosh, K. Miljković, R. J.

- Phillips, J. C. Andrews-Hanna, C. J. Bierson, J. W. Head, C. Milbury, et al. 2015. The fractured Moon: Production and saturation of porosity in the lunar highlands from impact cratering. *Geophysical Research Letters* 42(17), 6939–6944.
- Stern, L. A., W. B. Durham, and S. H. Kirby 1997. Grain-size-induced weakening of H₂O ices I and II and associated anisotropic recrystallization. *Journal of Geophysical Research: Solid Earth* 102(B3), 5313–5325.
- Sunder, S. S., and M. S. Wu 1990. Crack nucleation due to elastic anisotropy in polycrystalline ice. *Cold Regions Science and Technology* 18(1), 29–47.
- Suresh, S. 1998. *Fatigue of materials*. Cambridge university press.
- Vaughan, D. G. 1993. Relating the occurrence of crevasses to surface strain rates. *Journal of Glaciology* 39(132), 255–266.
- Wahr, J., Z. A. Selvens, M. E. Mullen, A. C. Barr, G. C. Collins, M. M. Selvens, and R. T. Pappalardo 2009. Modeling stresses on satellites due to nonsynchronous rotation and orbital eccentricity using gravitational potential theory. *Icarus* 200(1), 188–206.
- Wang, M., and P. Cao 2015. Numerical Study on Flattened Brazilian Test and Its Empirical Formula. *EJGE* 20, 12211–12224.
- Weber, L., and W. Nixon 1997. Fatigue of freshwater ice. *Cold regions science and technology* 26(2), 153–164.
- Weiss, J. 2004. Subcritical crack propagation as a mechanism of crevasse formation and iceberg calving. *Journal of Glaciology* 50(168), 109–115.
- Weiss, J., J.-R. Grasso, M.-C. Miguel, A. Vespignani, and S. Zapperi 2001. Complexity in dislocation dynamics: experiments. *Materials Science and Engineering: A* 309, 360–364.
- Weiss, J., and D. Marsan 2003. Three-dimensional mapping of dislocation avalanches: clustering and space/time coupling. *Science* 299(5603), 89–92.
- Weiss, J., and E. Schulson 2000. Grain-boundary sliding and crack nucleation in ice.

Philosophical Magazine A 80(2), 279–300.

5.10 Tables and Figures

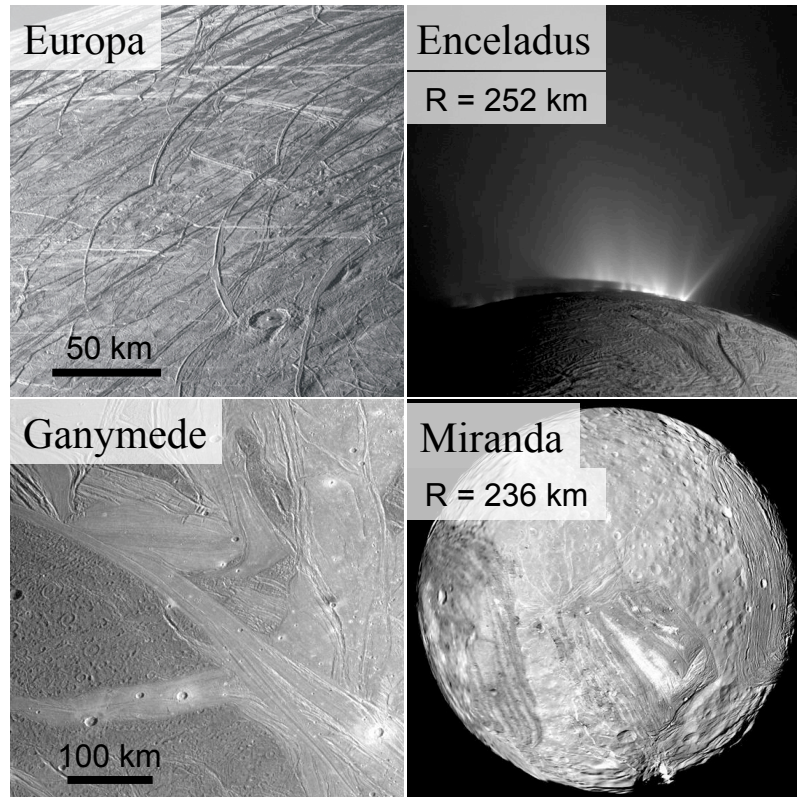


Figure 5.1: Examples of geologic features on icy satellites that suggest their surfaces might be weak. (top left) Arch shaped fractured (cycloids) on Europa form in response to tidal stresses ~ 50 kPa in magnitude. (top right) Eruptions of water vapor plumes on Enceladus are regulated by tidal stresses of < 100 kPa. Both light grooved terrain on Ganymede (bottom left) and coronae on Miranda (bottom right) are suggested to form by convection driven resurfacing, which would required the brittle yield stress of the surface to be less than ~ 50 kPa.

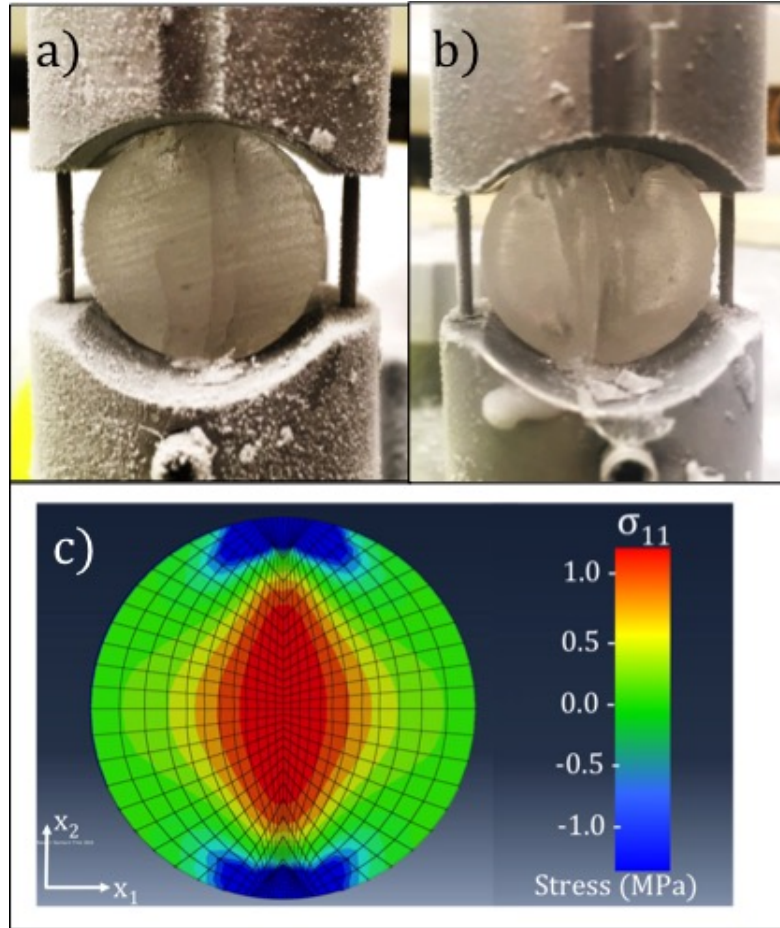


Figure 5.2: a,b) Example of failed ice samples after cyclic Brazil test. Ice samples are circular discs with a diameter of 2.54 cm and a thickness of 1.0 cm. Curved steel platens are used as contact pieces to alleviate stress concentrations at loading points. Failed samples show vertical fractures that propagate through the center of the sample. c) Numerical calculations of the σ_{11} stress during a Brazil test. Red colors show tensile stresses and blue show compressive stresses. Simulations were performed with the numerical model Abaqus and used a fixed boundary condition at the base of the disc and a fixed load of 100 lbs at the top of the disc.

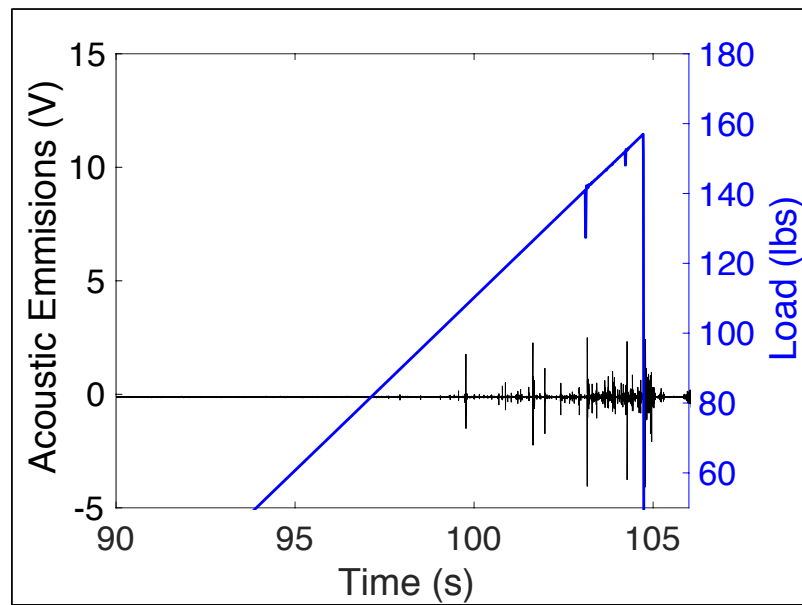


Figure 5.3: Acoustic emissions data measured during a Brazil test for monotonic failure. Small events prior to failure are interpreted as crack nucleation events. Initial load failure occurs at 650 N and load drop corresponds with acoustic events. The failure event coincided with the appearance of a vertical fracture in the sample.

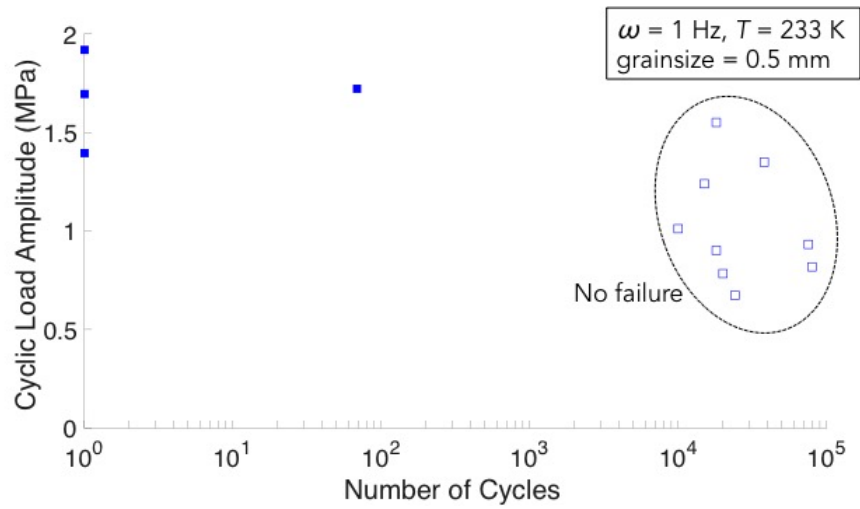


Figure 5.4: Results of our constant cyclic load amplitude fatigue tests performed at $T = 233 \text{ K}$ and $\omega = 1 \text{ Hz}$. Cyclic load amplitude is shown on the y-axis and number of load cycles is shown on the x-axis. Solid points show samples that fractured and open points show samples that did not fail (run outs).

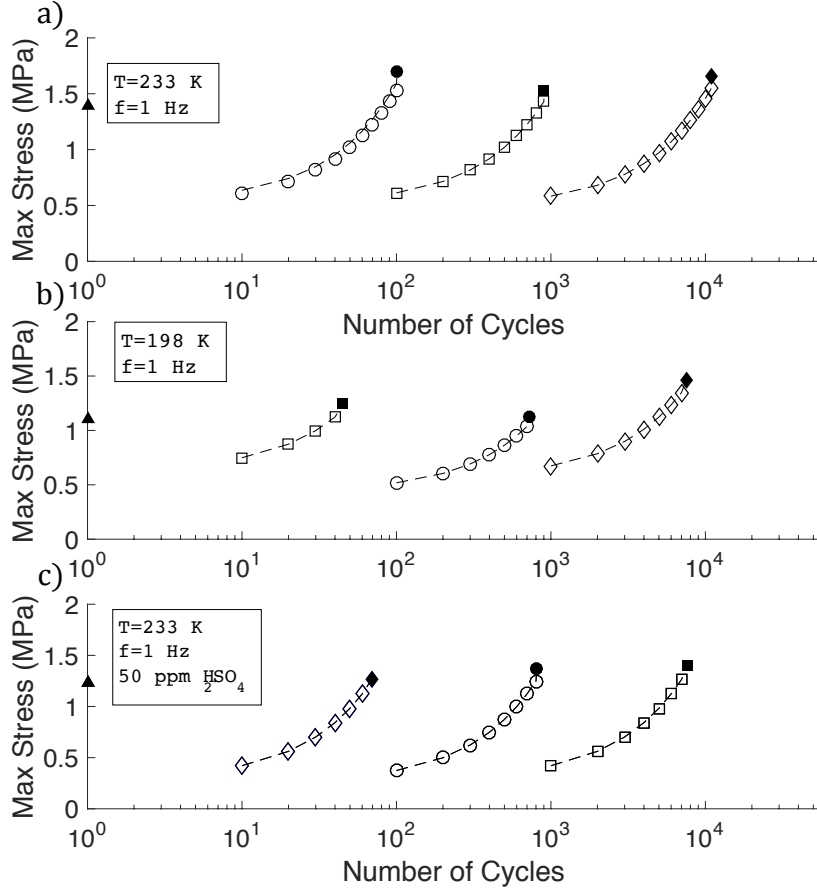


Figure 5.5: Results of our gradually increasing cyclic load amplitude fatigue tests shown for three different conditions. Open points represent samples that experience a certain cyclic load without failure which then were subjected to larger cyclic loads. Solid points represent fracture events. Dashed lines connect continuous points that were part of the same experiment. a) Results for fresh water ice at $T = 233$ K, b) results for fresh water ice at $T = 198$ K, c) results for ice doped with 50 ppm H_2SO_4 at $T = 233$ K

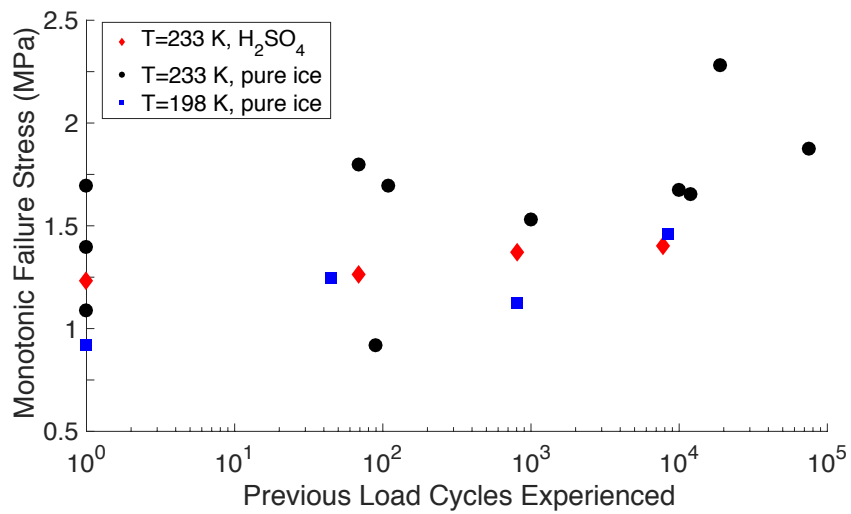


Figure 5.6: The tensile yield stress versus the number of load cycles experience for a variety of experimental conditions. Data from both types of fatigue tests are included, (gradually increasing cyclic load and the monotonic failure stress measured after run-out tests during constant cyclic load tests). Black circles and blue squares show experiments performed on pure ice at 233 K and 198 K, respectively. Red diamonds show experiments performed on ice doped with 50 ppm of sulfuric acid cyclically loaded at 233 K.

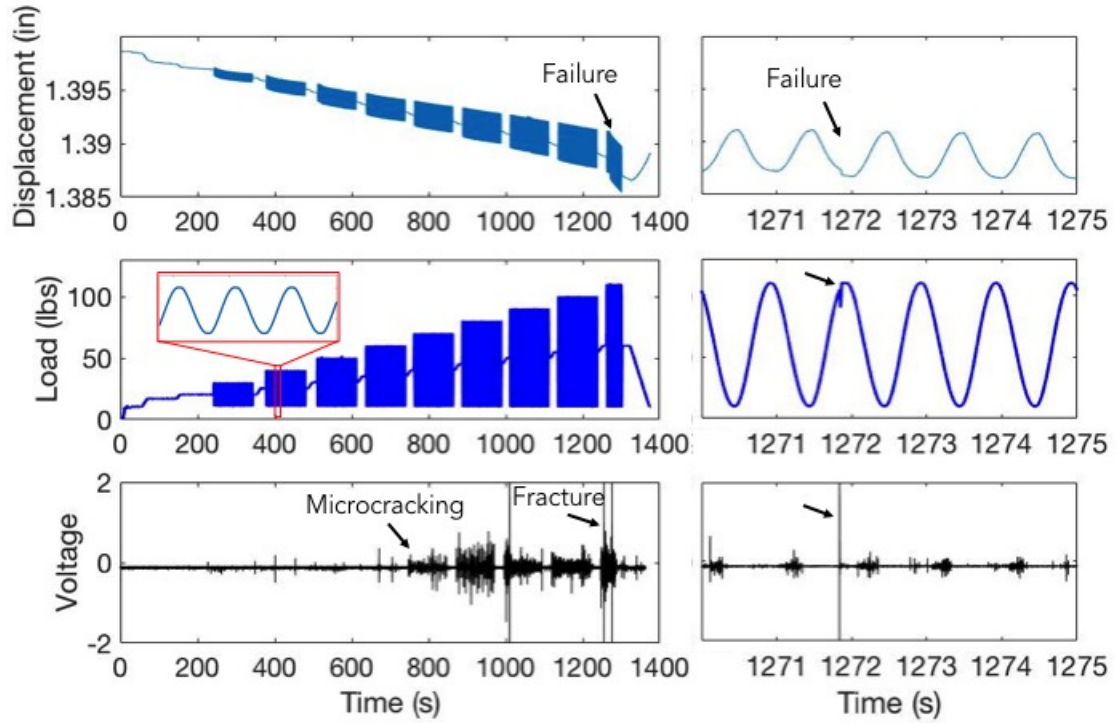


Figure 5.7: Mechanical data recorded during increasing amplitude fatigue tests on ice, showing vertical displacement (top), load (middle) and acoustic emissions (bottom) with time on the x-axis. Significant acoustic emission events begin at 750 seconds, interpreted as microcracking in response to cyclic loading. The sample fails at 1271.8 seconds, with a transition to tertiary creep, load drop and a large AE event. This failure event is interpreted as the formation of a vertical fracture through the middle of the sample (see figure 5.2).

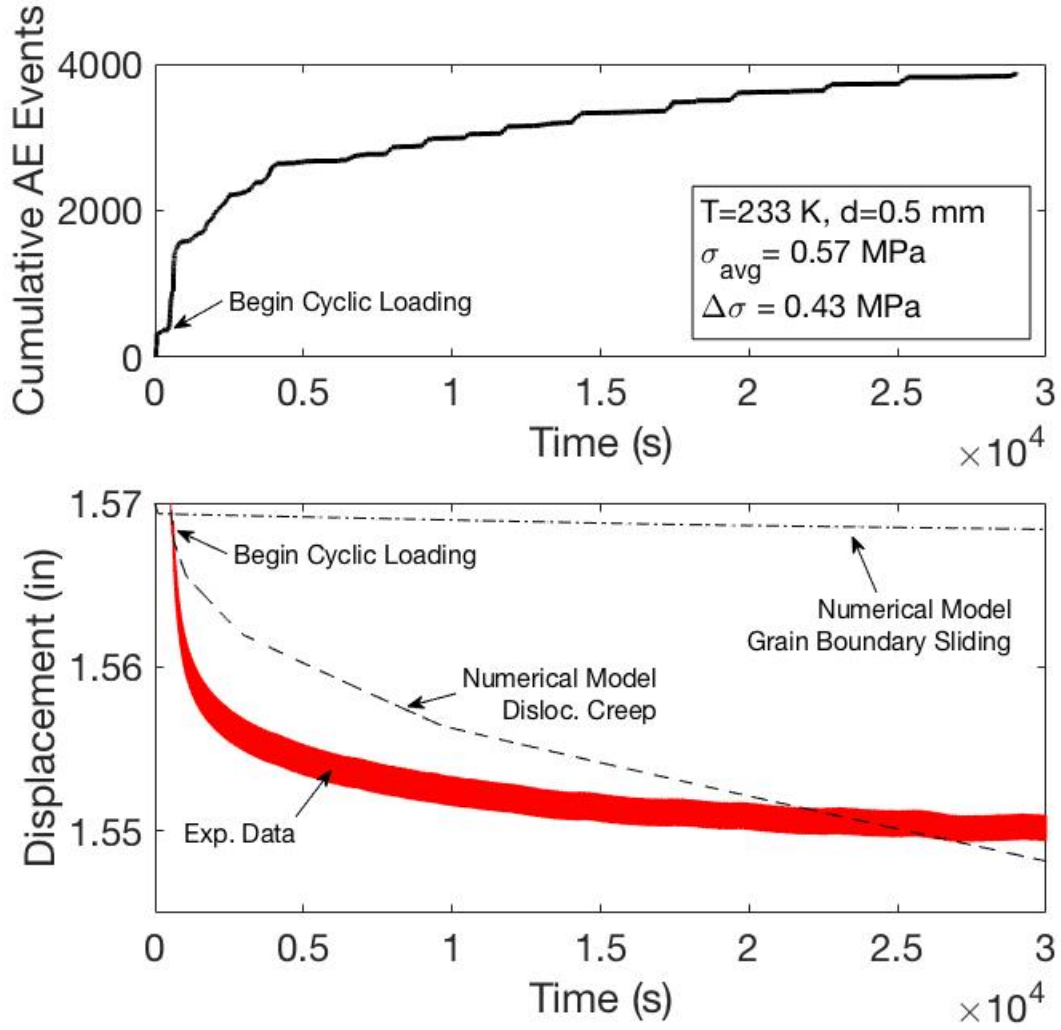


Figure 5.8: (top) The cumulative acoustic emission (AE) that occurred during a constant amplitude fatigue test with conditions described on the figure. This same fatigue test as discussed in figure 5.9. (bottom) The vertical displacement during cyclic loading test (shown in red) compared to the vertical displacement measured during our numerical simulation (dashed lines). Both a dislocation creep rheology and grain boundary sliding rheology are used independently.

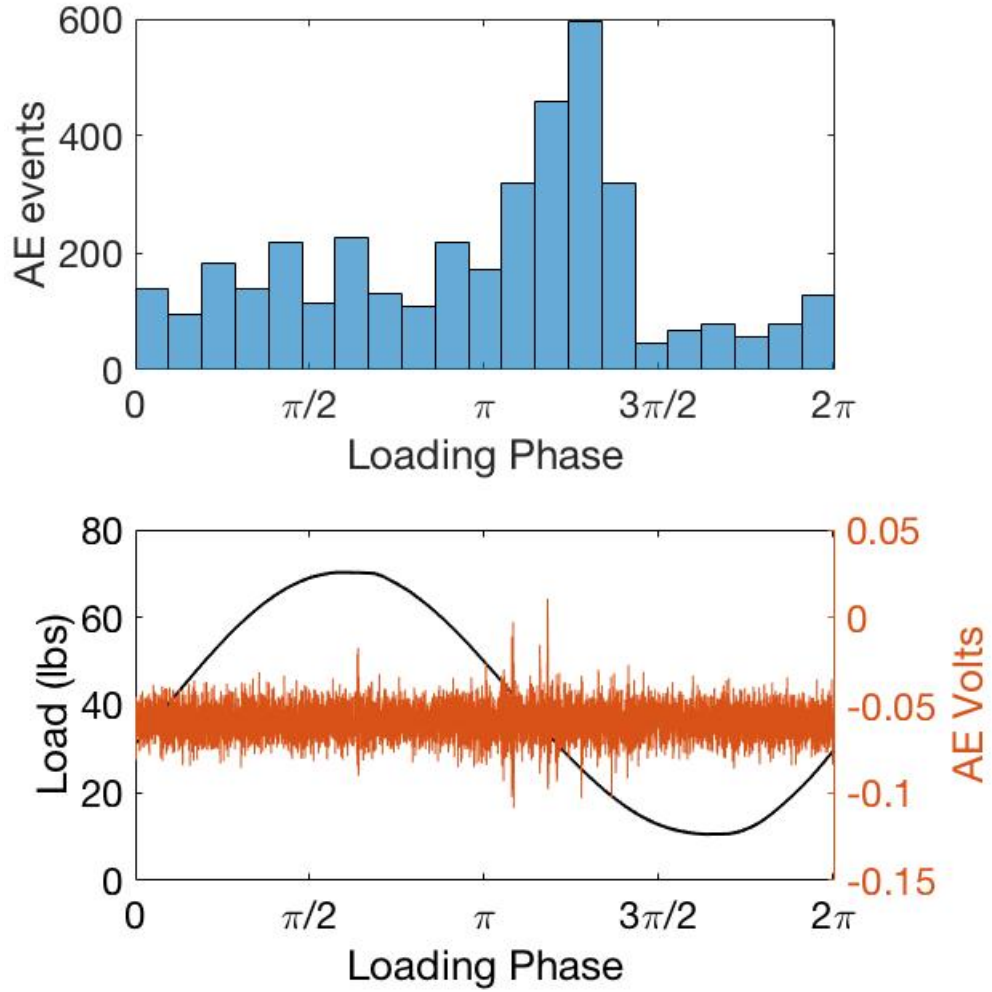


Figure 5.9: (top) Histogram of acoustic emission (AE) events during a fatigue test shown as a function of the loading phase the event occurred during. These AE events were recorded during a constant amplitude fatigue test out to 60000 cycles that did not result in failure. (bottom) The load and raw AE data that occurred during one loading cycles of the experiment.

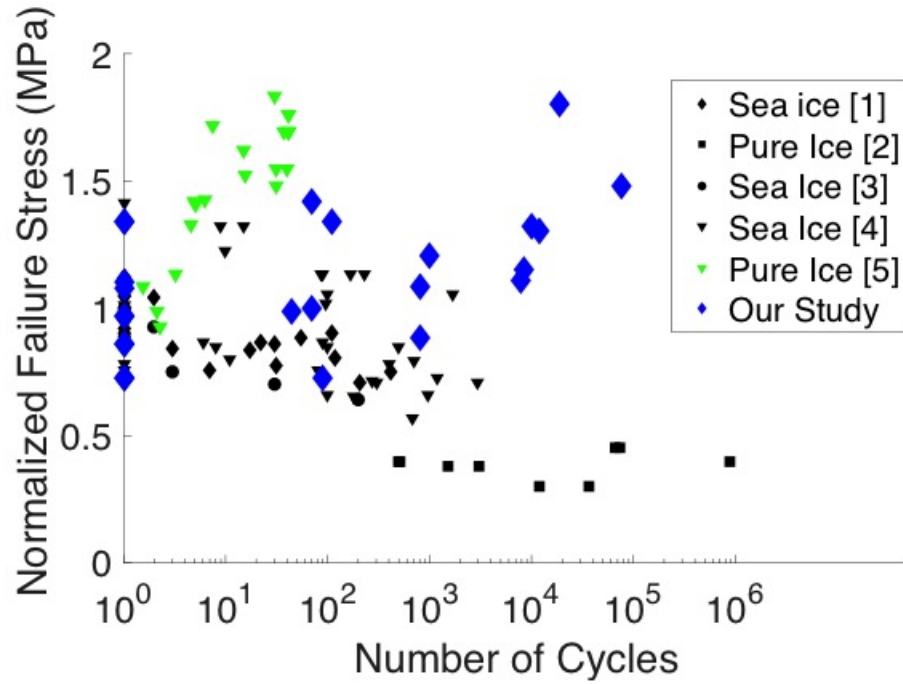


Figure 5.10: Comparison of our fatigue data to previous studies of ice fatigue. The y-axis shows the normalized failure stress and the x-axis shows the number of load cycles until failure. The failure stress was normalized by the monotonic failure stress measured in each study. Blue diamonds show all of our data shown in figure 5.7. In situ sea ice fatigue data are represented by black diamonds (Langhorne *et al.*, 1998), black circles, (Haskell *et al.*, 1996) and black triangles (Bond and Langhorne, 1997). Reverse bending tests on large grained fresh water ice from Nixon and Smith (1987) are shown as black squares. Green triangles represent data from Iliescu and Schulson (2002) who cyclically loaded columnar ice in uniaxial compression.

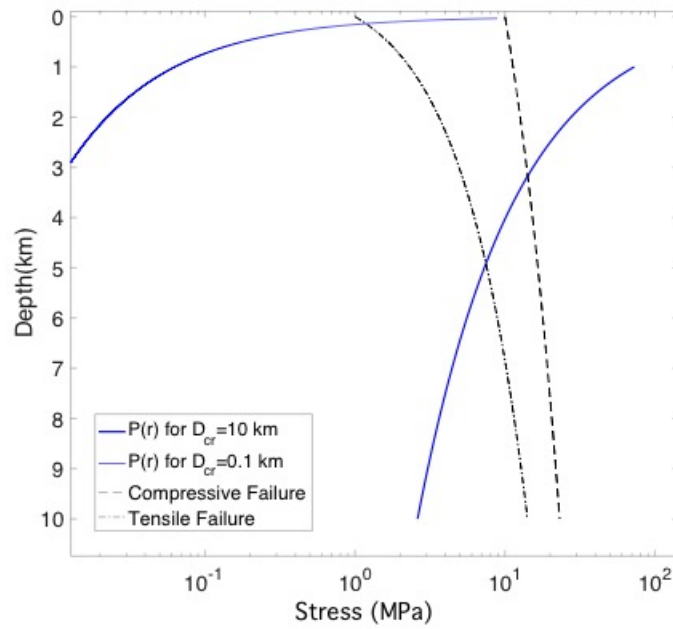


Figure 5.11: Example of the magnitudes of impact shock waves propagating into Europa's subsurface. Blue lines show the impact shock magnitude during the formation of a 10 km diameter crater and a 1 km diameter crater. Dashed and dot-dashed lines show the compressive brittle failure stress and the tensile brittle failure stress of water ice with depth.

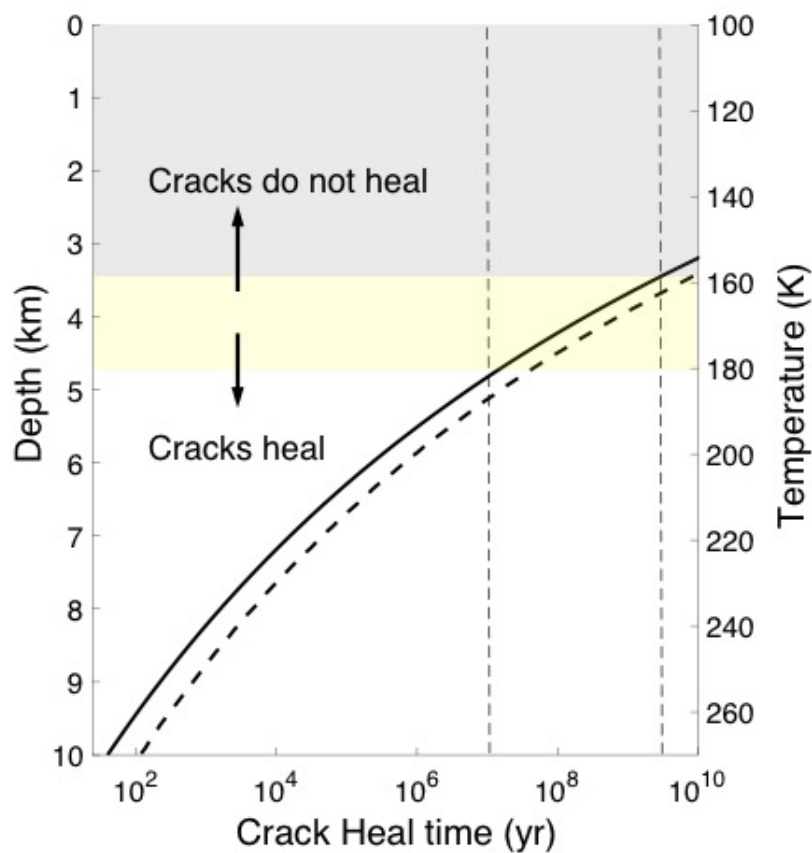


Figure 5.12: The x-axis shows the time scale to heal a 1 m long crack via diffusion accommodated crack tip retreat. Lines for crack tip radius $r = 0.01$ mm and $r = 0.1$ mm are shown as solid and dashed lines, respectively. The y-axis shows temperature and approximate depth in the ice shell of Europa assuming the ice shell is conductive with a thermal gradient of ~ 3 K/km. Vertical thin dashed lines show the approximate surface age of Europa and the age of the solar system. At temperature below 160 K, the time scale for crack healing is greater than 10^9 years.

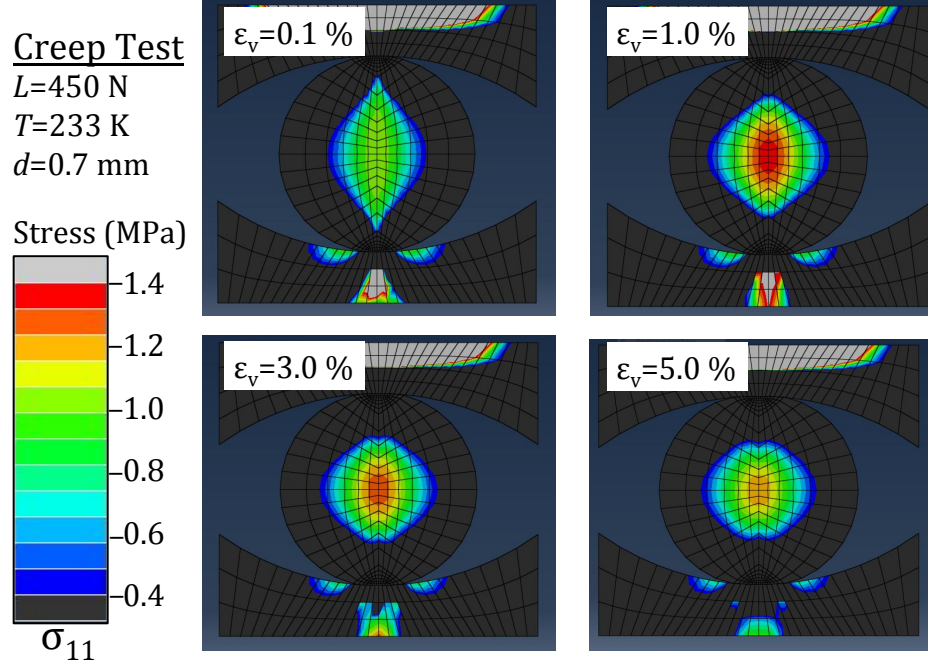


Figure 5.13: Final element model showing the stress in the sample during a creep test in Brazil geometry. Four panels show how the stress changes after increasing amounts of vertical strain ϵ_v . Black lines show the element mesh. Color contours show the tensile stress ranging from 1.4 MPa (red) to 0.4 MPa (blue). Elements with stresses outside this range are shown in gray. The load on the ice sample was $L = 450$ N. The curved top and bottom pieces are elastic with properties of stainless steel and the sample is viscoelastic with the properties of ice at 233 K and a grain size of 0.7 mm (see Supplemental Material).

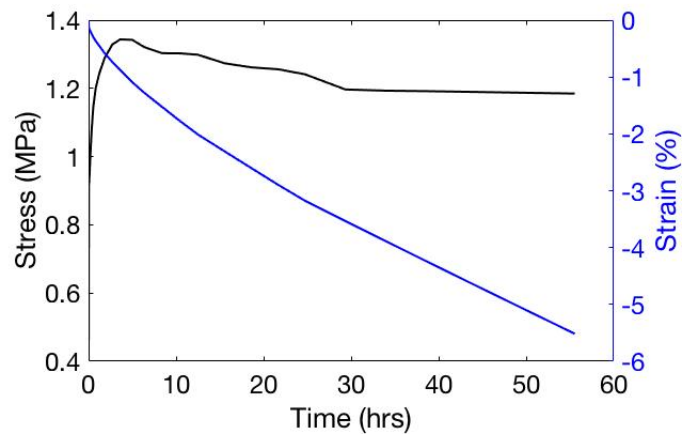


Figure 5.14: The stress and strain during the numerical simulation of a creep test, shown in figure 6.12. The black line shows the stress which increases at first then gradually decreases. The blue line shows the vertical strain which becomes more negative (compressive) with time.

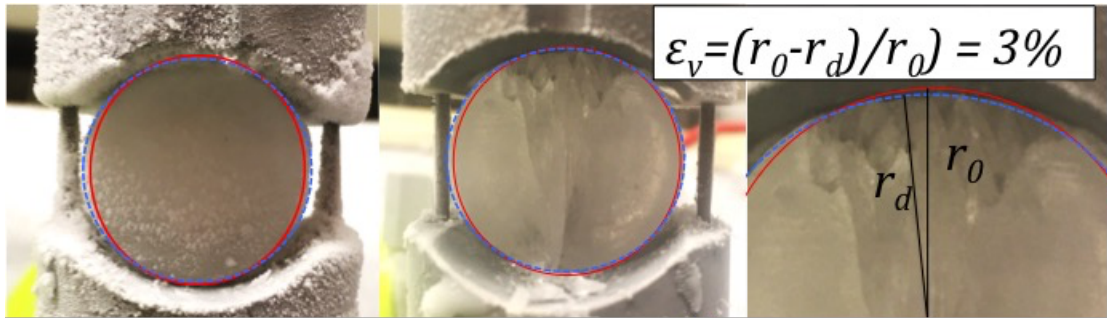


Figure 5.15: Vertical strain measurements shown for deformed samples. Red lines show perfect circles fit to represent the undeformed sample shape. Dashed blue lines show ellipses fit to the deformed sample shape. The image on the left shows a sample that was deformed to 24000 cycles at stresses of 0.8 MPa and did not fracture. The image in the center shows a fractured sample after an increasing amplitude test. The image on the right shows how the radius of the undeformed and deformed sample shape were used to calculate the vertical strain on the same sample shown in the center image.

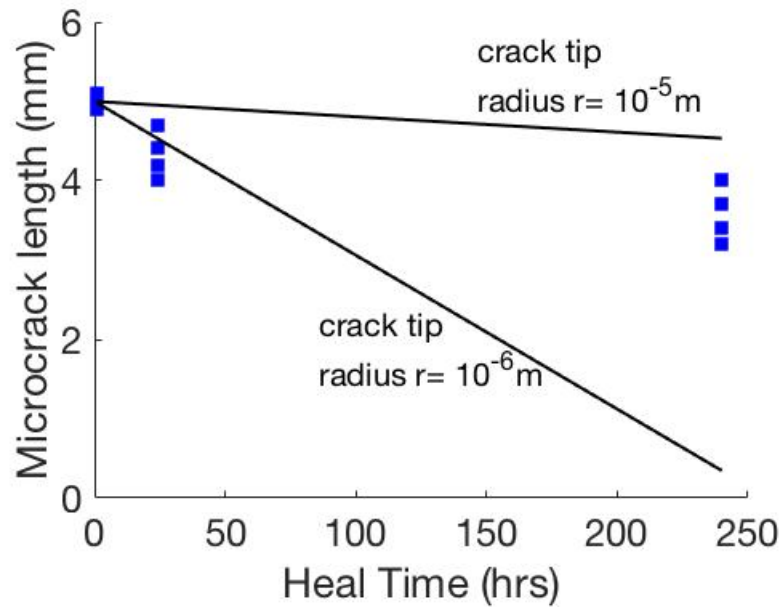


Figure 5.16: Comparison between our crack healing model and observations of microcrack healing made by (Cole, 1986). Blue squares represent microcrack length measurements of three different cracks made at four different times. The lengths have been adjusted so the initial length of each crack is 5 mm. Black lines show our model predictions for crack healing for a crack tip radius of $r = 1$ micron and 10 microns.

Chapter 6 | Compaction of a Partially Molten Ammonia-Water Ice Shell: Implications for the Geologic Evolution of Triton

N.P Hammond¹, E.M.Parmentier¹.

1. Department of Earth, Environmental and Planetary Science, Brown University, Providence, RI 02912, USA

6.1 Abstract

Icy satellite ice shells likely contain impurities that could lower the melting temperature and have dramatic consequences for their thermal and geologic evolution. A small amount of ammonia in the ice shell lowers solidus to 176 K and also reduces the liquidus by a smaller fraction, suggesting a large temperature range of thermally stable partial melt. The fate of this ammonia rich melt, whether it erupts to the surface, stays in the ice shell or sinks to the bottom, will depend on its relative buoyancy and the rate at which melt can move through the ice shell, both of which will be influenced by the melt fraction, ammonia concentration and temperature which can change over time. Here we model the dynamical evolution of a partially molten, ammonia-water ice system. We examine the case of a thickening ice shell above an ocean with 5% ammonia and model the compaction and thermal evolution of the partially molten layer as the ice shell thickens. We use our model to understand the early evolution of Neptune's moon Triton after its orbital eccentricity has decayed. We find that the ammonia rich melt will sink down into the ocean and that the rate of ice shell compaction is efficient compared to the rate of ice shell thickening. Very little melt or ammonia remains in the ice shell and the concentration of ammonia in the ocean will increase with time. The increased ammonia concentration can help prevent the ocean from freezing and increases the final ocean thickness by about 50 km. The ocean also gradually becomes less dense as it becomes enriched. If the ice shell is composed of $> 3\%$ silicates by mass, ammonia-dihydride melt can be gravitationally buoyant and may erupt at the surface.

6.2 Introduction

In this paper, we investigate the dynamics of melt migration within icy satellite ice shells. We construct models to understand the compaction and thermal evolution

of partially molten regions in the ice shell with an emphasis on understanding the ammonia-water system and its application to the geologic evolution of Neptune’s moon Triton.

Understanding melt migration on icy satellites is important. There is abundant evidence that melts may have migrated toward the surface of several icy satellites: 1) plumes of water vapor erupting from the south pole of Enceladus are thought to be sourced directly from a subsurface ocean ~ 30 km below (Postberg *et al.*, 2009; Porco *et al.*, 2014), 2) chaos terrains on Europa likely form in response to water only 1 – 2 km below the surface which may have been generated by convective plumes that cause partial melting of the lithosphere (Head and Pappalardo, 1999; Sotin *et al.*, 2002), 3) several icy bodies are covered by potential “cryovolcanic” terrains that are thought to form when water-rich fluids erupt onto and flow across the surface (Kargel, 1995; Schenk, 1991; Desch *et al.*, 2009).

In addition to resurfacing, the behavior of partial melt is also important in controlling the thermal evolution of icy bodies. Partial melt that remains stable in the ice shell could strongly reduce the viscosity and increase the likelihood of solid-state convection (c.f. Goldsby and Kohlstedt 2001; Hammond and Barr 2014). Volatile rich partial melts that sink into the subsurface ocean could reduce the melting temperature of the ocean and help prevent the ocean from freezing. To understand how quickly melts migrate and whether melts migrate toward the surface, remain in the ice shell or sink into the ocean, it is essential to understand what controls the physical evolution of partial melts.

The primary control on the relative buoyancy of the melt is its density and composition. Water is denser than ice and will sink unless there is over pressurization driving flow (e.g. Manga and Wang (2007)). However, melts on icy satellites could be enriched in volatiles that can significantly alter their density (e.g. Crawford and Stevenson 1988). While icy satellites are thought to be composed largely of water

and silicates, a wide variety of chemical compounds may be present in the ice shell, ranging in abundance from 0.1 – 15% mass fraction (McKinnon *et al.*, 2008). Possible volatiles in the ice shell include CH₄, CO, CO₂, MgSO₄, NaCl, and NH₃ (Kargel, 1992). The chemical evolution of these systems is complex and not fully understood (c.f. Sohl *et al.* 2010; Neveu and Desch 2015). Kargel (1992) investigated the likely chemical evolution of icy satellites and subsurface oceans. He found that melts on the icy satellites of Jupiter and Saturn would tend to be enriched in salts, such as MgSO₄, and methanol, and if ammonia is the most abundant volatile (as may be the case further out in the solar system), melts would be dominated by ammonia.

Ammonia, compared to other volatiles, may have the most interesting influence on icy satellite evolution. The presence of ammonia can lower the solidus of the ice ammonia system to 176 K (Leliwa-Kopystyński *et al.*, 2002), and ammonia is one of the only volatiles that reduces the density of the melt when dissolved (Croft *et al.*, 1988).

Ammonia is thought to be more abundant in the outer solar system, where temperatures during planetary formation favored the condensation of ammonia from the solar nebula (Lewis, 1972). Ammonia has been spectroscopically detected on the surface of Pluto’s moon Charon (Brown and Calvin, 2000) and on Uranus’ satellite Miranda (Bauer *et al.*, 2002) and has also been measured in the plumes erupting out of Enceladus (Waite Jr *et al.*, 2009). Due to damage from solar radiation, ammonia may not be stable on the surface of icy satellites, and its spectroscopic signature could be erased after 10⁴ years in the Saturnian system and 10⁹ years in the Pluto system (Moore *et al.*, 2007). Some authors have argued that the detection of ammonia on icy satellite surfaces today suggests a recent emplacement of ammonia by cryovolcanic resurfacing (Desch *et al.*, 2009).

Several investigators have explored the role of ammonia in cryovolcanic resurfacing and have modeled the physical evolution of ammonia melts throughout the thermal

evolution of icy bodies (Kargel *et al.*, 1991; Kargel, 1995; Hogenboom *et al.*, 1997; Desch *et al.*, 2009; Desch and Neveu, 2017). However, previous works have used simplified approximations for how melts migrate in the ice shell that neglect the influence of many physical parameters (viscosity, melt fraction, buoyancy) on the rate of melt migration. Perhaps the most detailed investigation of compaction and melt migration in icy satellite ice shells is by Kalousová *et al.* (2014). Using a 1-dimensional advection model they investigated the downward propagation of melt on Europa from the near surface to the subsurface ocean and looked at the timescale at which melt would be present in the ice shell. Our approach is similar except that we focus on the ammonia-water system and its application to the geologic evolution of Triton.

Here we focus on understanding the early evolution of an ice shell, from an initially warm and thin state, and simultaneously model the growth and thermal evolution of the ice shell and the migration of ammonia-rich melts. Our goal is to understand whether ammonia will freeze into the ice shell as the ice shell cools or whether the ammonia will sink into the ocean. We also wish to understand under what conditions ammonia-rich melts can be gravitationally buoyant in the ice shell.

6.3 Ice-Ammonia System

Figure 6.1a shows the ammonia-water phase diagram at atmospheric pressure. The liquidus of the system decreases with increasing mass concentration of ammonia X_{NH_3} , from 273.1 K for $X_{NH_3} = 0$, down to a eutectic of 176 K at $X_{NH_3} \approx 0.3$ (Kargel, 1992; Leliwa-Kopystyński *et al.*, 2002). Below the solidus temperature of 176 K (which is independent of ammonia concentration), the stable phases are water ice mixed with ammonia-dihydride ice. Above the solidus, the stable phases are solid water ice mixed with ammonia-rich fluids. For a given temperature T and bulk concentration of ammonia X_0 , the fluid fraction and the composition of the melt can be

calculated with the lever rule. The equilibrium melt fraction $f = X_0/X_l$, where X_l is the concentration of ammonia at the liquidus for a given temperature. We calculate X_l based on equations given in Leliwa-Kopystyński *et al.* (2002). The equilibrium melt fraction f is related to the dynamic melt fraction ϕ which can change due to melt-solid segregation.

Figure 6.1b shows the melt fraction, ϕ as function of temperature for a bulk ammonia concentration of $X_0 = 0.1$. Because there is a large temperature range in which partial melt is stable, there may be large partially molten regions in icy bodies whose thickness depends on the thermal gradient and ammonia concentration. How thick would these partially molten layers be and how would they evolve over time? Above some critical melt fraction ϕ_c solids will be suspended in the fluids, and for $\phi < \phi_c$, solids will be in contact with one another so that they form a porous solid layer. A schematic of this structure is shown in figure 6.2.

In the region where $\phi > \phi_c$, water crystals with a density $\rho_i = 0.92 \text{ g/cm}^3$ will be buoyant with respect to the ammonia-water solution $\rho_l = 0.945 - 1 \text{ g/cm}^3$ (Croft *et al.*, 1988) and the velocity of the water crystals can be described by Stokes settling. If the fluid is vigorously convecting, convective motions may be able to keep smaller crystals in suspension but segregations can still occur at convective boundary layers where fluid velocities are low (Solomatov, 2000). The segregation flux of crystals in this layer is approximated as (Solomatov, 2000)

$$S = 0.1 \frac{\Delta \rho g d^2}{18 \eta_l}, \quad (6.1)$$

where g is gravity, d is grain size, η_l is the viscosity of the fluid, and $\Delta \rho = \rho_i - \rho_l$. For $d = 1 \text{ mm}$, $\eta_l = 10^{-3} \text{ Pa s}$, and $g = 1.3 \text{ m/s}^2$, parameters appropriate for Europa, the separation velocity 3 mm/s or 10 km/year . This is much faster than the rate of ice shell thickening, so we assume that crystals in this region settle out immediately

compared to the ice shell thermal evolution.

Fluid-crystal separation in the region $\phi < \phi_c$ occurs by porous flow accompanied by compaction. The separation velocity can be described by Darcy's law in terms of the compaction viscosity of the solid ξ and the permeability K where

$$S = \frac{\Delta\rho g K}{\eta_f} \left[(1 - \phi) + \frac{1}{\Delta\rho g} \frac{d}{dz} \left(\xi \frac{dS}{dz} \right) \right]. \quad (6.2)$$

Here z is vertical distance, the first term represents flow driven purely by gravity and the second term represent flow driven by pressure differences between the solid and the liquid as a result of compaction.

We imagine an initial condition where the ice shell is very thin. As the icy satellite cools, water crystals are deposited at the base of the ice shell at a value ϕ_c and at a rate V_d , the depositional velocity. As melt sinks from the ice shell into the ocean, the solid region will compact at a rate V_c and the change in height of the ice shell will be $V_H = V_d - V_c$. At the same time, as the ice shell grows, the depth of the solidus temperature migrates downward, and if ammonia melt is still present in this region the ammonia will be frozen into the ice shell. This problem is similar to the accumulation of the anorthositic crust on the moon and the trapping of interstitial melt in the crust (Piskorz and Stevenson, 2014). It is also very similar (but upside-down) to compacting igneous cumulates at the base of a magma chamber (Shirley, 1986). In fact, we take a very similar approach to Shirley (1986), in our numerical treatment in the growth and compaction of the ice shell. We use the term ice shell to mean both the completely solid layer and the region where the melt fraction is less than 0.5.

6.4 Numerical Model

To model the advection of fluids in the growing ice shell we use a 1-D finite difference advection code with 512 nodes. We use an implicit advection scheme using a Van Leer flux limiter. At the top of the ice shell where $z = 0$, the fluid flux is set to zero and at the base of the ice shell $z = H$, the melt fraction is set to $\phi = \phi_c (= 0.5)$. We assume freezing is concentrated at the boundary $z = H$ and at the solidus boundary. We neglect the effect of any freezing between these regions on the thermal evolution and melt fraction. As is evident later, this assumption is appropriate for a thickening icy satellite ice shell. The change in fluid fraction is calculated as

$$\frac{\partial \phi}{\partial t} = -\frac{\partial}{\partial z} (v_f \phi), \quad (6.3)$$

where v_f is the velocity of the fluid. The velocity of the fluid can be calculated in terms of the separation flux S and ϕ by showing the solid-liquid mixture has a divergence of zero and accounting for the fact that the fluid velocity at the top boundary is zero.

Since the ice shell thickness H is growing with time, we define a new coordinate $y = z/H$, which remains between 0 and 1. The change in melt fraction with time in the y domain can then be calculated using the chain rule

$$\frac{\partial \phi}{\partial y} = \frac{\partial \phi}{\partial z} \frac{\partial z}{\partial y}. \quad (6.4)$$

This leads us to solving the equation

$$\frac{\partial \phi}{\partial t} = -\frac{1}{H} \frac{\partial}{\partial y} \left[(v_f - V_H y) \phi \right] - \frac{V_H}{H}. \quad (6.5)$$

The terms with V_H in this equation can be thought of as pseudo-advective terms to account for the growth of the ice shell in a domain where there is no change in length.

The segregation flux in the ice shell is computed from equation 6.2. We assume the permeability is a function of the melt fraction such that $K = K_{ref}\phi^m$, where K_{ref} is the reference permeability and m is the permeability exponent. Wark and Watson (1998) measured the permeability of various rock types at a range of melt fractions and found $m = 3$ and that the permeability was also proportional to the grainsize d^2 . Permeability exponents of $n = 2$ have also been predicted based on theoretical arguments (Bargen and Waff, 1986). Sea ice permeability measurements are well fit by an exponent $m = 3$ (Golden *et al.*, 2007) and the reference permeability in ice may range from $K_{ref} = 10^{-8}$ to 10^{-10} m² (Kalousová *et al.*, 2014). In our simulations we use $m = 3$ and $K_{ref} = \frac{d^2}{300}$.

The compaction viscosity may also be strongly dependent on the melt fraction, though the exact dependence is uncertain. The compaction viscosity ξ is related to the shear viscosity of the solid η_s . As the melt fraction becomes smaller melt channels become increasingly narrow and difficult to squeeze out, so the compaction viscosity is often formulated as (McKenzie, 1984)

$$\xi = \xi_{ref} \frac{\phi_c - \phi}{\phi}. \quad (6.6)$$

where $\xi_{ref} \approx \eta_s$. Additionally, partial melt can affect the shear viscosity. Goldsby and Kohlstedt (2001) showed that water ice with partial melt at the grain boundaries was one to two orders of magnitude less viscous, possibly due to enhanced diffusion along the grain boundaries. Reduced viscosity with melt content has also been observed in mantle rocks (Cooper and Kohlstedt, 1984; Hirth and Kohlstedt, 1995). Arakawa and Maeno (1994) studied the viscosity of partially molten ammonia water ice at various temperatures and ammonia concentrations. He observed that at a given temperature the strain rate increased about an order of magnitude for each increase in melt fraction of 0.1. He fit his viscosity data with the equation $\eta_s = A \exp(-B\phi)$, where A and $B \approx$

22 are empirical constants. This relationship is similar to certain compaction viscosity formulas for partially molten peridotites (Renner *et al.*, 2003). We experiment with several formulations of the compaction viscosity with different dependences on ϕ , including $\xi = \xi_{ref}$, $\xi = \xi_{ref}\phi^{-1}$, $\xi = \xi_{ref}\phi^{-2}$, and $\xi = \xi_{ref}(\phi_c - \phi)/\phi$. Later we discuss simulations where the compaction viscosity is also a function of temperature.

We compute our solutions in non-dimensional form, normalizing length by a reference compaction length

$$L_{ref} = \sqrt{\frac{K_{ref}\xi_{ref}}{\eta_f}}. \quad (6.7)$$

We use a reference compaction viscosity of $\xi_{ref} = 10^{13}$ Pa s, appropriate for ice near the melting temperature with a grain size of ~ 0.2 mm (Goldsby and Kohlstedt, 2001). Thus the reference compaction length scale for the ice shell $L_{ref} \approx 5$ km. Velocity is non-dimensionalized by $S_{ref} = \Delta\rho g K_{ref}/\eta_f$. For reasonable ranges of gravity and permeability on various satellites $S_{ref} = 10^{-7}$ to 10^{-4} m/s. Time t is normalized by S_{ref}/L_{ref} .

The only explicit non-dimensional parameter in our models is the depositional velocity $\bar{V}_d$. When $\bar{V}_d$ is large, the ice shell growth rate is fast compared to the compaction rate and when $\bar{V}_d$ is small, the compaction rate can be faster than the ice shell growth rate. As discussed below, $\bar{V}_d$ values appropriate for a thickening icy satellite ice shell are very small but will change with time as the ice shell thickens.

6.5 Results for constant $\bar{V}_d$

We ran simulations for constant values of $\bar{V}_d$ varying from 10^{-3} to 1 and compared our work to the solutions of Shirley (1986). Figure 6.3 shows how the melt fraction with depth evolves for $\bar{V}_d = 0.01, 0.1$ and 1. The depth in these plots is shown in non-dimensional lengths with the initial layer thickness where $\phi < 0.5$ being 0.05.

For $\bar{V}_d = 0.001$, a steady state melt fraction structure develops where there is a

thin boundary layer at the top and the interior has a constant melt fraction $\phi_i = 0.004$. In this region, the pressure gradient is not driving flow and flow is only driven by buoyancy. At the bottom of the ice shell, where particles are being deposited, there is a boundary layer $L_c = 0.08$ in thickness where the melt fraction is increasing from ϕ_i to ϕ_c . As the solid layer continues to thicken, ϕ_i and L_c remain constant and the region where $\phi = \phi_i$ becomes thicker. The melt fraction structure is similar for $\bar{V}_d = 0.1$, except that ϕ_i and L_c are larger. For $\bar{V}_d = 1$, the ice shell is growing faster than the compaction rate and the melt fraction remains near ϕ_c throughout the ice shell.

These solutions are consistent with the results of Shirley (1986) whose model identified two distinct modes of compaction in a thickening igneous cumulate pile. During fast deposition, compaction only occurs near the top boundary (bottom boundary in (Shirley, 1986)) and the system is limited by the permeability. For slow deposition rates a constant internal melt fraction develops and a boundary layer near the depositional boundary develops where flow is limited by the compaction viscosity.

Shirley (1986) showed that ϕ_i and L_c are functions of $\bar{V}_d$. Figure 6.4 shows the values of L_c (in non-dimensional form) and ϕ_i for different values of $\bar{V}_d$ measured in our simulations. Our values are consistent with his findings, although the dependence of the compaction viscosity on the melt fraction can change ϕ_i and L_c by a factor of two to three.

The effect of compaction viscosity is illustrated in Figure 6.5, which shows ϕ with depth for $\bar{V}_d = 0.1$ and four different compaction viscosities. A compaction viscosity with a larger inverse dependence on ϕ results in larger values of ϕ_i . The reason is that melt can move less efficiently from the top boundary to the sedimenting boundary because the compaction viscosity near the top boundary (where the melt fraction is low) is much higher.

6.6 Icy Satellite Ice Shell Formation

The non-dimensional parameter $\bar{V}_d$ controls the compaction behavior and melt distribution. If $\bar{V}_d$ can be calculated for icy satellites, we can understand how the partial melt distribution will evolve during the formation of an ice shell. Recall that $\bar{V}_d = V_d/S_{ref}$, where V_d is the dimensional ice shell thickening rate and $S_{ref} = \Delta\rho g K_{ref}/\mu_l$. The rate the ice ocean boundary moves depends on the heat flow into and out of the boundary, F_{in} and F_{out} , as well as how quickly latent heat from freezing can be conducted away, such that

$$V_d = \frac{F_{out} - F_{in}}{\rho_l L \phi_c} \quad (6.8)$$

where ρ_l is the density of liquid and $L = 330 \text{ kg/m}^3$ is the latent heat of water.

During the initial stages of ice shell formation $F_{out} \gg F_{in}$. If the ice shell transmits heat solely by conduction and the heat flux and at surface is similar to the heat flux and the ocean boundary, then $F_{out} \approx k\Delta T/H$, where $k \approx 5 \text{ W/mK}$ is the thermal conductivity of ice (Petrenko and Whitworth, 1999), $\Delta T \approx 200 \text{ K}$ is the difference between the surface and melting temperature and H is the ice shell thickness.

For a reference permeability $K_{ref} = d^2/300$, the non-dimensional sedimentation rate for growing icy satellite ice shells is approximately

$$\bar{V}_d = \frac{3 \times 10^5 k \Delta T \mu_l}{\rho_l L \Delta \rho g \phi_c} \left(\frac{1 \text{ km}}{H} \right) \left(\frac{1 \text{ mm}^2}{d^2} \right). \quad (6.9)$$

The parameters that vary the most significantly between icy satellites are the surface gravity ($g_{Europa} = 1.3 \text{ m/s}^2$ versus $g_{Miranda} = 0.08 \text{ m/s}^2$) and the grain size. The grain size on icy satellites is undetermined. The likely range determined by Barr and McKinnon (2007) is $d=1 - 80 \text{ mm}$, assuming grain growth is limited by either

pinning from silicate impurities or by dynamic recrystallization. Expressed only in terms of gravity, grain size and ice shell thickness the non-dimensional sedimentation rate is

$$\bar{V}_d = 4 \times 10^{-5} \frac{1\text{m/s}^2}{g} \frac{1\text{km}}{H} \frac{1\text{mm}^2}{d^2}. \quad (6.10)$$

The uncertainty in grainsize and its importance in controlling permeability leads to a wide range of values, as high as $\bar{V}_d = 5 \times 10^{-4}$ appropriate for Miranda with small grainsize and thin initial ice shell, to $\bar{V}_d = 3 \times 10^{-7}$ appropriate for Europa with a large grain size. Despite this range, since $\bar{V}_d \ll 1$, compaction and the flow of partial melt will be efficient compared to the ice shell thickening rate, even when the ice shell is initially thin. This suggests that very little volatile rich melt will get trapped in the ice shell during its formation. However, there are a few important effects that are neglected in this simple analysis, including that the surface heat flux is not the same as the heat flux away from ice shell ocean boundary, and the heat flow will not be constant with time. Below we model the coupled thermal evolution and compaction of the growing ice shell.

6.7 Coupling Thermal Evolution and Melt Migration in the Ice Shell

6.7.1 Model

We compute a semi-analytic solution to the temperature in the thickening ice shell where the melt fraction $0 < \phi < 0.5$. The temperature at the top boundary is always at the surface temperature T_0 , and the temperature at the moving boundary $H(t)$ is always at the liquidus T_l . Assuming H will grow as $\sqrt{\kappa t}$, temperature in the

thickening ice shell is

$$T(z, t) = T_0 + (T_l - T_0) \frac{\text{erf}(z/\sqrt{2\kappa t})}{\text{erf}(H(t)/\sqrt{2\kappa t})}, \quad (6.11)$$

where t is time, z is distance from the top boundary, and κ is the thermal diffusivity in the ice shell. We assume the thermal conductivity in the ice shell is constant. The net vertical flux of fluids is zero and we assume the temperature of the fluid is equal to the temperature of the solids near by.

We calculate the sedimentation rate V_d so that it is consistent with the freezing rate predicted from the thermal evolution. When the ice shell thickens the latent heat released must be conducted away from the boundary such that

$$V_d = \frac{k}{L\phi_c} \frac{dT}{dz} \Big|_{z=H(t)}. \quad (6.12)$$

At each time step, the melt advection code solves for V_H as a function of V_d and ϕ , and the new ice shell thickness H is computed. The new value of H is then used to calculate the temperature profile. As described above, the temperature profile is then used to calculate the sedimentation rate V_d in the next time step. In this way we consistently couple the thermal evolution and phase separation processes.

One of the most potentially important effects we try to incorporate is that the compaction viscosity will likely be a strong function of temperature. In our temperature-coupled model, we use the temperature in each element to calculate the compaction viscosity as

$$\xi = \xi_{ref} \exp \left(\frac{Q}{R} \left(\frac{1}{T} - \frac{1}{T_l} \right) \right) \frac{\phi_c - \phi}{\phi}, \quad (6.13)$$

where R is the gas constant and $Q = 60$ kJ/mol is the activation energy for diffusion creep in ice (Goldsby and Kohlstedt, 2001) and T_l is the temperature at the liquidus.

6.7.2 Results

Figure 6.6 shows an example simulations with an initially 1 km thick ice shell with a linear temperature gradient and a surface temperature $T_s = 40$ K. For these simulations we assume a reference compaction length of $L_{ref} = 5$ km, corresponding to a melting point viscosity of $\xi_{ref} = 10^{13}$ Pa s. An initial melt fraction distribution is calculated based on the phase diagram assuming a bulk ammonia concentration $X_0 = 0.05$, giving $\phi \sim 0.15 - 0.5$ at depths below 600 m. Figure 6.6a shows how the melt evolves with time after ~ 3000 years. Over this time the ice shell only accumulates 100 m of new ice at the base but the ice shell compacts by about ~ 150 m, so overall thickness decreases slightly. Since the ice shell does not change significantly in thickness the temperature is relatively constant over time. Regions in warmer locations with lower compaction viscosities begin compacting first, and the compaction gradually progresses to shallower depths where the viscosity is higher. The compaction viscosity near the solidus boundary is 10^6 times higher than the viscosity at the ocean boundary. After 3000 years there is a 50 m thick region near the solidus boundary where there is still some ammonia rich melt. When running this simulation for longer we find that some of this melt does end up getting trapped as the solidus front moves downward, but the average melt fraction in the ice shell ends up being less than 10^{-4} .

In the example show in Figure 6.6, the initial condition of a 1 km thick ice shell causes the ice shell to thicken slow compared to the compaction rate. We are also interested in how the melt fraction and thermal evolution might couple if the thickening rate and compaction rate are initially similar. Figure 6.7 shows the early evolution of an ice-ammonia layer initially 1 m thick. In the first 6 years the ice shell thickens rapidly enough to cause the average melt fraction in the ice shell to be close to 0.5. The ice shell begins to compact and after 28 years the melt structure has reached a steady state solution where the internal melt fraction ϕ_i has been reached. At this

time, the top 20 m has a considerable amount of melt trapped above the solidus front $\phi_{trapped} = 0.3$. In this simulation we limited the minimum melt fraction to be 0.001 in order to speed up the simulation, so the actual steady state ϕ_i that should have been reached would likely be much lower than 0.001. Figure 6.8 shows the same simulation but run until the ice shell reaches about 900 m in thickness, which takes about 4000 years. It is clear that at later times, the amount of melt that gets trapped above the solidus is equal to the internal melt fraction ϕ_i , which is close to zero.

6.8 Discussion

We find that over a period of ~ 1000 years, partially molten regions will compact and nearly 100% of volatiles in the melt should migrate into the subsurface ocean. A primary reason compaction is more efficient in ice compared to igneous cumulates is that the viscosity of liquid water is orders of magnitude lower than the viscosity of silicate melts (e.g. McKenzie (1984)). During the early evolution of an initially thin ice shell, only the upper 10 meters, traps a significant portion of volatiles. The compaction layer at the bottom of the ice shell is also only ~ 10 meters thick.

What do our results suggest about the geologic evolution of icy satellites? During the initial growth of an ice shell, volatiles in the melt will only get trapped in a thin region near the surface. The lack of volatiles in the bulk of the ice shell is significant, because it makes eutectic melting events, upon reheating of the ice shell, unlikely. Because compaction is efficient, it also suggests that partial melt is unlikely to reduce the viscosity of ice shells over geologic time scales. For partial melt to be present in icy satellite ice shells today, it likely has to be produced by melting as a result of tidal heating or shear heating along faults. Tidal heating concentrated in thermal upwellings and generating partial melt is a suggested mechanism for forming chaos terrain on Europa (Sotin *et al.*, 2002). Shear heating along tidally modulated strike slip faults is suggested to form partial melt and potential plume generation on Europa

(Han and Showman, 2008), Enceladus (Nimmo *et al.*, 2007; Kalousová *et al.*, 2016), and Triton (Prockter *et al.*, 2005). Kalousová *et al.* (2014) used their modeling results to argue that melt generated by tidal heating would migrate downward too quickly to accumulate at the near surface and form chaos terrain. Our melt migration timescales appear to be consistent with theirs but we do not include a melting source in any of our models, so we can't use our results to support this conclusion.

Since our work shows that ammonia-rich melt is transported into the ice shell very efficiently as the ice shell thickens, it suggests that subsurface oceans will become enriched in volatiles over time. We explore the significance of this result below with respect to the geologic evolution of Triton.

6.8.1 Application to Triton

Triton is the only large satellite of Neptune. Much of the surface is covered in “cantaloupe terrain”, which some authors have suggested has a cryovolcanic origin (Croft *et al.*, 1995). A dearth of impact craters on Triton suggests a relatively young surface age (Stern and McKinnon, 2000). Triton has a retrograde orbit around Neptune, implying it was likely gravitationally captured (Agnor and Hamilton, 2006). Just after orbital capture, Triton would have had a very high eccentricity, causing it to experience intense tidal heating (Ross and Schubert, 1990). The magnitude of tidal heating was likely large enough to cause differentiation and melt a majority of the ice shell (Ross and Schubert, 1990). Significant tidal dissipation causes orbital eccentricity to dampen, and Triton's current orbital eccentricity is near zero (Burns *et al.*, 1986). The timescale for the circularization of Triton's orbit is estimated to be between 10^5 to 10^8 yr, (Ross and Schubert, 1990; Čuk and Gladman, 2005). After Triton's orbit circularized, tidal heating would become insignificant compared to the radiogenic heat flux from the core.

We model the thermal evolution of Triton after its orbit has circularized and

explore the effect of increasing ammonia concentration in the ocean as the ice shell thickens. We use a 1-dimensional thermal conduction code similar to that used by (Hammond *et al.*, 2016) (also see section 4.3 for methods details). The H₂O layer thickness is 330 km and the silicate core has a radius of $R_c = 1023$ km. Both the core and the ice shell transmit heat purely by conduction and the ocean is assumed to be vigorously convecting such that the temperature in the ocean is constant. Radioactive isotopes in the silicate core produce heating and we assume the core has the same concentration of radioactive isotopes as CM chondrites (Lodders, 2003; Robuchon and Nimmo, 2011). We begin the simulations 30 Myr after CAI formation so short lived radioisotopes do not have a significant effect. The thermal conductivity in the core is $k = 3$ W/mK and the thermal conductivity in the ice shell $k_i = 0.48 + 488/T$ (Petrenko and Whitworth, 1999). We assume a simple initial condition, using a uniform internal temperature with an ice shell 30 km thick and $X_0 = 5\%$ ammonia mass concentration in the ocean.

To account for the process of ammonia partitioning into the ocean as the ice shell forms we increase the ammonia concentration in the ocean as

$$X_{NH_3} = \frac{X_0}{1 - H/H_{max}} \quad (6.14)$$

where H is the ice shell thickness and H_{max} is the total thickness of the H₂O layer. The equations for the melting temperature as a function of pressure and ammonia concentration are taken from Leliwa-Kopystyński *et al.* (2002).

Figure 6.7 shows how the temperature evolves in Triton’s ice shell with and without accounting for the increasing concentration of ammonia. After 4.5 Gyr of cooling the ice shell is 270 km thick and there is a 60 km thick ocean. The ammonia concentration in the ocean has increased to $X_{NH_3} = 25\%$. Without accounting for the increase in concentration, the ocean would only be 10 km thick. Thus accounting for the

advection of ammonia rich fluids makes a significant difference in Triton's thermal and geologic evolution.

6.8.2 Conditions for Buoyancy

As the subsurface ocean becomes more enriched in ammonia its density decreases. This raises the possibility that the ammonia rich ocean could become buoyant and lead to cryovolcanic resurfacing. After the ice shell thickens and the ocean reaches a critical ammonia fraction and density, it could become unstable and lead to an overturn. This idea is similar to the overturn that is thought to have occurred on Earth's moon after the crystallization of the magma ocean. The buoyant ocean could rise and lead to catastrophic resurfacing. This would be an interesting way to explain why Triton's surface appears to be extremely young and possibly cryovolcanically resurfaced (Croft *et al.*, 1995; Stern and McKinnon, 2000).

To examine this possibility we, calculate the density of the ocean as a function of pressure, temperature and ammonia concentration and compare this to the density of the ice shell. Croft *et al.* (1988) measured the density of ammonia-water solutions at various temperatures and ammonia concentrations. We fit their data with the formula below. We assume that the ammonia concentration in the ocean stays at the liquidus, thus the temperature and the ammonia concentration do not vary independently in this case. We use a constant bulk modulus for the water of $B = 2.1$ GPa (Petrenko and Whitworth, 1999). The density as function of pressure P and temperature is

$$\rho_l = \left(1.0 - 1.7 \times 10^{-3}(273.1 - T) - 1 \times 10^{-5}(273.1 - T)^2\right) \left(1 + P/B\right) \text{g/cm}^3 \quad (6.15)$$

For the density of the ice shell we use the temperature dependent values for the coefficient of thermal expansion and bulk modulus measured by (Leadbetter, 1965).

The density of the ice shell

$$\rho_i = 0.92 \left(1 + (273 - T)\alpha(T) \right) \left(1 + P(T)/B \right) \left(1 + \rho_r f_r \right) \quad (6.16)$$

where f_r is the volume fraction of rock in the ice shell and $\rho_r = 3.5 \text{ g/cm}^3$ is the rock density.

Figure 6.8 shows the liquid and ice shell density with depth in Triton's ice shell for the temperature result shown in figure 6.7 with increasing ammonia concentration. Although ammonia significantly reduces the density of the ocean, the pressure causes the ocean to increase in density with depth more than the ice shell. This causes the ocean to always be denser than the ice shell. Although water is often approximated as incompressible in some earth science and fluid dynamics problems, the compressibility is important in icy satellite oceans below thick ice shells.

If the ice shell is composed of $f_r = 3\%$ rock by mass, then the ammonia ocean can be buoyant relative to the ice shell. The ocean would become gravitationally unstable and could rise to the surface, leading to a geologically recent period of intense cryovolcanism. The ammonia rich ocean could migrate upwards either through the development of Rayleigh-taylor instabilities, porous flow, or by moving through fractures.

It is unclear how much silicate material can be incorporated into the ice shell. Compaction is efficient at removing impurities as the ice shell thickens and it has been argued that the dust influx due to impacts from micrometeorites can probably only increase the silicate fraction of the ice shell by less than 0.01% (see Moore *et al.* 2009). However, the bulk mass fraction of silicates on Triton is 80%, so trapping only 3% of those silicates in the ice shell does not seem unreasonable by comparison. Silicate particles may be more difficult to remove from partially molten regions in the ice shell, due to their larger sizes. If silicate grains are suspended in the ocean during ice shell

formation, rock particles could get mixed in with ice that is deposited at the base of the ice shell. Trapping suspended silicate grains into sea ice during sea ice formation is a process known to occur in areas of the arctic ocean (Sherwood 2000). Additionally, silicate particles have been observed in water vapor plumes erupting from the south pole of Enceladus (Postberg *et al.*, 2009), proving that silicate particles are present in at least one subsurface ocean in the solar system.

The maximum possible silicate grainsize that could be suspended in the subsurface ocean can be calculated with estimates for the average convective velocity in the ocean. Soderlund *et al.* (2014) modeled ocean circulation on Europa and found the average radial velocity to be 3 cm/s and that Coriolis force created a very fast equatorial current of 250 cm/s. This could entrain large silicate particles from the ocean floor into convective currents. We estimate (using equation 1) that silicate particles with $d = 3$ mm could be readily suspended in subsurface oceans, leading to the possibility of trapping a small fraction of silicates in the ice shell.

If there are no silicates in the ice shell, then ammonia-water volcanism can likely only occur on undifferentiated or partially differentiated satellites (Kargel, 1995; Hogenboom *et al.*, 1997), and would be unlikely on Triton. However, the ammonia-rich ocean may still become buoyant if there are exsolved gasses (Crawford and Stevenson, 1988) or if ocean pressurization occurs. Manga and Wang (2007) showed that ice shell thickening can cause over pressurization in subsurface oceans and that once an ocean becomes thin, pressurization can become significant enough to drive upward flow. It is possible that once Triton's ocean became thin, ocean pressurization and increased ammonia concentration combined to cause significant upward migration of fluids through the ice shell. However this hypothesis requires more analysis to determine its physical plausibility.

6.9 Conclusion

The large temperature range between the solidus and liquidus in the ammonia water ice system leaves the possibility that there could be partially molten regions on icy satellites that contain ammonia in their ice shells. However, our models show that compaction of this partially molten region should be extremely efficient compared to the thermal evolution of the ice shell. All of the ammonia is likely to sink into the ocean and decrease the density and freezing temperature. This increased concentration makes a significant difference in the thermal evolution of Triton and can help prevent Triton's ocean from freezing, increasing the final ocean thickness by 50 km. An initial ammonia concentration of 5% in the ocean can increase to 25% after the ice shell thickens. The reduced density of the ammonia rich ocean increases the likelihood of cryovolcanic resurfacing. If the ice shell contains at least 3% silicates by mass, the ammonia-rich ocean would be gravitationally buoyant. If the ice shell were pure, ocean pressurization or exsolved gasses would be required to cause cryovolcanic resurfacing from the ocean.

Here we applied our coupled thermal evolution and phase advection models to the narrow case of the ammonia-water ice system in a thickening ice shell. Our model could be expanded to understand the evolution of other icy satellite chemical systems or even to silicate planets. More work is needed to understand how melts migrate throughout icy and silicate planets and our non-dimensional advection model can be readily adapted to understand a wide variety of scenarios.

6.10 References

- Agnor, C. B., and D. P. Hamilton 2006. Neptune's capture of its moon Triton in a binary-planet gravitational encounter. *Nature* 441(7090), 192–194.
- Arakawa, M., and N. Maeno 1994. Effective viscosity of partially melted ice in the ammonia-water system. *Geophysical research letters* 21(14), 1515–1518.
- Bargen, N., and H. S. Waff 1986. Permeabilities, interfacial areas and curvatures of partially molten systems: results of numerical computations of equilibrium microstructures. *Journal of Geophysical Research: Solid Earth* 91(B9), 9261–9276.
- Barr, A. C., and W. B. McKinnon 2007. Convection in ice I shells and mantles with self-consistent grain size. *J. Geophys. Res. Planets* 112, 2012.
- Bauer, J. M., T. L. Roush, T. R. Geballe, K. J. Meech, T. C. Owen, W. D. Vacca, J. T. Rayner, and K. T. Jim 2002. The near infrared spectrum of Miranda: Evidence of crystalline water ice. *Icarus* 158(1), 178–190.
- Brown, M. E., and W. M. Calvin 2000. Evidence for crystalline water and ammonia ices on Pluto's satellite Charon. *Science* 287(5450), 107–109.
- Burns, J. A., M. S. Matthews, et al. 1986. *Satellites*. Number 77. University of Arizona Press.
- Cooper, R., and D. Kohlstedt 1984. Solution-precipitation enhanced diffusional creep of partially molten olivine-basalt aggregates during hot-pressing. *Tectonophysics* 107(3-4), 207–233.
- Crawford, G. D., and D. J. Stevenson 1988. Gas-driven water volcanism and the resurfacing of Europa. *Icarus* 73(1), 66–79.
- Croft, S., J. Kargel, R. Kirk, J. Moore, P. Schenk, and R. Strom 1995. The geology of Triton. In *Neptune and Triton*, pp. 879–947.
- Croft, S., J. Lunine, and J. Kargel 1988. Equation of state of ammonia-water liquid: Derivation and planetological applications. *Icarus* 73(2), 279–293.

- Ćuk, M., and B. J. Gladman 2005. Constraints on the orbital evolution of Triton. *The Astrophysical Journal Letters* 626(2), L113.
- Desch, S., and M. Neveu 2017. Differentiation and cryovolcanism on Charon: A view before and after New Horizons. *Icarus* **287**, 175–186.
- Desch, S. J., J. C. Cook, T. Doggett, and S. B. Porter 2009. Thermal evolution of Kuiper belt objects, with implications for cryovolcanism. *Icarus* 202(2), 694–714.
- Golden, K. M., H. Eicken, A. Heaton, J. Miner, D. Pringle, and J. Zhu 2007. Thermal evolution of permeability and microstructure in sea ice. *Geophysical Research Letters* 34(16).
- Goldsby, D. L., and D. L. Kohlstedt 2001. Superplastic deformation of ice: Experimental observations. *J. Geophys. Res.* **106**, 11017.
- Hammond, N. P., and A. C. Barr 2014. Global resurfacing of Uranus’s moon Miranda by convection. *Geology* 42(11), 931–934.
- Hammond, N. P., A. C. Barr, and E. M. Parmentier 2016. Recent tectonic activity on Pluto driven by phase changes in the ice shell. *Geophysical Research Letters* 43(13), 6775–6782.
- Han, L., and A. P. Showman 2008. Implications of shear heating and fracture zones for ridge formation on Europa. *Geophysical Research Letters* 35(3).
- Head, J. W., and R. T. Pappalardo 1999. Brine mobilization during lithospheric heating on Europa: Implications for formation of chaos terrain, lenticula texture, and color variations. *Journal of Geophysical Research: Planets* 104(E11), 27143–27155.
- Hirth, G., and D. L. Kohlstedt 1995. Experimental constraints on the dynamics of the partially molten upper mantle: Deformation in the diffusion creep regime. *Journal of Geophysical Research: Solid Earth* 100(B2), 1981–2001.
- Hogenboom, D., J. Kargel, G. Consolmagno, T. Holden, L. Lee, and M. Buyyounouski 1997. The ammonia–water system and the chemical differentiation of icy satellites.

- Icarus* 128(1), 171–180.
- Kalousová, K., O. Souček, G. Tobie, G. Choblet, and O. Čadek 2014. Ice melting and downward transport of meltwater by two-phase flow in Europa’s ice shell. *Journal of Geophysical Research: Planets* 119(3), 532–549.
- Kalousová, K., O. Souček, G. Tobie, G. Choblet, and O. Čadek 2016. Water generation and transport below Europa’s strike-slip faults. *Journal of Geophysical Research: Planets*.
- Kargel, J. 1995. Cryovolcanism on the icy satellites. In *Comparative Planetology with an Earth Perspective*, pp. 101–113. Springer.
- Kargel, J., S. Croft, J. Lunine, and J. Lewis 1991. Rheological properties of ammonia-water liquids and crystal-liquid slurries: Planetological applications. *Icarus* 89(1), 93–112.
- Kargel, J. S. 1992. Ammonia-water volcanism on icy satellites: Phase relations at 1 atmosphere. *Icarus* 100(2), 556–574.
- Leadbetter, A. 1965. The Thermodynamic and Vibrational Properties of H₂O Ice and D₂O Ice. In *Proceedings of the Royal Society of London A: Mathematical, Physical and Engineering Sciences*, Volume 287, pp. 403–425. The Royal Society.
- Leliwa-Kopystyński, J., M. Maruyama, and T. Nakajima 2002. The water–ammonia phase diagram up to 300 MPa: Application to icy satellites. *Icarus* 159(2), 518–528.
- Lewis, J. S. 1972. Low temperature condensation from the solar nebula. *Icarus* 16(2), 241–252.
- Lodders, K. 2003. Solar system abundances and condensation temperatures of the elements. *The Astrophysical Journal* 591(2), 1220.
- Manga, M., and C.-Y. Wang 2007. Pressurized oceans and the eruption of liquid water on Europa and Enceladus. *Geophysical Research Letters* 34(7).
- McKenzie, D. 1984. The generation and compaction of partially molten rock. *Journal of Petrology* 25(3), 713–765.

- McKinnon, W. B., D. Prialnik, S. A. Stern, and A. Coradini 2008. Structure and evolution of Kuiper belt objects and dwarf planets. *The Solar System Beyond Neptune* **1**, 213–241.
- Moore, J. M., G. Black, B. Buratti, C. B. Phillips, J. Spencer, and R. Sullivan 2009. Surface properties, regolith, and landscape degradation. *Europa, edited by RT Pappalardo, WB McKinnon, and K. Khurana, University of Arizona Press, Tucson*, 329–349.
- Moore, M. H., R. Ferrante, R. Hudson, and J. Stone 2007. Ammonia–water ice laboratory studies relevant to outer Solar System surfaces. *Icarus* **190**(1), 260–273.
- Neveu, M., and S. J. Desch 2015. Geochemistry, thermal evolution, and cryovolcanism on Ceres with a muddy ice mantle. *Geophysical Research Letters* **42**(23).
- Nimmo, F., J. R. Spencer, R. T. Pappalardo, and M. E. Mullen 2007. Shear heating as the origin of the plumes and heat flux on Enceladus. *Nature* **447**, 289–291.
- Petrenko, V. F., and R. W. Whitworth 1999. *Physics of ice*. OUP Oxford.
- Piskorz, D., and D. J. Stevenson 2014. The formation of pure anorthosite on the Moon. *Icarus* **239**, 238–243.
- Porco, C., D. DiNino, and F. Nimmo 2014. How the geysers, tidal stresses, and thermal emission across the south polar terrain of Enceladus are related. *The Astronomical Journal* **148**(3), 45.
- Postberg, F., S. Kempf, J. Schmidt, N. Brilliantov, A. Beinsen, B. Abel, U. Buck, and R. Srama 2009. Sodium salts in E-ring ice grains from an ocean below the surface of Enceladus. *Nature* **459**(7250), 1098–1101.
- Prockter, L. M., F. Nimmo, and R. T. Pappalardo 2005. A shear heating origin for ridges on Triton. *Geophysical research letters* **32**(14).
- Renner, J., K. Viskupic, G. Hirth, and B. Evans 2003. Melt extraction from partially molten peridotites. *Geochemistry, Geophysics, Geosystems* **4**(5).
- Robuchon, G., and F. Nimmo 2011. Thermal evolution of Pluto and implications for

- surface tectonics and a subsurface ocean. *Icarus* 216(2), 426–439.
- Ross, M. N., and G. Schubert 1990. The coupled orbital and thermal evolution of Triton. In *Bulletin of the American Astronomical Society*, Volume 22, pp. 1127.
- Schenk, P. M. 1991. Fluid volcanism on Miranda and Ariel: Flow morphology and composition. *Journal of Geophysical Research: Solid Earth* 96(B2), 1887–1906.
- Shirley, D. N. 1986. Compaction of igneous cumulates. *The Journal of Geology* 94(6), 795–809.
- Soderlund, K., B. Schmidt, J. Wicht, and D. Blankenship 2014. Ocean-driven heating of Europa’s icy shell at low latitudes. *Nature Geoscience* 7(1), 16–19.
- Sohl, F., M. Choukroun, J. Kargel, J. Kimura, R. Pappalardo, S. Vance, and M. Zolotov 2010. Subsurface Water Oceans on Icy Satellites: Chemical Composition and Exchange Processes. *Space Science Reviews* 153, 485–510.
- Solomatov, V. 2000. Fluid dynamics of a terrestrial magma ocean. *Origin of the Earth and Moon* 1, 323–338.
- Sotin, C., J. W. Head, and G. Tobie 2002. Europa: Tidal heating of upwelling thermal plumes and the origin of lenticulae and chaos melting. *Geophysical Research Letters* 29(8).
- Stern, S. A., and W. B. McKinnon 2000. Triton’s surface age and impactor population revisited in light of Kuiper Belt fluxes: Evidence for small Kuiper Belt objects and recent geological activity. *The Astronomical Journal* 119(2), 945.
- Waite Jr, J. H., W. Lewis, B. Magee, J. Lunine, W. McKinnon, C. Glein, O. Mousis, D. Young, T. Brockwell, J. Westlake, et al. 2009. Liquid water on Enceladus from observations of ammonia and 40Ar in the plume. *Nature* 460(7254), 487–490.
- Wark, D. A., and E. B. Watson 1998. Grain-scale permeabilities of texturally equilibrated, monomineralic rocks. *Earth and Planetary Science Letters* 164(3), 591–605.

6.11 Tables and Figures

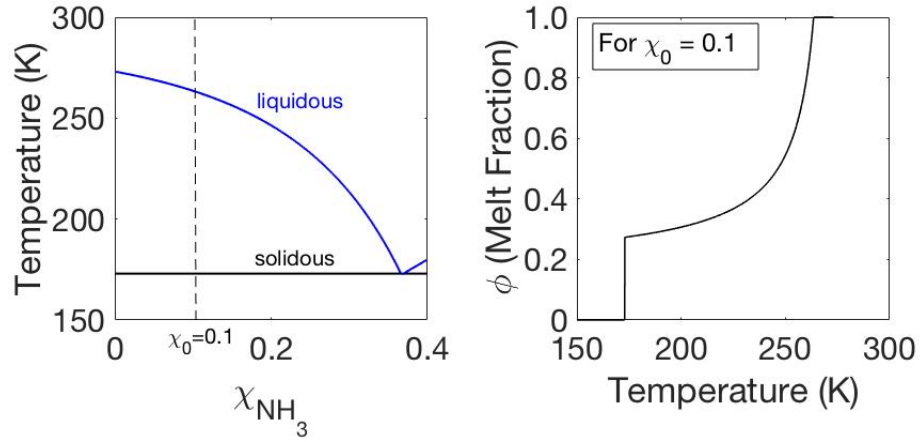


Figure 6.1: (Left) The ammonia-water phase diagram at atmospheric pressure, from experiments by (Leliwa-Kopystyński *et al.*, 2002). The blue line shows the liquidus as a function of the mass concentration of ammonia. The black solid line shows the solidus. (Right) The melt fraction (on the y-axis) as a function of temperature (on the x-axis) for a bulk ammonia concentration of $X_0 = 0.1$.

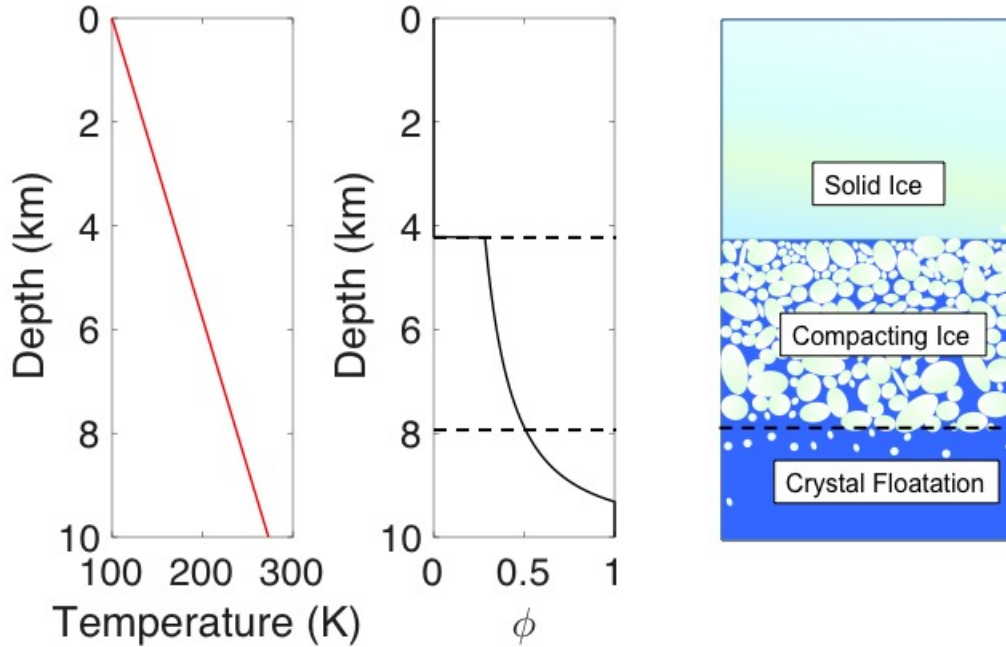


Figure 6.2: Example of the structure of an ice shell 10 km thick with a bulk ammonia concentration of $X_0 = 0.1$. In the middle panel, the top dashed line shows the depth of the solidus and the bottom dashed line shows the region when the melt fraction $\phi = \phi_c$. The region where $\phi < \phi_c$, the solids form a connected layer and solid liquid separation occurs by porous flow. When $\phi > \phi_c$ ice crystals float upwards in ammonia rich water and accumulate at the region where $\phi = \phi_c$.

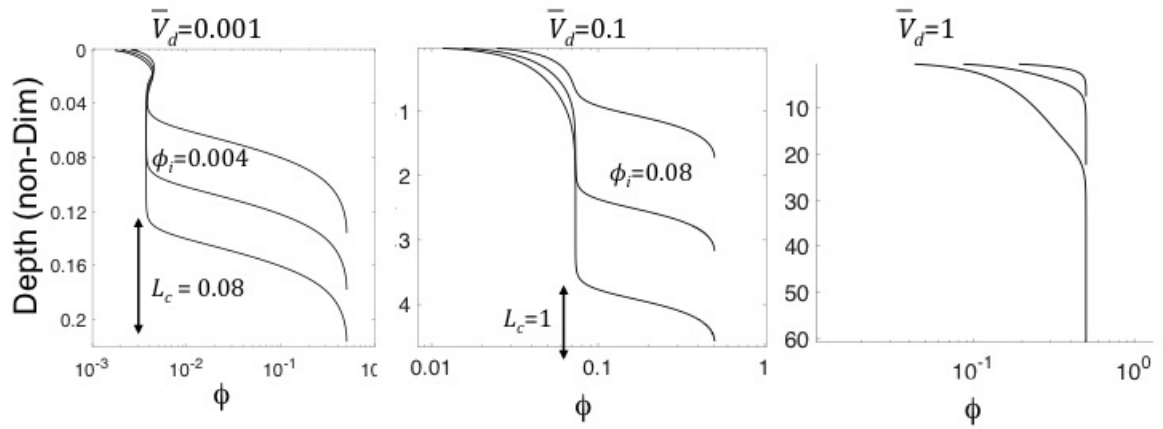


Figure 6.3: Results of our model for the compaction and growth of a partially molten layer for different non-dimensional deposition rates $\bar{V}_d$. The left, middle and right panel use a deposition rate $\bar{V}_d = 0.001, 0.1$ and 1 , respectively. These simulations use a compaction viscosity $\xi = \xi_{ref}\phi^{-1}$.

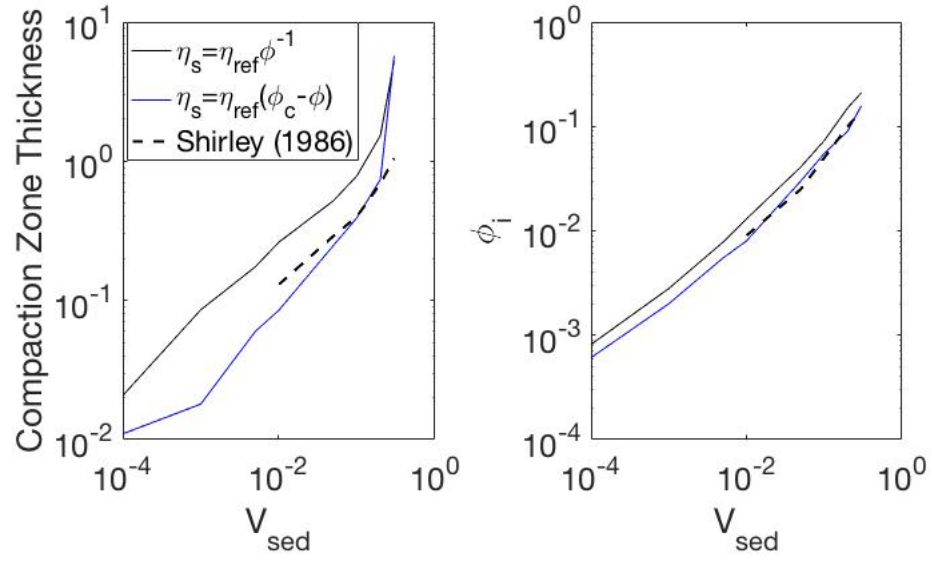


Figure 6.4: Our model results for the compaction zone thickness L_c (in non-dimensional units) and the internal melt fraction ϕ_i as a function of the deposition rate $\bar{V}_d$. The solid black and blue lines show our results using a compaction viscosity of $\xi = \xi_{\text{ref}} \phi^{-1}$ and $\xi = \xi_{\text{ref}} (\phi_c - \phi)$, respectively. The black dashed lines show the results of (Shirley, 1986).

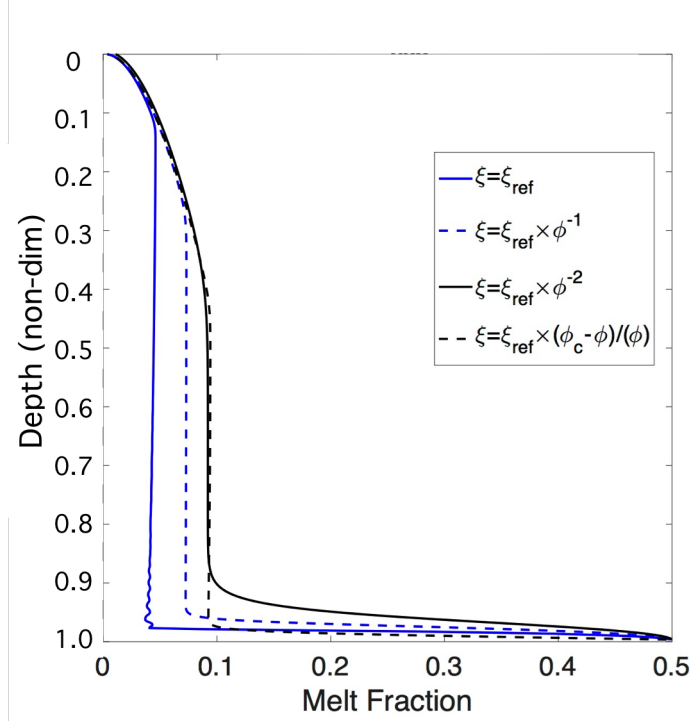


Figure 6.5: The melt fraction with depth for a non-dimensional deposition rate of $\bar{V}_d = 0.1$ and with compaction viscosities having different dependencies on the melt fraction ϕ .

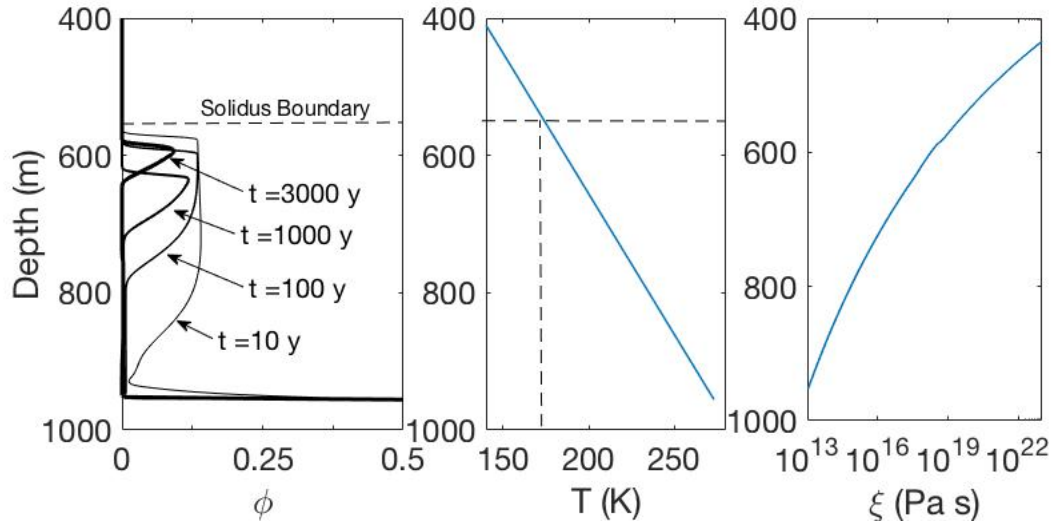


Figure 6.6: The results of our coupled advection-thermal evolution model showing the melt fraction evolution in a thin ice shell. We use a temperature dependent compaction viscosity. The panel on the left shows the melt fraction after 10, 100, 1000 and 3000 years. The dashed black line shows the solidus boundary. The center panel shows the temperature with depth, which does not change significantly over time. The right panel shows compaction viscosity which is coupled to the temperature.

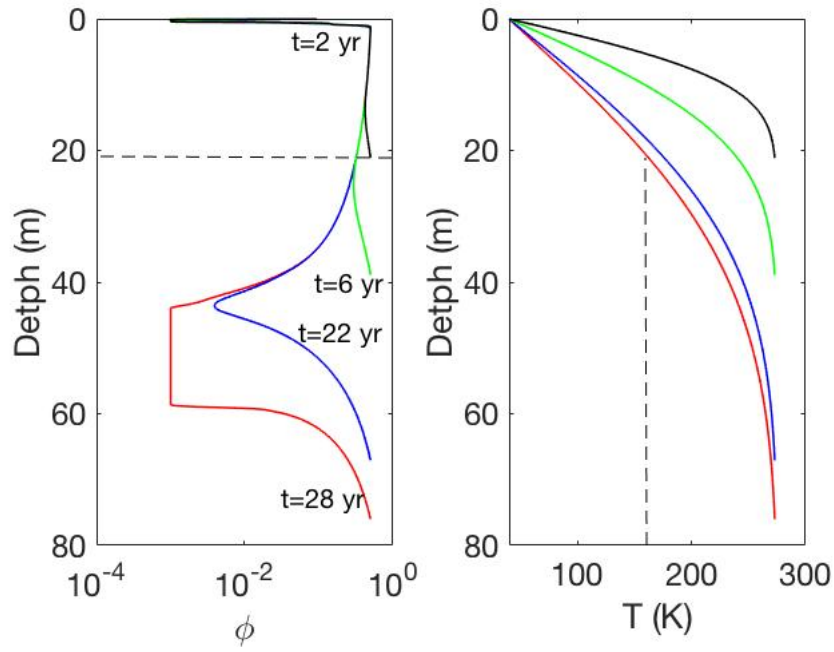


Figure 6.7: The thermal-chemical evolution of an ice-ammonia layer initially only 1 m thick. (top left) Black, green, blue and red lines show the melt fraction in the ice shell after 2, 6, 22 and 28 years. Depth below the surface is shown on the y-axis. The top panel on the right shows the corresponding temperature profile through the ice shell at these times.

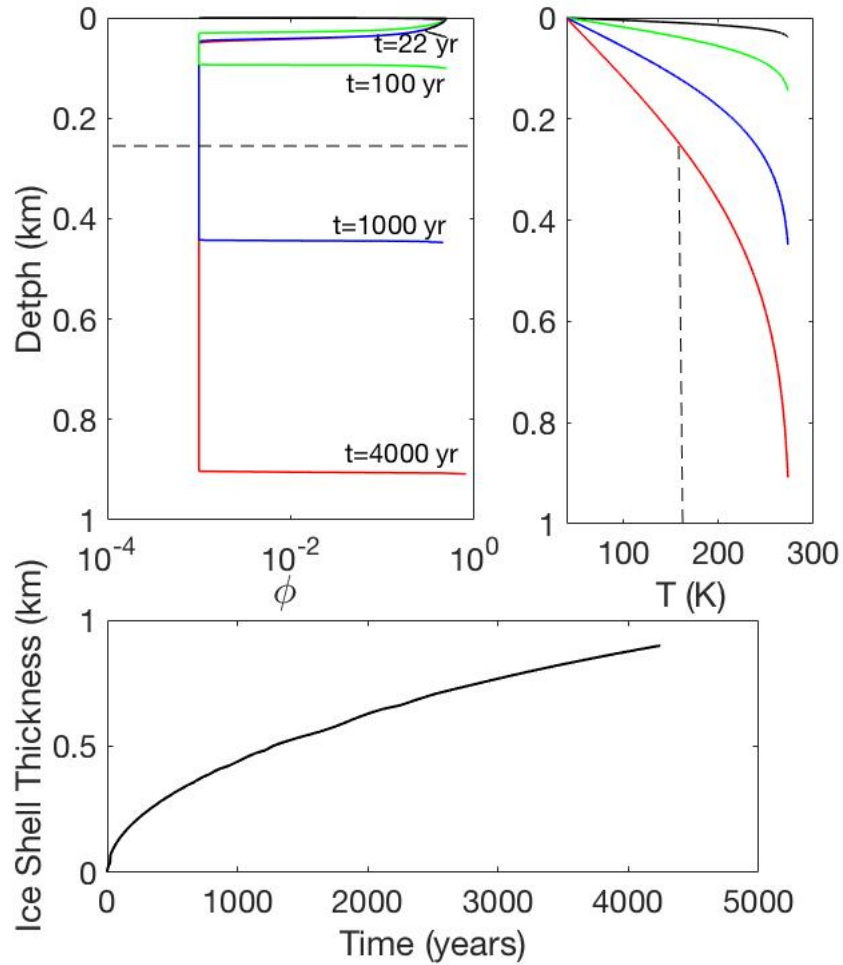


Figure 6.8: The thermal-chemical evolution of an ice-ammonia layer, over 4000 years. This is the same simulation as shown in figure 6.7 (top left) Black, green, blue and red lines show the melt fraction in the ice shell after 22, 100, 1000 and 4000 years. Depth below the surface is shown on the y-axis. The top panel on the right shows the corresponding temperature profile through the ice shell at these times. The vertical dashed line shows the solidus temperature and the horizontal dashed line shows the depth of the solidus boundary after 4000 years. (bottom) the ice shell thickness as a function of time.

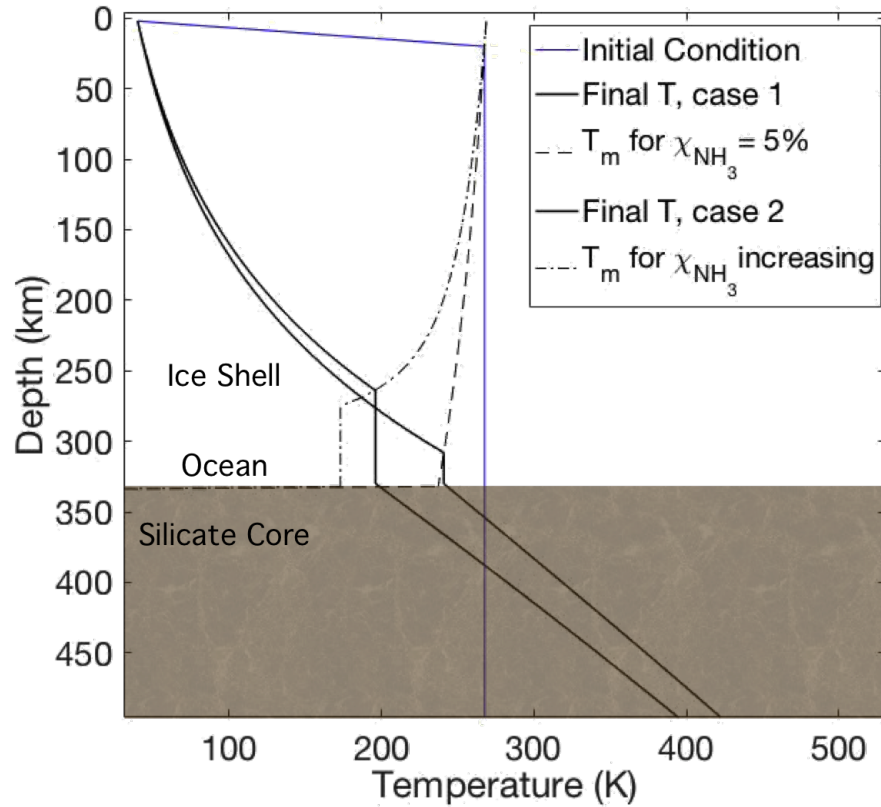


Figure 6.9: Results for the thermal evolution of Triton showing the temperature in the ice shell and the upper portion of the silicate core. The blue line shows the initial temperature profile. The solid black lines show the temperature after 4.5 Gyr of cooling. The dashed line shows the melting temperature assuming 5% ammonia in the ice shell and ocean. The dashed-dotted line shows the melting temperature assuming that ammonia becomes concentrated in the ocean over time.

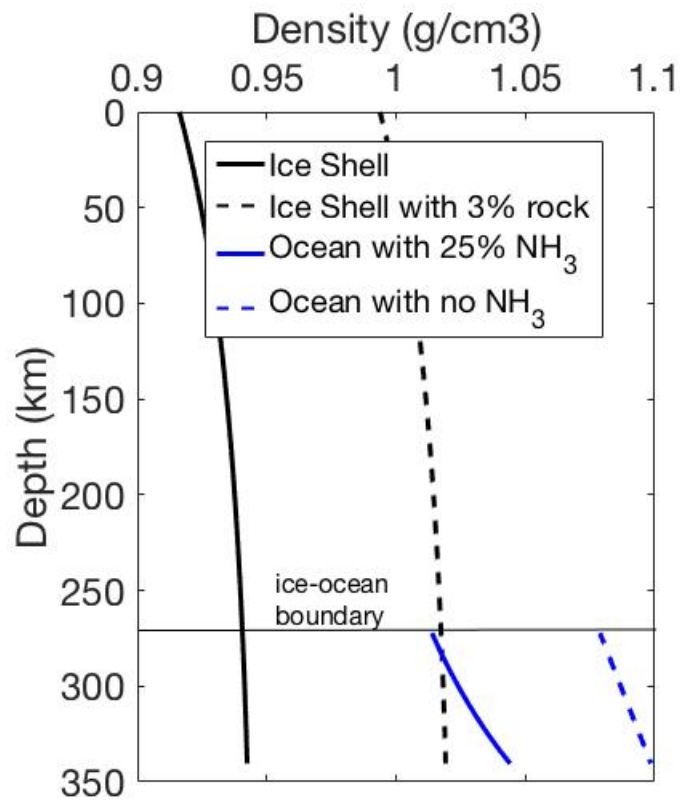


Figure 6.10: The density profile with depth corresponding to the temperature in case 2 for the thermal evolution of Triton shown in figure 6.7. The black solid and dashed lines shows the density for pure ice and ice with 3% mass fraction of silicates, respectively. The blue solid and dashed lines show the density of an ocean with 25% ammonia and no ammonia, respectively.