Linking short-term to long-term convergent tectonics using geodetic observations, geological constraints and numerical simulations

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Linking short-term to long-term convergent tectonics using geodetic observations, geological constraints and numerical simulations

Rishav Mallick

Asian School of the Environment

A thesis submitted to the Nanyang Technological University
in partial fulfillment of the requirements for the degree of
Doctor of Philosophy

2021

Statement of Originality

I hereby certify that the work embodied in this thesis is the result of original research, is free of plagiarised materials, and has not been submitted for a higher degree to any other University or Institution. I confirm that the investigations were conducted in accord with the ethics policies and integrity standards of Nanyang Technological University and that the research data are presented honestly and without prejudice.

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Rishav Mallick

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I have reviewed the content and presentation style of this thesis and declare it is of sufficient quality and grammatical clarity to be examined. To the best of my knowledge, it is free of plagiarism and the research and writing are those of the candidate except as acknowledged in the Author Attribution Statement. To the best of my knowledge, the investigations were conducted in accord with the ethics policies and integrity standards of Nanyang Technological University and that the research data are presented honestly and without prejudice.

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Emma M. Hill

Authorship Attribution Statement

This thesis contains material from three articles published in the following peer-reviewed journals, and one article in review.

Chapter 2 is published as [Mallick, R., Lindsey, E.O., Feng, L., Hubbard, J.A., Banerjee, P. and Hill, E.M., \(2019\). Active convergence of the India-Burma-Sunda plates revealed by a new continuous GPS network. *Journal of Geophysical Research: Solid Earth*. 124, 3155– 3171. <https://doi.org/10.1029/2018JB016480>](#)

- I designed the study with Eric Lindsey and Emma Hill, and carried out the analysis and modeling.
- Lujia Feng processed the GPS data and provided geodetic velocities.
- Judith Hubbard provided insight and inputs about interpreting the data and results.
- Emma Hill, Eric Lindsey and Judith Hubbard mentored me for this project.
- Paramesh Banerjee setup the GPS network and was instrumental in generating the data presented in the article.
- All co-authors contributed to writing the manuscript.

Chapter 3 is published as [Mallick, R., Meltzner, A.J., Tsang, L.L.H., Lindsey, E.O., Feng, L., and Hill, E.M., \(2021\). Long-lived shallow slow-slip events on the Sunda megathrust. *Nature Geoscience*. 14, 327–333. <https://doi.org/10.1038/s41561-021-00727-y>](#)

- I designed the study with Aron Meltzner, Louisa Tsang and Emma Hill, and carried out the analysis and modeling.
- Lujia Feng provided the GPS data.
- Aron Meltzner provided the coral data.
- Eric Lindsey and I developed the inverse modeling scheme.
- All co-authors contributed to writing the manuscript.

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- I designed the study with Judith Hubbard, Kyle Bradley and Eric Lindsey, and carried out the analysis and modeling.
- Aktarul Ahsan and A.K.M. Khorshed Alam maintained and collected the GPS data used in this article.
- James Moore and I developed the postseismic deformation code.
- All co-authors contributed to writing the manuscript.

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- I designed the study with Roland Bürgmann, compiled the data and developed the numerical models.
- Kaj Johnson and I developed the geodynamic modeling code.
- Judith Hubbard provided the structural geologist's interpretation for this study, and helped focus the content in the manuscript.
- All co-authors contributed to writing the manuscript.

We the undersigned agree with the above stated "proportion of work undertaken" for each of the above published (or submitted) peer-reviewed manuscripts contributing to the thesis:

Signed:

Rishav Mallick

Student

Asian School of the Environment
Nanyang Technological University
Date: 23-June-2021

Emma M. Hill

Supervisor

Asian School of the Environment
Nanyang Technological University
Date: 24 June 2021

Assoc Prof Benoit Taisne
Associate Chair (Research)
Asian School of the Environment
Nanyang Technological University
Date: 25 June 2021

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I did not always want to be a geophysicist or study tectonics. After my undergraduate studies, I thought I would study ocean acidification and work on the climate crisis. I credit Kusala Rajendran with introducing me to the exciting world of seismology and geophysics and enticing me with the physics of breaking rocks during my master's program at *IISc*. Fast forward 7 or 8 years and now my dreams have flowing and fractured rocks in them.

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Abstract

Faults are the brittle manifestation of plate tectonics. In accommodating the relative motion of tectonic plates, faults modify it into various steady and unsteady-state sliding phenomena. Earthquakes are generated by catastrophic and rapid unsteady fault phenomena, and are the typical association drawn between faults and society. However, faults are not just sources of disaster. They are major contributors to the processes that shape our landscape and climate (among other things) since they build geologic structure, such as mountain ranges and sedimentary basins. To gain a comprehensive understanding of faults, from their role as agents of plate tectonics to their contributions to seismic hazard and risk, I combined satellite-based observations of ground deformation with geological and mechanical constraints derived from the rock record and laboratory-derived constitutive relations.

The first step to quantify the seismic hazard posed by a fault or fault network is to estimate the rate of elastic strain accumulation; it is this elastic strain which will later be released by fault slip. I used a recently installed continuous GNSS network to estimate the geodetic velocity field of the Indo-Burma-Sunda convergent margin. Through developing a regional block motion model, I showed how oblique motion between the India and Sunda plates is partitioned between multiple blocks and that all the major faults in this system are locked. This means that the strain budget associated with the earthquake cycle on the Rakhine-Bangladesh megathrust is significantly greater than M_w 8, contrary to a long-held belief that this is an aseismic plate boundary.

When a regional plate motion model exists, deviations from the expected rigid body motion of plates arise from spatial and temporal variations in strain accumulation on active faults. I developed numerical models of strain accumulation between earthquakes and tested them against paleo-geodetic observations from Simeulue Island, offshore Sumatra, Indonesia. I demonstrated that the observations are consistent with the longest-ever recorded Slow-Slip-Event. This event lasted over 30 years, as opposed to the seconds or minutes it takes for typical earthquakes. I explain this phenomenon using a novel numerical model coupling frictional sliding with pore-fluid migration into and out of the fault zone. The spectrum of fault failure (stress drop duration) ranges all the way from seconds to decades, even though the underlying physics may differ depending on the timescale.

Embedded within geodetic observations are signals that allow us to understand how the crust can be deformed into mountains. I present one such case study from the Shillong Plateau, a ~ 1.5 km tall Precambrian basement-cored orogen that is neither a part of the Himalayas (to the north) nor is it part of the Indo-Burman mountains (to its southeast), making its origin enigmatic. On comparing our geodetic models with the exhumation history of the plateau, we find that the modern convergence across the orogen is too fast compared to what we expect given its observed long-term uplift rate. This difference is reconciled by realising that the Surma Basin, a 10-15 km foreland basin immediately

south of the plateau, may be the first documented example of ongoing passive margin subduction initiation in the Cenozoic. Subduction of old, cold crust pulls the Shillong Plateau horizontally, driving convergence and in the long-term uplifts the plateau. Though we uncovered what drives the convergence and topographic growth here, we still do not know when this growth occurs in relation to the earthquake cycle. I address this theme in the final section of this thesis.

Motivated by geodetic and geologic observations of fold-thrust belts, I developed numerical models to determine how much on-fault and off-fault permanent strain evolves over the earthquake cycle for an anticline. With this novel mechanically and geologically consistent model, we can now estimate the partitioning of permanent crustal strain into the different phases of the earthquake-cycle as it builds topography. This model can be used to make predictions of signals that exist in seismo-geodetic data in order to bridge timescales of deformation, as data acquisition and quality improve.

I conclude this thesis by noting that advances in our understanding of faulting are limited by a lack of high quality spatial and temporal observations, as well as the opaqueness of the underlying physics. Enhancements in the community knowledge base will occur mainly through a multi-pronged approach that promotes the development of physics-based predictions derived from laboratory data and then tested against field observations. The societally relevant aspiration is to use what we learn about the mechanics of faulting to better quantify the hazard and risk posed to high-density populations who live close to seismogenic faults.

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Chapter 1

Introduction

The typical image of an earthquake is one of damage, destruction and loss of life. And yet, a global estimate of high earthquake hazard regions with the population density map of the world shows a remarkable correlation. If earthquake-generating faults are agents of such catastrophe, why are we continually drawn to these geographic regions? The answer lies in an Earth system-scale view of plate boundary faults.

The plate tectonic revolution showed us that the surface of the Earth is broken up into a finite number of tectonic plates, or rigid lithospheric blocks, that float on top of a dense and viscous asthenosphere. Plate boundary faults separate these rigid lithospheric blocks and accommodate their relative motion. When localized, these faults can generate large earthquakes and trigger a host of other related natural disasters, like landslides and tsunamis. However, these faults also provide the resistive strength to hold up mountains and plateaus, which would otherwise collapse due to the action of gravity. Not only do faults keep the mountains tall, but the process of mountain building is itself primarily driven by progressive shear on faults. The hydrological and climate systems interact with the ever-changing landscape near active faults to create sources of fresh water, in glaciers and river systems, thereby enabling a cradle of civilization. Away from such a dynamic environment, it becomes more difficult for human civilization to thrive. If we are to make progress in understanding the ways in which global processes, like tectonics, can help sustain or devastate civilization, we need to elucidate the physics of faulting from the earthquake cycle time scale to the geological time scale.

Learning about active tectonics and faulting requires us to first recognise the massive scale of the operation. The spatial scale of faults range from minuscule cracks in rocks to the 1000 km structures that appear as gigantic, arcuate features on the Earth's surface. To develop a theoretical foundation for the mechanics of faulting, we generally look to laboratory-scale studies of cracks and fractures, and then extrapolate the results to the scale of crustal faults. But, we should not expect the Earth to simply conform to the predictions of fracture mechanics. Expanding our scale of interest to that of the crust, we see that it introduces a number of new physical agents that can partake in the fault slip process -

fluids infiltrated within the crust, off-fault and distributed deformation mechanisms, the presence of heterogeneities introduced by the deformation and transport history of rocks, and thermal feedbacks due to heat generation from sliding rocks, means that there are a number of diverse and complex effects that emerge in studies of faulting. While this makes the system spectacularly complicated, it also provides a richness in the observations that make this a very exciting field of study. To discover the underlying processes that govern faulting, I use simplifying assumptions from fracture mechanics to test against geophysical data. In places where they match, we can use our simplified models of the system to predict future behaviour; where they do not match, we learn something new about the Earth.

1.1 Themes addressed in the thesis

Understanding how faults slip and eventually build mountains requires us to first understand how the crust is loaded by tectonic stresses. It is the release of this stress that drives earthquakes; it is also the release of this stress that breaks and deforms the crust into various landforms. Therefore, it is imperative that we understand how the Earth's crust builds stress, and relaxes it.

The stress build up process is generally slow, evolving over decades and centuries, and can be observed using existing geodetic networks and methods. In the past, most studies of stress change in the crust were driven by observations of earthquakes, because that was when there was some signal to probe the underlying physics. With the advent of high-precision geodetic positioning, we now have direct observations of crustal motion and deformation preceding earthquakes. The geodetic timeseries during this period, called the *interseismic* period, is often observed to be linear (Figure 1.1), which allows us to capture the entire variability of such a timeseries by estimating the velocity. Away from plate boundaries, site velocities represent the rigid rotation of lithospheric plates, while close to plate boundaries, site velocities result from a combination of rigid plate motion and steady-state strain accumulation. The deviation from rigid plate motion can then be inferred as the source of steady stress build up in the crust.

The relaxation of the crust, on the other hand, is enabled by permanent (inelastic) deformation over a broad spectrum of failure modes - rapid and spontaneous slip in earthquakes, slow deformation in spontaneous Slow Slip Events (SSE), and through accelerated and steady-state stress-driven creep (Figure 1.1). Existing geodetic methods routinely identify various failure modes in the crust; in deformation timeseries, these events appear as step changes (for earthquakes) or as abrupt and significant velocity changes (SSE and accelerated creep). These events are most often interpreted as the response of an elastic crust to localized fault slip on the plate boundary.

Crossing over to longer timescales, the geological record contains only a time-integrated estimate of relaxation processes with no sensitivity to the elastic deformation history of the crust. The rock record

FIGURE 1.1: A schematic geodetic timeseries showing the combined effects of the earthquake cycle and long-term plate motion. Rapid velocity changes (shown in vertical dashes) are typically associated with stress drop events like earthquakes and SSE. With modern geodetic tools, such as GNSS (Global Navigation Satellite Systems) and InSAR (Interferometric Synthetic Aperture Radar), we only have access to a short observational window, generally the interseismic period. The zoomed in plot shows idealized displacement timeseries from the different periods of one earthquake cycle.

also shows us deformation length scales that range from localized fault slip to distributed shear, and may occur simultaneously (e.g., [Behr and Bürgmann \[2021\]](#), [Rowe and Griffith \[2015\]](#)).

Because of this, the simplifications that work for the earthquake cycle timescale do not carry to long-term deformation, and the geological record is insufficient to understand the elastic loading and release that drives the earthquake engine. Building a holistic view of crustal deformation therefore requires a multi-scale approach, both in space and in time. Below, I list the important questions I explore in this dissertation and subsequently I discuss how combining geodesy, geology and numerical observations can help tackle them.

1. At the earthquake cycle timescale, from seconds to years, we are interested in the elastic spring-like build-up and release of stress in failure episodes, specifically those that produce a net stress drop (earthquakes and SSE). These failure episodes show great diversity in style, duration and magnitude. The questions I try to answer are: *What can geodetic data tell us about an eventual failure episode? How do we know if a fault is failing and how long can this go on?*
2. Beyond the earthquake cycle, the long-term behaviour of the crust is to build geologic structure. This deformation is permanent, it is not limited or focused on a single fault, and is a consequence of inelastic relaxation processes. The questions I address are : *Why and how do faults build topography? When in the earthquake cycle does bulk deformation occur? How can*

we observe these processes in action? How do we build models that capture the various failure modes in the earthquake cycle while also being compatible with long-term kinematics?

1.1.1 Earthquakes and failure modes in the crust

An earthquake generating fault is generally thought of as a shear crack in an elastic medium (e.g., [Svetlizky and Fineberg \[2014\]](#)). Spontaneous frictional slip on a crack surface, and therefore an earthquake, is a mechanism to reduce the stress level of the entire body subjected to some external load. Typically, this phenomenon takes seconds to minutes and releases seismic waves that are the source of the violent shaking we fear. Compared to the centuries over which the stress builds up, an earthquake is a rapid stress drop process. However, there are stress drop events like SSE that are significantly slower, occurring over days to years, and do not generate the typical ground shaking that earthquakes do [[Bürgmann, 2018](#), [Saffer and Wallace, 2015](#)]. While both sets of events are thought to be governed by frictional instabilities, it is strongly debated if the same physics governs both types of failure processes (e.g., [Dal Zilio et al. \[2020\]](#), [Gomberg et al. \[2016\]](#), [Hulbert et al. \[2019\]](#), [Michel et al. \[2019\]](#), [Obara and Kato \[2016\]](#)). The common link between these different failure modes is that the section of plate boundary fault on which they will occur appears stuck in anticipation of the failure episode. The frictional locking of the fault builds stress and also creates a geodetically detectable signal. This gives us some predictive power in forecasting where slip episodes will occur in the future, even though we cannot observationally constrain the eventual slip velocity and total amount of slip (e.g., [Almeida et al. \[2018\]](#), [Avouac \[2015a\]](#)). However, this is still useful in terms of seismic hazard estimation since we can develop worst-case scenarios or use numerical simulations of earthquake sequences to estimate the probabilities of earthquake occurrence from estimated locked regions, and their associated magnitudes (e.g., [Kaneko et al. \[2010\]](#)). An example of developing such hazard-related products is in Chapter 2, where I use geodetic velocities to jointly estimate regional plate motion and the extent of frictional locking on major plate boundary faults within the Indo-Burma convergent margin.

Unlike earthquakes, SSE do not rupture catastrophically and hence do not draw much interest from a hazard perspective directly. However, these events have been observed in many different tectonic environments and tend to repeat periodically in some settings, making them excellent candidates for laboratory-like studies to understand stress evolution of plate boundary faults and the mechanics of frictional sliding at the crustal scale [[Obara and Kato, 2016](#)]. However, there is no obvious universal mechanism that explains the general characteristics of SSE (e.g., [Bernaudin and Gueydan \[2018\]](#), [Dal Zilio et al. \[2020\]](#), [Hawthorne and Bartlow \[2018\]](#), [Im et al. \[2020\]](#), [Segall et al. \[2010\]](#), [Tong and Lavier \[2018\]](#)). This is partly because the observational record of SSE is lacking, and many of the existing models appear to fit the limited data satisfactorily, thereby disallowing any statistical basis for choosing any single proposed mechanism. A related challenge faced in the field is that, as SSE get slower and possibly last longer, it becomes difficult to even distinguish an SSE from

steady-state interseismic processes. This is because detecting an SSE requires knowledge of the background sliding velocity before or after the slip event, and so we need data that spans longer than the SSE itself. Such information is not readily available for most faults due to a lack of geodetic data coverage in space, and mainly in time. In Chapter 3, I was able to overcome this issue by using non-traditional geodetic methods such as coral microatolls, as a record of tectonically-driven vertical motion. Using an observation timespan longer than a century, I show how we can combine observations with numerical simulations to detect a long-lived (32 years) SSE and investigate its operating mechanisms. Observations like these, that push the envelope of what is defined as an SSE, force us to rethink our understanding of the underlying mechanics at a more fundamental level. They also makes us question if we have been neglecting and missing these processes in existing data.

1.1.2 Permanent deformation and geologic structure

At timescales spanning multiple earthquake cycles, $10^3 - 10^6$ yrs, the elasticity of the crust is not as important as its finite transport and deformation - this occurs as slip on the main fault and secondary faults as well as bulk deformation away from the fault. At these timescales, the crust behaves more like a viscous fluid than an elastic solid [Turcotte and Schubert, 2002]. Typical geodynamic modeling techniques use this as an operating assumption and work with visco-plastic rheologies as opposed to the frictional-elastic setup more commonly employed in earthquake cycle models. On the observational side too there are open questions about how data from the earthquake cycle and the long-term deformation record can be reconciled. Geodetic data are most sensitive to a combination of plate kinematics and frictional effects in the horizontal component (Figure 1.1), while the vertical component captures the frictional locking of the plate boundary fault and some possible permanent deformation signal. However, the permanent deformation signal is generally an order of magnitude smaller than the elastic response of the crust to frictional sliding; this combined with noise in the data and imperfections in modeling choices, or *epistemic uncertainties*, obfuscate any attempt of estimating long-term vertical processes directly from data. The geological record, from rock outcrops, sub-surface imaging and structural geological interpretations, is only a snapshot of the cumulative deformation that gives us the structure that we see, but there is a vast non-uniqueness in inferring the path that parcels of rock took to get to where they are now (e.g., Lamb [2021]).

A common fix around this issue is to consider time-averaged behaviour from the separate observational timespans and techniques and then compare them. In Chapter 4, I present such a case study from the Shillong Plateau, a Precambrian basement-cored orogen that is growing within the foreland of the Eastern Himalayas. I used horizontal geodetic velocities from regional observation stations to estimate the convergence rate across the orogen, which I then combined with a geologically motivated kinematic model of plateau uplift and erosion rates to estimate rock uplift rates. On comparing these predicted rock uplift rates with geochronological and geomorphological uplift rate estimates, I found a mismatch which I then interpreted as an indication for an additional process, subduction of

the Indian slab south of the Shillong Plateau. While such comparisons of time-averaged properties can reveal many interesting phenomena, as I discuss in Chapter 4, they do not tell us about *how a mountain is built over the various parts of the earthquake cycle*. There is no easy answer to that question, but there are certain observations that can help us gain some insight into the underlying mechanics, even without an actual timeseries record of the entire process. The key here is to carefully analyse geodetic data from different parts of the earthquake cycle and demonstrate that the displacement field cannot be fit by a purely elastic crustal response to fault slip. It is important to note that this misfit cannot arise simply because of epistemic uncertainties, but must arise due to unidentified deformation sources away from the main fault. Carrying out work like this requires a combination of precise knowledge of active geological structures in the area, high fidelity seismological and geodetic observations from the earthquake cycle, and appropriate numerical simulations that can account for both the elastic and inelastic behaviour of the crust over timescales ranging from seconds to million years. I developed such a framework in Chapter 5 using a one-way coupling of a geodynamic crustal folding model with an earthquake sequence simulator. Armed with these tools and information, I discuss how to identify geodetic signals corresponding to ongoing permanent bulk deformation in fold-thrust belt settings, and how this could give us an estimate of crustal rheology.

1.2 Thesis roadmap

In the next four chapters I present research papers that are either published or in review, that expound on the themes raised in this chapter. In Chapter 2, I discuss how I constructed a regional block model using a new continuous GPS array deployed to study the convergence of the Indian plate with the Burma plate and the Eurasian plate. The goal here was to combine all the available geodetic, seismological and geological observations to develop an understanding of the tectonics of the Indo-Burma convergent margin, and assess if the major plate boundary faults were accumulating strain and presented a hazard to the millions of people living in the area. The short answer is that the faults are locked and loaded.

Deviations from the expected linearity in interseismic timeseries (Figure 1.1) appear in the geodetic record of the deformation history of the Sunda megathrust in Chapter 3. Here I used mechanical models of frictional sliding to detect the longest-recorded SSE yet, in the coral record from the 18th-19th century. I also showed that such long-lived events may be occurring on other parts of the Sunda megathrust, and the resulting signal may have previously been misinterpreted.

In Chapters 4 and 5, I focus on how geodetic observations can be used to study mountain building. Even though we know there is great diversity in failure modes in the crust, geodetic observations from the interseismic period of many tectonic environments do not reflect this complexity; they only tell us if a part of the fault is stuck or not. I exploit this ambiguity in Chapter 4 to estimate shortening rates from the geodetically-derived velocity field across a growing mountain, the Shillong Plateau, and

compare it against the long-term uplift rate recorded in the rocks and topography of this orogen. In this study, we are agnostic to when in the earthquake cycle the conversion of horizontal convergence to vertical uplift is accomplished. To elucidate how topography is built by the earthquake engine, I narrow my focus down to the scale of a single fold in Chapter 5, and develop numerical simulations that allow us to combine seismological and geodetic observations with geological interpretations of fold growth to infer the interplay of localized fault slip and bulk failure to build mountains, one fold at a time.

I end my thesis with remarks on what I learnt about the strength and limitations of various datasets and modeling techniques over the course of my time at *NTU* in Chapter 6, and discuss untackled possibilities for the future.

Chapter 2

Active convergence of the India-Burma-Sunda plates revealed by a new continuous GPS network*

Abstract: The Rakhine (Arakan)-Bangladesh megathrust, along which the Indian and Burma plates collide, is assumed by some to be inactive/aseismic due to the lack of notable interplate earthquakes in the modern instrumental catalog. However, geological and historical evidence of the great 1762 Arakan earthquake suggest the megathrust can produce $M \sim 8$ events that could adversely affect the lives of millions of people in the region. To investigate the seismogenic potential and determine the slip budget of the megathrust we first need to solve for India-Burma-Sunda relative plate motions. We present a new set of 24 GPS velocities (2011-2017) from the MIBB (Myanmar-India-Bangladesh-Bhutan) continuous GPS network. We use the new velocities and those from previously published studies to explore the geometries and slip rates of three major faults (Rakhine-Bangladesh megathrust, Churachandpur-Mao Fault (CMF), and Sagaing Fault) that accommodate the India-Burma-Sunda plate motion. Our results suggest that the three major faults we studied are likely fully coupled; the modern shortening rate across the Burma plate is 12-24 mm/yr, while the total dextral shear rate is 25-32 mm/yr. The possibly fully coupled shallow megathrust, and splay faults that may sole into it, are geodetically invisible while they are not slipping. However, we can identify the transition from coupling to steady creep on the deeper extension of the megathrust; we use this to show active oblique India-Burma convergence and to map along-strike and along-dip variations in dip-slip and strike-slip motion. This implies that the megathrust is currently accumulating strain which will eventually be released in earthquakes.

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2.1 Introduction

The Rakhine-Bangladesh megathrust (variously known as the northern Sunda megathrust south of 18° N, the Rakhine (or Arakan) megathrust from 18°–22° N, and the Bangladesh megathrust from 22°–25° N) is the plate boundary between the Indian and Burma plates (Figure 2.1). The megathrust, which accommodates the subduction of the Indian plate beneath the Burma plate, has resulted from the oblique collision between the Indian and Sunda plates since the Eocene [Meng et al., 2012]. The obliquity in India-Burma motion is accommodated by some combination of the megathrust, Churachandpur-Mao Fault (CMF) and other unmapped structures within the Indo-Burman Wedge (IBW). To the east of this oblique convergent margin, the 1200 km long right-lateral strike-slip Sagaing Fault forms the eastern boundary of the Burma plate and accommodates the remainder of the relative plate motion between the Indian and Sunda plates [Gahalaut et al., 2013, Maurin et al., 2010, Socquet et al., 2006, Steckler et al., 2016, Vigny et al., 2003]. This system of faults terminates northward into the Shillong Plateau and the Eastern Himalayan Syntaxis.

From south to north along the megathrust (18°–25° N), there is a dramatic increase in the cumulative sediment input from the Ganga and Brahmaputra river systems into the Bengal Basin, the thickness of which varies from 8 to 20 km [Radhakrishna et al., 2010]. The collision of these sediments with the Burma plate has subaerially exposed subduction/accretion structures within the overriding plate, such as the outer and inner IBW. The outer IBW is composed of Neogene clastic sequences affected by a series of asymmetric folds [Maurin and Rangin, 2009]; this is well expressed in Bangladesh as the Chittagong–Tripura Fold Belt (Figure 2.1). The inner IBW rises as the Indo-Burman Ranges (IBR), further divided into the western IBR, consisting of Cretaceous flysch overlain by Paleogene sandstone, and eastern IBR, comprising Triassic flysch and Jurassic ophiolites [Sloan et al., 2017].

The western IBR hosts the CMF, which is likely an over-steepened Neogene thrust [Uddin and Lundberg, 1998], currently thought to be reactivated as a dextral strike-slip fault [Gahalaut et al., 2013, Wang et al., 2014]. The CMF is clearly mapped at the surface near the Imphal Valley (Figure 2.1), south of which it either terminates or becomes obscured. This fault is hypothesized to accommodate a significant amount of dextral shear (18 mm/yr) resulting from the India-Burma plate motion [Gahalaut et al., 2013, Wang et al., 2014].

On the eastern edge of the eastern IBR, the east-dipping Kabaw Fault separates two blocks of contrasting material, the Triassic flysch and Jurassic ophiolites of the eastern IBR and the Cretaceous sandstones overlain by young fluvial sediments of the Myanmar Central Basin (MCB) [Nielsen et al., 2004, Sloan et al., 2017]. The contribution of the Kabaw Fault to the strain budget of the India-Burma-Sunda system is uncertain due to an absence of clear geomorphic indicators of recent activity [Wang et al., 2014]. The eastern edge of the MCB is marked by the Sagaing Fault, a dextral strike-slip fault accommodating 18–20 mm/yr of the Burma-Sunda motion in an almost purely north-south direction [Bertrand et al., 1998, Maurin et al., 2010, Vigny et al., 2003, Wang et al., 2014].

The description of the tectonic setting highlights that the deformation associated with the eastern boundary of the Burma plate, the Sagaing Fault, is well constrained from neotectonic studies [Bertrand et al., 1998] as well as modern geodetic investigations [Maurin et al., 2010, Vigny et al., 2003]. However, modern contributions of active structures within the MCB and the IBW to the India-Burma-Sunda plate motion are poorly understood. This makes it difficult to assess the current deformational field and has led to controversies and contrasting opinions about major structures in this tectonic area. Here, we focus on the question of ongoing convergence between the Indian and Burma plates, and the total slip budget available to faults in the IBW.

2.1.1 Convergence of India and Burma plates

The presence of a subducted Indian slab imaged beneath the Burma plate implies that subduction of the Indian slab occurred sometime in the past, and drove the convergence between the India and Burma plates [Bijwaard et al., 1998, Pesicek et al., 2010, Raoof et al., 2017]. However, the absence of notable seismicity on the interface between the India and Burma plates has led some authors to speculate that subduction of the Indian slab has ceased or continues purely aseismically [Le Dain et al., 1984, Ni et al., 1989]. The north-south directed compressive axes of well-located earthquakes ($M_w > 4$) within the downgoing Indian slab have been used to suggest that the greatest stressing rate is due to collision of India and Eurasia, not the subduction of India beneath Burma [Gahalaut et al., 2013, Le Dain et al., 1984, Ni et al., 1989]. Deep intra-slab earthquakes (> 90 km) display tensional axes aligned with slab dip and based on this some authors [Rangin et al., 2013, Rao and Kalpna, 2005] propose that the slab is beginning to break off and subduction has ceased and the convergence, if any, must be driven by upper plate processes.

The geology records the history of the upper plate. Youthful deformation of anticlines on Sandwip Island, and possibly the Comilla Tract (Figure 2.1) and uplifted marine terraces on the islands of Cheduba and Ramree (Figure 2.1) indicate continued inelastic deformation of continental shelf sediments in the Holocene [Aung et al., 2008, Brunnschweiler, 1966]. These observations indicate that the active deformation front of the megathrust extends at least to the mouth of the Ganga-Brahmaputra delta in Bangladesh, presumably accommodated by the mapped fold and thrust belt [Maurin and Rangin, 2009, Nielsen et al., 2004, Steckler et al., 2008]. Further south, in Myanmar, the Rakhine megathrust seems to slip on average every 500-1000 years and may have slipped all the way to the trench in the great 1762 Arakan earthquake [Wang et al., 2013].

In summary, the seismological observations have been interpreted to imply ceased or aseismic subduction while the geological evidence suggests there is continued convergence between the India and Burma plates that results in earthquakes.

FIGURE 2.1: Overview of the study area discussed in this paper. MIBB cGPS stations are shown as yellow diamonds, while other published GPS velocities [Gahalaut et al., 2013, Kreemer et al., 2014, Steckler et al., 2016] are shown as white circles. GCMT earthquake locations (1950-2017) are shown, and colour coded by depth. The Indo-Burman Ranges are divided into the western IBR and eastern IBR based on their geology. The rupture area of the 1762 M $\sim$ 8 Great Arakan earthquake is marked in purple. The black arrows show the direction and magnitude (in mm/yr) of motion of India with respect to the Shan block from our model. IBR – Indo Burman Ranges.

2.1.2 Plate kinematics from geodetic studies

Past geodetic studies of the India-Burma collision zone are antithetical about the convergence across the Rakhine-Bangladesh megathrust. These studies typically use computationally inexpensive, two-dimensional dislocation models, neglecting along strike variations in relative plate motion. [Gahalaut et al. \[2013\]](#) modeled campaign and continuous GPS in India and Myanmar using a two-dimensional elastic dislocation model. They infer a strike-slip rate of 18 mm/yr between the India and Burma plates accommodated on the CMF, with the shallow interface creeping interseismically at ~ 8 mm/yr. They note that there is no significant fault perpendicular motion across their 200 km-wide profile, implying no active convergence across the northern IBW. In this model, the India-Burma plate boundary is a strike-slip fault located within the IBW, with the rest of India-Sunda motion localized on the Sagaing Fault. More recently, [Steckler et al. \[2016\]](#) used two-dimensional models of horizontal GPS velocities from a wider area (comprising Bangladesh, India and Myanmar) to identify active convergence and elastic strain accumulation between the India and Burma plates. They infer a convergence rate of 13-17 mm/yr between the India and Burma plates within the central IBW, with the Rakhine-Bangladesh megathrust the most likely candidate for the fault accommodating this strain budget. [Steckler et al. \[2016\]](#) do not explicitly invert for the strike-slip component of relative plate motion but rather impose published slip rates for the Sagaing Fault and CMF to fit the data. They further hypothesize that oblique motion on the Kabaw Fault could accommodate up to 6 mm/yr of the India-Sunda plate motion.

This discrepancy in estimated convergence across the megathrust is primarily due to differences in data coverage. The addition of data from Bangladesh allows [Steckler et al. \[2016\]](#) to fit an elastic dislocation model to a data transect that is 1000 km wide and that shows an increase in the fold-belt-perpendicular velocities from Bangladesh to Myanmar (Figure A.2). However, both models impose relative plate motion as a simple translation, neglecting the fact that on a spherical earth there are no rigid-body translations; there are only rigid rotations. A possible way to tackle this problem is to use block modeling, which correctly accounts for spatially variable plate velocities by considering plate motions on a sphere [[McCaffrey, 2002](#), [Meade and Loveless, 2009](#)]. Block modelling of the India-Sunda motion with GPS data has been attempted in the past with sparse datasets [[Socquet et al., 2006](#)], and was used to infer 36-39 mm/yr of total relative India-Sunda motion between 18° N and 24° N. The Sagaing Fault takes up 18 mm/yr of this motion as dextral strike slip, while the rest is accommodated on the Rakhine-Bangladesh megathrust and/or faults in the IBW. [Socquet et al. \[2006\]](#) used mostly campaign stations in Myanmar and provided only a preliminary view of the India-Burma-Sunda system. We can now refine and update those results. In this paper, we complement previously published GPS data with the new continuous Myanmar-India-Bangladesh-Bhutan (MIBB) GPS network, and use these data to model the deformation field in this complex tectonic region. We model the convergence and dextral shear rates across the region using block rotations and elastic dislocation models, and use our results to answer the following fundamental questions:

FIGURE 2.2: Map of MIBB station velocities in the India reference frame presented in this paper are shown in yellow. The labels refer to their 4-character unique station ID. We combined these velocities with published velocities [Gahalaut et al., 2013, Kreemer et al., 2014, Steckler et al., 2016], shown in dark green. The uncertainty ellipses show the 95% confidence interval of the mean velocities associated with each station.

1. Is there geodetically observable convergence across the Rakhine megathrust and its northern extension, the Bangladesh megathrust?
2. How is strain partitioned between the megathrust and upper-plate structures?

2.2 GPS data and analysis

FIGURE 2.3: Three-component position time series for 24 MIBB stations in ITRF08 [Altamimi et al., 2011]. Each of the components of the time series have been assigned an arbitrary constant shift to aide in visualization. The black lines show the fit to the time series at every station using the method described in Feng et al. [2015]. The stations are arranged from top to bottom: 2 stations in Bhutan, 7 stations in Bangladesh, 7 stations in India and 8 stations in Myanmar.

2.2.1 MIBB GPS data

The Myanmar-India-Bangladesh-Bhutan (MIBB) GPS network is a continuously operating regional network that was established in 2011. In this paper, we report the time series and velocities of 24

stations in the network (2011-2017) (Figure 2.2-2.3). These are distributed as east-west and north-south transects crossing the Bangladesh megathrust, CMF, Sagaing Fault, Dauki Thrust and the Main Frontal Thrust of the eastern Himalayas (yellow vectors in Figure 2.2).

2.2.2 GPS data processing

We process data for all 24 stations for the period of 2011-2017 using the GPS-Inferred Positioning System and Orbit Analysis Simulation Software (GIPSY-OASIS) version 6.2 and NASA Jet Propulsion Lab (JPL) final precise satellite orbits and clocks. We solve for the daily positions in the International Terrestrial Reference Frame 2008 (ITRF2008) (Altamimi et al. [2011]). We use the procedure described in Feng et al. [2015] to calculate interseismic velocities. In short, we model the time series at each station as a combination of two main signals: (1) a steady interseismic velocity, which we assume is the linear rate at which plate motion is accommodated between earthquakes, (2) seasonal oscillations, modelled as annual and semi-annual sinusoids. For the stations SDWN and SWBO, which showed significant coseismic and postseismic signals from the M_w 6.8 Thabeikkyin earthquake in 2012 (Figure 2.3), we added signals related to earthquake deformation (i.e., heaviside functions for coseismic offsets and logarithmic decays for postseismic processes). We used a least-squares estimator to calculate the interseismic velocity at each station [Feng et al., 2015]. To calculate the uncertainties in the velocities we compute the inverse of the Jacobian, which is a matrix of partial derivatives of the data with respect to all the model parameters, and scale it using the temporal correlation of the residuals with the First Order Gauss-Markov EXponential (FOGMEX) algorithm [Herring, 2003].

2.2.3 Station velocities

We model interseismic velocities as the superposition of the velocity field from rigid block rotations and elastic strain accumulation on kinematically coupled plate interfaces. We exclude stations affected by co- and post-seismic processes (SDWN and SWBO) since we are not investigating the effects of the earthquake cycle in this study. We combined the velocities from our dataset with other published velocities for the region using the solutions obtained by Gahalaut et al. [2013], Kreemer et al. [2014] and Steckler et al. [2016] in ITRF08. We rotate all the velocities into the Indian reference frame using the India-ITRF08 Euler pole (latitude = 51.42° , longitude = 2.10° , rotation rate = 0.5146×10^{-6} %/yr) presented in Steckler et al. [2016] (Figure 2.2). We limit our analysis to strain accumulation in the Burma plate between 22° N - 25° N, which is where we have good station coverage. We do not use the vertical velocity components, since local site effects and extreme seasonal variations in atmospheric and hydrological loading likely contribute a larger signal than tectonics (Figure 2.3).

2.3 Inversion and results

FIGURE 2.4: (A) Comparison of data and model fits from our preferred block model (rigid blocks; dip-slip prior of 0 ± 4 mm/yr on the CMF; strike-slip prior of 20 ± 2 mm/yr on the Sagaing Fault). We find that a fully coupled fault model can explain the data reasonably well. This confirms our null hypothesis that the residual distribution (see Inset figure) is a standard normal distribution; this implies that our model is a statistically feasible representation of reality. HAKA and KANI are marked in this figure to show that we do not need to invoke slip on the Kabaw Fault (KBF) to fit the velocities. (B) Zoom into the CMF. The stations HENG and CCPR ~ 4 km apart but show dissimilar velocities, while sites to the north show a continuously varying velocity field indicating local site effects, not plate boundary creep. CMF - Churachandpur-Mao Fault; KBF – Kabaw Fault.

We parameterize our preferred model through an objective optimization of model parameters and prior constraints using Akaike's Bayesian Information Criterion (ABIC) following [Funning et al. \[2014\]](#). Our model parameters $\mathbf{m}^*$ are block translation rates and fault geometry parameters (for the 2-D problem); for the 3-D block modeling problem our model parameters are the angular velocity vectors for each block, the horizontal strain rate tensor for internally deforming blocks and the weight of the dip-slip penalty hyperparameter $\sigma_{\text{slip rate}}^2$ (Section A.4).

For a given dataset $\mathbf{d} = \begin{bmatrix} \mathbf{d}_{GPS} \\ \mathbf{d}_{\text{slip rate}} \end{bmatrix}$ consisting of observed GPS velocities, and a priori known slip rates, with data uncertainties embedded in the covariance matrix $C_d = \begin{bmatrix} C_{dGPS} & 0 \\ 0 & \sigma_{\text{slip rate}}^2 C_{d\text{slip rate}} \end{bmatrix}$ and

a linear model $G = \begin{bmatrix} G_{GPS} \\ G_{\text{slip rate}} \end{bmatrix}$, we define ABIC as follows:

$$ABIC(\sigma_{\text{slip rate}}^2) = N \log S(\mathbf{m}^*) + \log \left| G^T W(\sigma_{\text{slip rate}}^2) G \right| + \log \left| W^{-1}(\sigma_{\text{slip rate}}^2) \right| + \kappa \quad (2.1)$$

We define the weighting function W as C_d^{-1} . κ is independent of $\sigma_{\text{slip rate}}^2$ but depends on the number of model parameters and hyperparameters.

Here $S(\mathbf{m}^*) = (\mathbf{d} - G\mathbf{m}^*)^T W(\sigma_{\text{slip rate}}^2) (\mathbf{d} - G\mathbf{m}^*)$ and $\mathbf{m}^* = (G^T W G)^{-1} G^T W \mathbf{d}$ is the least-squares solution to the linear inverse problem [Menke, 2012]. We define $N = N_{GPS} + N_{\text{slip rate}} - M$ where N_{GPS} is the total number of GPS observations, $N_{\text{slip rate}}$ is the total number of slip rate constraint equations and M is the total number of model parameters. The operator $|A|$ on a square matrix A is defined as the product of the absolute values of its non-zero eigenvalues.

We also test for the significance of error improvement between different models A and B using the F-test [Menke, 2012]. The test statistic is defined as -

$$F_{AB} = \frac{\chi_A^2 / DF_A}{\chi_B^2 / DF_B} \quad (2.2)$$

We test the probability $p \left(F < \frac{1}{F_{AB}} \text{ or } F > F_{AB} \right)$ of the improvement in error having occurred by random chance using the asymmetric F-distribution. Here DF refers to the degrees of freedom in the model.

2.3.1 Constructing fault geometry

We assume that the shallow part of each fault is fully coupled (not slipping) while the deeper extension, below the brittle-ductile transition, is creeping steadily at the plate rate; this coupled-uncoupled transition, henceforth referred to as the coupling transition, has a significant geodetic signal associated with it which we can fit with an elastic model (Figure A.1-A.2). Our model has three major faults (the Rakhine-Bangladesh megathrust, CMF, and the Sagaing Fault), and therefore three coupling transitions. We find the best-fitting model geometry by first using two-dimensional semi-infinite dislocations (representing the creeping side of the transition) to investigate the locations of these coupling transitions, and the magnitude of strain accumulation. We use a Markov-Chain-Monte-Carlo method to sample a suite of possible models that can fit the GPS dataset (for a detailed description of the method see Section A.1).

We find that the horizontal locations of the deep coupling transitions for the megathrust and CMF are merely 50 km away from each other (Figure A.2), located below the inner IBW. Our constraints on depth are poor (Figure A.3-A.4; Section A.2.1), but our models suggest that the depths of the coupling transitions must be greater than 25 km, with a best fitting value of 30 km for both faults (Section A.2.1). The slip rate and deep coupling transition location of the Sagaing Fault are much

FIGURE 2.5: Results of the block modelling exercise; we show the results from our preferred model (Figure 2.4). The arrows show estimated plate motion vectors within the blocks relative to India in a regular grid. The inset figure shows the Euler poles computed for each of the blocks relative to India. IBW-Indo-Burman Wedge; MCB-Myanmar Central Basin Block. The formal uncertainties from the model covariance matrix is approximately 0.2 mm/yr for dip-slip and strike-slip rates on the fully-coupled faults, however a more realistic measure of uncertainty is ~ 2 mm/yr, and can be visualized from the range of mean slip rates in Figure A.6-A.8. CMF - Churachandpur-Mao Fault; IBW-Indo-Burman Wedge; MCB-Myanmar Central Basin Block.

better constrained (Figure A.6); it is consistently ~ 5 km east of its surface trace, implying that the fault dips $\sim 70^\circ$ towards the east. Using the location of the inferred coupling transition for the down-dip extent of the faults, and the mapped traces (CMF and Sagaing Fault) or the inferred trace of the deformation front (Rakhine-Bangladesh megathrust) for the up-dip extent of the fault, we create arcuate fault geometries for the shallow portion of the three faults (Figure 2.4). This step is necessary because block modelling treats elastic dislocations in the back-slip sense while our 2D models used deep dislocations. However, both methods are most sensitive to the horizontal location of the coupling transition, which creates the strongest surface velocity gradients. We use this three-dimensional geometry to set up a block model on a spherical earth using the Blocks software [Meade and Loveless, 2009].

2.3.2 Block modeling

The total width of the study area is ~ 1000 km. At this scale, the curvature of the Earth and resulting plate rotations play a significant role in the observed surface velocity field. The interseismic velocities

FIGURE 2.6: (A) and (B) Results of grid-search for dip and depth of the Rakhine-Bangladesh megathrust using 3-D rigid block models. We apply a dip-slip prior of 0 ± 1 mm/yr on the CMF (B) and no priors on the CMF in (B). The red star indicates the best fitting (lowest misfit) model parameter. The contours show that the model is not very sensitive to dip (δ) and depth (D) independently, but is sensitive to $x_d = \frac{\delta}{\tan D}$ which is the horizontal location of the coupling transition, just like in the 2-D models (Section A.1). (C) We compare the reduced χ^2 statistics of our different modelling choices. Golden yellow bars - 2D and 3D block models whose geometries are constructed using the geometries inferred from the 2D models; blue bars - 3D models in which we explicitly solved for the geometry of the Rakhine-Bangladesh megathrust. (D) We show the trade-off between slip on the CMF and strain-rates within the IBW. Golden yellow regions - 95% confidence interval of x_d inferred from 2D models; the blue regions - 95% confidence interval for x_d (only for Rakhine-Bangladesh megathrust) inferred from all our 3D block models. CMF - Churachandpur-Mao Fault; IBW - Indo-Burman Wedge; MCB - Myanmar Central Basin; DS prior - Dip-slip prior

for stations far from plate boundaries vary due to rigid body rotations about an Euler pole. Deviations from these block-like motions arise due to strain accumulation on faults (block boundaries), or to internal strain within the block itself [McCaffrey, 2002, Meade and Loveless, 2009]. We use the software Blocks, to carry out our modelling [Meade and Loveless, 2009].

We choose the Indian Block as our “reference block” with respect to which we define the velocity field. We include interseismic velocities from the IGS reference stations HYDE, IISC and BAN2 to define the stable Indian Block. We create a model with four blocks: the Indian Block (INDIA), Indo-Burman Wedge (IBW), Myanmar Central Basin Block (MCB) and Shan Block (SHAN). A point to note here is that our GPS stations do not directly sample the Sunda plate, and so we use the Shan block as a proxy for the Sunda plate.

These blocks are bounded by the megathrust, the CMF, and the Sagaing Fault (Figure 2.4-2.5). These faults accumulate strain and cause distortions in the surface velocity field that we model using the back-slip formulation [Savage, 1983]. We allow a uniform internal strain-rate field in some of our models as it acts as a proxy to describe deviations from the ideal “rigid elastic” description of a block, instead of breaking up the IBW into further smaller blocks (e.g., Thatcher [2007]). This deviation might be due to slip on multiple faults within the IBW or to diffuse deformation of the wedge material within the IBW; we cannot distinguish between these options given our current station density.

We compute the best fitting Euler poles from Blocks using the geometry we inferred from the 2D models. We then fix the geometry of the CMF and Sagaing Fault, and use the gridsearch algorithm to determine the best-fitting coupling transition depth and fault dip for the Rakhine-Bangladesh megathrust (Figure 2.6A, B), which are non-linear parameters, and compare it to the 2D case. We explore a suite of models with varying levels of dip-slip penalization on the CMF, for cases of a rigid IBW as well as an internally deforming IBW and test the statistical significance of these models.

2.3.3 Results

Our block modeling results are shown in Figure 2.4-2.7, A.9. All models show convergence between the Indian Block and the IBW, with at least 12 mm/yr of shortening/arc-perpendicular motion between the blocks (Figure 2.5, A.6-A.8). Our preferred model fits are shown in Figure 2.4, while the model predicted slip rates and plate motions (relative to India) are shown in Figure 2.5. This model is one of the statistically acceptable representations of this tectonic region (minima in ABIC in Figure 2.7A) but does not reveal all the complexities or trade-offs in the information we have learned from studying the data. We elaborate on the robust results we have gathered from a suite of models below.

The observed convergence can be explained by assuming full coupling on the shallow section of the Rakhine-Bangladesh megathrust; the estimated slip rate on the megathrust varies spatially from 12-24 mm/yr with an average of 18 mm/yr (Figure 2.7). The velocity field resulting from elastic dislocations is almost entirely east-west in the outer IBW (Figure 2.6D). Since we approximate the megathrust

as a north-south structure we obtain minimal strike-slip components, however this would change if we make the megathrust more arcuate (Figure A.8). We observe a significant difference in the geometry of the megathrust inferred from 2D and 3D models (Figure 2.6C, D). 2D models require an exaggerated arcuate shape for the megathrust, with the horizontal position of the coupling transition very close to the CMF; however, the misfit from such a model is significantly large compared to the block models in which we explicitly solved for fault geometry (Figure 2.6C), such that it can be ruled out at the 95% confidence level using an F-test. Block models with different modelling choices, such as different penalties on the CMF or allowing internal strain in the IBW, vary in the estimated dip and depth of the coupling transition but yield a robust estimate of the horizontal location of the coupling transition for the Rakhine-Bangladesh megathrust (Figure 2.6A, B, D). We chose a preferred block model using ABIC: the model with $\sigma_{CMF} = 4$ mm/yr and no internal-strain in the IBW (Figure 2.7A, A.9).

The slip rate estimates of the CMF are not very robust and trade-off significantly with possible internal deformation within the IBW (Figure 2.6D, 2.7). However, we observe that in all our models there needs to be a northward translation of the inner IBW even though the mechanism is unclear. Models in which we permit internal strain-rate within the IBW almost entirely shut off the CMF while producing a velocity field remarkably similar to that expected for a purely right-lateral strike slip fault (Figure 2.6D, 2.7). The ABIC values of these models are greater than that of the preferred model, but they cannot be rejected at the 95% confidence level of an F-test (Figure A.9). This means that even though we cannot be certain of the mechanism of deformation in the inner IBW, there must be dextral slipping fault(s) that accommodate this deformation. In models in which we do not impose any constraints on the rake of the CMF, the model adds some thrust and normal motion along segments of the fault, which appears to be unphysical (Figure 2.7, A.7-A.8). A more careful study of the variation of the inferred rake when varying the strength of the dip-slip penalization on the CMF also shows that the rake becomes more variable and tends to predict significant dip-slip when the penalization is weak (Figure 2.7). This is an artifact of fitting noisy data that we do not think is geologically relevant, and thus we use the requirement of limited fault-perpendicular motion on the CMF to place an upper bound on its possible strike-slip rate at ~ 8 mm/yr (Figure 2.7).

The slip rate on the Sagaing Fault (~ 19.5 mm/yr) falls within our prior assumption of 20 ± 2 mm/yr dextral strike-slip. We apply these priors because the slip rate on the Sagaing Fault can be independently constrained in the two-dimensional models (Figure A.4), and they improve the stability of the solution to our inverse problem. However, our model predicts a component of dip-slip motion on the Sagaing Fault as well, reaching ~ 4 mm/yr at 24° N. To test if this observation is robust, we ran an experiment with an extra block in the northern Shan plateau, similar to the geometry of Socquet et al. [2006] (Figure A.8B). The addition of this block does not significantly affect the fit to the data but removes all dip-slip motion from the Sagaing Fault. Thus, we suggest that this dip-slip motion is a consequence of the lack of data to constrain the rotation rate of the northern Shan block. Since the

stations in this block are close to faults in the Shan Plateau, these sites cannot independently capture the far-field velocity of the Sagaing Fault, requiring ad-hoc methods to account for the resulting distortion of the velocity field (Section A.2.2; Maurin et al. [2010]).

FIGURE 2.7: Trade-off curves for a priori constraints on the Churachandpur-Mao Fault (CMF). (A) Misfit statistics ($\Delta ABIC$ in blue, and reduced χ^2 in orange) and fault parameters vary with the standard deviation σ_{CMF} of the prior dip-slip constraint on the CMF for rigid block models. At values of $\sigma_{CMF} \geq 4$ mm/yr the inversion becomes unstable, as seen in the increase variability in λ_{CMF} (rake) and departure from strike-slip behaviour. This gives us a range of permissible values of slip rate, 0-8 mm/yr, for the CMF. (B) Model parameters show lesser variation with σ_{CMF} for block models where the IBW is allowed to deform internally compared to (A). Even as σ_{CMF} is increased to 100 mm/yr, the slip rate on the CMF does not exceed 6 mm/yr.

2.4 Discussion

In our modeling of the India-Burma-Sunda collision zone, we build complexity step-by-step: we start with two-dimensional models to find regions with large surface velocity gradients that likely correspond to coupling transitions on faults, and then use those results to construct a kinematically consistent block model that can fit the velocities. From these models, we show that there is a significant component of convergence between the India and Burma plates that is not being released as creep, and that the data are most consistent with a large locked zone on the Rakhine-Bangladesh megathrust. This locked area (shown in red in Figure 2.4 and 2.5) most likely releases the accumulated strain intermittently through earthquakes, as observed for other locked faults worldwide. Given that this fault

has been nearly seismically silent during the instrumental period we suggest that earthquakes on the megathrust and/or associated anticlinal structures may have a large (> 500 years) recurrence interval, and that as a result these events may be large.

2.4.1 Geodetically observable convergence across the Bangladesh megathrust

Previous kinematic studies that used two-dimensional dislocations report contrasting results regarding convergence across the IBW, i.e., from no convergence [Gahalaut et al., 2013] to 13-17 mm/yr convergence [Steckler et al., 2016]. Our 3-D block models of the India-Burma-Sunda system shows that a significant convergence rate across the Rakhine-Bangladesh megathrust is necessary to explain the India-Burma-Sunda plate motion vectors. The dip-slip component of the convergence across the Bangladesh megathrust decreases northward from 24 to 12 mm/yr from 22° N to 25° N (Figure 2.6, A.6-A.8) due to rotation of the IBW relative to India. While the total convergence rate is well constrained, we cannot be certain which faults will ultimately accommodate this strain. The megathrust in our models represents a basal décollement underlying the Chittagong fold-and-thrust belt, and if it were to rupture as a single event it would represent a large, Cascadia-type earthquake. However, there are also a series of thrust faults and anticlines within this belt with unknown levels of seismic activity, though the geodetic data presented here largely rule out the possibility of significant aseismic convergence on these structures, as this would represent sources of strain occurring closer to the fault trace than supported by our modeling. Thus, seismic events on out-of-sequence thrusts are also a possibility, as occurred in the 1999 M_w 7.6 Chi-Chi earthquake in Taiwan [Kao and Chen, 2000, Yue et al., 2005], and the 2008 M_w 7.9 Wenchuan earthquake in China [Hubbard et al., 2010].

2.4.2 Strain partitioning in the Burma plate

Our models cannot uniquely constrain the slip rate of the CMF due to the strong trade-offs between models that permit dip-slip on the CMF versus models that penalize dip-slip on the CMF but allow internal deformation of the IBW. In Figure 2.6C and A.9, we show the models which are difficult to distinguish statistically (using ABIC and F-tests); they allow anywhere between 0-13mm/yr of motion on the CMF (Figure A.7-A.8). The most common interpretation of the geodetic data has been that the CMF is an active plate/sliver boundary fault [Gahalaut et al., 2013, Steckler et al., 2016]; but in our models we show that this is not a robust result. We therefore ask: how do we know the CMF is active? Gahalaut et al. [2013] argue that there is surface creep observed in the GPS velocity field, and Wang et al. [2014] note that three to six river channels may be deflected by the CMF. However, our models demonstrate that there is no strong geodetic evidence for interseismic shallow creep in the current data, and that the velocity field can still be fit by a CMF that is fully coupled to a depth greater than 25 km (for a synthetic experiment to demonstrate the non-uniqueness of inferred shallow creep we refer the reader to Section A.3; Figure A.5). Shallow creep along the entire fault is not necessary, and even

unlikely, in a block model that considers the arcuate shape of the CMF and the megathrust (Figure 2.4, Figure A.5). We do note that the survey-mode GPS stations HENG and CCPR (Figure 2.4B), which are merely 4 km apart, show significantly different north-south velocities that are not well fit in models with a fault that is coupled to a depth greater than 25 km. This indicates that either there is very local shallow creep near these stations, or that one or both stations have unexplained site effects that are affecting the geodetic signal. Continuous shallow interseismic creep has been observed on strike-slip faults (for a review see Harris [2017]), but we caution against interpreting widespread creep based on observations at just one location, and particularly against then using this to interpret activity of a major, block-bounding fault. Other evidence for an active CMF is also limited: the river channel deflections observed by Wang et al. [2014] are few and display offsets which are less than 3 km; they could instead represent a river flowing around pre-existing topography, which follows bedding. Thus, invoking that the CMF is an active sliver boundary is difficult to defend. However, since our results indicate that there has to be some mechanism of northward transport (relative to India) in the inner IBW (Figure 2.6D, Figure A.7-A.8), we can state that the convergence is accommodated by the outer IBW while the inner IBW could be a shear zone that deforms in a north-south manner (Figure 2.8).

The IBW is juxtaposed against the MCB along the Kabaw Fault, which has been hypothesized to accommodate ~ 6 mm/yr of oblique slip [Steckler et al., 2016]. However, we find no need to invoke slip on this fault to fit our data in either our two-dimensional or block models. The MIBB network adds two new continuous GPS stations HAKA and KANI (Figure 2.4) on either side of the fault which show < 3 mm/yr difference in their velocities relative to India. This small difference can be fully reconciled as a combination of the Burma plate rotation and the effect of strain accumulation on the down-dip extension of the Rakhine-Bangladesh megathrust and/or CMF. The data do not require that the Kabaw Fault be inactive, but its slip rate must be within the modelling uncertainties, likely $< 2 - 3$ mm/yr. Geologic observations along the Kabaw Fault corroborate this inference and suggest that its slip rate is less than the sedimentation rate, so that if there is any young deformation, it is buried [Wang et al., 2014].

Along the eastern edge of the Burma plate, the Sagaing Fault takes up ~ 20 mm/yr of dextral strike-slip. This is consistent with previous geodetic studies of the northern and central transects [Maurin et al., 2010, Vigny et al., 2003] as well as Quaternary slip rate estimates of the Sagaing Fault [Bertrand et al., 1998]. The total shearing rate between India and the Shan Block, accommodated by fully coupled faults is 30 mm/yr, lower than previously inferred rates of up to 40 mm/yr [Gahalaut et al., 2013, Steckler et al., 2016], since these previous studies neglected block rotations (Section A.2.1, A.2.2; compare Figures A.2C and A.6-A.7).

From our models, we infer that the Rakhine-Bangladesh megathrust and the Sagaing Fault act as block-bounding faults for the Burma plate while the CMF may be a backstop to the IBW or a nascent block boundary (Figure 2.8). If the CMF is a major strike-slip fault partitioning some of the oblique convergence between India and Burma plates, then the slip rate cannot exceed an average of 8 mm/yr

along its entire length (Figure 2.7). We posit end-member models of this tectonic environment as follows:

1. E-W compressional strain is being accumulated within the outer wedge, most likely associated with the Rakhine-Bangladesh megathrust while N-S shearing is accommodated within the inner wedge. This shear may have localized onto a single fault, the CMF, and the upper bound of its slip rate is 8 mm/yr (Figure A.7C-D).
2. The IBW and MCB are a single rotating block relative to India, thereby defining the Burma plate (Figure A.6A-C). The IBW is the internally deforming portion of the Burma plate, while the MCB is rigid. The Rakhine-Bangladesh megathrust accommodates convergence between India and Burma while the obliquity in plate motion has not localized onto the CMF but is instead distributed within the inner IBW (Figure 2.8).

With a first-order understanding of the tectonics of the India-Burma-Sunda plate system (Figure 2.8), we comment on the subduction of the Indian slab beneath Burma as well as the earthquake likelihood of the Rakhine-Bangladesh megathrust.

2.4.3 Continued subduction of the Indian slab

Our ensemble of block models suggests that there is significant active inter-block convergence across the Rakhine-Bangladesh megathrust (Figure 2.5, A.6-A.8), but we are unable to describe the forces that cause this convergence since block modeling is only a kinematic description of the observations. However, we can test the validity of some first order geodynamic hypotheses based on combined insight from geodetic, geological and seismological observations.

Some authors have suggested that the subduction of the Indian slab beneath Burma has ceased in the Quaternary (or for even longer) and propose that gravitational collapse of the Tibetan plateau drives the observed shortening of the IBW (e.g., [Rangin et al. \[2013\]](#)). The hypothesis is that the observed slab parallel tensional axes in the Indian slab indicate ceased subduction while the observed foldbelt-perpendicular compressional axes in the IBW are assumed to be a consequence of topographically driven flow of material from the Tibetan plateau into the Myanmar Central Basin and thus shortening the IBW, or the IBW itself is over-steepened and is collapsing. The aforementioned interpretation of earthquake data is likely flawed, since slab parallel tensional axes are linked to cold, negatively buoyant slabs that have not yet reached the 660 km discontinuity [[Isacks and Molnar, 1971](#)] and the aforementioned stress axes orientations in the slab as well as upper plate (Figure 2.8B) have been observed in many active subduction zones, such as in the Andaman, Kyushu and Cascadia subduction zones [[Mallick et al., 2017](#), [Seno and Yoshida, 2004](#)].

The stresses from gravity driven tectonics are normally manifested as a series of strike-slip or normal faults in the back of the fold-and-thrust belt (e.g., as seen in the Niger delta [[Corredor et al., 2005](#)]).

However, we have two results from our study that challenge this notion. (1) We can fit the velocities in the Myanmar Central Basin using the rigid elastic block assumption implying that the velocities arising from gravitational potential energy collapse of either the Shan Plateau or the IBW do not extend to the MCB. (2) We do not observe normal faulting or extension within the inner IBW; even in models in which we allow internal deformation of the IBW, the sense of motion is north-south shear (strain-rate tensor in Figure 2.6D) implying that gravity tectonics is an unlikely explanation for the observed convergence in the outer IBW.

Thus, even though we cannot explicitly discriminate between the forces driving the active India-Burma convergence, the geodetic, seismological and geologic evidence can be explained by an active subduction zone.

2.4.4 Earthquake hazard due to the Rakhine-Bangladesh megathrust

The IBW hosts one of the world's flattest accretionary prisms, with $\sim 0.1^\circ$ slope for the outer wedge, and has a very shallow décollement [Betka et al., 2018, Burgi et al., 2016]. This implies that the detachment is very weak, possibly in part because fluid overpressures on the décollement seem to be nearly lithostatic [Steckler et al., 2016, Zahid and Uddin, 2005]. These observations, together with the lack of interplate seismicity in the instrumental catalog, have been used to suggest that the Rakhine-Bangladesh megathrust creeps continuously or periodically to relieve strain [Kundu and Gahalaut, 2012, Le Dain et al., 1984, Ni et al., 1989]. We note that a lack of microseismicity is not diagnostic of ceased fault activity; the Cascadia subduction zone is a good example of an active megathrust in a high sedimentation setting which exhibits very few moderate-sized earthquakes in the modern catalogue, but which is known to produce occasional $M \sim 9$ events [Nelson et al., 1995]. Here we demonstrate that the geodetic data in Bangladesh (Figure 2.2, A.2) do not show any evidence of creep, which would manifest as a sharp offset in velocity at the fault trace. Instead, there is a signal of interplate coupling and strain accumulation (Figure 2.4).

The lack of surface fault creep is not in conflict with observations of fluid overpressure and low friction at shallow depths [Steckler et al., 2016, Zahid and Uddin, 2005]. The absence of creep is an observation of strong interplate coupling, but it does not imply that the shallow megathrust is frictionally locked (we refer the reader to Table 1 in Almeida et al. [2018] for definitions). A down-dip locked zone is expected to create stress shadows on the shallow megathrust that will limit the creep rate to a small fraction of the plate rate, irrespective of the frictional properties [Almeida et al., 2018]. This results in accumulation of a slip deficit during the interseismic period, even on fault patches that are frictionally unlocked or unclamped.

The most recent great earthquake along the Rakhine megathrust is the $M \sim 8$ 1762 Great Arakan Earthquake [Halsted, 1841]; the average recurrence interval of major plate boundary rupturing events in this area may be 500-1000 years [Aung et al., 2008, Wang et al., 2013]. This suggests that even

FIGURE 2.8: (A) Perspective view of the study area with faults and earthquake epicenters from GCMT superimposed on top. AA' shows the location of the swath profile considered in (B). (B) Topographic profile and depth profile for a 100 km swath along AA'. The outer wedge is one of the flattest in the world with $\alpha \sim 0.1^\circ$ (topographic slope); the inner wedge is steeper with $\alpha \sim 0.5^\circ$. This transition also happens to be the point where we infer a coupling transition in our models. The plate interface, shown as the thin black line in the depth profile, is taken from Mitra et al. [2005]. We hypothesize that Indo-Burma plate motion is accommodated on the megathrust, CMF and/or other unmapped structures in the IBW. The red to blue colour transition in the IBW depth profile represents the likelihood of dip-slip (red) to strike-slip (blue) motion on faults within the IBW. We also show GCMT earthquake focal mechanisms, coloured by depth, in cross-sectional view along AA'. The extensional axes of the intra-slab earthquakes follow the slab dip while the upper plate appears to be in arc-normal compression. This suggests that subduction is active. CMF - Churachandpur Mao Fault; IBW - Indo-Burman Wedge.

if the shallow megathrust is frictionally unlocked, there may be locked patches at greater depths that can nucleate earthquakes that propagate into the shallow part of the fault. Such earthquakes may produce sudden coastal uplift and resulting tsunamis [Cummins, 2007], and be recorded by corals and terraces [Wang et al., 2013]. We can draw similar inferences about possible earthquakes on the subaerial Bangladesh megathrust, but we have minimal information about earthquakes on this fault. To date, the Great Bengal earthquake of 1548 is the only $M8+$ event hypothesized to have ruptured the Bangladesh megathrust, although it alternatively or in addition may have been caused by slip on the Dauki Thrust [Morino et al., 2011, Steckler et al., 2008].

Our models suggest that the megathrust is fully coupled all the way from the deformation front to an arcuate line beneath the high Indo-Burman Ranges (Figure 2.6D). The maximum strain accumulation on the Rakhine-Bangladesh megathrust is 20 ± 5 mm/yr, for models with high dip-slip penalization on the CMF (Figure 2.7A). This could be used as a limiting scenario for seismic hazard models for this area. The most likely estimate, supported by objective model selection, is 18 ± 4 mm/yr (Figure 2.7A, A.9). The continuous GPS time series do not show any obvious time-varying creep/slow-slip episodes during the period 2011-2017 (Figure 2.3). We rule out the possibility that the megathrust is creeping interseismically, although some limited patches within the megathrust could potentially host creep phenomena (time-invariant or time-varying) that are within the measurement uncertainties. Thus, there is almost certainly ongoing strain accumulation on the megathrust, which will eventually be released either in an earthquake or as postseismic creep following earthquakes on more limited patches.

2.5 Conclusions

The Indo-Burman Ranges are part of a fold-and-thrust belt that formed as a result of oblique India-Sunda plate collision. This collision manifests as a complex system of slip partitioning between multiple faults spread over the Burma plate. We use a new continuous GPS dataset from the MIBB network to model the horizontal deformation field over an area spanning 1000 km from Bangladesh to Myanmar. We use elastic dislocations, horizontal strain-rates and block-like motion on a sphere to model the horizontal velocities obtained from our stations, together with published datasets, and find:

1. We infer active convergence across the Indo-Burman Ranges at a rate of 12-24 mm/yr. The India and Burma plates are fully coupled to a depth of 30 km and the surface projection of the deep coupling transition follows the trend of the high Indo-Burman Ranges, 200 km east of the modern deformation front.
2. We estimate dextral shear of 8 mm/yr across the IBW, distributed between the CMF and/or other unresolved upper plate structures. The Sagaing Fault takes up 20 mm/yr of dextral shear, resulting in a total north-south shear of 30 mm/year between India and Sunda.

3. We fit the horizontal velocity field well using a block model with fully coupled faults and show that simplified modeling in two dimensions can cause significant biases in the results, such as implying creep on the CMF or large slip rates on the Kabaw Fault.
4. Based on geological observations, we suggest that, to a first order, the deep India-Burma plate interface accommodates 12-24 mm/yr of oblique India-Burma motion, which may further be partitioned among various upper plate structures in the seismic cycle.
5. The absence of notable interplate seismicity on the shallow part of the megathrust during the past 100 years does not imply that the megathrust cannot nucleate large earthquakes.

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Chapter 3

Long-lived shallow slow-slip events on the Sunda megathrust*

Abstract: During most of the time between large earthquakes at tectonic plate boundaries, surface displacement timeseries are generally observed to be linear. This linear trend is interpreted to result from steady stress accumulation at frictionally locked asperities on the fault interface. However, due to the short geodetic record it is still unknown whether all interseismic periods show similar rates, and whether frictionally locked asperities remain stationary. Here we show that two consecutive interseismic periods at Simeulue Island, Indonesia experienced significantly different displacement rates, which cannot be explained by a sudden reorganization of locked and unlocked regions. Rather, these observations necessitate the occurrence of a 32-year slow-slip event on a shallow, frictionally stable area of the megathrust. We develop a self-consistent numerical model of such events driven by pore-fluid migration during the earthquake cycle. The resulting slow-slip events appear as abrupt velocity changes in geodetic timeseries. Due to their long-lived nature, we may be missing or mismodelling these transient phenomena in a number of settings globally; we highlight one such ongoing example at Enggano Island, Indonesia. We provide a method for detecting these slow-slip events which will enable a substantial revision to the earthquake and tsunami hazard and risk for populations living close to these faults.

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3.1 Introduction

Tectonic slow-slip phenomena span a wide range of slip rates, from mm/yr to mm/s, and represent a mode of tectonic fault slip behavior that can be detected by geodetic instruments but do not release the seismic radiation typically associated with earthquakes. While post-earthquake accelerated fault creep (afterslip), earthquakes deficient in high frequency radiation (slow earthquakes), and spontaneous slow-slip events (SSEs) are all under the purview of slow-slip phenomena, in this study we are concerned only with the kinematics and dynamics of SSEs. It is unclear whether SSEs are simply earthquakes in slow motion [Michel et al., 2019] or a separate group of phenomena caused by significantly different underlying mechanisms [Bürgmann, 2018, Gombert et al., 2016]. The most common hypothesis regarding the mechanics of SSEs on faults governed by frictional processes involves large earthquake (instability) nucleation sizes on velocity-weakening fault zones [Chen and Lapusta, 2009, Liu and Rice, 2005, 2007, 2009, Segall et al., 2010] in the presence of low effective confining stresses, which means that instabilities can nucleate but not run away to seismic slip rates. Such low confining stresses generally result from elevated pore pressures [Audet and Kim, 2016, Audet and Schaeffer, 2018, Bürgmann, 2018, Saffer and Wallace, 2015]. Additional SSE mechanisms include a switch from velocity-weakening behavior at low velocities to velocity-strengthening behavior at higher velocities [Hawthorne and Rubin, 2013, Im et al., 2020], and the fluid-driven interplay of frictional and ductile processes at depths where rocks become hot enough that flow-like behavior is activated [Bernaudin and Gueydan, 2018, Lavier et al., 2013, Suzuki and Yamashita, 2009, Yin et al., 2018].

Here we show that a decadal scale SSE nucleated on a shallow velocity-strengthening section of the Sunda megathrust. Velocity-strengthening faults are generally thought to be stable, and therefore unable to nucleate frictional instabilities like earthquakes and SSEs [Marone et al., 1991, Scholz, 1998]. The observations and inferences we present challenge the typical model of SSE nucleation and therefore necessitate a different mechanism to explain its kinematics.

3.2 Transient deformation in the Simeulue coral record

In this study, we focus on the deformation history of Simeulue Island in the time period 1738-1861 [Meltzner et al., 2015, Tsang et al., 2015a] (Figure 3.1). This is thought to be an inter-seismic period for this section of the Sunda megathrust leading up to a plate boundary rupture of $M_w \sim 8.5$ in 1861 [Meltzner et al., 2012, 2015]. The magnitude and spatial pattern of the coseismic deformation in the 1861 event bears remarkable similarity to the 2005 M_w 8.6 Nias-Simeulue megathrust earthquake [Meltzner et al., 2012, 2015] indicating that fault zone properties have remained largely static for at least two earthquake cycles. This implies that it is reasonable to use inferences of frictional properties from the better documented earthquake cycle, the 2005 earthquake [Hsu et al., 2006, Qiu et al., 2019] (Figure 3.1A), to study fault zone mechanics from the 18th – 19th centuries.

FIGURE 3.1: Tectonic setting of the study (inset map). (A) The 2005 M_w 8.6 Nias-Simeulue Earthquake ruptured a portion of the Sunda megathrust, and slip is shown as green contours (at intervals of 2 m), while the post-earthquake creep (afterslip contours every 0.5 m) is shown in yellow [Qiu et al., 2019]. The three coral sites we use in this study are labeled as LBJ, SMB and UTG. We show computed velocities from 1738-1829 (grey) and 1829-1861 (red). The site PBK recorded no velocity change before or after 1829. The black line represents the cross-section we use in the 2-d approximation in Figure 3.2.

(B) Timeseries of the coral growth at three sites (1738-1861), flipped so that the timeseries is now in a geodetic reference frame, and the estimated timing of the velocity change (1829 ± 3). Open triangles represent highest level of growth of the coral; only the filled circles (the highest level of growth prior to each diedown) [Meltzner et al., 2015] are considered in the velocity estimation (predicted positions from a piecewise linear model – thick line; 95% confidence interval – thin envelope). The record ends in 1861 when there was a great earthquake (similar in magnitude and spatial extent as the 2005 M_w 8.6 Nias-Simeulue earthquake), which uplifted and killed all the corals.

The shallow marine corals on Simeulue record the combined effects of long-term sea-level and land-height change. These corals recorded a linear subsidence at a rate of 1-2 mm/yr from 1738 to 1829 (Figure 3.1B, 3.2A). Around 1829 (± 3 years), an abrupt velocity change occurred at each of the coral sites (Figure B.1, Section B.1), causing significantly greater subsidence rates of up to 10 mm/yr (Figure 3.1B, 3.2A), which persisted until the great earthquake in 1861. Previous studies have shown that these velocity changes are tectonic in origin, and interpreted them as spontaneous variations in the spatial extent of frictionally locked asperities and stably creeping regions of the Sunda megathrust [Meltzner et al., 2015, Tsang et al., 2015a]. Specifically, the locked section of the megathrust was thought to have expanded down-dip to a depth of ~ 55 km [Tsang et al., 2015a]. While variations in dimensions of locked asperities have been interpreted to occur on megathrusts in relation to stress transfer from nearby earthquakes [Loveless, 2017, Melnick et al., 2017], there is no evidence to suggest moderate/great earthquakes occurred within ~ 500 km of Simeulue around the time of the abrupt velocity change in the coral records [Meltzner et al., 2012, 2015]. Additionally, frictional locking is not expected to persist beyond the brittle-plastic transition, which is thermally controlled in subduction zones, and is estimated to be around 30 km depth for this location [Hippchen and Hyndman, 2008, Klingelhoefer et al., 2010, Simoes et al., 2004]. In light of these inconsistencies, we reevaluate the observations and inferences from the paleogeodetic record for Simeulue using a physics-based inverse method to relate fault slip rates to subsidence rates, and test the hypothesis: can steady frictional locking/sliding on a megathrust explain the observed vertical deformation history?

3.3 Stress-driven steady inter-seismic behavior cannot explain the observations

We develop an inverse method in which we solve for the locked ($v = 0$) and unlocked ($v > 0$) distribution of fault segments during each time period separately, for steady frictional behavior on a megathrust driven by the stresses arising from subducting an elastic slab into the mantle (in plane strain - see Section B.3). In our formulation, fault slip rates in the unlocked segments are governed by the stressing rates afforded by the nearby locked segments [Almeida et al., 2018] i.e. we seek a simply connected locked domain and the surrounding region is treated as a crack driven at constant stress (Figure 3.4A); this is inherently different from typical over-parameterized inter-seismic coupling inversions which solve directly for slip rates and are forced to use slip rate regularization to stabilize the inverse problem. We pose the inverse problem in terms of the misfit between the observed coral subsidence rates and the predicted rates as a function of the geometric parameters that describe the locked and unlocked domains. This is a non-linear problem for which we employ a Markov Chain Monte Carlo sampling scheme (see Section B.3 for details).

From the inferred fault segment locking/unlocking distributions we show that under the assumption of piece-wise steady-state inter-seismic processes, a down-dip migration of the inter-seismically locked

FIGURE 3.2: Observations and modeling results for the 18th-19th century coral record in Simeulue projected along the cross-section shown in Figure 3.1A. (A) Subsidence rates for 1738-1829 (gray) and 1829-1861 (red). We assume the velocities from each epoch collectively show the average response of southern Simeulue Island to tectonic changes (filled circles). The individual site vertical velocities are plotted with error bars, while the model predictions are shown as polygons (67% confidence level, with darker colors showing envelopes closer to the median). (B) Estimated slip rate for 1738-1829 is shown as a gray polygon. We model the slip rate for 1829-1861 as an SSE (red) superimposed on the existing locking from 1738-1829. (C) We compare the probability distribution of the locked region (gray) and SSE location (red) with qualitative estimates of velocity-weakening (green – coseismic) and velocity-strengthening (yellow – afterslip) regions inferred from the 2005 M_w 8.6 Nias earthquake [Qiu et al., 2019], along the transect shown in Figure 3.1a. (D) Probability distributions of slip rate, slip of the shallow SSE (black) and the equivalent moment release (orange) assuming a shear modulus of 30 GPa and along-strike width of 100 km.

FIGURE 3.3: Observations and modelling results for the 18th-19th century coral record in Simeulue. (A) Subsidence rates for the two time periods, 1738-1829 (gray) and 1829-1861 (red). We assume the velocities from each epoch collectively show the average response of southern Simeulue Island to tectonic changes (filled error bar). The individual site vertical velocities are plotted with error bars, while the model predictions are shown as polygons (67% confidence level, with darker colours showing regions closer to the median). (B) Estimated slip rate for 1738-1829 is shown as a gray polygon (67% confidence level). The slip rate for the period 1829-1861 is estimated using two different models: (1) steady interseismic processes with a change in locked/unlocked regions (blue polygon), (2) SSE (orange polygon) superimposed on the existing locking from 1738-1829.

zone is necessary to fit the 1829-1861 coral data (Figure 3.3,3.4C); this feature has been noted by previous studies [Meltzner et al., 2015, Tsang et al., 2015a]. Additionally, the previously locked shallow megathrust needs to unlock and remain that way, in order to improve the fit to the 1829-1861 subsidence rates (Figure 3.4B). However, this configuration is physically implausible for three reasons:

1. The transition from down-dip frictional locking to creep at the full plate subduction velocity would need to occur over infinitesimally narrow widths on the megathrust (Figure 3.4D) in order to fit the 1829-1861 vertical velocities. This would create nearly infinite stressing rates which cannot be supported by the elastic medium.
2. If the down-dip extent of the megathrust were to relock and remain in that configuration from 1829 to 1861, it needed to have completed within 6 years (uncertainties in the timing of the velocity change) (Figure 3.1B). This implies sustained creep deceleration of at least $\sim 10 \text{ mm/yr}^2$

on the megathrust at depths of $\sim 30 - 50$ km, followed by ~ 3 decades of elastic strain accumulation within a domain which is at temperatures greater than 500°C (Figure 3.4C) [Hippchen and Hyndman, 2008, Simoes et al., 2004]. We are currently unaware of any mechanism that can cause such a process spontaneously on a plate boundary fault or shear zone under these conditions; rock deformation at these temperatures is dominated by either viscous or plastic creep, not stick-slip.

3. Earthquake cycle simulations show that unlocking of previously locked fault segments, as we infer for the shallow megathrust from our piece-wise steady-state inverse models (Figure 3.4B), does not occur abruptly. Gradual unlocking of frictional asperities is seen in numerical models when creep from the fault segments surrounding the asperity erode its boundaries [Mavrommatis et al., 2017]. However, this process leads to nonlinear displacement timeseries [Mavrommatis et al., 2014, 2015, 2017] which is again inconsistent with our observations.

FIGURE 3.4: PDFs for spatial extent of locked and creeping regions on the megathrust. (A) Geometric setup and terminology used to describe the 4 segments of the megathrust: shallow creeping, frictionally locked, unlocked/transition zone and freely sliding plate boundary. (B) PDFs of the spatial extent of ζ_{up} for two time periods (1738-1829 and 1829-1861) assuming steady frictional locking and creep governs the evolution of slip on the megathrust. Maximum a posteriori probability (MAP) of ζ_{up} for 1738-1829 is shown as green thick line (95% confidence interval - thin green lines); MAP of ζ_{up} for 1829-1861 is shown as red thick line (95% confidence interval - thin red lines). To explain the 1829-1861 coral observations with only steady interseismic processes, the upper limit of the locked zone ζ_{up} would have to migrate down-dip by 50-100 km while the (C) deeper transition zone ζ_{down} would have to migrate to a depth of 50-60 km. (D) The transition from ζ_{down} to ζ_{free} (W) would have to narrow to infinitesimal widths (pdf is maximum at $W = 0$) making this model unphysical.

To explain the abrupt change in vertical velocities in the observational record, our preferred model is one where the locking configuration from 1738 to 1829 persisted on the fault, while a multi-decadal SSE spontaneously nucleated and was sustained on the shallow megathrust from 1829 to 1861 (Figure 3.2, 3.3). We model this SSE as uniform velocity fault creep over a simply connected domain (Section B.3). We infer the plausible range of transient fault slip rates to be 0.7-2.5 times the long-term trench-normal convergence rate and the SSE released 0.8-3.0 m of slip (Figure 3.2D). If we assume that this deformation occurred uniformly over the width of our transect (100 km), we

predict a moment release equivalent to an event of $M_w 7.2 - 7.9$ (Figure 3.2D). This event, along with the Banyak Islands SSE [Tsang et al., 2015b], SSEs in the Mexican subduction zone [Radiguet et al., 2012], Alaskan SSEs [Li et al., 2016] and the pre-Tohoku SSE [Bürgmann, 2018, Mavrommatis et al., 2014], represent some of the largest recorded SSEs. As a result of the low slip rates, such events can be difficult to detect in geodetic time series, since they may not be fast enough to cause a reversal in motion or show the obvious nonlinear patterns in the time series typically expected for SSEs. However, if we apply mechanical constraints on kinematic models (Section B.3) we can test if the observations are consistent with steady plate motion or if they are more compatible with a long-lived transient slip event. A similar method was used to detect slow, sustained creep migration along the Cascadia megathrust [Bruhat et al., 2016]. Such an approach has the potential to lower the threshold at which we can detect transient slip behavior on a megathrust.

3.4 SSEs on velocity-strengthening faults

Comparing the region inferred to slip in the 1829-1861 SSE with coseismic and postseismic slip from the 2005 $M_w 8.6$ Nias earthquake (Figure 3.2C), it appears that the 1829-1861 SSE occurred in a velocity-strengthening region [Qiu et al., 2019]. Rate-and-state friction modeling of the postseismic slip for the 2005 earthquake suggests that the shallow megathrust is likely under low effective confining pressures and exhibits weakly velocity-strengthening behavior [Hsu et al., 2006].

We constructed a rate-and-state frictional model (Figure 3.5) which shows that the observations of ~ 100 years of quiescence of shallow fault motion (1738-1829) can be explained by the response of a velocity-strengthening fault to a sudden fluid expulsion event early in the earthquake cycle (Figure 3.5A, 3.6). As faults can be regulatory agents for the transfer of fluids (fault-valve behavior) [Perfettini and Ampuero, 2008, Sibson, 1992], the release of fluids due to the breakage of permeability seals abruptly increases the confining pressures on the fault segment; the disruption of fluid networks may be caused by the propagation of a past earthquake into this section of the megathrust [Hughes et al., 2010, Husen and Kissling, 2002, Sibson, 2013] (at some time before 1738) or by strong shaking caused by passing earthquake waves [Materna et al., 2019]. As a result, the fault is clamped (Figure 3.5A). We model fault clamping as an instantaneous drop in pore fluid pressure ($\Delta P_F = -2$ MPa) with a resulting normal stress increase. This is followed by a period of strain accumulation on the partially drained fault for up to centuries (Figure 3.5B, 3.7). As the stresses and the stressing rates on the clamped fault increase over the following ~ 100 years, the fault is perturbed away from steady-state (due to elastic interaction from the creep front migrating inwards) and nucleates a transient slip event with slip rates comparable to the plate velocity [Perfettini and Ampuero, 2008] (Figure 3.5). Assuming normal stress recovery at a kinematically imposed linear rate, as a proxy for slip rate modulated gouge compaction and associated pore-pressure rise [Sibson, 1990, 2013, Sleep and Blanpied, 1992], the fault which had just initiated accelerated creep is further destabilized as it

FIGURE 3.5: (A) Temporal evolution of fault velocity leading up to and during a shallow SSE (colors show normalized slip velocity; the vertical axis shows normalized distance along the fault and the horizontal axis is time). Model parameters used in the simulation (values of frictional parameters a-b in blue and effective confining stress (MPa) in black) are shown on the left. Fluid expulsion driven clamping is prescribed at $t = 50$ yrs. This is followed by ~ 150 yr long strain accumulation. A SSE nucleates spontaneously and continues to slip for ~ 25 years which corresponds to the period over which fluid pressures are restored (inset figure). (B) Representative slip rate profiles (horizontal axis - normalized slip velocity) drawn for the pre-SSE at $t = 150$ yrs (gray line) and SSE at $t = 210$ yrs (red line).

returns to its undrained state (Figure 3.7). The resulting SSE continues for decades with near-constant slip rates comparable to those observed between 1829-1861 at Simeulue (Figure 3.7).

SSEs are typically modelled in the rate-and-state frictional framework as spontaneously nucleating instabilities on velocity-weakening faults [Chen and Lapusta, 2009, Liu and Rice, 2005, 2007, Segall et al., 2010]. The instability in these models is limited to an aseismic slip rate by a host of processes which ensure that with increasing slip, the elastic energy release exceeds frictional strength loss [Bürgmann, 2018]. Other models proposed to explain SSE, not dependent on velocity-weakening friction, are typically applicable at depths hot enough that the rheology of the rocks are brittle-ductile [Bernaudin and Gueydan, 2018, Lavier et al., 2013, Yin et al., 2018] and hence are not relevant for our observations. Alternate models of SSE in the brittle domain include nucleation on a fault with velocity-weakening friction at low velocities and a switch to velocity-strengthening friction at increased velocities [Hawthorne and Rubin, 2013, Im et al., 2020]. While this model is plausible, such behavior is not ubiquitous – it has mainly been seen in halite and certain types of serpentinite gouges [Moore et al., 2007]. The shallow velocity-strengthening section of the Sunda megathrust may have localized within layer silicates in the subducting sedimentary package [Bradley et al., 2019].

FIGURE 3.6: Snapshots through time of a frictional instability on a velocity strengthening fault. A steady-state creeping fault was perturbed early in the simulation ($t = 50$ yrs) by a pore-fluid expulsion event and allowed to evolve. In this simulation we do not allow additional weakening from pore-pressure recovery on the fault. The colors represent different time periods ($t = 190$ to 230 yrs) - from the initial acceleration of the SSE (blue) until the instability is arrested and the fault resumes creeping at its steady rate velocity (red). The SSE nucleates at the 12-14 km position on the simulated fault, as the fault is trying to recover to its steady state creep rate. However, an overshoot of slip rates occurs leading to a pulse of high slip rate at the center of the simulated fault. The transition from creep at below 10^{10} m/s to the accelerated slip pulse of $\sim 10^8$ m/s occurs in a short period of time (1-2 yrs). This instability is then damped out and is smeared over the available fault area over 10-15 years.

The frictional properties of these rocks have not been tested at the slip rates relevant to this study; however, they have been noted to be purely velocity-strengthening over the ranges ($0.1 - 10 \mu\text{m/s}$) testable with experimental setups [Moore et al., 2007, Saffer and Marone, 2003]. In this study, we provide a new mechanism to explain SSEs, by destabilizing a velocity-strengthening fault with normal stress variations likely related to pore fluid migration from the subducting sedimentary package into fault gouge (Figure 3.7). Recent theoretical work on this topic shows that the full incorporation of poroelastic effects into frictional behavior on faults indeed leads to mildly unstable behavior on velocity-strengthening faults, which may explain the mechanics of SSEs [Heimisson et al., 2019].

The results of our study force us to rethink our understanding of the nucleation and propagation of SSEs in a frictional framework. Given that the habitat of long-term SSEs at the down-dip edge of subduction zones is a region of high fluid pressure originating from dehydration reactions releasing fluids from sediments which are sealed by silica precipitation near the mantle wedge corner [Audet and Kim, 2016, Audet and Schaeffer, 2018, Audet et al., 2009, Hyndman and Peacock, 2003, Sibson,

FIGURE 3.7: Illustration of the pore-fluid driven SSE model presented in this paper. The fractured fault gouge (governed by velocity-strengthening friction) is sealed and compacted by slow shear [Sibson, 1990, 2013, Sleep and Blanpied, 1992] during an SSE, and leads to an increase in pore fluid pressure (P_F) on the megathrust (decrease in $\bar{\sigma} = \sigma - P_F$). Following the SSE period, we show two possibilities: (1) an earthquake nucleates elsewhere and breaks permeability seals as it propagates into the shallow megathrust; this process drains the fault, increasing $\bar{\sigma}$ (bold line), (2) the SSE terminates and the unlocked section of the fault resumes its steady-state creep, with negligible change in $\bar{\sigma}$ (dashed line).

2013], it seems plausible that earthquake-cycle driven drainage and restoration of pore fluids along the fault may control the occurrence of megathrust SSEs (Figure 3.7) without any need to invoke complicated competing mechanisms to nucleate and then stabilize frictional instabilities on velocity-weakening faults.

A test for the applicability of this mechanism is to see if regions hosting long-term SSEs overlap with afterslip following large earthquakes, since afterslip is a typical sign of velocity-strengthening behavior. Sections of the megathrust in Nankai [Johnson and Tebo, 2018, Yagai and Ozawa, 2013] as well as in Ecuador [Rolandone et al., 2018, Tsang et al., 2019] appear to demonstrate this overlap of afterslip and SSE, thereby adding observational credence to our hypothesis. A different way to look for observations of this mechanism at play in the modern instrumental record is to compare locations and timing records of seismicity within the forearc and/or the subducting oceanic crust

with megathrust creep rates. Compared to the SSE period, we expect a cessation of plate boundary creep, due to expulsion of pore-fluids, to preclude an increase in off-fault microseismicity (caused by released fluids travelling through the host rock). In other words, we expect an anti-correlation between off-fault seismicity and megathrust creep rates.

3.5 Implications for megathrust earthquakes

FIGURE 3.8: A possible ongoing long-lived SSE near Enggano Island. (A) The horizontal velocities from a campaign site (engg 1992-2001) and a continuous site (MLKN 2005-2015) show a marked change in amplitude and direction, which we infer as the joint effect of frictional locking and an SSE. In the inset we show a schematic cross-section of a locked seismogenic zone (cyan) and a shallow SSE (red) and their respective velocity predictions at Enggano. (B) Displacement timeseries for MLKN (blue – trench-normal motion; black – vertical motion). (C) Schematic to show how the surface displacement timeseries from the locked and SSE region interact together to produce the observations at MLKN.

SSEs occurring on shallow megathrusts load neighboring seismogenic sections of the fault, bring them closer to failure, and pose a major earthquake and tsunami hazard [Saffer and Wallace, 2015]. There is precedent, in the case of the Sunda megathrust, for a close temporal relationship between multi-year SSEs and great plate boundary earthquakes. The 1829-1861 SSE discussed in this study

was followed by a $M_w \sim 8.4$ earthquake [Meltzner et al., 2015, Morgan et al., 2020], and the Andaman segment (further north) experienced a SSE from 2001-2004 along the shallow megathrust which was followed by the 2004 M_w 9.2 Sumatra-Andaman earthquake [Paul and Rajendran, 2015]. Creeping faults by themselves may not present a significant hazard, but their interactions with neighboring segments make them unexpectedly dangerous [Harris, 2017].

Looking forward, a potential area of concern is the southernmost section of the Sunda megathrust, near Enggano Island. There, we interpret the GPS data as indicative of a multi-year SSE along the shallow megathrust while the seismogenic domain is locked and accumulating strain (Figure 3.8). This section of the megathrust was previously assumed not to accumulate any strain due to the observation of negligible trench-normal motion in the horizontal velocities derived from a continuous GPS site [Prawirodirdjo et al., 2010]. However, once we account for the vertical velocities, which shows rapid subsidence (Figure 3.8B), we find that such a signal is incompatible with steady aseismic creep on the megathrust. Instead, we show that frictional locking of the seismogenic zone of the megathrust in concert with a long-lived shallow SSE may be canceling out landward motion while enhancing the subsidence of the island (Figure 3.8C). With an improved understanding of the interplay between the locked seismogenic zones, shallow SSEs and the associated time series, we conclude that the hazard to the communities living near this section of the megathrust is potentially higher than what was previously thought, and models of risk and mitigation strategies need updating.

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Chapter 4

Subduction initiation and the rise of the Shillong Plateau*

Abstract: Discrepancies between geodetically and geologically estimated thrust fault slip rates are generally viewed as a methodological problem. Even when slip rate is steady over geological time, a discrepancy may exist because each method is sensitive to different deformation processes. However, this offers a tool to estimate the partitioning of convergence between footwall and hanging wall deformation, and therefore a way to discriminate among orogenic styles. Here we investigate one such discrepancy for the Shillong Plateau, a basement-cored contractional orogen within the Himalayan foreland basin. Using a regional block model to explain the modern geodetic velocity field and explicit uncertainty analysis of the geologic rates, we show that this discrepancy cannot be reconciled simply by invoking uncertainties in individual methods.

Our results indicate that the Shillong Plateau is not an ongoing forward break of the Bhutan Himalayas, as was believed until recently. Instead, the observed inter-plate convergence and plateau uplift in this region may be driven primarily by an attempt of the negatively buoyant Indian passive margin lithosphere (the Surma Basin), south of the plateau, to initiate subduction. As a result, the uplift history of the plateau, which constrains the geologic rate, is significantly lower than expected given the geodetic convergence rate. We propose that this convergence is largely accommodated by the transport of the footwall into the mantle. This geodynamic scenario has important regional seismotectonic implications: (1) the cold and brittle sinking passive margin may have enabled the deep extent (~ 30 km) and therefore large magnitude of the $M_w 8+$ Shillong Earthquake of 1897; (2) the collapse of the Indian lithosphere into the mantle may have created the anomalously deep (~ 0 km) Surma Basin; and (3) this subsidence may also drive accelerated post-Miocene westward propagation of the Indo-Burman Wedge. We propose that the Shillong Plateau is the only modern

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example of passive margin collapse, and can serve as a natural laboratory to study the earliest phase of subduction.

4.1 Introduction

The past few decades have seen an increasing interest in comparing geodetically inferred short-term fault slip rates to geologic slip rates across convergent margins [Allmendinger et al., 2009, Avouac, 2015b]. Discrepancies in slip rate estimates have been explained by (1) time-varying fault slip rates, (2) systematic errors in the data (geologic and/or geodetic), (3) earthquake cycle effects and incomplete temporal sampling of modern rates, and/or (4) sensitivity of geodetic and geologic rates to different geodynamic processes.

FIGURE 4.1: Tectonic overview of the study area discussed in the paper. (A) Seismicity and focal mechanisms of earthquakes coloured by depth. Major plate boundaries and faults discussed in the paper are labelled. Inset map shows a zoomed out view of the India, Burma and Eurasia plates and plate boundaries. Brm Bn – Brahmaputra Basin, MFT – Main Frontal Thrust, SP – Shillong Plateau, MCB – Myanmar Central Basin, IBW – Indo-Burman Wedge, MH – Mikir Hills. (B) Lithologies in the Shillong Plateau; redrawn from Yin et al. [2010]. We note the existence of crystalline basement outcrops within the Brahmaputra Basin. (C) The upper panel shows a topographic cross section (TT' in (A)) across the Shillong Plateau (black line with standard deviation in gray). The normalized river steepness (K_{SN}), calculated using TopoToolbox [Schwanghart and Scherler, 2014], is shown in red. High K_{SN} values are present near the Dauki Thrust and the Oldham Fault. The lower panel shows seismicity (dots) and focal mechanisms occurring within 100 km of TT', coloured by depth. Earthquakes south of the Shillong Plateau occur almost exclusively below 15 km, indicating brittle failure of the transitional-oceanic crust.

In this paper, we are concerned with the last process. At convergent margins, slip on the plate boundary thrust results in (1) uplift of the hanging wall (due to overthrusting) and/or (2) transport of the footwall into the mantle in the form of subduction (underthrusting). In this situation, geologic slip rates derived from exhumation histories in the hanging wall (in the hinterland) represent the component of convergence that goes into overthrusting, and are insensitive to underthrusting. Modern interseismic geodetic velocity fields, on the other hand, record horizontal plate motions and the overall convergence between the hanging wall and footwall, but do not predict the partitioning between overthrusting and underthrusting. Therefore, a mismatch in rates from different time scales can carry information about the regional geodynamic processes at play.

We investigate deformation associated with the Shillong Plateau, a 1.5 km-high region of crystalline basement, deforming and growing in what should have been the Himalayan foreland basin, south of the Bhutan Himalaya (Figure 4.1). The plateau is genetically a part of the Indian (continental) plate but appears to be deforming relative to stable India along the north-dipping Dauki Thrust (Figure 4.1) as a new micro-plate [Bilham and England, 2001, Lindsey et al., 2018, Vernant et al., 2014]. If the convergence here is accommodated by overthrusting, all of the accrued strain must result in topographic growth of the plateau. This has been the working hypothesis to describe the growth of the plateau [Bilham and England, 2001, Clark and Bilham, 2008].

This hypothesis has been tested with a combination of modern GPS shortening rates and long-term thermochronometry-derived cooling rates. The long-term shortening rate estimate of the plateau (0.5–2.5 mm/yr) comes from its uplift history recorded in the cooling ages of zircon and apatite grains since 8–14 Ma [Biswas et al., 2007, Clark and Bilham, 2008], while the modern GPS-derived convergence rates have a large range (2.5–9 mm/yr) owing to poor spatial and temporal data coverage [Bilham and England, 2001, Jade, 2004, Jade et al., 2007, Paul et al., 2001, Vernant et al., 2014]. Some authors have suggested that the overlap in the ranges of estimated shortening rates from both methods (at least in part) implies that all of the convergence is transformed into plateau uplift, i.e. complete overthrusting. This has led to speculation that Shillong Plateau uplift is driven by a forward break of the Bhutan Himalaya, since the total India-Tibet shortening rate is higher here than to the west [Bilham and England, 2001, Clark and Bilham, 2008, Johnson and Nur Alam, 1991] (Figure 4.2A).

In this study, we evaluate the statistical significance of the discrepancy in rates by updating the regional plate motion model for the Shillong Plateau and comparing the resulting 3-D velocity field to published geological and thermochronological results. We demonstrate that the difference in rates is statistically significant, which invalidates the hypothesis that overthrusting is the principal driver of this system. This implies that Shillong Plateau uplift is not driven by a forward break of the eastern Himalaya, but rather by a process that promotes underthrusting. The Surma Basin south of the Shillong Plateau is underlain by transitional-oceanic lithosphere, which is ideal for initiating subduction [Faccenna et al., 1999, Nikolaeva et al., 2010]. We propose that the discrepancy in short-term and long-term rates can be explained by subduction initiation along the Dauki Thrust (Figure 4.2B, C). This process would begin with subsidence of the crust south of the Shillong Plateau, which would

pull the plateau to the south and generate convergence with limited topographic uplift. We discuss the implications of this process for regional tectonics and seismic hazards.

FIGURE 4.2: Comparison of two candidate conceptual models that could be used to describe the tectonics of the Shillong Plateau. (A) The Main Himalayan Thrust (MHT) break forward hypothesis requires a 300 km connection between the deep decollement in the Bhutan Himalaya to the Dauki Thrust at ~ 30 km depth. This will also imply dismemberment of the underthrusting Indian crust beneath the Himalayas. (B) Subduction Initiation of the Surma Basin along the pre-existent weak fabric of the Dauki Thrust due to the action of 15-20 km of sediment overburden and negatively buoyant transitional-oceanic crust. (C) Subduction Initiation model for the Surma Basin. We interpret the numerous near vertical fault plane solutions (dashed lines) for earthquakes within the transitional-oceanic crust (Figure 4.1) as evidence of the mechanical resistance of the old rigid plate being overcome. SP – Shillong Plateau; BH – Bhutan Himalaya; MFT – Main Frontal Thrust; MBT – Main Boundary Thrust; MCT – Main Central Thrust.

4.2 Data and methods

4.2.1 GPS data

We compiled GPS velocities from multiple sources in the ITRF08 reference frame [Barman et al., 2017, Jade et al., 2017, Kreemer et al., 2014, Mallick et al., 2019, Marechal et al., 2016, Vernant et al., 2014] and transformed these velocities to an India-stable reference frame using the Euler pole of Steckler et al. [2016]. For datasets not in ITRF08 [Vernant et al., 2014], we first used a 6-parameter Helmert transformation to convert these velocities to ITRF08 using parameters for rotation and translation computed for sites common to the given dataset and Kreemer et al. [2014]. While combining

overlapping datasets we identified duplicate entries (less than 500 m apart) and computed a mean velocity and uncertainty (weighted by variance) to represent each of these stations. We used the uncertainties for the data as a weighting function, with a minimum threshold of 0.5 mm/yr to avoid overfitting.

FIGURE 4.3: (A) Data and model predicted velocities from our preferred block model (maximum posterior probability). Major block boundaries are labelled and coloured as in Figure 4.1. The data (blue vectors), modelled velocities (red vectors) and block boundary slip rates (black vectors showing relative motion of the hanging wall) are all shown with the same scale. Station velocities are shown with their 95% confidence intervals as ellipses. Block names are labelled in black. Black box is the area shown in Figure 4.4. (B) Residuals, shown as red circles in a polar plot with uncertainty as crosses, show no systematic spatial bias which indicates the model can satisfactorily explain the data (contours every 2.5 mm/yr; outer contour is 5 mm/yr). IBW – Indo-Burman Wedge; MCB – Myanmar Central Basin; CMF – Churachandpur-Mao Fault

4.2.2 Block modelling

We used the MATLAB-based Blocks software [Meade and Loveless, 2009] to model the GPS data $\mathbf{d}_{GPS}$. We assume that the GPS data represent a steady-state linear interseismic process, thereby allowing the multi-year timeseries to be described completely by a single value, velocity.

The velocity at any station is modeled as a sum of rigid plate/block rotations on a sphere and elastic strain accumulation on block boundaries [Meade and Loveless, 2009]. We test for the residual post-seismic velocity field from the last great earthquake in the region, the 1897 M 8 Shillong earthquake, and find that it may be safely neglected given commonly used parameter ranges for viscoelastic processes (Section C.2; Figure C.3-C.4).

We use 6 plates/blocks to model the GPS velocities: the Shillong Block (consisting of the Shillong Plateau, Mikir Hills, Brahmaputra Basin and Assam Valley), Indo-Burman Wedge, Myanmar Central Basin, Northern Shan Block, Shan Block and Indian plate/block (Figure 4.3). We assume that some of these boundaries (Dauki Thrust, Rakhine-Bangladesh megathrust, Churachandpur-Mao Fault (CMF) and Sagaing Fault) are locked and accumulating strain at the full plate rate using a back-slip model to maintain continuity of the velocity field [Savage, 1983]. In the following section, we briefly describe our block modelling method; for a detailed description we refer the reader to Text S1.

Each block can be described using its 3-component Euler vector, $\boldsymbol{\omega}$, which is linearly related to the GPS data for a given fault geometry, $G_{GPS}(\delta: \text{dip}; D: \text{locking depth})$. In addition to GPS data, we also use a priori information on fault slip rates (G_{SR}) as a set of Gaussian constraint equations with unknown weights $\mathbf{d}_{SR} \sim N(\boldsymbol{\mu}_{SR}, \boldsymbol{\sigma}_{SR}^2)$, specifically slip rate constraints on the CMF (assumed to be

pure strike-slip) and the Dauki Thrust (assumed to be pure thrust) $\left(\overbrace{\sigma_{CMF}^2, \sigma_{Dauki}^2}^{\sigma_{SR}^2} \right)$.

$$\begin{bmatrix} \mathbf{d}_{GPS} \\ \mathbf{d}_{SR} \end{bmatrix} = \begin{bmatrix} G_{GPS}(\delta, D) \\ G_{SR}(\delta, D) \end{bmatrix} \boldsymbol{\omega} \quad (4.1)$$

Our goal is to find the range of values for $\boldsymbol{\omega}, \delta, D, \sigma_{SR}^2$ that can sufficiently explain the data given the priors. We do this using the combined covariance matrix (C_d) of the GPS data and slip rate priors as a way to weight the relative contribution of each dataset [Fukuda and Johnson, 2010].

$$C_d = \begin{bmatrix} C_d^{GPS} & 0 \\ 0 & C_d^{SR}(\sigma_{SR}^2) \end{bmatrix} \quad (4.2)$$

We can write this as a problem in Bayesian parameter estimation, where we want to maximize the posterior probability density function (PDF) of the model parameters given the observed data, $p(\boldsymbol{\omega}, \delta, D, \sigma_{SR}^2 | \mathbf{d}_{GPS})$.

$$p(\boldsymbol{\omega}, \delta, D, \sigma_{SR}^2 | \mathbf{d}_{GPS}) = \frac{p(\mathbf{d}_{GPS} | \boldsymbol{\omega}, \delta, D, \sigma_{SR}^2) p(\boldsymbol{\omega}, \delta, D, \sigma_{SR}^2)}{p(\mathbf{d}_{GPS})} \quad (4.3)$$

$p(\mathbf{d}_{GPS} | \boldsymbol{\omega}, \delta, D, \sigma_{SR}^2)$ is the data likelihood function, which we choose as a multivariate Gaussian distribution, while $p(\boldsymbol{\omega}, \delta, D, \sigma_{SR}^2)$ is the prior distribution on each of the model parameters. We assume uniform priors for $\boldsymbol{\omega}$, uniform positive priors for σ_{SR}^2 , and uniform priors in the interval $[0, 90]^\circ$

and [0,50] km for δ, D respectively. We sample from Equation 4.3 using a MATLAB-based slice sampler [Neal, 2003] to obtain a discrete representation of the posterior PDF of model parameters.

4.2.3 Uncertainties in long-term uplift rates

The geologic shortening rates for the Shillong Plateau come from estimates of long-term rock uplift rates that are converted to shortening through several assumptions (complete overthrusting, 2-d area balance, fixed detachment depth, etc.). In this paper, we deal directly with the thermochronological observables and derive confidence bounds for the inferred uplift rates by propagating the uncertainties in each assumption through Monte-Carlo sampling.

Thermochronology studies have shown the rocks sampled within the Shillong Plateau, close to the Dauki Thrust, reset the U-Th-[Sm]/He apatite clock but not the U-Th/He zircon thermochron [Biswas et al., 2007, Clark and Bilham, 2008]. Based on closure temperatures for apatite and zircon varying uniformly between [55, 85]° C and [170, 200]° C, respectively, and a geothermal gradient modeled as a truncated Gaussian distribution $30 \pm 10^\circ \text{C/km}$ (truncated at 10°C/km and 50°C/km), we estimate the probability distribution of minimum and maximum Cenozoic sedimentary cover initially deposited on the Shillong Plateau basement.

$$\text{Thickness} = \frac{T_{\text{closure}}}{\text{Geotherm}} \quad (4.4)$$

From apatite dating, the initiation of rock uplift ($t_{\text{initiation}}$) is estimated between 8-14 Ma [Biswas et al., 2007, Clark and Bilham, 2008]. The sedimentary cover was eroded away by 3-5 Ma to expose an erosion resistant crystalline basement which then began gaining elevation. The timing of this event (t_{topo}) is inferred from a change in sedimentary facies in the Surma Basin [Alam et al., 2003], the onset of flexural loading in the Surma Basin sediments [Najman et al., 2016], a volume balance of eroded material in catchments affected by the Dauki Thrust [Rosenkranz et al., 2018] and deflections in the paleodeposits of the Brahmaputra River as it moved north of the plateau due to the rising topography [Govin et al., 2018]. If we assume that uplift was a steady linear process prior to exposure of the basement, we can use these constraints to stochastically estimate the Miocene rock uplift rate $\dot{U}_R^m$.

$$\dot{U}_R^m = \frac{\text{Thickness}}{t_{\text{initiation}} - t_{\text{topo}}} \quad (4.5)$$

The Pliocene-Present surface uplift rate $\dot{U}_S^p$ is estimated from the rate of elevation gain of a bevelled paleo surface cut into the basement rocks; its present elevation is 1500 ± 100 m and was thought to have been at sea level 3-5 Ma [Alam et al., 2003, Grujic et al., 2006]. For the case of the paleo surface, the erosion rate $\dot{\epsilon} = 0$ because the surface is bevelled and smooth, indicating that it has been preserved since its last surface exposure in the Cretaceous [Alam et al., 2003]. As a result, we can estimate the Pliocene rock uplift rate $\dot{U}_R^p$ from the topographic uplift rate and the erosion rate

[England and Molnar, 1990].

$$\dot{U}_R^p = \dot{U}_S^p + \dot{\epsilon} \quad (4.6)$$

This surface appears to have been transported vertically with minimal tilting, meaning that rock uplift and erosion have acted uniformly on this plane; therefore Equation 4.6 is applicable over the entire area.

FIGURE 4.4: Results of block modeling, focused on the Shillong Block area (black rectangle in Figure 4.3). The orange vectors show the predicted long-term horizontal motion of the block relative to stable India. Inter-block slip rate vectors (motion of hanging wall relative to foot wall) are shown in black. The posterior PDF of the horizontal position of the locked-creeping transition in the Shillong Block is shown in blue as $p(\xi_d)$. DCFZ indicates the Dhubri Chungthung Fault Zone while the major thrust faults are marked in red (dashed red lines are the inferred traces of the Oldham Fault and Brahmaputra Fault). The estimated slip rate for the Main Frontal Thrust (MFT) is from Lindsey et al. [2018].

4.3 Results

We show the results of our plate motion models for the Shillong Block, Indo-Burman Wedge, Myanmar Central Basin, Northern Shan Block and Shan Block relative to the Indian Plate/block (Figure 4.3). Strain accumulation is assumed to have localized onto block boundary faults; their slip rates and orientations are shown as black arrows, while their geometries are shown as dashed lines. From the maximum a posteriori (MAP) block model, we estimate 30-35 mm/yr of oblique India-Shan convergence (azimuth $\sim 40^\circ$); ~ 20 mm/yr of this north-south motion is accommodated by dextral strike-slip on the Sagaing Fault, while the rest (7-9 mm/yr) results in either localized dextral shear on the CMF and/or overall shear distributed among the megathrust and other faults within the Indo-Burman Wedge [Mallick et al., 2019]. The Rakhine-Bangladesh megathrust accommodates between 10-20 mm/yr of east-west convergence, with the magnitude increasing southwards. The Dauki thrust accommodates between 5-8 mm/yr of north-south convergence (Figure 4.4), with a significant change in rake on either side of our chosen triple junction between the Indian Block, Indo-Burman Wedge and Shillong Block (Figure 4.4, S2).

In the following sections we focus on the Shillong Block and discuss the detailed results of our study in two parts: (1) geometry and kinematics of the Shillong Block and (2) geometry and deformation associated with the Dauki Thrust.

4.3.1 Geometry and Kinematics of the Shillong Block

We define the Shillong Block as the region encompassing the Shillong Plateau, Mikir Hills, Brahmaputra Basin and Assam Valley (Figure 4.1, 4.3A), as this single block can sufficiently explain the geodetic velocity field (RMSE = 1.7 mm/yr; Figure 4.3B). While a second block [Vernant et al., 2014] can improve the fit due to the addition of model parameters, the improvement is not statistically significant. For example, station NMRP in the Naga Hills is the easternmost point that we can use to constrain the rotation of the Shillong Block relative to the Myanmar Central Basin (Figure 4.3A). We satisfactorily fit the station velocity (model velocity pierces the uncertainty ellipse), making it unnecessary to invoke additional blocks or block boundaries such as the Assam Block and Kopili Fault [Barman et al., 2017, Vernant et al., 2014] or the Badapani-Tryshad Shear Zone [Yin et al., 2010].

The northern boundary of the Shillong Block is defined by the Main Frontal Thrust (MFT) in Bhutan. While we do not model data from the Bhutan Himalaya in our block modelling framework, we note that in 2-D the velocity field across the MFT is continuous, indicating a locked shallow plate boundary (Figure C.7; also see Lindsey et al. [2018]). Previous interpretations of shallow creep on the Bhutan MFT at a rate of 5-7 mm/yr were based on station velocities resolved in an Indian plate reference frame but interpreted in a Shillong-fixed reference frame [Marechal et al., 2016]. Here, we show that

the Shillong Block has a coherent and finite southward velocity relative to India (Figure 4.4), and thus shallow creep on the MFT is not necessary to explain the data (Figure C.7).

4.3.2 Dauki Thrust

We jointly infer the slip rate, geometry and weight of strike-slip penalization on the Dauki Thrust in our models (Figure 4.3-4.4, C.1-C.2). The estimated fault slip rate increases from the west (~ 5 mm/yr) to the center (~ 7 mm/yr), and then decreases eastward (~ 6 mm/yr), indicating that the Shillong Block and Indo-Burman Wedge are co-rotating in a clockwise manner relative to the Indian plate (Figure 4.5). We find that the depth and dip of the Dauki Thrust varies between $D \in [21, 37]$ (MAP = 27 km), $\delta \in [23, 40]$ (MAP = 28°) for our elastic block models (parameter ranges in brackets are 95% confidence intervals; Figure C.1). We note that the dip may represent an average dip for the system, i.e. if the Dauki Thrust flattens at depth as part of a ramp-flat system, our estimated dip is only a minimum value for the ramp. Thus, any predicted uplift rate (Slip rate $\times \sin \delta$) using this geometry is a minimum uplift rate.

The surface velocity field is most sensitive to the horizontal location of the locked-creeping transition of the block boundary $\zeta_d = \frac{D}{\tan \delta}$ [Lindsey et al., 2014, Mallick et al., 2019]. Although the concept of a spatial step change in fault slip rate is nonphysical, this is a common kinematic modelling simplification that exploits the fact that geodetic data cannot resolve the true transition width when using surface velocity gradients to infer fault slip rates at > 20 km depth [Lindsey et al., 2014]. We estimate ζ_d to be [30,55] km north of the trace of the Dauki Thrust (Figure 4.4), close to the down-dip projection of the south-dipping blind Oldham and Brahmaputra Faults. Since the Dauki Thrust dips to the north while the Oldham and Brahmaputra Faults are thought to dip to the south [Bilham and England, 2001, Rajendran et al., 2004], it is likely that they would intersect close to our inferred locked-creeping transition and are part of the same system accumulating tectonic strain. Thus, even though we can estimate the rate of strain accumulation, we cannot distinguish between the Dauki Thrust or a combination of the Dauki Thrust and faults along the northern margin of the Shillong Plateau as the structures that will eventually accommodate this strain.

The estimated rake of the Dauki Thrust, which is a measure of the inter-block convergence obliquity, appears to vary from west to east along the fault. We explore the full range of penalizations of strike-slip motion on the Dauki Thrust (σ_{Dauki} in Figure C.1) and find that a variation in convergence obliquity from predominantly thrust (west of the triple junction in Figure 4.4) to oblique with significant dextral motion (east of the triple junction), is necessary for our block models to fit the data. We posit that the dextral motion east of the triple junction may represent distributed shear within the Indo-Burman Wedge close to the Shillong Plateau rather than reflecting oblique slip on the Dauki fault itself (Section C.3, Figure C.5).

FIGURE 4.5: Comparison of mean topography (black with standard deviation in gray) for a transect along the southern front of the Shillong Plateau, and predicted slip rate of the Dauki Thrust, in blue (with standard deviation). We also show the predicted (overthrusting) uplift rate for each of the rectangular Dauki Thrust segments in Figure 4.4 calculated from two models: ramp advection ($\text{Slip rate} \times \sin \delta$) in light blue, and plateau pop-up ($\frac{\text{Slip rate} \times \tan \delta}{2}$) in green. The uplift rates are provided with ± 1 (bold error bar) and ± 2 (light error bar) standard deviations. All quantities increase from west to the center and then taper to the east.

4.4 Discussion

From our geodetic block model, we estimate the modern horizontal plate motions that drives the shortening of the Shillong Plateau. We explore if this horizontal convergence of 5-8 mm/yr (Figure 4.5) is congruent with the geologic observables over the million-year timescale. Past studies relating short-term and long-term rates across plate boundary faults find that discrepancies in rates can often be attributed to methodological biases in both geological estimates [Donnellan et al., 1993] and geodetic models [Thompson et al., 2015]. Viscoelastic relaxation of the mantle following earthquakes may also systematically lead to overestimates of modern rates [Perfettini and Avouac, 2004]. In this study, we show that post-seismic relaxation effects are negligible (Figure C.3,C.4). We formally quantify the uncertainties in the geologic and geodetic slip rate estimates to show that the discrepancies in slip rate estimates for the Dauki Thrust are statistically significant, and evaluate the seismotectonic implications.

FIGURE 4.6: Uplift history of the Shillong Plateau. (A) At 8–14 Ma, the top of the crystalline basement (gray rectangle does not carry any information on thickness of the basement) is at a depth of 2–5 km. The probability distribution functions (PDF) of the minimum and maximum depth of the top of the crystalline basement, which are actually estimates of the sedimentary overburden are shown as brown lines. This sedimentary overburden is removed from $t_{initiation}$ to t_{topo} at $\dot{U}_R^m$ mm/yr. Subsequently the plateau gained topography at $\dot{U}_R^P$ mm/yr. (B) Probability distribution of uplift rates in the Miocene (brown) vs. Pliocene-present (gray) from propagation of uncertainties in sedimentary thickness and timing information ($t_{initiation}$ and t_{topo}). Rock uplift rates predicted from the modern convergence rate assuming complete overthrusting are shown for two different uplift mechanisms: blue (ramp advection) and green (pop-up). There is almost no overlap between the modern uplift rate PDFs and either geological uplift rate PDF, indicating that shortening must be in part accommodated by underthrusting.

4.4.1 Steady Rock Uplift?

Has rock uplift of the Shillong Plateau been steady through time? In order to assess this, we compare the probability distributions of the Miocene uplift rate $\dot{U}_R^m$ and the Pliocene-present uplift rate $\dot{U}_R^p$ (Figure 4.6). We find that the probability distributions overlap, and uplift could have been constant or decreased over time, but did not accelerate. We estimate the probability $p(\dot{U}_R^m > \dot{U}_R^p) = 80\%$ (Figure 4.6B), indicating that the uplift rate likely decelerated from the Miocene to the Pliocene-present.

To compare the long term record with geodetic observations, we convert the estimated modern horizontal convergence rate (V_{pl}) to rock uplift at the locations where the thermochronological samples were obtained (Figure 4.5) using two different overthrusting models. In Model 1, material advects upward along a δ -dipping ramp connected to a horizontal décollement, with a predicted uplift rate of 3.0 – 4.1 mm/yr (Figure 4.5, 4.6B). In Model 2, long-term horizontal shortening is converted to uplift by rigid body transport of a trapezoidal plateau (details are in Section C.4 and Figure C.6) bounded by oppositely dipping faults [Bilham and England, 2001], with a predicted uplift rate of 1.5 – 2.9 mm/yr (Figure 4.5, 4.6B). There is insignificant (at the 99% confidence level) overlap of the probability distributions of the predicted modern uplift rate with either $\dot{U}_R^m$ or $\dot{U}_R^p$ (Figure 4.6B), demonstrating that the discrepancy between the geologic and geodetic uplift rates is statistically significant. Reconciliation of these rates requires a process that produces lower uplift rates than either overthrusting model. A model that predicts a decrease in uplift rate over geological time would be preferred but not required, given the probability distributions that we calculate.

4.4.2 Evidence Against Break-Forward of the Bhutan Himalaya

The break-forward model linking the Bhutan Himalaya to the Shillong Plateau has been the popular model of choice to describe the uplift of the plateau [Bilham and England, 2001, Clark and Bilham, 2008, Johnson and Nur Alam, 1991]. It combines an assumption of complete overthrusting with the observation that the southward components of the velocities (in an Indian reference frame) show a west to east increase in southern Tibet, rising from 15 to 20 mm/yr [Lindsey et al., 2018, Vernant et al., 2014]. Superficially, it looks like the increased convergence is partitioned between the Main Himalayan Thrust (MHT) and the Dauki Thrust; these faults are assumed to be somehow linked by a (sub) horizontal detachment (Figure 4.2A) or by transmission of collisional in-plane stresses into the foreland [Bilham and England, 2001]. This model implies that over multiple earthquake cycles, the Indian plate is stationary, and all inter-block convergence is accommodated by thrusting of the Shillong Plateau over the Surma Basin along the Dauki Thrust, either as a crustal scale fault-propagation fold [Clark and Bilham, 2008], or as a pop-up structure [Bilham and England, 2001].

The exhumation history demonstrates that some overthrusting has occurred. This is consistent with the Bouguer gravity anomaly over the Shillong Plateau, which is positive (~ 50 mGal), indicating

uncompensated topography (i.e. a lack of crustal root to support the ~ 1.5 km topography) [Nayak et al., 2008]. Joint inversions of receiver functions and surface wave dispersion do not show thickened crust supporting the Shillong Plateau [Mitra et al., 2018, Priestley et al., 2019]. This lack of a crustal root implies that the plateau is being thrust on top of the Indian plate and is held up by the elastic strength of the lithosphere [Lyon-Caen and Molnar, 1985]. However, these observations do not imply break-forward of the Bhutan Himalaya due to the following reasons:

1. The long-term uplift rates are significantly lower than what we predict from the modern convergence rate (Text S4, Figure 4.6, S6), implying that overthrusting is only a minor component of the convergence. The only overthrusting model that can match the geological uplift rates is one where the Shillong Plateau deforms in isostatically balanced pure shear (Section C.4, Figure C.6). However, this model requires the existence of a crustal root, which is not seen in geophysical observations of the Moho [Mitra et al., 2018, Priestley et al., 2019].
2. The Moho beneath the plateau is 35 km deep [Mitra et al., 2018], while the MHT is only 10-20 km deep [Coutand et al., 2014]; connecting these systems would require dismembering the Indian lower crust underthrusting Tibet (Figure 4.2A). An alternative model of plateau pop-up resulting from in-plane compression does not require a physical connection between the MHT and the Shillong Plateau [Bilham and England, 2001]. However, the predicted geological uplift rate from this model is still significantly greater (at the 99% confidence level) than the observed rates (Figure 4.6, S6).
3. Uplift of the Shillong Plateau has been synchronous with the development of the Main Central Thrust (MCT), Main Boundary Thrust (MBT) and Main Frontal Thrust (MFT) in the Himalaya. Given the amount of shortening that has accumulated on these faults and the overall southward propagation of the MHT [Avouac, 2015b], the Himalayan deformation front would likely have been greater than 400 km further north from the Dauki Thrust in the Miocene than it is today, making it that much harder to physically link these systems.

Together, these points demonstrate that kinematic linkage between the Himalaya and the Shillong Plateau through break-forward is implausible.

4.4.3 Subduction Initiation of Transitional-Oceanic Indian Lithosphere

We propose that subduction initiation of the lithosphere underlying the Surma Basin can explain the mismatch in long-term uplift rates compared to modern shortening rates. Subduction initiation at passive margins is difficult to achieve due to the rigidity and bending resistance of old lithosphere [Gregg Erickson, 1993, Mueller and Phillips, 1991], and passive margin subduction initiation has not thus far been identified in the Cenozoic [Stern, 2004]. However, the crust underlying the ~ 15 km

sedimentary pile of the Surma Basin displays a number of characteristics that prompt us to suggest that it may indeed have overcome any resistive forces and is attempting to subduct.

FIGURE 4.7: Perspective view of the study area discussed in this paper. The Surma Basin (delineated by Bouguer anomaly contours (mGal) in green) is a passive margin that has been transformed to an active margin by possibly initiating a buoyancy instability in the Miocene. This has promoted some overthrusting of the Shillong massif, allowing it to build up ~ 1.5 km of topography. The Moho in the cross-sectional view is inferred from seismological investigations of crustal structure (dotted regions – receiver functions; black boxes – surface wave dispersion; Mitra et al. [2018], Priestley et al. [2019]). The collapse of the Surma Basin into the mantle as well as the rigid Shillong Plateau affect the westward propagation of the Indo-Burman Wedge. We show the edge of the outer fold belt as a red continuous line, and the blind fold belt as a dashed line (Bangladesh megathrust). The cross-sectional view of crustal structure beneath the Indo-Burman Wedge is redrawn from Wang et al. [2019].

The crust beneath the Surma Basin is thought to have formed in the Cretaceous (~ 130 Ma), when India-Antarctica rifting began [Alam et al., 2003, Johnson and Nur Alam, 1991]. Although this rift zone is now buried under sediments, its signal remains as a Bouguer gravity anomaly high around the paleo margin/hinge, which separates oceanic crust to the east from Precambrian continental crust to the west [Mitra et al., 2018, Talwani et al., 2016] (labelled as the Hinge Zone in Figure 4.1A). East of the hinge zone and beneath the 15 km pile of sediments in the Surma Basin is a volcanic rifted margin; this was created by magmatic intrusion of basaltic material within the stretched and thinned continental crust [Menzies et al., 2002]. Hot spot volcanism from the Kerguelen plume, which postdates the initial rifting by 5-10 Ma, may have further densified and simultaneously weakened the crust during eruptive episodes such as those that produced the Sylhet Traps [Alam et al., 2003, Talwani et al., 2016]. Modern observations of crustal structure in this region show that the Surma Basin crust has

a thickness of 15-20 km and shear wave velocity ~ 4 km/s, both of which are significantly different from values expected for continental or oceanic crust, and hence it is called transitional-oceanic crust [Mitra et al., 2018] (Figure 4.7). We take these observations to imply that the transitional-oceanic crust beneath the Surma Basin is negatively buoyant and can subduct.

Moderate sized earthquakes along steeply dipping faults, oriented approximately parallel to the hinge zone, systematically break the transitional-oceanic crust [Kumar et al., 2015, Mitra et al., 2018] (Figure 4.1A, C). This seismicity may represent reactivation of crustal weaknesses along inherited structure [Kumar et al., 2015, Talwani et al., 2016] as the Indian slab tries to collapse into the mantle. These earthquakes are almost exclusive to the basement (not the sediment pile) (Figure 4.1C), and may be assisted by fluids in the lower crust and lithosphere [Regenauer-Lieb et al., 2001], as active crustal faulting can create conduits for fluids to permeate to greater depths and enable aqueous metamorphic transformations [Jamtveit et al., 2018]. Such metamorphic transformations are crucial in the ensuing stress balance, which determines whether the resisting forces present in an old, cold, and rigid lithosphere of a passive margin can be overcome by buoyancy or if the subduction zone freezes [Nikolaeva et al., 2010, Regenauer-Lieb et al., 2001].

Receiver functions image a deepening of the Moho from ~ 35 km at a station positioned directly above the Dauki Thrust, to ~ 50 km at a site within the Surma Basin [Mitra et al., 2018] (Figure 4.7), which suggests that the Dauki Thrust may have already facilitated at least 15 km of transport of the Indian transitional-oceanic crust into the mantle.

Together, the geologic history of the Indian transitional-oceanic crust, active faulting along pre-existing weak zones in the crust and deepening of the Moho, serves as evidence for the collapse of a passive margin due to a buoyancy instability that could eventually result in the creation of a new subduction zone [Nikolaeva et al., 2010]. Thus, the Surma Basin may be the first Cenozoic example of subduction initiation at a passive margin (Figure 4.7). However, given the three-dimensional aspect of the crustal geometry, it is unclear if this subduction initiation is spontaneous or whether it partially results from momentum transfer from the nearby active subduction of the Indian plate beneath the Burma plate [Mallick et al., 2019, Steckler et al., 2016].

Analog and numerical experiments suggest that as a negatively buoyant slab initiates the process of sinking into the mantle, it will pull the continental upper plate towards itself, leading to horizontal convergence and overthrusting (Figure 4.2C; Nikolaeva et al. [2010]). In our study area, this initial convergence would promote overthrusting of the Shillong Plateau, as observed for the Miocene (Figure 4.6). Experiments further indicate that if the convergent margin were to evolve into a self-sustained subduction zone, this overthrusting would slow down while horizontal shortening and basin subsidence would continue [Faccenna et al., 1999, Gurnis, 1992]. In our case, this is consistent with the possible decrease in the Pliocene uplift rate of the plateau (Figure 4.6) and the continued subsidence of the Surma Basin. This implies that the horizontal convergence across the Dauki Thrust has

been mostly accommodated by significant footwall transport into the mantle (Figure 4.7). This mechanism may also explain the increased India-Tibet modern GPS-velocities in the eastern Himalaya (compared to neighbouring Nepal) if the horizontal pull generated by the sinking transitional-oceanic plate is transmitted through the Indian continental lithosphere in a rigid manner (i.e. without stretching it like in a back-arc setting) (Figure 4.2C).

The process of sinking of the Surma Basin due to subduction initiation can explain its great depth (> 15 km) [Alam et al., 2003, Johnson and Nur Alam, 1991, Najman et al., 2016]. This basin is typically explained as being flexurally loaded by the topography of the Shillong Plateau [Johnson and Nur Alam, 1991]. However, the Surma Basin is significantly deeper than the 3 – 8 km foredeep of the Himalaya [Lyon-Caen and Molnar, 1985, Mitra et al., 2018]. Given that the Himalaya are on average three times taller than the Shillong Plateau, this implies that the subsidence is in excess of what is expected from topographic loading by the nearby orogen. However, excess subsidence is expected if the Surma Basin is sinking into the mantle. This also has implications for the Indo-Burman Wedge (Figure 4.7): the high mass storage ability of the Surma Basin may affect the growth and propagation of the outer Indo-Burman wedge by flexurally increasing its topographic slope. As sinking perturbs the wedge from its critical taper, it is forced to propagate westwards to reduce its taper and equilibrate. The long-term record appears consistent with this, showing at least 38 km of arc-perpendicular shortening over the past 8 Ma with an accelerated westward propagation of the deformation front in the Late Miocene [Betka et al., 2018].

4.4.4 Seismic Hazard Implications for the Shillong Plateau

A cold, sinking slab depresses the geotherm, which should promote brittle behaviour to greater depths along the nascent subduction interface [Syracuse et al., 2010]. This could explain the 21-37 km locking depth of the Dauki Thrust (Figure 4.7), which is deeper than the typical brittle-ductile transition depth of 10-15 km observed in continental fold-thrust belts. This extended seismogenic thickness should permit earthquake ruptures to reach greater depths. This may explain the unusual dimensions of the $M \sim 8+$ 1897 Great Shillong Earthquake (100 km along strike, reaching 30 km depth, up to 25 m slip) [Bilham and England, 2001, England and Molnar, 2015].

From our geodetic block model, we estimate a moment accumulation rate of $\dot{M}_0 = 1.4 - 2.1 \times 10^{18}$ N-m/yr (potency rate for the Shillong Plateau = $48-71 \times 10^6$ m³/yr), comparable to previous estimates [England and Molnar, 2015]. This accumulated elastic strain must be released in earthquakes and/or post-seismic processes along active faults (Dapsi Thrust, Dauki Thrust, Oldham Fault, Brahmaputra Fault, and Chedrang Fault). In the 122 years since the last great earthquake in this region, strain equivalent to a $M_w 7.3-7.6$ earthquake has already accrued within the Shillong Plateau. This represents a significant seismic hazard and risk to the populations of eastern India, Bangladesh and western Myanmar and yet it is only the minimum hazard estimate, since we assumed the 1897 event fully relaxed the crust. However, when we consider the ~ 300 km length of the Dauki Thrust, together with

the Dapsi Thrust has not ruptured in at least the last 500 years, if not longer [Rajendran et al., 2004, Steckler et al., 2008], the accumulated strain reservoir grows almost by an order of magnitude.

4.5 Conclusions

Sinking of the negatively buoyant transitional-oceanic lithosphere below the Surma basin may represent the only Cenozoic example of passive margin subduction initiation. This process likely began 8-14 Ma. In a first stage, slip on the Dauki Thrust caused by sinking of the footwall pulled the Shillong Plateau to the south and caused it to rise; erosion kept pace with uplift, ensuring no topography was built. In a second stage, once the Cretaceous sedimentary package was eroded away, the decreased erodibility of the exposed crystalline basement of the plateau allowed it to gain elevation. In the present day, the horizontal pull generated by this sinking of the Indian plate south of the Shillong Plateau must be dragging the Indian continental lithosphere southwards and leading to the increased geodetic convergence rate between India and the Bhutan Himalaya, compared to the neighbouring Nepal Himalaya. The collapsing Surma Basin may also affect the propagation of the Indo-Burman Wedge, and may be the cause of a documented acceleration in westward propagation of the wedge since the Miocene. The cold and brittle nature of the transitional-oceanic Indian lithosphere widens the seismogenic zone, presenting an increased seismic risk to the people of India, Bangladesh and Myanmar.

4.6 Acknowledgements

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Chapter 5

A unified framework for earthquake sequences and the growth of geological structure in fold-thrust belts*

Abstract: Observations of fold growth in fold-thrust belt settings show that brittle deformation can be localized or distributed. Localized shear is associated with frictional slip on primary faults, while distributed brittle deformation is recognized in the folding of the bulk medium. The interplay of these processes is clearly seen in fault-bend folds, which are folds cored by a fault with an abrupt change in dip (e.g., a ramp-décollement system). While the kinematics of fault-bend folding were described decades ago [Suppe, 1983], the dynamics of these structures remain poorly understood, especially the evolution of fault slip and off-fault deformation over different periods of the earthquake cycle. In order to investigate the dynamics of fault-bend folding, we develop a numerical modeling framework that combines a long-term elasto-plastic model of folding in a layered medium with a rate-state frictional model of fault strength evolution in order to simulate geologically and mechanically consistent earthquake sequences. In our simulations, slip on the ramp-décollement fault and inelastic fold deformation are mechanically coupled processes that build geologic structure. As a result, we observe that folding of the crust (like fault slip) does not occur steadily in time but is modulated by earthquake cycle stresses. We suggest combining seismological and geodetic observations with geological fault models to uncover how elastic and inelastic crustal deformation generate fault-bend folds. We find that distinguishing between the elastic and inelastic response of the crust to fault slip is possible only in the postseismic period following large earthquakes, indicating that for most fault systems this information currently remains inaccessible.

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5.1 Introduction

Geophysical models often assume that the lithosphere is broken into elastic blocks along plate boundary faults, and all inelastic (permanent) deformation is focused within the fault zone [Reid, 1911]. This simplification is typically applied to interpret short-term observations related to the earthquake cycle. For example, many fault models are composed of a dislocation plane inserted into an elastic medium, with geodetic and seismological observations interpreted as the response of the medium to localized shear (slip) on the fault (e.g., Kanda and Simons [2012, 2010], Okada [1985], Savage [1983]). The success of this computationally inexpensive approach in matching observations relative to the data uncertainty has reinforced its use (e.g., Savage [1983], Van Zwieten et al. [2013], Vergne et al. [2001]).

However, this description of tectonics is unable to reproduce a fundamental observable: the ability of fault slip to permanently deform the lithosphere and generate relief. While planar dislocations may sufficiently approximate faults over some timescales, additional considerations must be incorporated into tectonic models in order to bridge timescales (from the seconds to years of geodetic and seismic data, to the tens of thousands to millions of years associated with mountain building) and develop geologically consistent representations of crustal deformation.

FIGURE 5.1: Shear stress inside an elastic medium grows with slip on a non-planar fault. (A) Schematic of a fold-thrust belt. Shortening causes faulting and folding of the initially horizontal the sedimentary package. (B) Model of the frontal fault and fold. For 1m of slip on the ramp-décollement system, the surface uplift due to a purely elastic response of the crust is shown in blue. Background colors show maximum shear; black crosses show the orientation of the deviatoric part of the stress tensor (inward arrows - compression, outward arrows - extension).

5.1.1 Fault bends and unbalanced stress

The issue of crustal deformation is particularly relevant in convergent margins, where shortening of the crust is accommodated along thrusts with associated folding. In these settings, a gently dipping subduction megathrust or continental thrust separates the down-going plate from the overriding plate (Figure 5.1A). Typically, the hanging wall of the megathrust will host a series of more steeply dipping splay faults or ramps, with the younger thrusts near the toe of the system. When shortening is accommodated across this system, slip on the megathrust will be transferred onto one of the steeper splays, resulting in an active fault bend (Figure 5.1).

In most modeling studies of crustal deformation based on seismological and geodetic data, both the footwall and the hanging wall are assumed to behave as linear elastic solids over short timescales (seconds – years). Such an assumption is appropriate for planar faults, where uniform slip on the interface results in a net translation between the hanging wall and footwall. However, slip across a geometric bend leads to unbalanced shear stresses within an elastic medium (Figure 5.1B). These stress components grow with accrued slip on the main fault, and cannot be supported indefinitely by a purely elastic body. In order to generate an internally consistent modeling approach, some mechanism(s) must be added to relax these stresses. Ideally, the mechanisms that are added will also produce structures consistent with geological observations of fault bends in fold-thrust environments (Figure 5.2). By implementing geologically consistent mechanisms into elastic dislocation modeling approaches, we can therefore solve two problems at once: (1) balance shear stress loads due to slip on non-planar faults, and (2) create permanent crustal deformation and build relief.

5.1.2 Off-fault deformation

In order to relax unbalanced stresses within the crust, inelastic deformation must occur away from the main fault i.e. off-fault deformation (OFD). Geological observations of anticlines in fold-thrust belts show that permanent OFD occurs through inelastic shear, and can manifest in various ways [Butler et al., 2020, Cosgrove, 2015, Shaw et al., 1999, Suppe, 1983, Suppe et al., 2005]. These include localized slip on a backthrust (conjugate to the main thrust) to create a pop-up structure [Gregg Erickson et al., 2005], slip on bedding plane contacts within the fold [Johnson and Berger, 1989, Suppe, 1983], layer buckling and viscous/plastic flow of weak layers [Berger and Johnson, 1980, Biot, 1964, Butler et al., 2020, Fletcher, 1977, Honea and Johnson, 1976, Ramberg, 1963, 1970], and distributed deformation by thickening/thinning of individual layers within a thrust sheet [Honea and Johnson, 1976, Johnson and Pfaff, 1989, Poblet and McClay, 1996]. In this article, we specifically address folding of the hanging wall accommodated by bedding plane slip (which is equivalent to layer buckling); this mechanism relaxes the unbalanced elastic stresses in the hanging wall that arise from slip on the ramp-décollement system.

FIGURE 5.2: Surface deformation due to fault-bend folding. Fault slip leads to the formation of an axial surface/hinge within the hanging wall, at the junction of the ramp and décollement. This hinge is a shear band that is stationary with respect to the ramp and décollement and does not advect with the thrust sheet (shown by the flux of material passing through the hinge). The horizontal (red) and vertical (blue) instantaneous displacement fields are shown in response to unit slip (assuming infinitesimal strains) for fault-bend folding (solid) and elastic (dashed) deformation. The displacement gradients are sharp across the hinge as we assume that the fault bend is infinitesimally narrow, and hence the shear band appears as a backthrust.

5.1.3 Fault-bend folding and displacement fields

The kinematic theories of fault-bend folding and buckle folding describe the growth of a fault-cored fold by bedding plane slip within an incompressible hanging wall [Brandes and Tanner, 2014, Johnson and Berger, 1989, Suppe, 1983]. Since bedding plane thickness is typically conserved as a thrust sheet is advected through a fault bend, the amount of slip on the fault and the shape of the resulting fold are directly linked [Suppe, 1983].

For example, consider a ramp-décollement system with initial horizontal layering (Figure 5.2). As the thrust sheet is advected through the fault bend, a shear band appears in the hanging wall; this is the active axial surface or the hinge zone. If fault bend occurs over infinitesimal width, then the axial surface will be infinitesimally narrow, and the instantaneous surface displacement field will be identical to that above a backthrust (Figure 5.2).

However, axial surfaces and backthrusts are fundamentally different in their long-term kinematics [Gregg Erickson and Jamison, 1995, Gregg Erickson et al., 2005, Suppe, 1983]. While a backthrust will be progressively advected up the main thrust ramp, an active axial surface is fixed relative to the underlying fault geometry. The active axial surface is paired with an inactive axial surface; the latter represents the initial lower limit of the hanging wall above the fault bend that has been advected up the ramp. The spatial separation of the inactive and active axial surfaces therefore represents the slip

on the fault ramp after finite shortening. Although the geometries of the two axial surfaces appear similar, active deformation occurs only at the active axial surface (Figure 5.2).

The deformation in the hinge zone (active axial surface) can be considered to represent either slip on the bedding plane contacts or focused viscous/plastic shear of thin mechanically weak beds (while thicker and stronger beds remain undeformed). This type of deformation occurs because the transverse anisotropy in a stack of sediments allows mechanically stronger beds to retain their shape and structure, with shear localizing either within weaker beds or along the interfaces between beds (Figure 5.2). Most numerical models of fault-related deformation neglect this type of deformation because it requires knowledge of the orientations of the bedding planes in the hanging wall, information that is typically difficult to access, and in the case of imbricated fold-thrust belts, may be complex.

The net surface displacement field associated with this mechanism has large displacement gradients at two locations (Figure 5.2): (1) where the thrust ramp intersects with or comes closest to the free surface, and (2) where the active axial surface intersects the free surface. A model that represents only the elastic response of the medium to slip on a ramp-décollement system cannot reproduce the second displacement gradient at the back of the fold (Figure 5.1B, 5.2). This implies that there is diagnostic power in looking for such hinge-related OFD signals in geodetic displacement fields over growing folds, in order to study how the crust deforms in response to fault slip and infer its constitutive properties.

5.1.4 Geodetic observations of fault-bend folding

While geological mapping and sub-surface imaging provide information about the long-term growth of anticlines (integrated over multiple earthquake sequences), geodetically derived deformation fields are sensitive to the earthquake cycle and contain information regarding the time-varying nature of deformation [Allmendinger et al., 2009]. It is therefore somewhat surprising that geodetic observations of convergent margins over the interseismic period show a pronounced uplift zone in the same location as uplift recorded in geological and geomorphological proxies representing average behaviour over millions of years [Grandin et al., 2012, Jackson and Bilham, 1994, Meade, 2010, Saillard et al., 2017]. These observations, along with the common assumption that folding occurs as a viscous process [Biot, 1961, Ramberg, 1970, Ramberg and Johnson, 1976], has led to speculation that the growth of geological structure may be a time-invariant process occurring mostly during the slow and long-lived interseismic period, at rates below the detectability limit [Jolivet et al., 2020, Malatesta et al., 2020, Meade, 2010, Saillard et al., 2017].

However, observations of earthquakes demonstrate that inelastic hanging wall deformation can occur during the coseismic and postseismic phases of the earthquake cycle (Figure 5.3). The coseismic displacement field from the 1991 M_w 7.6 Chi-Chi earthquake, Taiwan (data from Kuo et al. [2014]), shows meter-scale displacements associated with OFD (Figure 5.3A-B) that produced a meter-scale

FIGURE 5.3: Observations of fault-bend folding modulated by the earthquake cycle. (A) The frontal rupture of the Chelungpu Fault in the 1999 M_w 7.6 Chi-Chi earthquake, western Taiwan. (B) Horizontal coseismic displacement field [Kuo et al., 2014] from optical imagery. We highlight four transects exhibiting significant displacement gradients likely due to hinge deformation, as predicted in Figure 2. (C) Postseismic deformation, average velocity field from 1996-1999, of the frontal folds of the Tabas foldbelt following the 1978 M_w 7.3 Tabas-e-Golshan earthquake. (D) Shallow fold growth (Kurit fold) has been imaged in the line-of-sight direction (mostly vertical) from 1996-2017 by multiple InSAR missions [Copley, 2014, Zhou et al., 2018]. The data are acquired in descending mode unless explicitly specified. We observe temporally decaying deformation with a significant displacement gradient in the back limb of the fold, likely caused by hinge deformation.

coseismic fold scarp [Chen et al., 2007]. Furthermore, time varying growth of the Kurit fold in Iran documents likely postseismic OFD. These observations, based on ~ 20 years of InSAR line-of-sight displacement fields (Figure 5.3C-D), show a signal on the order of 1-5 mm/yr, decaying with time; the signal is presumably a response to an initial stress perturbation caused by the 1978 M_w 7.3 Tabas-e-Golshan earthquake (data from Copley [2014], Zhou et al. [2018]). The coseismic displacement field for the Chi-Chi earthquake and the postseismic velocity field over the Kurit fold both show two large spatial gradients, similar to the prediction for axial surface deformation (Figure 5.2); we interpret this to be the result of fault-bend folding above a ramp-décollement system. These observations demonstrate that inelastic deformation, likely associated with bedding plane slip, occurs over a finite length scale, is a time-varying process, and can be modulated by the earthquake cycle.

To understand how brittle faulting and folding can occur simultaneously, we develop a framework to incorporate bedding-plane slip at earthquake timescales, by estimating the down-dip width (length-scale) of the actively deforming hinge zone, and incorporating hinge-related OFD into earthquake sequence simulations and related inverse problems.

We first use a Boundary Element Method (BEM) framework to model a ramp-décollement system in an elasto-plastic medium, incorporating the anisotropy of sedimentary layers [Huang and Johnson, 2016, Johnson, 2018]. We impose shortening in the model and calculate the evolution of geometry and strain-rate within the hanging wall. This model replicates the long-term kinematics of a fault-bend fold while maintaining the stresses in the fault and bedding plane system within the plastic yield envelope. Since hinge deformation in natural ramp-décollement systems has been observed both co- and post-seismically (Figure 3), we take our study a step further to assess how hinge deformation might appear in typical surface observations over timescales associated with seismo-geodetic methods. We do this by calculating the shear stress change from incremental deformation in the elasto-plastic models, and using it to drive earthquake sequence simulations; this computed shear stressing rate serves as the long-term loading rate for the fault and the inelastic regions within the hanging wall. With this loading rate for the fault and hanging wall system, and using the laboratory-derived rate-state friction law to describe the evolution of fault strength [Dieterich, 1979, Rice and Ruina, 1983, Ruina, 1983], we show that it is now possible to evaluate geologically-consistent earthquake sequence scenarios. The model recreates all parts of the earthquake cycle, from slow interseismic loading to the nucleation of frictional instabilities and an eventual earthquake, as well as the associated OFD in the hinge zone. We discuss the impact of hinge deformation on kinematic inversions of inelastic strain (slip on faults) based on geophysical data from different parts of the earthquake cycle, and evaluate our ability to detect hinge-related OFD using typical seismo-geodetic methods.

FIGURE 5.4: (A) Initial geometry for the elasto-plastic folding models. Slip is imposed on the ramp and décollement as a slip boundary condition; far-field loading is simulated by extending the décollement horizontally for 1000km (outside the model domain). We consider three different levels of smoothness of the fault bend (1, 2, 3). (B) Elastic-plastic schematic. The medium acts as an elastic body until the yield condition is reached, beyond which inelastic strain is accommodated by slip on the bedding planes shown in (A). (C) The hinge zone is the region within the hanging wall where inelastic deformation occurs (shown in lighter colours, representing more slip on the bedding planes), while the rest of the hanging wall behaves purely elastically (black - no slip). The outputs from the folding model allow us to explicitly define the inelastically deforming domain by its geometry and effective loading rate. In the schematic, we show bedding plane deformation in the hanging wall using double couples to indicate that a series of bedding plane contacts line up to appear as an equivalent thrust fault, in line with our expectations from fault-bend folding (see Figure 5.2).

5.2 Methods

We study the effect of maintaining a stress balance/equilibrium over the scale of a single fault-bend fold (Figure 5.4A). Specifically, we are interested in structures that form close to the free surface, where slip on a décollement is fed onto a ramp; they are therefore thought to be governed by brittle mechanics [Dahlen, 1990]. Since we are interested in the combined effects of elastic stress transfer and structural evolution of the system, we are working within a timescale ranging from seconds to 106 years. Beyond this timescale, the nucleation and the growth of new faults cannot be neglected, and other mechanisms of deformation and stress redistribution associated with significant changes in geometry occur [Avouac, 2015b]. Thus, our approach evaluates the short-to-intermediate timescale evolution of the system. At the earthquake cycle timescale, the geometry of the system is essentially static, while at the fold-evolution timescale, we consider the deformation of the hanging wall but ignore both the creation of new faults and any weakening or re-strengthening effects of frictional sliding (Figure 5.4B).

5.2.1 Elasto-plastic model of fold growth

We develop a boundary element framework (BEM) following [Johnson \[2018\]](#), in order to investigate how bedding plane slip away from a primary fault can accommodate the development of a fold. Unlike a typical continuum model, we do not evaluate the stress field at every point within the medium. Instead, we assume that bedding planes act as frictional contacts or narrow shear zones, and mesh these bedding contacts using 1-d fault segments (Figure 5.4A). This reduces the computational expense by exploiting the anisotropy inherent to a stack of sedimentary layers. In our model, initially horizontal strata can deform by passive translation or rotation, as well as by slip on the bedding plane contacts. We prescribe the rheology of the bedding plane contacts to be elastic-perfectly plastic, governed by Mohr-Coulomb elasto-plasticity (Figure 5.4B). Specifically, we use the Mohr-Coulomb criterion to model brittle failure (not temperature-dependent plastic flow or ductile creep), governed by a yield condition (Figure 5.4B) as follows:

$$|\tau^\infty + \tau| \leq \mu \bar{\sigma} \quad (5.1)$$

Here, τ^∞ is the remote load (in the shear direction; see Figure 5.4A) applied to the system. This load can arise from kinematically imposed horizontal shortening across the entire system, topographic loading, or slip on the décollement. In our model, uniform slip on the ramp-décollement fault is imposed as a boundary condition, which loads the hanging wall material. τ is the elastic shear stress field resulting from an arbitrary distribution of slip on the bedding plane contacts. The right-hand side of Equation 5.1 represents the brittle yield condition: μ is the coefficient of friction on each boundary element, while $\bar{\sigma}$ is the normal stress (normal to the boundary element). This implies that the yield condition on every boundary element is prescribed as a function of normal stress, and the elastic stress within the medium (due to loading) cannot exceed this yield condition. Once it does, the boundary elements will deform by frictional slip to ensure that the residual stress field is within the yield envelope (Figure 5.4B). In addition to the stress evolution due to slip on various structures, we also consider finite deformation of the medium due to the advection of material and rotation of the bedding plane contacts (Figure 5.5). This is done by calculating incremental displacements of all boundary elements and remeshing [[Johnson, 2018](#)]. This change in geometry also modifies the stress field, since the shear stresses resolve differently for different fault and bedding plane dips.

We assume that slip on faults does not change the elastic stress field in the fault-normal direction ($\Delta \bar{\sigma} = 0$). This is primarily to simplify the computations, as including $\Delta \bar{\sigma}$ would make the system of equations difficult to solve with our methodology (discussed later in this section). However, it also allows for the possibility that inelastic perturbations in the fault-normal stress component (deviations from the lithostatic condition) may not persist over long timescales, and are accommodated by other physico-chemical means, such as tensile cracking, fluid flow, pressure solution, and vein formation [[Gratier et al., 1999, 2013](#), [Otsubo et al., 2020](#), [Sibson, 2019](#)].

Assuming linear elasticity, we describe the change in elastic shear stress, $\Delta\tau$, arising from incremental slip Δs on the boundary elements as a linear operation [Segall, 2010, Singh and Rani, 1993], given a traction kernel K_τ :

$$\Delta\tau = K_\tau\Delta s \quad (5.2)$$

Applying Equation 5.2 in the failure criterion (Equation 5.1) gives the condition that must at all times be satisfied for all boundary elements. Because time is absorbed into increments of far-field loading, we write the evolution of slip on the boundary elements as a response to incremental loading, $\Delta\tau^\infty$, as follows,

$$\begin{aligned} |\tau^\infty + K_\tau s_0| &\leq \mu \bar{\sigma} \text{ at } t_0 \\ |\tau^\infty + \Delta\tau^\infty + K_\tau(s_0 + \Delta s)| &\leq \mu \bar{\sigma} \text{ at } t_0^+ \end{aligned} \quad (5.3)$$

If we define $\tau_+ = \tau^\infty + \Delta\tau^\infty + K_\tau s_0$ in Equation 5.3, as the instantaneous unbalanced elastic stress field at a given time, the solution for Δs can be formulated as a convex optimization problem. This linearity in the system, despite the use of absolute values, appears because we do not consider changes in the fault-normal stress (as mentioned previously), and can instead use the sign of τ_+ to decompose the solution space of Δs as

$$\begin{aligned} \tau_+ > 0 \quad & K_\tau\Delta s \leq \mu \bar{\sigma} - \tau_+ \\ & \tau_+ + K_\tau\Delta s \geq 0 \end{aligned} \quad (5.4)$$

$$\begin{aligned} \tau_+ \leq 0 \quad & -K_\tau\Delta s \leq \mu \bar{\sigma} + \tau_+ \\ & \tau_+ + K_\tau\Delta s \leq 0 \end{aligned} \quad (5.5)$$

$$\text{minimization} \rightarrow \Delta s \sim 0 \quad (5.6)$$

The set of equations derived from the failure criterion act as a bounding function for the solution Δs . To uniquely solve for Δs , we apply a least-squares minimization on the incremental slip (Equation 5.6), since we assume that the physical system wants to do minimal irreversible work to satisfy the stress budget (Equation 5.1). We use the MATLAB function *lsqmin* to do the convex optimization.

Given this framework, we conduct 14 numerical experiments (with varying numbers of bedding layer contact surfaces - Table 5.1) for three fault geometries (with different degrees of curvature at the fault bend) (Figure 5.4A). The underlying fault geometry consists of a 5 km long, 30° ramp splaying upward from a horizontal décollement, and is motivated by the observed length scales shown in Figure 5.3. We are interested in how the width of the hinge zone (hinge width, Figure 5.4C) is affected by both the sharpness of the fault bend and the number of bedding planes (equivalent to the thickness of the bedding layers), since this will be directly relevant in discretizing the domain of inelastic deformation for the earthquake-sequence simulations. In our simulations, we measure the hinge width by fitting the slip distribution on each bedding layer to a Gaussian distribution and finding the 95% confidence interval of the associated shape function.

FIGURE 5.5: (A) Geometry and model setup for the earthquake sequence simulations (see Table 5.2 for details). (B) Three scenarios of frictional properties on the fault: red – velocity weakening, black – velocity strengthening.

FIGURE 5.6: (A) Snapshots of incremental deformation (shown as colored bedding planes) of the hanging wall with increasing shortening. The blue lines approximately separate flat layers from tilted layers. The first snapshot has not yet reached a steady-state geometry; the next three snapshots show steady-state deformation, i.e. the hinge width and slip rate do not vary with incremental deformation. (B) and (C) Equivalent snapshots of simulations with different material properties, compared to the simulation outlined with a dashed box in (A), show that the fold form does not depend on G or μ (layers).

Property	Value
Number of bedding planes (N_{BP})	5-18
Total shortening (km)	2
Number of slip increments	40
Remote loading slip increments, $\Delta s^\infty(m)$	50
Shear Modulus, G (GPa)	10,30
Friction coefficient, μ	0.01,0.1
Poisson's ratio, ν	0.25

TABLE 5.1: Input parameters for the elasto-plastic fold evolution models.

5.2.2 Earthquake sequence simulations

In order to evaluate the impact of hinge-related OFD on earthquake cycle deformation, we develop a set of dynamic rupture simulations informed by the geometries and stress states arising from our models of fold growth. At the timescale of earthquakes and the earthquake cycle, we assume that the geometry of the main fault and the surrounding medium is invariant, since each earthquake accommodates meters of slip while the fold is kilometers wide. We use the fault geometry from Figure 5.4A (number 2) in the earthquake sequence simulations, since it creates reasonably narrow hinge zones (Figure 5.7C), but is smooth enough that the kink, which violates the principle of interpenetration of an elastic body [Romanet, 2020], does not strongly influence the results. We take the geometry and instantaneous slip rate values from a 15-layer model run (Figure 5.5A) after 1.5 km of shortening (the hinge width and slip rates have reached steady state – see Figure 5.6). We only extract the boundary elements with slip rates greater than 1% of the imposed slip on the ramp-décollement (the slip boundary condition), as this helps to reduce the dimensionality of the simulation while still capturing the deforming regions of the model (Figure 5.5A).

Although our long-term elasto-plastic fold growth model is able to reproduce geologically validated geometries, that model assumes that fault strength is constant in time, which is not realistic (Figure 5.4B). In fact, faults weaken and heal with loading conditions, and earthquakes occur when a fault weakens faster than it releases stress by slipping (see review by Scholz [1998]).

Since fold growth with hinge-related OFD has been observed in both co- and post-seismic natural settings, this variation in friction must be considered. In our simulations, we recreate the same cumulative slip and deformation as in the long-term models, while also evaluating the partitioning of slip into the *coseismic*, *postseismic* and *interseismic* phases of the earthquake cycle.

5.2.2.1 Rate-state frictional earthquake sequences

We treat the boundary elements that make up the ramp-décollement and the hinge zone as elastically coupled faults [Johnson, 2018]. These elements are loaded at a known long-term rate, $\dot{\tau}^\infty = -K_\tau \dot{s}^\infty$

(Figure 5.5A). We derive $\dot{s}^\infty$ from the outputs of the long-term simulation. The elasto-plastic long-term simulations have no explicit time dependence since incremental slip drives the system. In order to incorporate time into stress evolution over the earthquake cycle, we need to make a choice for the average slip rate of the system (10^{-9} m/s). The relative variation in Δs can be used to modify $\dot{s}^\infty$ accordingly, giving $\begin{bmatrix} \dot{\tau}^\infty \\ \dot{\tau}_{OFD}^\infty \end{bmatrix}$. We use a boundary integral formulation such that the linear momentum balance on these elements is satisfied as the simulation evolves as,

$$\begin{bmatrix} \dot{\tau}^\infty \\ \dot{\tau}_{OFD}^\infty \end{bmatrix} + K_\tau \begin{bmatrix} v \\ v_{OFD} \end{bmatrix} = \dot{\mu} \bar{\sigma} + \frac{G}{2v_s} \begin{bmatrix} \dot{v} \\ \dot{v}_{OFD} \end{bmatrix} \quad (5.7)$$

The left side of Equation 5.7 is the instantaneous elastic stressing rate as a function of slip velocity v (resolved in the shear direction for each boundary element). This is a quasi-dynamic approach where we neglect wave-mediated stress transfer through an elastic medium. The right side contains two terms: the first is the rate of change of the frictional strength of each element, and the second the radiation damping rate (a proxy for the energy lost to the system, at velocities comparable to the shear wave velocity v_s). We use rate-state friction and the ageing law to describe the evolution of fault strength with velocity v and a state parameter θ over a critical slip distance L [Dieterich, 1979, Ruina, 1983].

$$\begin{aligned} \mu &= \mu_0 + a \log \frac{v}{v_0} + b \log \frac{v_0 \theta}{L} \\ \frac{d\theta}{dt} &= 1 - \frac{v\theta}{L} \end{aligned} \quad (5.8)$$

where μ_0, v_0 are reference values of the friction coefficient and velocity, respectively, while a, b are material-specific constants. Equations 5.7 and 5.8 represent a coupled system of ordinary differential equations in v, θ that we solve using a Runge-Kutta 4th order adaptive time-stepping solver – MATLAB based *ode45*.

We choose representative values of a, b and L from typical ranges of laboratory-derived studies (Table 5.2; also see Blanpied et al. [1995]). Fault sections with $a - b < 0$ are considered velocity-weakening (i.e. a frictional instability could nucleate on it), while $a - b > 0$ represents a frictionally stable section of a fault, where any slip perturbation is damped back to steady-state. $\bar{\sigma}$ follows a depth-dependent lithostatic condition assuming hydrostatic pore-fluid infiltration; we choose a minimum threshold value for $\bar{\sigma}$ to be 20 MPa (equivalent to a depth of ~ 1 km) to stabilize the response of the fault close to the free surface. We do not allow $\bar{\sigma}$ for any element to change with slip even though elastic stress calculations would suggest that $\bar{\sigma}$ for boundary elements close to the fault-bend is altered by fault slip. This earthquake-sequence implementation of normal-stress change is a known issue but does not have a physically reasonable and computationally convenient fix as yet. The issue arises because the amplitude of normal stress perturbations are a geometric effect which grow linearly with fault

Property	Value
a	2
$a - b$ velocity weakening	40
$a - b$ velocity strengthening	50
v_0 (m/s)	10,30
L (m)	0.01,0.1
$\dot{\gamma}^\infty$ (m/s)	0.25
Fault dip ($^\circ$)	ramp - 30, décollement - 0
Fault length (km)	ramp - 5, décollement - 20
N_{BP}	15
G (GPa)	10
ν	0.25
v_s (m/s)	3000
ρ (kg/m ³)	1600
$\bar{\sigma}$ (MPa)	$\rho g x_3$ (depth-varying)
h^* (m)	667
Cohesive zone (m)	174
Mesh size (m)	50
W_{VW} (km)	13,16,20
Instability ratio	20,24,30
v_{dyn} (m/s)	0.1

TABLE 5.2: Input parameters for the earthquake-sequence simulations.

slip [Dunham et al., 2011, Romanet, 2020, Tal et al., 2018]; such an issue does not arise in planar fault simulations. Relaxing these stress components may occur through fluid-driven and visco-plastic failure processes [Dunham et al., 2011, Gratier et al., 1999, Otsubo et al., 2020, Sibson, 2019]; these are mechanisms that we do not include in this study, and could be considered in future studies.

In our numerical experiments, we fix all material properties (Table 5.2) and vary only the location of the up-dip transition from velocity-weakening to velocity-strengthening on the ramp-décollement system (Figure 5.5B), while the hinge region is considered only to be velocity-strengthening (Figure 5.5). We choose velocity-strengthening friction for the bedding plane faults that make up the hinge for two reasons: (1) this is one of the first earthquake-sequence studies where we explicitly account for inelastic strain associated with off-fault bedding plane slip, and so, do not want to explore too many complicated unknowns simultaneously, and (2) this is a quasi-dynamic simulation, which implies that we may be introducing numerical errors due to neglecting wave effects in the coseismic stress feedback between the fault and the hinge region, an effect which is exacerbated when the hinge can behave in a frictionally unstable manner.

We define the instability ratio, $\frac{W_{VW}}{h^*}$, as the ratio of the width of the velocity-weakening region, W_{VW} (Figure 5.5B), to a critical nucleation length-scale defined following Rubin and Ampuero [2005] as,

$$h^* = \frac{GL}{(1 - \nu)(b - a)\bar{\sigma}} \quad (5.9)$$

The instability ratio is a key parameter that is thought to determine the complexity of earthquake sequences, with low values (1-10) showing quasi-periodic slow-slip and system-wide ruptures, to high values (> 100) showing aperiodic Gutenberg-Richter earthquake statistics [Cattania, 2019]. In this study, we investigate the instability ratio range of 20-30, which is large enough to produce partial ruptures of the seismogenic region and is computationally convenient to evaluate. We also choose this range because it produces interseismic signals consistent with those observed in typical convergent margins: low to no interseismic shallow creep updip of locked regions [Almeida et al., 2018], and only a gradual unlocking of otherwise locked regions i.e., locked regions being eroded by deep creep in the years building up to an earthquake induce minimal curvature to the surface displacement timeseries [Bruhat et al., 2016, Mavrommatis et al., 2014, 2017].

We run three earthquake sequence experiments, for different W_{VW} , on the given fault geometry with hinge-related OFD for a period of 5000 years (Figure 5.5B); we report the results in the following section.

5.3 Results

In the following sections, we first discuss the length scale associated with the axial surface (hinge width) and the factors that control it, based on our elasto-plastic fold evolution models. We then consider the effects of incorporating hinge-related OFD, using a 15-layer model, into numerical simulations of earthquakes and aseismic processes. We present the results of our elasto-plastic fold evolution models in Figures 5.6-5.7, and important results from the earthquake sequence simulations in Figures 5.8-5.9.

5.3.1 Hinge localization

Previous work on multi-layer folding suggests that folding in the brittle crust varies from sinusoidal (large hinge width) to kink bands (narrow hinge width), depending on a combination of material properties – the rigidity of the medium, frictional strength of the layer interfaces, and bed thickness [Biot, 1961, 1964, Honea and Johnson, 1976, Huang and Johnson, 2016, Johnson, 2018, Ramberg, 1970, Ramberg and Johnson, 1976]. In our experiments, we find that the dominant control on hinge width (Figure 5.7A-B) is instead exerted by the number of bedding planes and the angularity of

FIGURE 5.7: Results of the elasto-plastic folding simulations. (A) Visualization of the length-scale and strain-rate at steady-state for three different experiments associated with Fault 2 (for geometry see Figure 5.4A). The darker colors of the bedding plane contacts represent no inelastic deformation, while the lighter colors indicate finite slip (normalized by slip imposed on the décollement). The ramp-décollement is marked in red. (B) Slip on each bedding plane is plotted against the horizontal distance x_2 . Top and bottom refer to the layers marked in (A). (C) Localization of the hinge (at steady state) is shown as a function of the number of bedding planes (N_{BP}) for the three fault geometries. The estimated hinge width is plotted with some scatter in the x-axis to aid visualization. (D) Comparison between the moment (potency) released by the hinge as a percentage of the total moment released by structures on the order of the length-scale of the fold (~ 10 km) for each of the fault geometries. Also plotted is the expected moment release by a fully localized shear band based on kinematic fault-bend folding theory.

the primary fault bend (inset Figure 5.4A), rather than the shear modulus of the medium, friction coefficient of the bedding plane contacts or the number of slip increments (Figure 5.6).

This result contrasts with some theoretical studies of buckle folding, which suggest that the final fold form depends on the elastic and frictional properties of the deforming medium [Honea and Johnson, 1976]. The discrepancy may be due to the different boundary conditions in our approaches – where Honea and Johnson [1976] considered remote deviatoric loading without any fault, and solved for equilibrium conditions within thin elastic beams, we explicitly drive folding in an elasto-plastic medium with a bent localized slip surface (ramp-décollement fault). In addition, the fault-normal stress in our simulations is depth-dependent pressure and is independent of fault slip. In contrast, fault-cored elasto-plastic simulations that permit the fault-normal stress to change with slip show a dependence on elastic and frictional properties [Johnson, 2018]. Specifically, in those models, sharp kink bands tend to form for higher shear strength of the layer interfaces or lower shear modulus of the medium [Johnson, 2018].

Despite the dissimilarities between these approaches and results, in all of the models, the number of bedding planes (equivalent to the bed thickness) exerts a dominant control on folding. Specifically, the hinge narrows as the number of bedding planes increases (Figure 5.7A-B). Since the number of bedding planes determines the thickness of the bedding layers, increasing the number of bedding planes decreases the bending resistance of the thrust sheet, which may follow the power-law relationship: $\tau_{\text{bending}} \propto \left(\frac{1}{N_{BP}}\right)^\alpha$ (Figure 5.7C).

This interpretation implies that as $N_{BP} \rightarrow \infty$, the hanging wall will essentially act as a viscous incompressible fluid being transported past a rigid corner [Turcotte and Schubert, 2002]. The hinge width in that case will depend entirely on the length scale over which the dip angle changes between the ramp and the décollement. With increasing width of the fault bend (Faults 1-3 in Figure 5.4A), the hinge width also increases (Figure 5.7C). At $N_{BP} = 18$ (the maximum in our model, corresponding to ~ 100 m thick beds), Fault 3 appears to have reached its asymptotic value, while Fault 1 (kinked fault, see Figure 5.4A) can likely localize further. We do not explicitly verify the high N_{BP} case, as models with $N_{BP} > 18$ require increasingly higher resolution of boundary element meshing, which increases the computational cost considerably.

To reinforce the general solution offered by our method, we compare the partitioning of moment between on- and off-fault processes in our models with the kinematics of a fault-bend-fold from velocity vector analysis [Johnson and Berger, 1989, Sathiakumar et al., 2020]. To compute the percentage of moment released by hinge-related OFD, we consider 10 km of the ramp-décollement (consistent with the wavelength of the fold) as the on-fault moment component, while the cumulative moment released by all the individual bedding plane faults is the off-fault component. The values from our simulations match the kinematic results satisfactorily (Figure 5.7D), indicating that the inelastic component of deformation in our simulations is only a function of the difference in dips of the ramp and décollement. This is consistent with our expectation from the kinematics of fault-bend folding, which depends only

on the geometry of the fault and the orientation of the hanging wall rocks [Brandes and Tanner, 2014, Johnson and Berger, 1989, Suppe, 1983].

5.3.2 Earthquake sequences with hinge-related OFD

From the incremental inelastic strain computed from the elasto-plastic folding models, and using laboratory-derived friction laws, we show how fault slip and hinge-related OFD evolve in time (Figure 5.8). Using an adaptive time-stepping solver, we resolve the coseismic, postseismic and interseismic periods of the earthquake cycle. Our results show that earthquakes nucleate near locations where the fault transitions from velocity-weakening to velocity-strengthening friction, and propagate unilaterally in a crack-like manner (Figure 5.8A-C). As expected for instability ratios of 20-30, we see both system-wide ruptures and partial ruptures of the velocity-weakening region [Barbot, 2019, Cattania, 2019, Rice, 1993]. This variability emerges, even on a planar fault, due to the pre-stress state on the fault: when a previous rupture leaves part of the velocity-weakening section under-stressed, propagating ruptures may not be able to overcome the local fracture energy and will terminate. More complicated sequences of earthquakes appear when the up-dip extent of the velocity-weakening region and the ramp-décollement transition overlap (Figure 8b), which may be a consequence of geometric effects of the fault-bend [Li et al., 2018, Ong et al., 2019, Qiu et al., 2016, Sathikumar et al., 2020].

Following an earthquake, the velocity-strengthening parts of the fault display afterslip, i.e. stress-driven accelerated creep, in response to the coseismic stress change [Marone et al., 1991, Perfettini and Avouac, 2004]. Here, we focus only on afterslip on the shallow velocity-strengthening section of the fault, but note that the deep velocity-strengthening section also creeps postseismically (Figure 5.8A-C). Since afterslip depends on the magnitude of stress transfer [Marone et al., 1991], only large earthquakes and earthquakes that rupture close to the velocity-strengthening sections can drive a detectable acceleration in the creep rate. This creep may initiate during the earthquake or immediately after the event, and will then decay nearly logarithmically with time (linear on a log-log plot, Figure 5.8D-F).

In addition to on-fault processes, inelastic strain also occurs within the hinge zone. By varying the width of the velocity-weakening region, we can observe how propagating ruptures activate deformation in the hinge region. Hinge-related OFD appears to initiate when the rupture front reaches the ramp-décollement transition, and then evolves similarly to afterslip (Figure 5.8D-F). This stress-driven evolution tells us that the aggregate behavior of the hinge follows from slip on the individual bedding layer interfaces, which are governed by velocity-strengthening friction (Table 5.2).

Typical interpretations of earthquake sequence deformation are based on surface observations. To mimic this, we visualize the vertical displacement rate of the ground surface in response to fold growth over a 1000-year period (Figure 5.9). In all scenarios, we see peak deformation rates during and immediately after the earthquake. This rapid deformation is followed by a signal that decays in

FIGURE 5.8: (A-C) Slip evolution (on-fault only) for model geometries shown in Figure 5.5. The horizontal axis is the cumulative slip on the fault while the vertical axis is the along-fault distance (0 represents the free surface, marked by a red dashed line). The location of the dip change from ramp to décollement is highlighted in gray. The frictional instability nucleation size (h^*) and the width of the velocity-weakening region are shown on the right-hand side. Slip is coloured by slip velocity and contoured by slip increment during each phase of the earthquake cycle, where red lines show coseismic increments plotted every 3 s, green lines show postseismic increments plotted every 3 days, and blue lines show interseismic increments plotted every 5 years. (D-F) Maximum slip velocity during a full rupture of the seismogenic zone shown at two scales: the coseismic timescale is shown in seconds (linear scale), while the postseismic timescale is shown with a logarithmic timescale.

time until the next event that breaks the frontal section of the fault (Figure 5.9B). Deeper partial ruptures appear as sharp features in the velocity field, but their post-earthquake slip evolution produces only a feeble response in the surface deformation field. We observe large spatial gradients, in the velocity field, near the hinge zone early in the postseismic period following large earthquakes; this gradient diffuses away in time (Figure 5.9). Late in the interseismic phase, the velocity field appears nearly time-invariant, with a broad uplift peak (long wavelength) over the deeper velocity-weakening to velocity-strengthening transition region on the fault that decays towards the fault extremities (Figure 5.9). When the shallow velocity-strengthening section of the fault spans the length of the ramp, there is substantial steady interseismic uplift of the anticline (Figure 5.9B-C). However, in simulations where the velocity-weakening section extends up the ramp and close to the free surface, there is negligible uplift during the long interseismic period (Figure 5.9A).

5.4 Discussion

We have developed a numerical framework to investigate how hinge-related OFD, in the hanging wall of a fault-cored anticline, evolves with fault slip to grow a fold over timescales ranging from milliseconds to thousands of years. Using elasto-plastic models, we estimate the length-scales and strain rates of hinge-related inelastic deformation (Figure 5.6, 5.7A). We then use this geometry and a long-term loading rate to run earthquake sequence simulations (Figure 5.5), with the goal of predicting deformation that can be observed by seismology and geodesy during the various phases of the earthquake cycle (Figure 5.9). In this section, we use the results of our simulations to discuss why hinge-related OFD is only rarely observed in geodetic data despite its documented role in long-term geological deformation, and how we can modify existing inverse methods to be more sensitive to these processes. We also discuss the impact of hinge-related OFD on the growth of topography.

5.4.1 How well can we detect ongoing inelastic hinge zone deformation?

Elastic and inelastic (hinge-related OFD) fold growth has been observed at coseismic and postseismic timescales in fold-thrust belt settings (Figure 5.3; Ainscoe et al. [2017], Béon et al. [2017], Copley [2014], Copley and Reynolds [2014], Daout et al. [2021], Gold et al. [2019], Kuo et al. [2014], Zhou et al. [2016, 2018]). Flexural slip processes have also been identified related to both dynamic triggering due to passing seismic waves [Kaneko et al., 2015], and steady interseismic deformation [Daout et al., 2019, 2021, Le Béon et al., 2019, Mackenzie et al., 2018, Mariniere et al., 2020]. However, such observations are infrequent, which is surprising given how ubiquitous fault bends are in sub-surface imaging of fold-thrust belts [Dahlen, 1984, Hubbard, 2015]. And yet, many geodetic observations from thrust settings can be fit satisfactorily by slip on a fault, most often planar, with a purely elastic response of the crust (e.g., Ingleby et al. [2020], Moreno et al. [2009], Page et al. [2011]).

FIGURE 5.9: Timeseries of vertical velocity at the surface over a 1000-year period for each of the simulations in Figure 5.8A-C. Vertical red lines indicate timing and the horizontal extent of coseismic ruptures.

From our modeling, we suggest that a significant part of this discrepancy arises due to the existence of an (approximate) equivalent elastic solution to most fold growth surface deformation fields (Figure 5.10). In other words, the sparsity and noise in the surface deformation data as well as the non-unique solutions in the numerical tools we employ hide the inelastic response of the hanging wall, and therefore allow a purely elastic hanging wall response to fault slip to fit the data sufficiently well. We expound on this in the subsequent paragraphs.

5.4.1.1 High density spatial sampling

In order to discriminate between the elastic and inelastic response of the crust to slip on a ramp-décollement fault based on surface deformation, two basic criteria must be satisfied. First, the surface measurements must be spatially dense enough to capture the displacement gradient near the back-limb of the fold (Figure 5.10A). Second, these measurements must have a high enough signal-to-noise ratio to successfully discern between the elastic response of the crust to slip on an underlying fault and the combined deformation of the crust due to fault slip and hinge zone deformation (Figure 5.10B).

Point measurements by many geodetic methods (GNSS, levelling, geodetic surveying), while precise, are unable to capture the spatial scale of hinge-related OFD. Without sufficient observation locations in the hanging wall, the displacement fields due to a purely elastic crust, a narrow hinge, and a wide hinge zone will all appear identical [Allmendinger et al., 2009]. This problem can be overcome by using remote sensing methods such as InSAR or cross-correlation of optical images [Avouac and Leprince, 2015, Elliott et al., 2016, Simons and Rosen, 2007]. These methods provide estimates of surface deformation with high spatial density sampling but are limited to observations over some arbitrary time period in the earthquake cycle, which can be a problem as fault slip is likely non-uniform in space and time.

5.4.1.2 Accounting for slip gradients and fault geometry

To identify significant inelastic hinge deformation, when we have sufficiently dense spatio-temporal observations, fitting the surface observations using a purely elastic crust (no hinge zone) must produce spatio-temporally coherent residuals. In other words, it should not be possible to explain the observed surface displacements by either – (1) spatial gradients in fault slip or (2) a modified fault geometry.

Encountering surface displacement fields due to spatial gradients in slip are inevitable. There are no geodetic datasets with sufficient spatio-temporal coverage that can provide us with displacement fields arising from uniform fault slip, a process that would require observations spanning an entire earthquake cycle (100-1000 years). However, the issue arising from alternate fault geometries can be addressed if we use structural information from geomorphic, geological and seismic imaging studies to construct and fix fault geometries (e.g., Avouac [2015b], Chen et al. [2007], Le Bon et al. [2014],

Yue et al. [2005]). We note that geologically consistent deformation fields may differ from the predictions from our simulations, due to topographic effects and surface processes (Figure 5.10D).

Combining such an apriori fault geometry model with surface displacement observations from some time period in the earthquake cycle would allow us to assess if there was any inelastic OFD. The key here is that hinge-related OFD would produce large surface displacement gradients in the back limb of the fold, a feature that would generate coherent residuals when the data is fitted with a purely elastic crust model and thereby allow us identify inelastic crustal deformation.

In the following section we expand on how we can detect hinge-related OFD from geodetic observations that do not span an entire seismic cycle but are limited to a narrow snapshot in time.

FIGURE 5.10: (A) Surface displacement field for growth of a fold over a complete earthquake cycle. The different colored lines show the results of models with different hinge widths. (B) Decomposition of the displacement field into the elastic response of the medium to slip on the ramp-décollement, and due to inelastic strain in the hinge zone. For sufficiently narrow hinges, the surface deformation signal is obvious. However, for a wide hinge, an equivalent elastic solution exists that masks the effect of hinge-related OFD. (C) Underlying geometry of the fault-bend fold. (D) Alternate fold growth models and associated surface uplift. The gray rectangle indicates the uplift pattern expected from the surface erosion model shown in (C).

5.4.2 Surface observations of OFD modulated by the earthquake cycle

High spatio-temporal resolution geophysical observations only go back a couple decades, which is insufficient to observe the full earthquake cycle. Thus, in order to detect hinge-related OFD, we need to not only have appropriate observations, but also a geological setting with the right conditions (Figure 5.10), and the right time interval. For example, consider the 1999 Chi-Chi earthquake: the static displacement field derived from remote sensing [Kuo et al., 2014] shows clear elastic and inelastic hanging wall deformation (Figure 5.3A-B). However, without information about the temporal evolution of strain, we cannot distinguish OFD that occurred coseismically from OFD that occurred in the hours and days after the event.

In this section, we discuss how our numerical simulations of earthquake sequence deformation can guide the use of multiple observational methods to detect hanging wall deformation processes. To visualize these effects, we plot the surface displacement field during different parts of the earthquake cycle, along with the potency (moment) released by the hinge for three representative cases (Figure 5.11; earthquake sequence simulations in Figure 5.8).

5.4.2.1 Coseismic and early postseismic

Not all earthquakes drive hinge-related OFD in the same way. For example, partial ruptures generally do not drive any hinge-related OFD, whereas earthquakes that rupture the entire fault cause significant folding and produce an observable displacement gradient near the hinge both coseismically and postseismically (Figure 5.8-5.9). This is because earthquakes that are small or far from the hinge, transfer negligible stress to the bedding plane faults that constitute the hinge. Since fault slip rates are controlled by their stress state, due to velocity-dependent frictional strength [Dieterich, 1979, Scholz, 1998], the hinge remains unperturbed by partial ruptures. For large earthquakes that can transfer sufficient stress to the hinge zone (on the order of MPa), the optimal timescale at which displacement gradients associated with OFD are observable ranges from immediately after the earthquake to months later (Figure 5.11A-C). The coseismic period, however, is unsuitable for OFD observations, because the deformation due to on-fault slip is orders of magnitude greater than hinge-related OFD (Figure 5.11).

In light of our results, the coseismic folding example from the 1999 Chi-Chi earthquake we highlighted earlier (Figure 5.3A-B), may in fact be the effect of rapid early postseismic deformation (Figure 5.11A). Armed with only a snapshot of deformation computed from optical images obtained days before and after the earthquake, we can only set bounds on how fast hinge-related OFD may have occurred; OFD may have occurred at a rate between meters/day to meters/second. Detecting the evolution of such rapid hinge deformation during the minutes and hours following the earthquake may require the development of new techniques. While waveform data from seismometers and high rate GNSS are sensitive to the coseismic rupture process, high rate GNSS observations decimated

at the minute to hourly scale contain additional information about slower processes that deform the medium, but do not release seismically detectable radiation (Figure 5.11). Thus, both types of data contain the rupture process, but they differ in their sensitivity to slower deformation. Joint kinematic inversions of earthquake source parameters typically pool waveforms and geodetic observations to obtain average parameters [Fukuda and Johnson, 2010, Funning et al., 2014, Yabuki and Matsu'ura, 1992], which diminishes our ability to estimate early afterslip. Instead, we suggest partial pooling of these datasets, which is a popular machine learning method to take advantage of datasets that contain similar information on average but differ in statistically coherent ways [Gelman and Hill, 2006]. Alternatively, it may also be possible to detect hinge deformation by explicitly inverting for rapid early afterslip in addition to coseismic slip [Milliner et al., 2020, Ragon et al., 2019].

5.4.2.2 Postseismic

For cases where hinge-related OFD is only detectable at the week to month timescale, GNSS observations at the daily scale and InSAR may help constrain the rates and styles of OFD (Figure 5.11C). This is the case for décollement ruptures (Figure 5.11C; slip evolution is shown in Figure 5.8C). Both, daily GNSS and InSAR are sensitive to the same process and can be pooled to reveal the temporal variation of deformation associated with the fault and the hinge. At timescales longer than years and decades, the postseismic deformation field decays to a background low amplitude interseismic velocity (Figure 5.9, 5.11).

From a theoretical perspective, these observational timescales vary inversely with the long-term loading rate ($\dot{s}^\infty$) and the shear modulus of the medium (G), which we chose as 10^{-9} m/s and 10 GPa in the simulations. The relaxation timescale of afterslip for velocity-strengthening faults can be estimated from spring-slider analysis as $t_R = \frac{(a-b)\bar{\sigma}}{\dot{\tau}^\infty}$ [Perfettini and Avouac, 2004]; $\dot{\tau}^\infty$ is the long-term loading rate, and has a linear dependence on $\dot{s}^\infty$ and G (Equation 5.2, 5.7). This means that the timescales shown in Figure 11 need to be adjusted based on the effective loading of the fault in question.

Take for example the case of the Kurit fold (Figure 5.3C-D), where we observe temporally decaying afterslip and hinge deformation forty years after the last known earthquake [Zhou et al., 2016, 2018]. While this may appear surprising compared to typical detectable afterslip (1-5 years, Helmstetter and Shaw [2009], Ingleby and Wright [2017]), it can be explained by the fact that the long-term slip rate on the frontal faults in the Tabas fold-belt is only 1-2 mm/yr ($< 10^{-10}$ m/s), and the surrounding materials have low rigidity [Walker et al., 2015, Zhou et al., 2018]. These properties result in an effective loading rate that is an order of magnitude lower than our simulations, and therefore the afterslip signal remains discernable over decades rather than the years that we predicted in our simulations.

FIGURE 5.11: Ease of detection of hinge-related OFD signals during different phases of the earthquake cycle for three representative cases (A-C) highlighted in Figure 5.7-5.8. Left panel shows incremental surface displacements over a number of timespans: immediately after the earthquake (black), 1-day afterslip with coseismic slip removed (gray), differential postseismic deformation in 1 month (purple) since the earthquake (from 1 day to 1 month after the event) and the 50 year interseismic deformation from 10 to 60 years after the earthquake (red). Right panels show % moment or potency rate released by the hinge (blue) and the absolute potency rate (green) versus time. The percentage potency is calculated by considering the total potency released by 10 km of the fault, from the free surface down, and the hinge region. Also highlighted are the typical timescales of observation of different seismo-geodetic methods.

5.4.2.3 Interseismic

During the interseismic period, detecting slow fold growth requires a shallow ramp that is able to creep (in order to generate displacement gradients). In most fold-thrust belts, the steady-state creep rates on ramps are limited by the stress shadow cast by down-dip fault locking [Almeida et al., 2018, Herman et al., 2018], which significantly reduces the stress accumulation rate updip, on both primary faults and within a hinge zone. While there are some notable local examples of steady creep-driven fold growth [Daout et al., 2019, Le Béon et al., 2019, Mackenzie et al., 2018, Marinier et al., 2020], frontal folds in large fold-thrust belts do not usually show substantial interseismic creep (e.g., in the Himalayas and the Indo-Burman fold-belts, Grandin et al. [2012], Lindsey et al. [2018], Mallick et al. [2019, 2020]). Since steady creep rates are typically slow, detecting OFD also requires a long time-window to accrue detectable displacements (Figure 5.11).

Even when interseismic OFD is present, it can be hard to detect. This is because the long wavelength signals related to fault locking and deep creep dominate the surface deformation signal during the interseismic period (Figure 5.9). If the interseismic creep signal is time-invariant, it mostly reflects the background loading process; as a result, such a signal is unlikely to provide any information about the evolution of the stress state or material properties. In general, detailed inferences about creeping or frictionally locked regions during the interseismic period are difficult due to the combined effects of the stress shadow from frictional locking on the fault and the limit of resolution relative to the noise floor [Almeida et al., 2018, Herman et al., 2018].

5.4.3 Building geologic structure and topography

Slip over a fault bend necessarily builds topography in the long term. This is apparent even for a purely elastic medium responding to slip over a ramp-décollement system (displacement profile in Figure 5.1B). In this case, deep interseismic creep (representing far-field horizontal shortening) causes long-wavelength uplift of the medium (Figure 5.9), while slip on the ramp-décollement causes uplift of the hanging wall above the ramp, with subsidence both downdip of the ramp and in the footwall close to the fault (Figure 5.10B). In addition to large-scale uplift of the hanging wall, OFD in the hinge zone can permanently deform the hanging wall over a narrow length-scale, producing a fold scarp (Figure 5.10).

Based on our modeling, a significant portion of topographic growth due to hinge-related OFD is coseismic (Figure 5.11: the hinge potency rate is highest during earthquakes). However, the work associated with this process is orders of magnitude smaller than concurrent fault slip, and hence it is difficult to disentangle this signal from a purely elastic crustal response to fault slip using observational techniques. Permanent deformation near the hinge only becomes detectable as on-fault processes slow down to match the rate of moment release by hinge-related OFD, but these OFD processes and related topographic growth occur continuously throughout the earthquake cycle. Our

simulations show that more than 50% of the total potency associated with hinge zone OFD occurs within the first month after the earthquake (Figure 5.11).

Long wavelength interseismic uplift has been detected in a number of geodetically monitored convergent margins, peaking over the coastline or shelf break (in subduction zones), or where topography is high (collisional boundaries). As a result, studies have interpreted that mountain building may be a relatively time-invariant process largely occurring during the interseismic period [Grandin et al., 2012, Jolivet et al., 2020, Malatesta et al., 2020, Meade, 2010, Saillard et al., 2017]. Based on our results, we suggest a more complicated model of mountain building. Different physical mechanisms operate over the different spatial zones that constitute a convergent margin, with accretion and fold growth within the fold-thrust belt, following brittle mechanics [Dahlen, 1984, 1990], and friction to flow-like processes beyond the backstop of the fold belt and above the deepest parts of the megathrust [Agard et al., 2018]. Topography building due to fault slip and hinge-related OFD represents how mountains are built in the brittle regime: one-fold at a time. It may be true that coastal uplift and the growth of high topography does occur interseismically, but further study is needed to understand how friction and flow processes operate together on the deep megathrust to build topography at the larger scale [Menant et al., 2019, Van Dinther et al., 2013].

5.4.4 Including hinge-related OFD in numerical simulations and inverse models

The past few years have seen a push towards incorporating geological and geodynamic considerations into earthquake simulations and seismic hazard models (e.g., Dal Zilio et al. [2019], Sathiakumar et al. [2020], van Zelst et al. [2019]). However, resolving the complexity of earthquakes, which evolve on the millisecond timescale, with a fold-thrust belt growing over million years is a daunting task [van Zelst et al., 2019]. The major difficulties that arise are: resolving the complexity of frictional instabilities with an appropriate spatiotemporal grid, and accounting for pervasive inelastic deformation within the bulk medium.

In this study, we separate the two timescales and use a one-way coupling from the geological timescale to the earthquake sequence timescale (Figure 5.5), thereby accounting for both timescales while preserving kinematic and mechanical consistency. We show that permanent deformation of the hanging wall occurs in a narrow band, with length-scales that are controlled by the angularity of the fault-bend and the flexibility/bending resistance of the hanging wall (Figure 5.7). Our computational cost is only mildly greater than for typical earthquake sequence simulations, since we need only explicitly define regions of inelastic strain, allowing the rest of the bulk medium to remain a linear elastic solid with no meshing. The off-fault response effectively becomes a time-varying loading rate term for the on-fault stress evolution in Equation 5.7. Variations in loading rates have been noted to affect earthquake nucleation sites and sizes in recent earthquake sequence simulations for fold-thrust belt

settings [Sathiakumar et al., 2020]. It remains to be investigated if including time-varying loading due to hinge related OFD significantly impacts the statistical behavior of earthquake sequences and the properties of aseismic phenomena, such as scaling relationships between stress drop and magnitude, rupture velocities and the radiation efficiency of fast seismic events, and the duration and propagation velocity of afterslip and other postseismic processes.

For kinematic inverse problems, hinge-related OFD can be approximated by borrowing concepts from plate motion kinematics, i.e. closing the velocity vector diagram of a triple junction formed by a ramp, décollement and hinge, collapsed onto a fault plane (e.g., Daout et al. [2016], Sathiakumar et al. [2020], Souter and Hager [1997]). These studies take advantage of the ambiguity in direction of shear (inelastic strain) associated with shear bands. In other words, the displacement and stress fields generated by backthrust deformation are nearly identical to those produced by slip on a series of bedding plane contacts in a hinge zone (see focal mechanisms in Figure 5.4C). This approximation is a solution for inverse problems aimed at inferring slip/strain from surface displacements, but creates a number of issues in the stress formulation needed to simulate earthquake sequences: (1) An infinitesimally narrow fault bend, which is a pre-requisite for the backthrust approximation, creates an impossible physical situation in a purely elastic medium – interpenetration of the medium around the fault junction [Romanet, 2020]; (2) as a result, the on-fault traction kernel (Equation 5.2) has a singularity at the fault bend that dramatically alters the tractions on the fault; and (3) it is difficult to choose an appropriate constitutive relation for the backthrust, considering that it does not exist. As a result, we recommend using this approximation for inverse problems but not for earthquake sequence simulations.

5.5 Conclusions

Fault bends create unbalanced elastic stresses in the medium that grow with slip on the fault. These stresses are relaxed by permanent or inelastic off-fault deformation (OFD) of the hanging wall. In the layered sedimentary rocks typically observed in fold-thrust belts, this inelastic OFD takes the form of a hinge zone, a process described by fault-bend folding theory. Although evidence for inelastic hinge-related OFD is ubiquitous in geologic and geomorphological data, it is difficult to observe the OFD process in seismo-geodetic datasets.

In this article, we present a unified modeling strategy for earthquake sequence simulations and long-term fold growth in the brittle crust that accounts for the mechanical coupling between fault slip and inelastic strain in the hanging wall. We show that it is difficult to identify inelastic behaviour of the crust in seismological and geodetic data, largely because of the existence of equivalent elastic solutions to surface deformation data. This problem and the associated non-uniqueness issue arise from (1) poorly known fault geometries and elastic properties of the crust, and (2) insufficient spatial and temporal sampling of earthquake-cycle modulated surface deformation data.

Hinge-related OFD and fault slip are sources of permanent crustal deformation, leading to the growth of topography. Topography and geologic structure, at least at the scale of individual anticlines, are built throughout the earthquake cycle. The rate of this process is strongly modulated by the rate of on-fault moment release, which means hinge-related OFD is fastest during and immediately after large earthquakes and the rate decays logarithmically in time. However, concurrent fault slip generates a significantly larger signal in observational datasets, thereby obfuscating the inelastic deformation near the hinge. Observing and isolating geodetic signals of hinge-related OFD is possible when the on-fault moment release decreases to values comparable to the off-fault moment release; this is most likely in the postseismic period. We highlight this for the two examples we have considered in this study – (1) rapid hinge-related OFD, at velocities possibly exceeding meters/day, following the 1991 M_w 7.6 Chi-Chi earthquake in Taiwan, captured by optical imagery acquired days before and after the event; (2) slow and long-lived hinge-related OFD, at rates of millimeters/year, following the 1978 M_w 7.3 Tabas-e-Golshan earthquake in Iran, imaged by InSAR over decades (Figure 5.3). Therefore, to infer the mechanics of how earthquake-cycle stresses fold the brittle crust, we suggest combining structural geological methods of inferring fault geometry with seismo-geodetic observations from the postseismic period.

5.6 Acknowledgements

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Chapter 6

Concluding remarks

I conclude this thesis by noting that the depth of our understanding of geological systems follows a three-tier hierarchy - *geometry*, *kinematics* and *dynamics*. In the case of active tectonics and faulting, our understanding of *geometry* comes from answers to - where are the faults? what are their strikes, dips and length-scales? With geometries established, we now try to infer the *kinematics* of the system - fault slip rates, their temporal evolution and the deformation history of various rock parcels. Inferring the geometry and kinematics of faults and the surrounding rocks is typically in the realm of observationalists and has resulted in tremendous discoveries over the past 60 years, ever since plate tectonics became an accepted theory. But now, when we demand predictive power from tectonic models, only relying on kinematics is insufficient. Elevating our understanding to the next tier necessitates the incorporation of what we expect from the forces and stresses in the system. However, inferring *dynamics* is an extremely challenging task!

This difficulty arises in part because the governing equations necessary to describe the dynamical state of the system are computationally challenging to solve, and partly because of the non-uniqueness in estimated parameters in the geosciences, derived from observations, makes it near impossible to develop a statistical basis for choosing one model over another. I do not provide a panacea, but instead have shown, in the last four chapters, how we can combine observations with simplified dynamical models to reduce this non-uniqueness as we try to achieve an improved understanding of (1) stress build up and the various failure modes in the earthquake cycle, (2) how earthquake cycle processes build mountains over the long-term.

In Chapter 2, I used a combination of geodetic data and geological arguments to infer regional plate motion of the India-Burma-Sunda plate system. This is a chronically understudied part of the world, where oblique subduction appears to be ongoing under the world's thickest sedimentary blanket. I simplified the mechanics of this system into elastic blocks with forced localization along their boundaries to estimate where elastic stress is being built up, and where earthquakes may eventually occur. Previous studies that disputed the hazard posed by the plate boundary faults in this region were either

mismodelling the data, or did not have access to the new velocities we published in Chapter 2. In this chapter, I highlighted that steady interseismic geodetic velocities are most sensitive to transitions from locked to creeping on plate boundary faults. Additionally, I also discuss how to use geologically informed constraint equations in block models to improve the kinematic descriptions of the major faults here. This gives us some estimate of regions accumulating strain, but contain no information about how this strain will be accommodated (as earthquakes, SSE, or other non spontaneous failure styles).

To learn how stuck sections of a fault release strain, we need to find observations of various failure processes, as well as develop physics-based numerical simulations to ascertain what is and what is not possible. By studying multi-decadal timeseries, using the coral record from Simeulue (Indonesia) in Chapter 3, I found that long-lived SSE can go on for decades. Previous studies that employed typical kinematic methods to study this data understood that a long-lived change had occurred in the data, but assumed it was simply a different steady-state process compared to the one preceding it. Combining a mechanical model of interseismic behaviour with an inverse (parameter estimation) routine, I showed that the coral data had recorded a 32-year failure mode on the shallow brittle section of the Sunda megathrust. This section of fault has been inferred to be velocity-strengthening or frictionally stable by other studies. Considering how models of SSE and earthquakes suggest that spontaneous instability nucleation occurs on velocity-weakening or frictionally unstable regions, our observation from Simeulue requires a new paradigm. Not only does it imply that earthquakes and SSE are not governed by the same physics, but it also indicates a need for further classification among SSE themselves. When it comes to earthquakes, the nucleation of the instability comes from the velocity-weakening behaviour of the fault, but the majority of the fault weakening and associated rapid slip is controlled by fluid-driven processes [Rice, 2006, Viesca and Garagash, 2015]. It appears that fluids also play an important role in the dynamics of SSE [Bürgmann, 2018], but how they affect fault stability during SSE is unclear. Given that SSE seem to nucleate and occur on velocity-weakening as well as velocity-strengthening faults, and the role of fluids appear contradictory among various models, it is unclear if there can even exist a universal mechanism that explains all SSE. Currently an SSE refers to any kinematically detected spontaneous creep event that does not reach seismic slip rates; a future goal may be to derive a clearer definition with a more focused scope.

Failure modes, like earthquakes and SSE, are collectively just how plate boundaries accommodate inter-plate motion; these events are powered by the elastic stress build-up preceding them, something that we can estimate geodetically. Over the long-term, only the inelastic effect of these events are recorded in the geology and landforms exposed at the surface. This means that short-term and long-term observations record different processes, and reconciling them requires developing a model of how geodetically-derived plate motion generates geologic structure. In Chapter 4, I compared geodetically-derived shortening rates with geological uplift rates for the Shillong Plateau in India, using a kinematic model based on available structural and seismological information. I note a disagreement in the result between the two methods, and show that this does not arise from methodological errors, but rather reflects an ongoing missing component of deformation in the geological uplift

rates - the subsidence of the footwall. I interpret this result as the first observation of passive margin subduction initiation in the Cenozoic, noting that the rise of the Shillong Plateau is possibly driven by sinking of the old, cold and fractured crust to its south. This process may have been going on since the Miocene and is evidence for a stage of the Wilson Cycle that is commonly assumed to exist, but never documented before.

In the final section of my thesis, Chapter 5, I combine the insight I gained from observations and models of the earthquake cycle with the kinematics of crustal folding to develop a mechanical model coupling fault slip and distributed bulk deformation for fold-thrust belt settings. The kinematics of folding were described almost 40 years ago, but by adding dynamics to the system, we can start to understand how faulting and folding interplay with one another. One question that has bothered those of us working with geodesy and geology is that even though folding and bulk deformation are seen in most geological outcrops as well as sub-surface seismic images of anticlines, why do we not find compelling evidence for this in geodetic data? The simple answer is (1) the lack of high spatial resolution geodetic data until recently, and (2) the difficulties in telling apart the elastic deformation of the crust from the combined elastic and inelastic deformation of the crust. In Chapter 5, I discuss the observation conditions necessary to make convincing interpretations of crustal folding and note that the postseismic period following large earthquakes may be the most appropriate period when inelastic bulk deformation can be distinguished from purely elastic behaviour. Future studies can consider using this study as a guide for how to collect data and what kinds of signals to look out for following large earthquakes, and then even try to estimate the rheology of the shallow crust, something that we have almost no constraints on currently.

This collection of studies illuminate that we need to be thinking about faults across scales to improve our understanding beyond geometry and kinematics. The complexity of fault mechanics combined with the non-uniqueness and epistemic uncertainties that arise from inferring sub-surface processes from surface data means that addressing some of the grand challenges in geoscience i.e., improved hazard estimation and understanding how topography is built, can only be done with a combined and concerted interdisciplinary effort. Constitutive laws derived by laboratory studies need to be implemented in physics-based simulations and the predictions tested against seismological, geodetic and geological data. If there is a suite of models (and model parameters) that can explain the observations, then we need to delve deeper into the physics to predict what kind of observations we would need to further discriminate between models. A societally useful goal of this model of faulting is of course to have significantly better predictive power for hazard than we currently do. We live here, we would like to worry a little less.

Appendix A

Supplementary materials for the India-Burma-Sunda plate kinematics study

Here we describe our inversion methodology used to infer fault geometries for the Rakhine-Bangladesh megathrust, Churachandpur-Mao Fault (CMF) and Sagaing Fault which are used in the block models. Section [A.1](#) is a description of the methodology for how we infer construct the fault geometry while Section [A.2.1](#) and [A.2.2](#) highlights some of the caveats in this exercise. In Section [A.3](#) we describe a synthetic experiment to demonstrate that the inference of shallow creep on the CMF, as has been noted by previous researchers, is not a robust result. In Section [A.4](#) we review the fundamentals of block modelling and describe how we use it to study the India-Burma-Sunda plate system.

A.1 Two-dimensional inversion

Interseismically the shallow part of the fault is stationary (coupled) while the deeper extension is freely creeping (uncoupled). We model the fault using a step change from coupled to uncoupled; therefore, we only need to solve for a deep dislocation that slips at the plate rate which represents the uncoupled part of the fault. For a two-dimensional approximation of a fault, there exist analytical expressions relating fault parameters for semi-infinite deep dislocations to observable surface velocity field (Equation [A.1](#) and [A.2](#)) [[Mansinha and E.Smyli, 1971](#), [Segall, 2010](#)]. We model the deep dislocation related to the megathrust as pure dip-slip motion which results in east-west convergence on the surface, while the CMF and Sagaing Fault are modelled as pure strike-slip faults that manifest

as a purely north-south surface velocity field (Figure A.1).

$$v_{ss}(x) = \frac{V_{ss}}{\pi} \tan^{-1} \left(\frac{x - x_s}{D} \right) \quad (\text{A.1})$$

$$v_{ds}(x) = \frac{V_{ds}}{\pi} \left(\cos \delta \tan^{-1} \left(\frac{x - x_d}{D} \right) + \frac{\sin \delta - \left(\frac{x - x_d}{D} \right) \cos \delta}{1 + \left(\frac{x - x_d}{D} \right)^2} \right) \quad (\text{A.2})$$

v_{ss} and v_{ds} represent the horizontal velocity field due to a strike slip deep dislocation, and dip slip deep dislocation, respectively, at a location x along the transect. V_{ss} , δ , D are the fault slip rate, fault dip and depth of dislocation respectively. x_s and x_d refer to the horizontal position of the strike-slip and dip-slip dislocation respectively. v_{ss} and v_{ds} are the data for our model, while V_{ss} , V_{ds} , D , x_0 and x_d are our model parameters. δ is fixed as 0° so that V_{ds} represents pure horizontal convergence, but even a 10° variation of δ does not alter our results noticeably. We exploit this property to connect the coupled-uncoupled transition that we infer from our models to the surface trace of a fault. Horizontal velocities are very sensitive to $x_d = \frac{D}{\tan \delta}$, but are relatively insensitive to the dip angle of the deep dislocation (for e.g, Figure A.6B).

We create two east-west transects: a southern transect (width between 22°N - 23.5°N : Transect A) and a northern transect (width between 23.5°N - 25°N : Transect B) (Figure A.2). This allows us to model the combined effects of along-strike and along-dip variation of the position of elastic strain accumulation. We approximate the faults as north-south structures, which allows us to decompose the surface velocity field as follows: east-west velocities as convergence velocities due to the megathrust and north-south velocities as associated with dextral shearing of CMF and the Sagaing Fault. We add a constant term, c , as a model parameter in the forward models to account for shifting the velocity field to a suitable reference frame (Equation A.3).

$$\mathbf{v}(x_i) = (v_{ds}(x_i) - c_{ew}) \hat{\mathbf{e}}_{EW} + (v_{ss}(x_i) - c_{ns}) \hat{\mathbf{e}}_{NS} \quad (\text{A.3})$$

$\mathbf{v}$ is the vector velocity at each station x_i . c_{ew} and c_{ns} are the constant offsets for the convergence velocities and shearing velocities respectively. $\hat{\mathbf{e}}_{EW}$ and $\hat{\mathbf{e}}_{NS}$ are the unit vectors in the east-west and north-south directions respectively.

The constant offset has an additional advantage in that it allows us to fit our data without being forced to explicitly pin our modeled velocity field to a pre-defined reference frame. We also ran models where we pinned the velocity field to an arbitrary point deep in the stable Indian plate but do not show them in this paper since the results do not vary significantly. This method forces the westernmost points in the transect to have zero velocity and assumes that the Indian reference frame is perfectly defined by the data. We do not think such a hard constraint is necessary since we are considering a transect that is almost 1000 km long, and block rotations are likely to create spatially variable plate rates.

FIGURE A.1: Two-dimensional model setup for inversion of the horizontal velocity field. (A) Eastward (red boxes), and northward velocity field (blue boxes) for an arbitrary station distribution in the India reference frame. The red and blue lines show theoretical velocity fields due to dislocations in the bottom panel (B). The distance and velocity scales are arbitrary, and only provide a visual guide to the method used in Section A.1. (B) Modelled fault geometry. We use a 0° dipping dip-slip semi-infinite dislocation (red line) to fit the eastward velocity field; we use two 90° strike-slip semi-infinite dislocations (blue lines) to fit the northward directed velocity field. The dashed lines indicate fully coupled sections which have no geodetic signal associated with them, and hence we are insensitive to their geometry. (C) The eastward surface velocity field is not very sensitive to the depth of the dislocation. The coloured lines (blue to green to yellow) show the predicted velocity field for a model in which we fixed the depth of the dislocation between 25 to 70 km.

Since we have non-linear dependencies between our observables and model parameters we use a Monte Carlo sampling method to find the best-fitting parameters and uncertainties for the forward models (Equations A.1, A.2 and A.3) that can explain our data. Our chosen sampling technique, Slice sampling [Neal, 2003], generates maximally uncorrelated samples by means of a sequential Gibbs sampling along each model parameter, thereby avoiding the long convergence time experienced with Simulated Annealing or Metropolis-Hastings samplers [Lindsey and Fialko, 2013, Neal, 2003]. We maximize the likelihood function to fit the data and evaluate the results in a Bayesian framework. For a given prior model distribution $p(\mathbf{m})$, where p is the probability distribution function for model $\mathbf{m}$, the likelihood that the data $\mathbf{d}$ is explained by this model is $p(\mathbf{d}|\mathbf{m})$. The posterior distribution $p(\mathbf{m}|\mathbf{d})$ is given by Bayes' Theorem in Equation ???. We choose uncorrelated Gaussian error statistics (Equation ??) for our data likelihood function and a combination of uniform and Gaussian distributions for the priors. We discuss the significance of these choices in Section A.2.1.

$$p(\mathbf{m}|\mathbf{d}) \propto p(\mathbf{m})p(\mathbf{d}|\mathbf{m}) \quad (\text{A.4})$$

$$p(\mathbf{m}|\mathbf{d}) \propto p(\mathbf{m}) \exp\left(-\frac{1}{2}(\mathbf{v}_m - \mathbf{d})W(\mathbf{v}_m - \mathbf{d})^T\right) \quad (\text{A.5})$$

$\mathbf{v}_m$ is the modeled velocity field, while W is the weighting function derived from the data covariance matrix. The weighting function W , is a normalized diagonal matrix constructed from the inverse of the data covariance matrix. To avoid overweighting GPS stations with extremely low reported measurement uncertainties (0.5 mm/yr or less), we imposed a minimum uncertainty of 0.5 mm/yr in the construction of the data covariance matrix. After calculating the posterior distribution of the model parameters (Figure A.3-A.4) we use the Jacobian J , which is the matrix of partial derivatives of the forward model with respect to model parameters, and the model covariance matrix C_m , constructed by using a best-fit multivariate Gaussian distribution to approximate the posterior distribution, to estimate the confidence intervals for the mean of the predicted velocity field as $T_{\alpha,d_f} \hat{\sigma} \sqrt{JC_m J^T}$. Here, T_{α,d_f} is the value of the student's T distribution calculated for $\alpha = 95\%$ level of confidence and d_f degrees of freedom. We use the reduced χ^2 error, $\hat{\sigma}^2$, as an estimator for the scalar uncertainty multiplied to every member of the unit model-covariance matrix.

The results of the inversion and the posterior distribution of the model parameters are shown in Figure Figure A.3-A.4. We see that the sources of strain, for strike-slip and dip-slip sources, are shifted significantly eastward in Transect B compared to Transect A. We take these results to imply that the coupling transition of the plate interface is arcuate.

We next fit the data using a Block model, which unlike the two-dimensional model requires explicit parametrization of the geometry of the locked fault, and solves for the back-slip rate on this finite fault. In an ideal case of a long and perfectly planar fault, the back-slip and semi-infinite methods would predict identical horizontal and vertical surface velocities (e.g. [Meade and Loveless, 2009]). In our case, the faults are of finite length and bend slightly at the dislocation edge, hence the need for a more fully three-dimensional model to obtain the most robust results.

To define the fault geometry for the block model, we take the inferred dislocation edge (parameters x_d or x_s , and D) for each dislocation and connect it to the assumed surface trace of the associated fault to create a shallow fault which now has a non-zero dip. For example, we use the inferred values of x_d , D for the Bangladesh megathrust to connect the down-dip coupling transition to the surface trace marked by the edge of growing anticlines in Bangladesh (Figure 2.1), and create a planar shallow fault (Figure 2.4). This 2D-to-3D approach is defensible because both methods are most sensitive to the down-dip coupling transition, which is the source of strong surface velocity gradients.

A.2 Points to note in the inversion strategy

A.2.1 Uncertainties in the depth of coupling transitions

Most two-dimensional approximations for relative plate motion assume that inter-block motion is accomplished by a simple translation. This would imply that the easternmost data points in both the transects, located within the Shan Plateau, are directly responsible in calculating the east-west block velocity of the Shan block with respect to India. However, it is well known that block-like motion on the surface of the Earth manifests as rigid-body rotations about Euler poles. Therefore, the total slip rate calculated on the two-dimensional dip-slip dislocations may be biased due to the neglect of this rotation. If we were to model areas where the width of the transect were small compared to the length of the fault and the Euler pole was sufficiently far away, the errors arising from neglecting rotational effects would be minimal. However, neither of these assumptions holds well for the arcuate fault geometry in the Indo-Burman Range.

There is a strong trade-off between the inferred dip-slip rate and the locking depth of the semi-infinite dislocation. To fit surface velocities that show long wavelength signals with a single dislocation, the locking depth is generally required to be very deep. This is a problem we encounter while modeling both the dip-slip dislocation possibly associated with the megathrust, as well as the strike-slip dislocation possibly associated with the CMF. When we use uniform priors on the locking depth, the posterior distribution displays very large variance and the best-fitting depths are greater than 40 km. This is unrealistically large for convergent margins and not expected given that the brittle-ductile transition is likely much shallower [Syracuse et al., 2010]. Therefore, we instead apply loosely constrained Gaussian priors 25 ± 10 km for the locking depth of both dislocations. This gives us slightly more realistic best-fitting locking depths, but we do not have confidence in this model estimate. Instead, we interpret this information as the locking depths for the dislocations in the IBW are at least 25 km deep, but we cannot constrain it unless we can first infer the correct fault geometry using other independent techniques.

FIGURE A.2: Results of the inversion for the horizontal velocity field using a two-dimensional approximation. (A) Mapview of data and model predictions with the inferred horizontal location of the coupling transition marked as gray rectangles for each transect (Section A.1). (B) Data and model predictions of the eastward and (C) northward directed velocities. The red and green lines show the predicted velocity field calculated from the maximum a posteriori estimate of model parameters (Section A.1) while the semi-transparent polygons indicate the 95% confidence intervals of the mean of the forward modelled velocities. (D) shows the residual velocities.

FIGURE A.3: 1-D and 2-D marginal aposteriori distributions for model parameters associated with pure dip-slip in the 2-D models in (A) Transect A in Figure A.2, and (B) Transect B in Figure A.2. The red line in the 1-D marginal distributions show the maximum aposteriori estimate.

V - plate rate, ζ - horizontal location of coupling transition, D - coupling transition depth, C - reference frame correction.

FIGURE A.4: 1-D and 2-D marginal aposteriori distributions for model parameters associated with pure strike-slip in the 2D models in (A) Transect A in Figure A.2, and (B) Transect B in Figure A.2. The red line in the marginal distributions show the maximum aposteriori estimate.

V_{CMF} - CMF plate rate, ζ_{CMF} - horizontal location of coupling transition for CMF, D_{CMF} - coupling transition depth for CMF, V_{SF} - Sagaing Fault plate rate, ζ_{SF} - horizontal location of coupling transition for Sagaing Fault, D_{SF} - coupling transition depth for Sagaing Fault, C - reference frame correction.

A.2.2 Effect of the Shan Plateau on slip rate estimates of the Sagaing Fault

The stations in the Shan Plateau also strongly influence the north-south velocities of inter-block motion. Bookshelf faulting and possibly asthenosphere-driven intra-block deformation in the Shan Plateau distorts the velocity field, causing deviations from rigid block-like motion. This makes robust estimation of the fault slip-rate of the Sagaing Fault difficult since these stations are meant to provide far-field constraints. Previous studies have used ad-hoc arguments such as removing a forward-modeled best-fitting velocity surface of India-Eurasia motion to correct the velocities at these stations [Maurin et al., 2010]. We use the velocities at these sites with and without the correction of Maurin et al. [2010]. Without the correction, the posterior distributions in our two-dimensional models are largely unconstrained and we observe very strong trade-offs between all the model parameters for the Sagaing Fault. In the results presented above, we apply the correction to estimate the model parameters only in the two-dimensional transects but use the uncorrected velocities in the three-dimensional model to test the effect of these stations on the robustness of the estimate of convergence across the IBW.

A.3 Mismodelled creep on the CMF

To test the biases that may arise from improper projection of data along transects for the CMF, we ran synthetic model inversions for the case of a curved dextral strike-slip fault that is fully coupled to a depth of 25 km. The fault is infinite along strike, dips at 90° and is divided into 3 segments (Figure A.5). Two segments have a strike of 0° while the central segment has a strike of $\sim 30^\circ$ similar to what is observed for the CMF. We generate a velocity field for this fault on a uniform grid, and select a specific subset of 30 stations close to the bend in the fault (Figure A.5B). We project this data onto a transect of east-west strike and use only the north-south velocities as a proxy for the fault-parallel velocity field (Figure A.5A). We then invert this velocity field for two scenarios using the two-dimensional approximation as in Section A.1.

1. We solve for a step change in coupling on the fault. We recover the full plate rate, and locking depth correctly in the inversion. However, we obtain an eastward step of 20 km in the inferred horizontal location for the coupled-uncoupled transition.
2. We solve for a step change in coupling on the fault as well the possibility of creep on the shallow section of the fault. In this model, we recover the full plate rate, and the locking depth and observe an eastward step of 20 km for the coupling transition just like the previous case. However, we also infer ~ 9 mm/yr of shallow creep that extends to a depth of ~ 5 km.

FIGURE A.5: Mismodelled creep due to improper projection of velocities near a bend in a strike-slip fault. (A) Velocities (in a fault-centred reference frame) are projected onto an east-west transect and inverted for two separate scenarios (Section A.3). The blue triangles show the velocities used in the inversion while the gray circles show the true scatter of the velocities. The red line is the predicted velocity field for a fault that is locked to 25 km, while the green line is the model of a fault that is locked to 25 km but displays shallow creep. (B) Map view of surface velocity field and the orientation of a strike-slip fault with a bend. The fault is infinite along strike. The gray arrows show the true horizontal velocity field, and the blue triangles show the location of the sampled stations.

This exercise demonstrates that the inference of shallow creep for a curved fault can be an artifact of improper projection of velocities across a transect, and does not indicate if the CMF is actively creeping at the surface.

A.4 Block modelling

Block modelling provides a convenient method to simultaneously solve for block rotations and strain accumulation on faults. The plate velocities for stations far from plate boundaries in a block are due to rigid body rotations about an Euler pole. Deviations from such block-like motions arise from strain accumulation on faults (block boundaries), or due to internal strain within the block itself. In this study we use the Blocks software [Meade and Loveless, 2009] to conduct our modelling.

The use of Euler poles to describe velocities on the surface of a sphere has some elegant advantages. It implicitly forces circuit closure for any plate system (i.e., the closed loop integral of block angular velocities in any arbitrary path sums to zero), which is what we expect for a system with no net rotation. It allows for spatially varying velocities (due to rotation) without invoking more model parameters such as active faults or treating transects of data independently. We chose the Indian Block as our “reference block” with respect to which we defined the velocity field, however this exercise is not dependent on the choice of data reference frame. We add in interseismic velocities from IGS reference stations HYDE, IISC and BAN2 to define the stable Indian Block [Steckler et al., 2016].

We create a model with four closed blocks: the Indian Block (INDIA), Indo-Burman Wedge (IBW), Myanmar Central Basin Block (MCB) and Shan Block (SHAN) (Figure 2.5). These blocks are bounded by the Rakhine-Bangladesh megathrust, CMF and Sagaing Fault. These faults are allowed to accumulate strain and distort the surface velocity field using the back-slip formulation [Okada, 1985, Savage, 1983]. The northern and southern boundaries are chosen so as to encapsulate stations in their associated blocks, and we set the locking depths of these segments to zero so that they do not affect the inversion, since there is no strain field associated with these segments [Meade and Loveless, 2009]. We also impose Gaussian a priori constraints on the slip rate of the Sagaing Fault to stabilise the inversion (20 ± 2 mm/yr right-lateral strike-slip), as this slip rate is well constrained [Maurin et al., 2010, Vigny et al., 2003, Wang et al., 2014] (Figure A.4); without these priors the best fitting model attempts to overfit the data in the Shan block and results in spurious slip rates on the Sagaing Fault. To deal with a similar block modelling problem in Sumatra, which also includes a gently dipping megathrust with a strike-slip fault in its hanging wall, [Bradley et al., 2017] used the slip vector azimuths of earthquakes on the Sunda megathrust to impose a rake constraint on the megathrust, and thus were able to resolve the elastic contributions of the megathrust and the Sumatran Fault. However, in our study area, the paucity of earthquakes does not permit us to impose such a constraint. Instead, we impose a priori constraints on the slip rate of the CMF; we penalize any dip-slip component on

the CMF, since we assume that the CMF is currently a right-lateral strike-slip fault [Gahalaut et al., 2013, Wang et al., 2014]. To test our bias, we also run experiments allowing the rake of the CMF to vary freely in the inversion.

We use the formulation and software of [Meade and Loveless, 2009], which allows us to obtain a best-fit series of Euler poles for plate pairs and the associated model covariance matrix through a set of normal equations when the fault geometry is predefined. We define the interseismic velocity $\bar{v}_{int}$ as a linear combination of the rotational field $\bar{v}_{rot}$, back-slip velocities $\bar{v}_{bs}$ and finite internal strain-rate field within a block $\bar{v}_{\dot{\epsilon}}$. Each of these fields can be written at a site i as linear operations on either an angular velocity vector or strain rate tensor. $\bar{v}_{rot}$ and $\bar{v}_{bs}$ are modeled as partial derivatives of rigid body rotation $\mathbf{R}$ and elastic dislocations $\mathbf{R}_f$, with respect to the angular velocities of each plate pair about its Euler pole. The angular velocity Ω , is a function of the latitude (θ), longitude (ϕ) and rotation rate (ω) of the Euler pole for the respective block pair, evaluated at the site location (θ^i, ϕ^i). $\bar{v}_{\dot{\epsilon}}$ can be written as $\mathbf{L}\dot{\epsilon}$ where $\mathbf{L}$ is a matrix that contains the distance of every observation point from a reference point (θ^0, ϕ^0) within its respective block, while $\dot{\epsilon}$ is the uniform strain-rate tensor $[\dot{\epsilon}_{\phi\phi} \ \dot{\epsilon}_{\phi\theta} \ \dot{\epsilon}_{\theta\theta}]$

$$\begin{aligned}\bar{v}_{rot}^i &= \Omega \times r^i = \mathbf{R}(\theta^i, \phi^i) \Omega(\theta, \phi, \omega) \\ \bar{v}_{bs}^i &= \mathbf{R}_f(\theta^i, \phi^i) \Omega(\theta, \phi, \omega) \\ \bar{v}_{\dot{\epsilon}}^i &= \mathbf{L}\dot{\epsilon}\end{aligned}\tag{A.6}$$

$$\begin{aligned}\bar{v}_{int}^i &= \bar{v}_{rot}^i - \bar{v}_{bs}^i + \bar{v}_{\dot{\epsilon}}^i \\ &= [(\mathbf{R} - \mathbf{R}_f) \quad \mathbf{L}] [\Omega \ \dot{\epsilon}]^T \\ &= \bar{\mathbf{R}} [\Omega \ \dot{\epsilon}]^T\end{aligned}\tag{A.7}$$

The problem setup in equation A.7 is overdetermined. The inversion for model parameters $[\Omega \ \dot{\epsilon}]^T$ is performed in the least-squares sense as $(\bar{\mathbf{R}}^T W \bar{\mathbf{R}})^{-1} \bar{\mathbf{R}}^T W d$, where $\bar{\mathbf{R}}$ is the design matrix, while W and d refer to the weighting function and the data vector, respectively. The weighting function is a normalized diagonal matrix constructed from the inverse of the data covariance matrix. In order to avoid overweighting GPS stations with extremely low reported measurement uncertainties (0.5 mm/yr or less), we imposed a minimum uncertainty of 0.5 mm/yr in the construction of the data covariance matrix. The covariance matrix of model parameters is given by $\hat{\sigma}^2 (\bar{\mathbf{R}}^T W \bar{\mathbf{R}})^{-1}$, where $\hat{\sigma}^2$ is the reduced χ^2 error.

Our preliminary model runs were using the geometry defined by the 2D models. To verify and further investigate the geometry of Bangladesh megathrust we conducted a gridsearch for the depth of the coupling transition, and fault dip for multiple cases. (1) A set of model runs where we penalized dip-slip on the CMF with varying weights, (2) allowed free-rake on the CMF and (3) penalized dip-slip on the CMF with varying weights but permitted a uniform internal strain-rate tensor within the

Indo-Burman Wedge. The robust result that emerged was the horizontal location of the down-dip coupling transition of the Bangladesh megathrust, which turned out to be significantly different from the 2D models.

FIGURE A.6: Comparison of block models with different parameterizations. (A)-(C) are results from a model in which we penalize dip-slip on the CMF but allow uniform internal strain-rate within the Indo-Burman Wedge (IBW). (D)-(E) are the results from a model run in which we do not impose any constraints on the CMF, nor do we permit internal deformation of the IBW. In all the models the transparent red patches show the geometry of the shallow fault we use in our models. (A) and (D) show the data and model fits using the model while the inset shows the residual velocity field. In (B) we display the results of the gridsearch used to estimate the best fitting parameters for the coupling transition of the Bangladesh megathrust in the case where we penalize dip-slip on the CMF, but permit internal deformation of the IBW. (C) and (E) show the estimated fault slip-rates and plate motion vectors while the inset shows the location of the euler-poles of each block relative to India.

FIGURE A.7: Comparison of block models with different weights of hyperparameters i.e. the weight of the apriori penalization of dip-slip on the CMF. (A)-(B) are results from a model in which we allow uniform internal strain-rate within the Indo-Burman Wedge (IBW) and use a weak dip-slip penalty on the CMF of 10mm/yr. (C)-(D) are the results from a model run in which we do not permit internal deformation of the IBW but use a dip-slip penalty weight of 4 mm/yr calculated from the trade-off curve in Figure 2.7. In all the models the transparent red patches show the geometry of the shallow fault we use in our models. (A) and (C) show the data and model fits using the model while the inset shows the residual velocity field.(B) and (D) show the estimated fault slip-rates and plate motion vectors while the inset shows the location of the euler-poles of each block relative to India.

FIGURE A.8: Estimated fault slip-rates and plate motion vectors from block model runs with fault geometry inferred from 2D models in Section A.1. In all models, the inset shows the locations of euler poles for each block pair. (A) Rigid block model with a dip-slip prior of 0 ± 1 mm/yr on the CME. (B) Rigid block model with an added N. Shan block with a dip-slip prior of 0 ± 1 mm/yr on the CME. The additional block removes any dip-slip component on the Sagaing Fault. (C) Rigid block model with no constraints on the CME. (D) Block model with internal deformation of the IBW and a dip-slip prior of 0 ± 1 mm/yr on the CME.

FIGURE A.9: Model statistics for models shown in Figure 2.6-2.7. (A) Results of the F-test (blue) is the probability that error reduction occurred due to random chance. We also show the reduced χ^2 of the residuals (gray bars). The dashed blue line is at the 5% confidence level i.e. only if the F-statistic is below 5%, we accept the error reduction as significant. The F-statistic is calculated against the optimum/preferred model ($\sigma_{CMF} = 4$ mm/yr). (B) We show the ABIC for our model runs. The optimum model is the one with the lowest ABIC value.

Appendix B

Detailed methods for long-lived SSE study

B.1 Coral timeseries analysis

The coral timeseries for sites LBJ, UTG, SLR and SMB are taken from published data [Meltzner et al., 2015]. The measurement uncertainty at each point used in the velocity extraction is 2 cm. This uncertainty estimate comes from an observation that the largest measurement bias between two corals from the same site was 8 cm [Meltzner et al., 2015]; we interpret that as the 99% confidence level extrema (3σ).

It is obvious that a simple linear least-squares fit cannot explain the timeseries, and so we explore 3 different hypotheses: (1) a quadratic least-squares fit, invoking a continuous acceleration, (2) piece-wise linear fits with the timing of the change in velocities (T_{change}) being common to all the sites, (3) piece-wise linear fits treating each site independently.

We then assess the error improvement using the F-test. For the piece-wise linear fits, we add an equality constraint such that the fitted function is continuous across T_{change} . We estimate the mean and covariance of the fitted parameters using simple matrix inversion since the addition of equality constraints to the linear inverse problem can be thought of as a typical Least Squares problem with additional Lagrangian multipliers, which means the posterior distribution is still Gaussian. For a given dataset d , design matrix G (a function of T_{change}) and weighting function W , we estimate the distribution of the model parameters $m \sim N(\hat{m}, C_m)$ subject to an equality constraint $Am = b$ as,

$$\hat{m} = (G^T W G)^{-1} \left[G^T d - A^T \left(A (G^T W G)^{-1} A^T \right)^{-1} \left(A (G^T W G)^{-1} G^T d - b \right) \right] \quad (\text{B.1})$$

$$C_m = \sigma^2 \left((G^T W G)^{-1} - (G^T W G)^{-1} A^T \left(A (G^T W G)^{-1} A^T \right)^{-1} A (G^T W G)^{-1} \right) \quad (\text{B.2})$$

The uncertainties in m associated with T_{change} are non-Gaussian but are an order of magnitude lower

FIGURE B.1: Testing multiple hypotheses to explain the trends in the coral timeseries. (A) Assuming a piecewise linear fit for the timeseries with a common timing for the velocity change, T_{change} , we evaluate the misfit (with Gaussian error statistics - thick blue line). (B) We evaluate the reduced χ^2 and the significance of error reduction using the F-test for the following hypotheses: piecewise linear fits with an abrupt velocity change at T_{change} (blue line - this is the preferred model), continuous acceleration (red line) and piecewise linear fits with an abrupt velocity change whose timing varies for each station (red diamonds).

than C_m , which is why we report only C_m . σ^2 is the least-squares misfit when we use $\hat{m}$ to fit the data.

The reduced- χ^2 from the different hypotheses is shown in Figure B.1B. From the F-test, we conclude that piece-wise linear fits for a common $T_{change} = 1829$ across all sites is the most likely hypothesis. We do not use the site SLR in subsequent analysis since the uncertainties in the velocities for the site are too large for the velocities to be meaningful.

B.2 Slow Slip Events (SSE) on faults governed by rate-state friction

B.2.1 Governing Equations

We can describe the variation of the friction coefficient on a thrust fault as a function of fault velocity v , and a state parameter θ which represents the average duration of asperities, using rate-and-state dependent friction and the ageing law [Dieterich, 1979, Ruina, 1983].

$$\mu = \mu_0 + a \log \left(\frac{v}{v_0} \right) + b \log \left(\frac{v_0 \theta}{L} \right) \quad (\text{B.3})$$

$$\frac{d\theta}{dt} = 1 - \frac{v\theta}{L} \quad (\text{B.4})$$

μ_0, v_0 are reference values of the friction coefficient and velocity respectively; a, b are material properties that describe the evolution of the friction coefficient to changes in velocity and its evolution in time (and slip, by extension) respectively, while L is the critical slip distance over which a population of frictional contacts evolve.

Consider the force balance on the fault, we equate the stresses arising from elastic interaction with friction and radiation damping as follows:

$$\dot{\tau}^\infty t + K\delta = \mu \bar{\sigma} + \frac{G}{2v_s} v \quad (\text{B.5})$$

Here $\dot{\tau}^\infty$ is the far field stressing rate on the fault (derived from loading the fault at a slip rate of V_{pl}), K is the elastic interaction kernel computed from the analytical solutions for stress due to a shear crack in an elastic medium (plane strain) (Segall et al. [2010]). The right hand side contains the frictional stress ($\bar{\sigma}$ is the normal stress) and the last term is the radiation damping approximation (Rice 1993, Rice and Tse 1986) (G is the shear modulus and v_s is the shear wave speed in the medium).

The time derivative of the force balance (Equation B.5) gives us the momentum balance. For simplification we use the following variable transformations:

$$\begin{aligned}\zeta &= \log \frac{v}{v_0} \\ \phi &= \log \frac{v_0 \theta}{L}\end{aligned}\tag{B.6}$$

Using these transformations in the time derivative of Equation B.5 with Equation B.3 and B.4, we obtain the governing equation (momentum balance) to generate earthquake sequences:

$$\begin{aligned}\dot{\tau}^\infty + K v_0 \exp \zeta &= \left(a \dot{\zeta} + b \dot{\phi} \right) \bar{\sigma} + \frac{G}{2v_s} (v_0 \exp \zeta) \dot{\zeta} + (\mu_0 + a \zeta + b \phi) \dot{\bar{\sigma}} \\ \dot{\bar{\sigma}} &\rightarrow 0 \text{ or kinematically imposed}\end{aligned}\tag{B.7}$$

We solve for unknown variables $\zeta, \phi, \bar{\sigma}$ by numerically integrating the above coupled ODEs (Equation B.7 and B.4) using a 4th order Runge-Kutta adaptive time-stepping method, subject to an arbitrary initial condition ($\zeta^0 = 10^{-3}, \phi^0 = 0, \bar{\sigma}^0 = 60 \text{MPa}$).

B.2.2 SSE on velocity-weakening faults

To generate SSE on velocity-weakening (VW implies the friction coefficient decreases with an increase in velocity, thereby localizing strain and promoting unstable fault slip) faults, we use an instability nucleation size comparable to the dimension of the the VW region (R). We assume the earthquake cycle is driven only by shear tractions and thus neglect any temporal variation in the fault normal stress ($\dot{\bar{\sigma}} = 0$).

Simple stability analysis of this system of equations gives us a nucleation size (Rice [1993]) for a VW frictional instability as:

$$h^* = \frac{GL}{(b-a)\bar{\sigma}}\tag{B.8}$$

For $\frac{R}{h^*} \sim 1.1$, we generate frictional instabilities which correspond to SSE (Figure B.2A) where with increasing slip, the rate of elastic energy release exceeds the frictional weakening thereby preventing the instability from running away to seismic slip rates.

A characteristic feature of these kinds of SSE are that they have a long nucleation phase, where fault slip rate accelerates over a time interval comparable to the interval over which the SSE shows almost linear displacement timeseries (Figure B.2B). This means that for an SSE to have an average slip rate of 1-1.5 V_{pl} over a period of 10-20 years, it must be preceded by an acceleration on the fault lasting 10-20 years. However this is not observed in coral timeseries, and so is an unsatisfactory explanation.

Instead, we observe an abrupt velocity change and so the underlying mechanism needs to produce a sudden change in velocity (at a timescale significantly shorter than the SSE duration itself) while still creeping at a rate comparable to the plate loading rate.

FIGURE B.2: Comparison between SSE on velocity-weakening (VW) and velocity-strengthening (VS) faults. (A) and (C) show the slip rate evolution during the pre-SSE and SSE periods for the VW and VS case respectively. (B) and (D) show the mean slip rate and cumulative fault slip for both cases. The VS fault shows a short transition time between being stationary to creeping at the plate rate or faster; this key feature matches the coral data.

B.2.3 SSE on velocity-strengthening faults

Frictional instabilities do not typically nucleate on velocity-strengthening (VS) faults because an increase in fault strength with increase in fault slip rates implies that instabilities are damped out.

However, unstable frictional behaviour on VS faults can occur due to two reasons: (1) A decrease in the effective normal stress on the fault, due to evolution of fluid pressures, weakens the fault and counters the velocity damping the VS material would have otherwise executed. (2) When a VS fault is perturbed away from its steady-state creep behaviour, by a past negative stress drop event, the evolution of the state-parameter (θ) can promote mildly unstable frictional behaviour.

We exploit both mechanisms and propose a hybrid mechanism where the nucleation of an instability initiates as the fault recovers from a previous negative stress drop event; this velocity increase is coupled with gouge compaction and the release of pore fluids onto the fault (Perfettini and Ampuero [2008]), which further weaken the frictional strength and cause the instability to last up to 3 decades.

We solve the governing equations for velocity, state and normal stress evolution detailed in Equations B.4, B.7 subject to an initial condition. We model the initial condition with an instantaneous pore-fluid expulsion event ($t^- \rightarrow t^+$), possibly caused by a previous great earthquake, which drained the fault and dropped $\bar{\sigma}$ by ~ 2 MPa (Figure 3.5, B.2C).

This condition is obtained by computing an effective change in ζ, ϕ from Equation B.5 for an instantaneous change in normal stress from $\bar{\sigma}^- \rightarrow \bar{\sigma}^+$.

$$\mu_0 (\bar{\sigma}^+ - \bar{\sigma}^-) + a (\bar{\sigma}^+ \zeta^+ - \bar{\sigma}^- \zeta^-) + b (\bar{\sigma}^+ \phi^+ - \bar{\sigma}^- \phi^-) = 0 \quad (\text{B.9})$$

Since this is an instantaneous event, we assume the change in elastic interactions is negligible in Equation B.5 implying the only change in the force balance is in the frictional strength (Equation B.9).

The instantaneous event also implies that the state variable has not evolved from $t^- \rightarrow t^+$ i.e. $\phi^- = \phi^+$. Simplifying this, we can write:

$$(\mu_0 + b\phi) (\bar{\sigma}^+ - \bar{\sigma}^-) + a (\bar{\sigma}^+ \zeta^+ - \bar{\sigma}^- \zeta^-) = 0 \quad (\text{B.10})$$

We provide an expression for ζ^+ which serves as the initial condition for the momentum balance.

$$\zeta^+ = \frac{\sigma^-}{\sigma^+} \zeta^- - \left(\frac{\mu_0 + b\phi}{a} \right) \left(1 - \frac{\sigma^-}{\sigma^+} \right) \quad (\text{B.11})$$

We impose Equation B.11 on Equation B.7 at $t = 50\text{yrs}$ (Figure B.2C) and numerically integrate it using a Runge-Kutta 4th order finite difference method which allows adaptive time stepping.

The instability initiates spontaneously around 200 yrs as the velocity on the fault tries to reach its steady state value (Figure B.2D, 3.6). In the absence of any additional weakening such an instability arrests itself and is limited to a slip rate indistinguishable from steady-state fault creep (Figure 3.6). We add slip rate driven pore-pressure rise as an additional source of weakening to reproduce the multi-decadal nature of the SSE (Figure B.2).

B.2.3.1 Pore-pressure evolution

Our exploration of pore-pressure driven fault weakening is motivated by the *fault-valve* model in [Sibson \[1990\]](#). The basic premise is that frictional sliding is a self-sealing process where slow fault shear promotes the formation of a low-permeability fluid barrier through precipitation/dissolution of insoluble/soluble phases or pressure solution creep that closes pore space (porosity) in the gouge [[Gratier et al., 2002](#)], thereby building overpressure and decreasing fault strength. At the earthquake cycle timescale, it is thought that on-fault fluid pressure is relieved during earthquakes due to cracking and fracturing of the fault gouge during fast slip; slow interseismic processes are thought to promote compaction and build pore-pressure [[Marone et al., 1990](#), [Sibson, 1990](#)].

Motivated by laboratory observations and such a conceptual model, typical velocity dependant porosity evolution models are derived to have a logarithmic functional form (porosity $\propto \log(v)$). This porosity evolution equation is then used in a fluid mass balance as an unsteady-state (membrane) diffusion problem to solve for the evolution of pore-fluid pressure [[Segall and Rice, 1995](#)]. This formulation creates a strong dependence of pore-pressure on velocity and the state variable (v, θ), but erases any influence of cumulative slip (or shear). However, experimental investigations of the variation of gouge porosity with shear is observed to be more complex, as gouges follow an initial shear-driven compaction phase which is then replaced by a shear rate-dependance [[Marone et al., 1990](#)]. Additionally the porosity evolution equations are calibrated for sliding experiments run at $\sim 10^{-6}$ m/s [[Marone et al., 1990](#)], which is orders of magnitude faster than the rates at which we observe the long-lived SSE at Simeulue.

With these considerations in mind, and being mindful of the fact that we know very little about the lithology and the strain history of the Sunda megathrust at the location where we infer the shallow SSE, as well as the exact form of pore-pressure evolution at very low slip rates, we chose to model the rate of pore-fluid pressure rise during the spontaneously-nucleated SSE period as a linear function in time with the normal stress at the end of the SSE restored to its pre-perturbation value (60 MPa in this case). This is conceptually similar to the numerical model of [Sleep and Blanpied \[1992\]](#), where they allowed porosity reduction due to gouge compaction during interseismic creep to drive earthquake cycles.

$$\begin{aligned} \dot{\sigma} &= 0 & t &\in [t^+, t_{SSE}] \\ \dot{\sigma} &= \frac{2(\bar{\sigma}^+ - \bar{\sigma}^-)}{\Delta t_{SSE}} \left(1 - \frac{t - t_{SSE}}{\Delta t_{SSE}} \right) & t &\in [t_{SSE}, t_{SSE} + \Delta t_{SSE}] \end{aligned} \quad (\text{B.12})$$

An SSE produced by this mechanism displays piece-wise linear displacement timeseries (Figure B.2D), and is preceded by an undetectable fault acceleration (Figure 3.6), since this occurs at values below the detectable threshold of surface data. Therefore, we consider this as a likely mechanism for shallow long-lived SSE since it can reproduce all the salient features of the coral timeseries and

agrees with the frictional properties inferred for this segment of the Sunda megathrust (abrupt velocity change, the magnitude of the velocity change is comparable to the plate velocity, the SSE lasts for 3 decades, and it occurs on the same part of the fault that showed significant afterslip following the 2005 M_w 8.6 Nias earthquake [Qiu et al., 2019]).

B.3 Inverse method to estimate SSE slip rates

We use physical constraints from Equation B.7 to develop our inverse method as follows. In the steady inter-seismic period, locked parts of the fault accumulate strain while creeping parts of the fault slip at time invariant slip rates determined by the net shear stress rate at that location (Equation B.7), with $\dot{\mu} = \dot{\sigma} = 0$. We can write this as:

$$\begin{aligned} \dot{\tau}^\infty + K_\tau v &= 0 & \text{steadily creeping at constant stress-rate} \\ \dot{\tau}^\infty + K_\tau v &> 0 & \text{locked} \end{aligned} \tag{B.13}$$

We segment the fault into 4 regions, using 3 free parameters $\zeta_{up}, \zeta_{down}, \zeta_{free}$ (Figure 3.4A). Region 1 is a creeping zone which extends from the free surface to ζ_{up} ; region 2 is a locked zone between ζ_{up} and ζ_{down} ; region 3 is a creeping zone between ζ_{down} and ζ_{free} ; region 4 creeps at the plate rate below ζ_{free} .

The data for the inverse problem is the vertical velocities we computed in Section B.1. Since we use a plane strain approximation and the stations do not exactly fall on a 2-d transect, we calculate weighted average velocities and uncertainties from the 3 sites to represent the vertical motion of Simulue Island in a representative cross-section (Figure 3.2).

The slip rate to surface velocity relations are taken from analytical solutions for edge dislocations in plane strain for elastic media [Segall, 2010] and a geometric setup applicable to the Sunda megathrust [Bradley et al., 2017, Hayes et al., 2018, Kanda, 2010]. The misfit between the data and model-predicted velocities is assumed to follow an unbiased multivariate Gaussian distribution with unknown variance. We sample the posterior distribution of the free parameters using a Markov-Chain-Monte-Carlo method (slice sampling) [Neal, 2003]. We use this method to estimate the steady-state slip rates on the Sunda megathrust for 1738-1829 as well as 1829-1861 (Figure 3.2, 3.3).

For the case where we assume a transient slip event occurs as a quasi-steady process, we parameterize a continuously connected section of fault to have a constant slip rate (within its domain, as well as over the time period 1829-1861). This velocity is superimposed on the jointly estimated 1738-1829 slip rate distribution from the steady-state inversion. We use slice sampling to sample the up-dip and down-dip extent of the transient as well as its slip rate $(\zeta_{up}^{trans}, \zeta_{down}^{trans}, V_{trans})$ in addition to the steady-state locking/creeping parameters $(\zeta_{up}^{lock}, \zeta_{down}^{lock}, \zeta_{free}^{lock})$ for the 1738-1829 period (Figure B.3).

FIGURE B.3: 1-d and 2-d marginal PDFs of the spatial parameters (ζ_{up} , ζ_{down} , ζ_{free} in Figure 3.4A) describing (A) frictional domains for a model where we assume the 1738-1829 velocity field is attributed to steady frictional behaviour on the Sunda megathrust followed by (B) the occurrence of a long-lived transient slip event (SSE) from 1829-1861. The SSE is estimated to occur between ζ_{up}^{trans} and ζ_{bot}^{trans} with an average slip rate of V_{trans} (mm/yr) (we normalize this by the plate rate V_{pl}). (See Methods for a more detailed description of all parameters). (C) We show the joint distribution of SSE slip rate and the along-dip location where this slip occurred. The darker colours show a higher value of the PDF; the red line in the 1-d marginal PDFs is the maximum a posteriori estimate.

Appendix C

Supplementary materials for the Shillong Plateau study

We describe a Bayesian approach to combining observed station velocities with a priori constraints on fault slip rates in a block modeling framework in Section C.1. In Section C.2, we use a visco-elastic model of the lower crust/upper mantle to show that the post-seismic effects of the 1897 $M \sim 8$ Shillong earthquake is negligible in the modern velocity field. In the third section of this document C.3, we discuss the convergence obliquity of India-Shillong Block and the Indo-Burman Wedge-Shillong Block along the Dauki Thrust, and possible scenarios in which this obliquity is accommodated. In the final section C.4, we present various models of rock uplift in response to an imposed horizontal contraction, assuming deformation is limited to the upper plate (hanging wall).

C.1 Bayesian Inference in Block Modeling

We use 6 micro-plates/blocks to model the GPS velocities: the Shillong Block (consisting of the Shillong Plateau, Mikir Hills, Brahmaputra Basin and Assam Valley), Indo-Burman Wedge, Myanmar Central Basin, Northern Shan Block, Shan Block and Indian plate/block (Figure 4.2). We assumed that some of these boundaries (Dauki Thrust, Rakhine-Bangladesh megathrust, Churachandpur-Mao Fault and Sagaing Fault) are locked and accumulating strain at the full plate rate to maintain continuity of the velocity field. Each block can be described using its 3-component Euler vector $[\omega_x, \omega_y, \omega_z]$; these parameters are linearly related to the GPS data for a given fault geometry (δ , dip; D , locking depth). This parametrization of a fault as a binary locked-creeping response allows us to simplify the inverse problem; it is a justifiable assumption because the interseismic velocity field is most sensitive to the location of this locked-creeping transition (e.g., Lindsey et al. [2014], Mallick et al. [2019]).

$$\mathbf{d}_{GPS} = G(\delta, D)\boldsymbol{\omega} \quad (\text{C.1})$$

The fault geometry of all blocks and block boundaries except the Shillong Block and the Dauki Thrust are taken from [Mallick et al. \[2019\]](#). In this paper δ and D exclusively refer to the dip and locking depth of the Dauki Thrust as the model parameters we estimate.

In addition to GPS data, we also use a priori information on fault slip rates as a set of Gaussian constraint equations with unknown weights $\mathbf{d}_{SR} \sim N(\boldsymbol{\mu}_{SR}, \boldsymbol{\sigma}_{SR}^2)$. We assume that the Churachandpur-Mao Fault is a dominantly strike-slip fault ($\mu_{CMF}^{\text{dip-slip}} = 0$) [[Mallick et al., 2019](#)], while the Dauki Thrust is predominantly a thrust fault ($\mu_{Dauki}^{\text{strike-slip}} = 0$). Since this set of constraints is linearly related to the block Euler vectors, they can be combined with the observations to give a forward model as follows:

$$\mathbf{d} = \begin{bmatrix} \mathbf{d}_{GPS} \\ \mathbf{d}_{SR} \end{bmatrix} = G(\delta, D)\boldsymbol{\omega} + \begin{bmatrix} \boldsymbol{\epsilon}_{GPS} \\ \boldsymbol{\epsilon}_{SR} \end{bmatrix} \quad (\text{C.2})$$

where $\boldsymbol{\epsilon}$ is a vector of uncertainties associated with $\mathbf{d}$, drawn from a zero-mean Gaussian distribution. If we assume that $\mathbf{d}_{GPS}$ and $\mathbf{d}_{SR}$ are uncorrelated sets, we can combine their uncertainties into a data covariance matrix:

$$\sigma_{GPS}^2 C_d = \begin{bmatrix} \sigma_{GPS}^2 C_d^{GPS} & 0 \\ 0 & \gamma^2 C_d^{SR}(\sigma_{CMF}^2, \sigma_{Dauki}^2) \end{bmatrix} \quad (\text{C.3})$$

where $\gamma^2 = \frac{\sigma_{SR}^2}{\sigma_{GPS}^2}$ is the relative weighting between the a priori slip rates and the observed data. Since γ^2 depends exclusively on σ_{GPS}^2 , σ_{CMF}^2 and σ_{Dauki}^2 , we simply write the covariance matrix as:

$$C_d = \begin{bmatrix} C_d^{GPS} & 0 \\ 0 & C_d^{SR}(\sigma_{CMF}^2, \sigma_{Dauki}^2) \end{bmatrix} \quad (\text{C.4})$$

The covariance matrix of the slip rate priors is a function of the unknown σ_{SR}^2 , which are hyperparameters we estimate in this study. This is commonly referred to as a *hierarchical model* in machine learning literature [[Fukuda and Johnson, 2010](#), [Malinverno and Briggs, 2004](#)].

This gives us a set of linear ($\boldsymbol{\omega}$) and non-linear parameters ($\delta, D, \sigma_{CMF}^2, \sigma_{Dauki}^2$) that we want to estimate from the data. We invoke Bayes' Theorem and write this out as a problem in Bayesian parameter estimation.

$$p(\boldsymbol{\omega}, \delta, D, \sigma_{CMF}^2, \sigma_{Dauki}^2 | \mathbf{d}_{GPS}) = \frac{p(\mathbf{d}_{GPS} | \boldsymbol{\omega}, \delta, D, \sigma_{CMF}^2, \sigma_{Dauki}^2) p(\boldsymbol{\omega}, \delta, D, \sigma_{CMF}^2, \sigma_{Dauki}^2)}{p(\mathbf{d}_{GPS})} \quad (\text{C.5})$$

$p(\boldsymbol{\omega}, \delta, D, \sigma_{CMF}^2, \sigma_{Dauki}^2 | \mathbf{d}_{GPS})$ is the posterior probability density function (PDF) of the model parameters given the observed data. $p(\mathbf{d}_{GPS} | \boldsymbol{\omega}, \delta, D, \sigma_{CMF}^2, \sigma_{Dauki}^2)$ is the data likelihood function, while

$p(\boldsymbol{\omega}, \delta, D, \sigma_{CMF}^2, \sigma_{Dauki}^2)$ is the prior distribution. We assume uniform priors for $\boldsymbol{\omega}$; uniform positive priors for $\sigma_{CMF}^2, \sigma_{Dauki}^2$; and uniform priors in the interval $[0, 90]^\circ$ and $[0, 50]$ km for δ, D respectively.

To speed up computation, we take advantage of the fact that we have linear and non-linear parameters; PDFs of linear parameters for a Gaussian data likelihood function (first term on the right in Equation C.5) have analytical closed form descriptions because they are Gaussian. We split the PDF of the model parameters into linear and non-linear model PDFs as follows,

$$p(\boldsymbol{\omega}, \delta, D, \sigma_{CMF}^2, \sigma_{Dauki}^2 | \mathbf{d}_{GPS}) = p(\boldsymbol{\omega} | \delta, D, \sigma_{CMF}^2, \sigma_{Dauki}^2, \mathbf{d}_{GPS}) p(\delta, D, \sigma_{CMF}^2, \sigma_{Dauki}^2) \quad (C.6)$$

The posterior PDF of the linear parameters given the non-linear model parameters is what we evaluate in a standard block modelling problem.

$$p(\boldsymbol{\omega} | \delta, D, \sigma_{CMF}^2, \sigma_{Dauki}^2, \mathbf{d}_{GPS}) \propto p(\mathbf{d}_{GPS} | \boldsymbol{\omega}, \delta, D, \sigma_{CMF}^2, \sigma_{Dauki}^2) p(\boldsymbol{\omega} | \delta, D, \sigma_{CMF}^2, \sigma_{Dauki}^2) \quad (C.7)$$

For a given set of values of $\delta, D, \sigma_{CMF}^2, \sigma_{Dauki}^2$, uniform priors on $\boldsymbol{\omega}$ and Gaussian error statistics for the data likelihood function, this is akin to a least-squares minimization to estimate $\boldsymbol{\omega}^*$:

$$\begin{aligned} s(\boldsymbol{\omega} | \delta, D, \sigma_{CMF}^2, \sigma_{Dauki}^2, \mathbf{d}) &= (\mathbf{d} - G\boldsymbol{\omega})^T C_d^{-1}(\sigma_{CMF}^2, \sigma_{Dauki}^2) (\mathbf{d} - G\boldsymbol{\omega}) \\ \boldsymbol{\omega}^* &= (G^T C_d^{-1} G)^{-1} G^T C_d^{-1} \mathbf{d} \\ C_{\boldsymbol{\omega}^*} &= (G^T C_d^{-1} G)^{-1} \end{aligned} \quad (C.8)$$

Thus, for any given value of the non-linear model parameters, we can fully describe the PDF of $\boldsymbol{\omega}$.

In this framework, we have neglected the effect of the distribution of σ_{GPS}^2 , as we are only interested in the relative weights of the other hyperparameters (relative to σ_{GPS}^2). We can estimate $\hat{\sigma}_{GPS}^2$ using an analytical solution by maximizing the joint posterior PDF with respect to σ_{GPS}^2 [Yabuki and Matsu'ura, 1992]:

$$\frac{\partial p(\sigma_{GPS}^2, \boldsymbol{\omega}, \delta, D, \sigma_{CMF}^2, \sigma_{Dauki}^2 | \mathbf{d}_{GPS})}{\partial \sigma_{GPS}^2} = 0 \quad (C.9)$$

$$\hat{\sigma}_{GPS}^2 = \frac{s(\boldsymbol{\omega}^*)}{N_{GPS} + N_{SR}} \quad (C.10)$$

This allows us to evaluate the joint PDF of the other model parameters $(\delta, D, \sigma_{CMF}^2, \sigma_{Dauki}^2)$ given $\hat{\sigma}_{GPS}^2$, similar to the philosophy of the Akaike Bayesian Information Criterion [Akaike, 1987, Fukuda and Johnson, 2010, Fukuda et al., 2008].

FIGURE C.1: 1-d and 2-d marginal posterior probability distributions of non-linear model parameters used in our regional block model. The diagonal plots are the marginal distributions of each parameter while off diagonal plots show the joint distribution of pairs of parameters. δ, D are the dip, depth of the Dauki Thrust, while $\sigma_{ds}^{CMF}, \sigma_{ss}^{Dauki}$ are the inverse weights of the hyperparameters applied on the Churachandpur-Mao Fault and Dauki Thrust respectively. We note that δ, D are an approximation of fault geometry. If the Dauki Thrust demonstrates a ramp-flat geometry, δ is the minimum dip of the ramp. The distribution of hyperparameters can be interpreted as follows: the distribution of σ_{ss}^{Dauki} corresponds to a weak strike-slip penalization on the Dauki Thrust which implies that the variation in convergence obliquity along the Dauki Thrust is necessary to explain the data (see Section C.3); the distribution of σ_{ds}^{CMF} corresponds to a relatively strong penalization of dip-slip on the CMF resulting in 2-10 mm/yr of purely strike-slip motion between the Indo-Burman Wedge and Myanmar Central Basin.

$$\begin{aligned} \log p(\delta, D, \sigma_{CMF}^2, \sigma_{Dauki}^2 | \mathbf{d}_{GPS}, \hat{\sigma}_{GPS}^2) = \\ - ((N_{GPS} + N_{SR}) \log s(\boldsymbol{\omega}^*) + \log |G^T C_d^{-1} G| + \log |C_d| + \kappa) \end{aligned} \quad (C.11)$$

Here, $|A|$ is the absolute value of the product of eigenvalues of a matrix A ; and N_{GPS}, N_{SR} are the number of observation equations and a priori slip rate constraint equations, respectively. κ is a constant that depends on the data distribution and the number of free model parameters. We sample the joint posterior PDF of the non-linear model parameters using a MATLAB based slicesampler (10^5 samples), which is a Markov Chain Monte Carlo (MCMC) method that generates maximally uncorrelated samples by means of sequential Gibbs sampling along each model parameter [Neal, 2003].

This gives us the discrete posterior PDF of the non-linear model parameters, $\delta, D, \sigma_{CMF}^2$ and σ_{Dauki}^2 , which we show in Figure C.1. We compute the posterior marginal PDF of the linear parameters from MCMC sampling of Equation C.11 and the Gaussian approximation in Equation C.8.

$$p(\boldsymbol{\omega}|\mathbf{d}_{GPS}) \propto \sum_i (p(\boldsymbol{\omega}^*|\delta_i, D_i, \sigma_{CMFi}^2, \sigma_{Dauki_i}^2, \mathbf{d}_{GPS})) \quad (C.12)$$

We use this PDF of Euler poles, $\boldsymbol{\omega}$, to compute the PDF of slip rate of the Dauki Thrust using linear propagation of Gaussian uncertainties (Figure 4 of the main text) which is 0.5-1.0 mm/yr different from the slip rate PDF produced by running the Blocks software over the entire suite of posterior samples. This is because the fault dip-slip rates calculated in the Blocks software from the distribution of non-linear parameters $(\delta, D, \sigma_{CMF}^2, \sigma_{Dauki}^2)$ is the back-slip velocity [Savage, 1983] and needs to be multiplied by the corresponding $\cos \delta$ to recover the dip-slip rate derived from horizontal forcing. We show these corrected results in Figure C.2.

FIGURE C.2: Posterior distribution of Dauki Thrust slip rates. (A) Geographical location of each fault segment coloured according to the colour of the PDF shown in (B) and (C). (B) PDF of strike-slip rate of each fault segment in mm/yr; (C) PDF of dip-slip rate of each fault segment in mm/yr. The dip-slip rate increases from the west (black) to centre (brown-green) and then tapers to the east (yellow-blue).

C.2 Post-seismic modeling of the 1897 $M \sim 8$ Shillong earthquake

We model 120 years of deformation following the 1897 $M \sim 8$ Great Shillong earthquake using two scenarios of coseismic stress perturbation.

- The Dauki Thrust slipped in the event (Figure C.3)
- The Oldham/Brahmaputra Fault slipped coseismically (Figure C.4)

We compute the response of the lower crust and upper mantle from 1897 to present using uniformly gridded shear zones (with Maxwell rheology: $\mu = 30$ GPa and the following viscosity structure)

$$\eta = \begin{cases} 10^{18} \text{ Pa s} & : N \geq -50 \text{ km} \quad \forall E, Z \\ 10^{20} \text{ Pa s} & : N < -50 \text{ km} \quad \forall E, Z \end{cases}$$

Our gridded domain is 500 km (E) $\times$ 300 km (N) $\times$ 50 km ($Z \geq 40$ km) encompassing both the Dauki Thrust and Oldham Fault, while the displacement and velocity field is computed only at the surface ($Z = 0$) of a uniform, elastic half-space.

FIGURE C.3: Postseismic surface velocities calculated 120 years after the rupture of a M_w 8.1 event on the north-dipping Dauki Thrust. The surface velocity field (shown in blue) is calculated on a regular grid. The Dauki Thrust is shown as a red line, while the Oldham and Brahmaputra Faults are shown as red dashed lines. The gridded shear zones are shown as boxes with their centers marked as black plus marks (+). The colour within the shear zones shows the relative magnitude of the second invariant of the inelastic strain (yellow: low; brown: high).

We compute the elastic stress interactions (K) and surface displacement kernels for the shear zones using closed form expressions (analytic solutions from [Barbot et al., 2017]), while the stresses from

fault slip are computed using Okada [1992]. We use the MATLAB-based ode45 solver to compute the time evolution of deviatoric stresses, $\boldsymbol{\tau}$, and inelastic strains, $\boldsymbol{\epsilon}^i$, in the domain for the initial value problem (at $t = 0$, the earthquake provides an elastic stress perturbation ($\Delta\boldsymbol{\tau}_{cs}$)).

$$\left. \begin{aligned} \frac{d\boldsymbol{\epsilon}^i}{dt} &= \frac{\boldsymbol{\tau}}{\eta} \\ \frac{d\boldsymbol{\tau}}{dt} &= K \frac{d\boldsymbol{\epsilon}^i}{dt} \end{aligned} \right\} t > 0$$

$$\left. \begin{aligned} \boldsymbol{\epsilon}^i &= 0 \\ \boldsymbol{\tau} &= \Delta\boldsymbol{\tau}_{cs} \end{aligned} \right\} t = 0$$
(C.13)

120 years after the earthquake, it appears that the shear zones have fully relaxed the stress imparted by the 1897 event. The velocity field due to lingering post-seismic effects is under 1 mm/yr, which is at the noise threshold of most station velocities, and thus not detectable (Figure C.3,C.4).

In the case of homogeneous shear zone viscosities, the surface velocity field is even smaller. We use this laterally variable viscosity structure to show that even if post-seismic inelastic strain is focused beneath the Shillong Plateau, the surface velocity field 120 years later is an order of magnitude smaller than modern steady plate motion and thus does not bias our results.

FIGURE C.4: Same as Figure C.3, except co-seismic slip on a modelled south-dipping fault which represents either the Brahmaputra Fault or Oldham Fault. The magnitude of the velocity field does not change based on the choice of either fault.

C.3 Convergence Obliquity along the Dauki Thrust

The data do not allow us to treat the Dauki Thrust as a pure thrust fault. Our models predict a sharp change from mostly thrusting west of the triple junction (India - Indio-Burman Wedge – Shillong Block), to oblique thrusting (dextral strike-slip) to the east. This represents a T-T-T style stable triple

FIGURE C.5: Schematics of inter-block motion at the India-Shillong Block-IBW triple junction. (A) The relative motion between India-Shillong Plateau and IBW-Shillong Plateau is localized onto the Dauki Thrust and as a result the fault slips in an oblique sense east of the triple junction. Plate circuit closure is shown as a triangle. The triple junction is marked; it is stable and will migrate westward with the propagation of the IBW through time. The black lines in the IBW are traces of anticlines [Wang et al., 2014]. (B) We consider localized thrusting along the Dauki Thrust; however, all strike-slip motion is transferred onto the IBW and this is distributed over some width, similar to viscous flow past a rigid boundary [Turcotte and Schubert, 2002]. As a result, anticlines in the IBW, close to the Shillong Plateau, are deflected compared to further south.

junction configuration with the triple junction moving progressively westward (Figure C.5A). We note that such a change in obliquity along the Dauki Thrust does not predict stretching or extension within the Shillong Plateau, nor does it predict the rake of the Dauki Thrust; it suggests instead that the westward motion of the Indo-Burman Wedge relative to India leads to dextral shearing at the boundary between the Indo-Burman Wedge and the Shillong Plateau. If we assume this shear has localized onto the block boundary, the eastern segment of the Dauki Thrust is predicted to be an oblique thrust fault (Figure C.2B). However, in the absence of any geological or geomorphic evidence for strike slip motion at the southern edge of the plateau, we consider non-rigid deformation of the Indo-Burman Wedge as a likely scenario.

This alternate explanation comes from observing the deflection of the anticlinal structures in the Indo-Burman Wedge near the Dauki Thrust (Figure C.5B). In a broad sense, the Shillong Plateau acts a strong rigid block that hinders the westward transport of thrust sheets (relative to India) in the Indo-Burman Wedge [Maurin and Rangin, 2009, Wang et al., 2014]. This could result in internal

deformation of the wedge sediments close to the eastern section of the Dauki Thrust. The velocity field arising from this distributed deformation would be limited in its spatial extent and could be modeled explicitly as a separate, internally deforming block in our framework (Figure C.5B), given a denser GPS network. In this study, we only consider this possibility schematically as a model that can explain the overall block-like motion of India, the Shillong Block and the Indo-Burman Wedge with predominantly thrust motion on the Dauki Thrust (Figure C.5).

C.4 Kinematic Models of Plateau Uplift due to Horizontal Shortening

Here we consider 3 models of rock uplift (and ramp advection i.e shortening $\times \sin(\text{dip})$) driven by horizontal contraction, and apply it to the Shillong Plateau. We compare the predicted uplift rates from each model with the observed long-term uplift rates and ascertain the statistical significance of their overlap (Figure C.6). This allows us to discriminate between possible orogenic styles that built the Shillong Plateau since the Miocene.

C.4.1 Pure Shear Thickening

We assume a 2-d rectangular crustal block of dimensions $W \times T$ (width, height) deforms in pure shear when a horizontal shortening s is applied, in the absence of gravitational effects (Figure C.6A). The rheology of this crustal parcel is unimportant as long as it remains incompressible. Here, and in subsequent calculations, we use $W \sim U[50, 100]$ km and $T = 35$ km based on the details of the crustal structure presented in Mitra et al. [2018].

We use conservation of mass to estimate the topographic growth, as follows:

$$WT = (W - s)(T + (dT + r)) \quad (\text{C.14})$$

dT, r are the topographic growth and root respectively. We assume that $dT = r$ for pure shear. This gives us,

$$dT = \frac{T}{2\left(\frac{W}{s} - 1\right)} \quad (\text{C.15})$$

Assuming time invariant processes gives us (the dot represents time derivatives),

$$d\dot{T} = \frac{T}{2\left(\frac{W}{s} - 1\right)} \quad (\text{C.16})$$

FIGURE C.6: Geometric setup (A)-(C) for each uplift model considered. (D) Comparison of PDFs of long-term geological uplift (gray and light brown) with modelled uplift rates assuming $T = 35\text{km}$, $W \sim U[50, 100]$ km. Slip rate information is taken from the Monte Carlo samples generated in the Bayesian inverse problem. Only the PDFs for pure shear (red) and isostatic pure shear (light green) overlap with the geological uplift PDFs at the 95 % confidence level. However these models predict the existence of a crustal root beneath the Shillong Plateau; this is not seen in geodetic [Nayak et al., 2008] and seismological investigations of crustal structure [Mitra et al., 2018, Priestley et al., 2019]

C.4.2 Isostatically Compensated Pure Shear Thickening

Here we consider a 2-d rectangular crustal block of dimensions $W \times T$ deforms in pure shear, while also maintaining isostasy. The governing equation for pure shear is Equation C.14; the difference is the relation between dT and r is computed by maintaining isostatic equilibrium (Figure C.6B).

For a continent of density ρ_c floating on top of a mantle of density ρ_m , isostatic equilibrium at an arbitrary compensation depth gives,

$$\rho_c(dT + r) = \rho_m r \quad (\text{C.17})$$

Simplifying this, we get

$$r = \frac{\rho_c}{\rho_m - \rho_c} dT \quad (\text{C.18})$$

Combining the two equations, we obtain an expression for dT in terms of the geometry, densities and shortening. Assuming temporal invariance, we obtain the uplift rate,

$$dT = \left(1 - \frac{\rho_c}{\rho_m}\right) \frac{T}{\frac{W}{s} - 1} \quad (\text{C.19})$$

C.4.3 Plateau Pop-Up

In this section we consider two methods of accommodating horizontal shortening by plateau pop-up. The first method is a geometric consideration, where the plateau is treated as a rigid body that is translated vertically upwards along symmetric faults (dip = δ) in response to horizontal shortening (Figure C.6C). The uplift rate is,

$$dT = \frac{s}{2} \tan \delta \quad (\text{C.20})$$

An alternate method involves using mass balance to account for the conversion of shortening to uplift (Figure C.6C). The dimensions of the plateau in the initial configuration are W_t, W_b (width of top and base respectively) $\times T$ (thickness). Due to crustal shortening, this plateau is deformed such that its surface gains an elevation of dT while its base mirrors this uplift pattern. Mass conservation is written as follows,

$$sT = \frac{W'_b + W_b}{2} dT + \frac{W'_t + W_t}{2} dT \quad (\text{C.21})$$

Assuming the initial and final configurations have not changed significantly, $W'_b \approx W_b$ and $W'_t \approx W_t$.

$$sT \approx (W_b + W_t) dT \quad (\text{C.22})$$

Assuming time invariance, the uplift rate is

$$dT = \frac{T}{W_b + W_t} s \quad (\text{C.23})$$

Here we assume $W_t = 100$ km, while we allow $W_b \sim U[0, 50]$ based on the crustal structure presented in Mitra et al. [2018].

FIGURE C.7: Results of a 2-d Bayesian inversion to fit the southward velocity field (in Indian reference frame) for stations in India, Shillong Block and Bhutan Block. We show that the velocity is fit by a continuous velocity field which implies creep on the Bhutan Main Frontal Thrust is unlikely (see Section 3.1 of the main text). (A) Map view of the data used (only data within the black box is considered for the inversion). We use a modeling method similar to that described in Section C.1 (Equations C.1, C.8), except we treat plate motion as a horizontal translation similar to [Marechal et al. \[2016\]](#). A full description of the method is available in [Mallick et al. \[2019\]](#). (B,C) Transects corresponding to data shown in (A). We sample the posterior distribution of model parameters ($\zeta_d^{\text{Dauki}}, D_{\text{Dauki}}, V_{\text{Dauki}}, \zeta_d^{\text{MFT}}, D_{\text{MFT}}, V_{\text{MFT}}, c$) using slice sampling [\[Neal, 2003\]](#) and plot the horizontal positions of the ζ_d (in grey) for the Dauki Thrust and Main Frontal Thrust (MFT) in (B) and (C). Since the dataset has a large number of outliers, we choose Laplacian error statistics for our data likelihood function in Equation C.5 [\[Tarantola, 2005\]](#). We also show the distribution of the predicted southward velocity field (orange) in (B) and (C). The darker colours indicate a relatively greater probability of a parameter value while the lighter colours are from the tail end of the distribution. ζ_d - Horizontal location of the locked-creeping transition on a fault; D - Locking depth of the fault; V - Slip rate; c - reference frame constant.

FIGURE C.8: Results from our best fitting elastic block model, (A) and (B), (discussed in the main text) compared with a block model in which we allow internal strain within the Shillong Block, (C) and (D), to approximate velocity perturbations from other structures such as the Oldham Fault/ Brahmaputra Fault/ Kopili Fault etc. The data and model predictions are shown in (A) and (C), with the inset showing the distribution of residuals. The individual components of the model predictions, block rotations, elastic effects of fault locking and velocities due to uniform internal strain, are shown in (B) and (D).

The misfit reduction in the internal strain block model (compared to the elastic block model) is insignificant, and so we do not think we are missing any signal with the use of a simpler elastic block model. Additionally, the estimate of convergence across the Shillong Plateau in both models are indistinguishable; this implies that 5-8 mm/yr is a robust estimate of the convergence rate.

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