Inferring upper mantle rheology from numerical modeling of experimental and geodetic observations

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**INFERRING UPPER MANTLE RHEOLOGY
FROM NUMERICAL MODELING OF
EXPERIMENTAL AND GEODETIC
OBSERVATIONS**

SAGAR MASUTI

Asian School of Environment

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

2020

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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Authorship Attribution Statement

*(B) This thesis contains material from two paper(s) published in the following peer-reviewed journal(s) and third one to be submitted soon / from papers accepted at conferences in which I am listed as an author.

Chapter 2 is published as Masuti, S., Barbot, S., Karato, S., Feng, L., & Banarjee, P. (2016). Upper mantle water stratification inferred from observations of the 2012 Indian Ocean earthquake. *Nature*, 538, 373-377.

The contributions of the co-authors are as follows:

- Asst. Prof. Sylvain Barbot provided the initial project direction.
- I performed all the numerical experiments using Komodo computing cluster and wrote the manuscript with the help from Asst. Prof. Sylvain Barbot and Prof. Shun Karato.
- Dr. Lujia Feng processed the raw GPS data and prepared the time series and Dr. Paramesh Banerjee develops SuGAR network.
- All the co-authors contributed in finalizing the manuscript.

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The contributions of the co-authors are as follows:

- I and Prof. Shun Karato discussed and came up with problem statement.
- Dr. Jennifer Girard provided high-pressure data.
- I conducted the numerical calculations and wrote the manuscript with help from Prof. Shun Karato. All the co-authors contributed in finalizing the manuscript and its revisions

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Dedication

To my parents, my wife, and my chachu.

Abstract

As a major mantle mineral with volume fraction $>60\%$, olivine is assumed to control the rheology of the upper mantle. While the deformation of olivine and olivine bearing rocks has been extensively researched over the past half century, not all aspects of olivine rheology are well understood. In this thesis, I investigate the rheology of olivine using two types of data, viz., geodetic data following the 2012 M_w 8.6 Indian Ocean earthquake and data from experiments on olivine under various thermodynamic conditions. I determine the flow law parameters for the soft- and hard-slip systems, and propose a constitutive relation for the rheology of olivine which consistently captures the transient and steady-state creeps.

Using the proposed rheology of olivine, I explain the geodetic observations following the Indian Ocean earthquake with the wet dislocation creep, indicating a minimum water content of ~ 0.01 wt% or $1,600 \text{ H}/10^6 \text{ Si}$ in the asthenosphere. However, the earthquake broke the entire oceanic lithosphere down to great depths, indicating dry olivine in the lithosphere. I explain the steep rheological contrast by dehydration across the lithosphere-asthenosphere boundary, presumably by buoyant melt migration while forming the oceanic crust.

Also, using the same proposed rheology, I determine the transient creep flow law parameters from mechanical data of experiments performed under anhydrous conditions on natural dunites. Using a Markov Chain Monte Carlo method, I find different properties for the transient and steady-state components of dunites, with activation energies of $E_K = 430 \pm 20 \text{ kJ/mol}$ and $E_M = 535 \pm 15 \text{ kJ/mol}$, and stress exponents $n_K = 2.0 \pm 0.1$ and $n_M = 3.6 \pm 0.1$, respectively. These differences suggest that a physical mechanism controlling the transient creep differs from the steady-state creep of olivine aggregates. The lower activation energy of the transient creep could be due to a higher jog density of the corresponding soft-slip system.

Furthermore, in order to quantify the effect of water on the transient creep of olivine, I gathered experimental data on the deformation of hydrated single-crystal olivine, where experiments were performed at different pressures and with varying

water content. I determine water and pressure effects in olivine are anisotropic: while the water effect on the soft-slip system (i.e., $[110]_c$ orientation) is small (i.e., $r_{[100]}=0.35 \pm 0.08$), it is large (i.e., $r_{[001]}=1.3 \pm 0.30$) for the hard-slip system (i.e., $[011]_c$ orientation). Assuming a model of soft-slip controlling the transient creep and hard-slip controlling the steady-state creep proposed for ice is true for olivine, I propose that short-term time-dependent small strain deformation associated with post-seismic period is strongly dependent on the water content.

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Chapter 1

Introduction

Earthquakes can be viewed as giant rock deformation experiments in which a large coseismic stress is applied on the rocks of the lower crust and/or the upper mantle. The response of the lower crust and/or the upper mantle can be recorded as postseismic deformation using geodetic instruments such as Global Navigation Satellite System (GNSS) and Interferometric synthetic aperture radar (InSAR). Such postseismic observations following many earthquakes have been explained using a variety of physical mechanisms such as afterslip on and below the coseismic faults (e.g., [Hsu et al., 2006](#); [Perfettini & Avouac, 2004](#)), pore fluid migration in the crust (e.g., [Fialko, 2004](#); [Jonsson et al., 2003](#); [Peltzer et al., 1998](#)), and/or viscoelastic relaxation of the lower crust or the upper mantle (e.g., [Deng et al., 1998](#); [Nur & Mavko, 1974](#)).

Most of the time, all these mechanisms work in parallel, and determining their individual contributions accurately is challenging. Nevertheless, these mechanisms

can sometimes be distinguished using geodetic observations. For example, the deformation following the 1999 Mw 7.1 Hector mine earthquake was explained by the viscoelastic relaxation model, as an afterslip model suggested opposite vertical deformation ([Pollitz et al., 2001](#)). Such viscoelastic relaxation models are particularly helpful in understanding the rheology of the upper mantle.

The rheology of the upper mantle can be studied using three broad approaches. First, conducting experiments on minerals/rocks that are representative of the Earth's upper mantle under in-situ conditions, such as deformation experiments on olivine and dunites (e.g., [P. Chopra & Paterson, 1984](#); [S. Karato & Jung, 2003](#); [S. Karato et al., 1986](#); [Mei & Kohlstedt, 2000a,b](#)). Second, there is modeling of geophysical observations, such as postglacial rebound, postseismic deformation studies (e.g., [Freed & Bürgmann, 2004](#)). In the second category, there are many studies that concentrate on estimating the viscosity structure of the upper mantle from postglacial rebound and postseismic observations using mechanical analogue models such as Maxwell models (e.g., [Muto et al., 2016](#); [Peltier et al., 1980](#); [Sun et al., 2014](#)). The third approach, which is adopted in this thesis, combines these two methods, that is, modeling observations (e.g., postglacial and postseismic deformation) captured using geophysical methods with knowledge from laboratory studies, such as constitutive relationships and flow law parameters of the rocks and minerals representative of the upper mantle. Specifically, flow law parameters of olivine/dunites can be used to model postglacial rebound data ([R. L. Post &](#)

Griggs, 1973) and postseismic observations (e.g., Freed & Bürgmann, 2004). Applying the third approach requires some discussion of the current understanding of the rheology of the upper mantle from laboratory and modeling perspectives.

1.1 Deformation of the upper mantle from laboratory perspective

The deformation in the upper mantle is controlled by olivine, which is the major mantle mineral (with volume fraction $> 60\%$). A magnesium iron rich olivine mineral usually has orthorhombic crystal structure. Consequently, the deformation of olivine and olivine-bearing rocks has been intensively studied over the past half century. Olivine has a orthorhombic crystal system and in laboratory experiments are conducted on three different orientations i.e., $[110]_c$, $[011]_c$, and $[101]_c$ and these orientations activate three different slip systems viz., $(010)[100]$, $(010)[001]$, and $(001)[100]$ or $(100)[001]$. The $(010)[100]$ slip systems is activated in the laboratory by applying stress in direction bisecting (010) and $[100]$ as shown in Figure 1.1

Early studies concentrated on the olivine bearing rocks such as dunites and peridotites (e.g., Blacic, 1972; Carter & Ave'Lallemant, 1970; P. Chopra & Paterson, 1981, 1984; W. Durham & Goetze, 1977; Poumellec & Jaoul, 1984; Raleigh, 1968). Since dunites and peridotites contain other minerals such as chromite and pyroxene, scientists have turned their attention to pure San Carlos olivine, which is a

FIGURE 1.1: Schematic of the olivine single crystal with slip on (010) plane in the [100] direction indicating the activation of [100](010) i.e., easy slip system or $[110]_c$ orientation.

iron rich member with Fo_{90} , where Fo stands for forsterite (e.g., [S. Karato & Jung, 2003](#); [S. Karato et al., 1986](#); [Mei & Kohlstedt, 2000a,b](#)).

1.1.1 Deformation mechanisms

At steady-state, olivine under upper mantle conditions deforms mainly by thermally activated motion of dislocation, which is divided into three mechanisms viz.,

Peierls mechanism, dislocation creep, and diffusion creep. In general, Peierls mechanism has an exponential relationship between the stress and strain-rate, and is accepted to be dominant in the lithospheric mantle (i.e., under low temperature conditions). Steady-state dislocation creep has a nonlinear relationship between the differential stress (σ) and the strain rate ($\dot{\epsilon}_{\text{dis}}$) of the form

$$\dot{\epsilon}_{\text{dis}} = A \sigma^n (C_{\text{OH}})^r \exp \left(-\frac{E + pV}{RT} \right), \quad (1.1)$$

where A is the pre-exponential factor; n is the stress exponent; E is the activation energy; R is the universal gas constant; T is the temperature; V is the activation volume; p is the confining pressure; C_{OH} is the water concentration; and r is the water content exponent (G. Hirth & Kohlstedt, 2003; S. Karato & Jung, 2003). Steady-state diffusion creep has a linear relationship between the differential stress and the strain rate ($\dot{\epsilon}_{\text{dif}}$) of the form

$$\dot{\epsilon}_{\text{dif}} = A \sigma d^{-q} (C_{\text{OH}})^r \exp \left(-\frac{E + pV}{RT} \right), \quad (1.2)$$

where d is the grain size and q is the grain size exponent. Note that the flow law parameter values differ for diffusion creep and dislocation creep. For example, the activation energy of olivine in the dislocation creep regime is ~ 530 kJ/mol, while for the diffusion creep it is ~ 350 kJ/mol (G. Hirth & Kohlstedt, 2003).

1.1.2 Identification of deformation mechanism

In order to understand the deformation of the upper mantle, one needs to, first, create conditions of high temperature and high pressure; second, control the thermodynamic environment (such as water and oxygen fugacity); third, apply deviatoric stress to the sample; and fourth, record the mechanical response as stresses and strains. After the experiments, the deformed samples can be examined in the laboratory for deformation microstructures. By comparing microstructures from the laboratory to microstructures of naturally deformed rocks, interesting conclusions can be drawn about the deformation of the upper mantle. Mechanical data (i.e., stress vs. strain) can be analyzed using the above-described flow laws to identify the deformation mechanism.

Analysing the deformed samples by comparing it to naturally deformed samples is difficult as a naturally deformed rock might have gone through more than one type of deformation mechanism, depending on the temperature and pressure conditions. Therefore, it is difficult to identify the deformation mechanism in such samples. Additionally, deformation by dislocation creep leaves a strong signature in terms of lattice preferred orientation. However, the diffusion creep does not leave any such microstructural signatures ([S.-i. Karato, 2010](#)). As rocks coming from the deep interior of Earth (lower crust or below) have complicated history of deformation, interpretations of microstructural analysis of natural rocks may differ across laboratories.

Analyzing mechanical data (i.e., stress and strain measurements) also poses several challenges. The first comes from the fact that in order to precisely measure mechanical data, the chemical environment of the experiment has to be properly set up. Secondly, different apparatuses have different uncertainties in stress and strain measurements (e.g., [Masuti et al., 2019](#)). For example, a gas-medium apparatus can measure stress precisely, but the confining pressure is limited to 0.5 GPa. On the other hand, the measurements errors of stress in solid-medium apparatus are large, although the confining pressure can go up to 3.6 GPa (e.g., [Jung et al., 2009](#); [S. Karato & Jung, 2003](#)).

Depending on the grain size and the differential stress, the dominant mechanism of deformation in olivine may be either dislocation creep or diffusion creep. A deformation mechanism map drawn at different strain rates with the grain size on one axis and the differential stress on the other helps to visualize the dominant mechanism ([M. F. Ashby, 1972](#)). Dislocation creep is usually considered the dominant mechanism under asthenospheric conditions based on microstructural analysis of samples in deformation experiments (e.g., [S. Karato & Wu, 1993](#)), seismic anisotropy studies (e.g., [Montagner & Tanimoto, 1990](#)), and mantle xenolith microstructure analysis (e.g., [Mercier, 1985](#); [Park & Jung, 2015](#)). However, it is known that deformation of olivine under both laboratory conditions and in the asthenosphere is very close to the transition between dislocation creep and diffusion creep. Due to this diffuse boundary, a number of recent studies have found a region

which has characteristics of both dislocation creep and diffusion creep: the region has a nonlinear relationship between the stress and strain rate, and depends on the grain size. This region is called dislocation accommodated grain boundary sliding, or disGBS for short (e.g., [Hansen et al., 2011](#); [G. Hirth & Kohlstedt, 1995](#)).

1.1.3 Effect of water and pressure on the deformation of olivine

Water influences the rheological properties of the upper mantle, and hence promotes many geological processes, such as plate tectonics and mantle convection. Water reduces the strength of olivine-bearing rock, and its effect on olivine has been quantified i.e., $r = 1.2$ for dislocation creep and $r = 1.0$ for diffusion creep (e.g., [G. Hirth & Kohlstedt, 2003](#); [S. Karato & Jung, 2003](#); [Mei & Kohlstedt, 2000a](#)). To better understand how varying water content and other thermodynamic parameters (e.g., temperature and pressure) impact the strength of the upper mantle, we define

$$\dot{\gamma} = \dot{\gamma}_0 \left(\frac{\sigma}{G} \right)^n, \quad (1.3)$$

where $\dot{\gamma}$ is the strain rate, σ is the deviatoric stress, G is rigidity, and $\dot{\gamma}_0$ is the reference fluidity given by

$$\dot{\gamma}_0 = A G^n C_{\text{OH}}^r \exp \left(-\frac{E + pV}{RT} \right). \quad (1.4)$$

Using equation (1.4), we can plot fluidity vs. depth by varying the water content, temperature, and the activation volume for an oceanic upper mantle. As water content has a major effect on the strain rate ($r = 1.2$) in the dislocation creep regime, we can see that higher water content will produce higher reference fluidity (Fig. 1.2). The effect of water is important as it dissolves in olivine and other mantle minerals as a point defect. Due to its abundance, olivine constitutes a water reservoir in the upper mantle. It has been estimated that the amount of water in the mantle may exceed the mass of the oceans (e.g., [Hirschmann, 2006](#)).

The activation volume defines the lower boundary of the asthenosphere. Physically, it represents the partial molar volume of mobile vacancies and the partial molar volume of activated complexes, for example, complexes such as SiO_4 (e.g., [Ross et al., 1979](#); [Sherby & Simnad, 1961](#)). Over the years, determination of activation volume has been one of the challenging problems in the mineral physics community and the consensus over its value has not been reached yet (e.g., [Bürgmann & Dresen, 2008](#); [Korenaga & Karato, 2008](#)). The challenge has been partly due to non-availability of deformation apparatus and partly due to measurement challenges (see Chapter. 3 and references there in).

1.1.4 Different stages of deformation: Transient vs. steady-state creep

When a material is subjected to external stress, the material deforms by an instantaneous strain. If the material returns to its initial state when the stress is removed,

FIGURE 1.2: Effect of water content, temperature, and pressure on fluidity. a) Temperature profiles for an oceanic lithosphere of ages 40, 50, 60, and 70 Myr calculated using a cooling half-space model with reference temperature of $T_m = 1350^\circ\text{C}$. b) Effect of plate age and temperature on fluidity. c) Effect of water content on fluidity. d) Effect of pressure on fluidity (i.e., activation volume). The activation volume controls the lower boundary of the asthenosphere, whereas temperature controls the upper boundary of the asthenosphere.

the strain is elastic. However, if the material is subjected to stress for a longer duration or at a higher temperature, then a time dependent strain occurs and the material does not return to its initial state when the stress is removed. Such a strain is called anelastic or inelastic (i.e., thermodynamically irreversible). Anelastic strain in most materials starts with a transient period. During the transient period, microstructures such as dislocation density or grain size evolve, with each

achieving steady-state at different times (S. Karato, 2008). Consequently, before achieving steady-state, the microstructures will pass through multiple quasi-steady-states, which are exhibited as transient creep. Hence, transient creep is defined as a period when either the strain rate at a constant stress or stress at a constant strain rate changes with time (Fig. 1.3).

FIGURE 1.3: Transient creep and steady-state creep of any material under a constant strain rate and constant stress conditions.

1.2 Deformation of the upper mantle from a modeling perspective

An alternative to rock deformation experiments is to model postseismic transient deformation and postglacial rebound geodetic data. These data can be modelled using the viscoelastic constitutive equations given by phenomenological models such as Maxwell, Kelvin, standard linear solid (SLS), and/or Burgers models. These models usually contain linear elastic and linear viscous elements connected either in series or parallel (Bürgmann & Dresen, 2008). Out of these models, the linear

Maxwell and linear Burgers models are the ones used most commonly when explaining geodetic observations following an earthquake (Figure 1.4).

The linear Maxwell model consists of a spring and a dashpot connected in series. The spring represents the elastic component, where stress (σ) and strain (ε) have a linear relationship with a proportionality constant, known as rigidity (i.e., $\sigma = G\varepsilon$, where G is the rigidity of the material), whereas the dashpot represents viscous deformation, where stress and strain-rate ($\dot{\varepsilon}$) have a linear relationship with a proportionality constant, known as viscosity (i.e., $\eta = \sigma/\dot{\varepsilon}$, where η is the viscosity of the material). When a stress is applied to a Maxwell body, the spring accommodates instantaneous elastic deformation, which is fully recoverable. However, the dashpot responds slowly to the stress applied, accommodating viscous deformation, which is non-recoverable, and usually represents long-term steady-state deformation of the lower crust/upper mantle.

The linear Burgers model has a Maxwell element connected in series with a Kelvin element. The Kelvin element has a spring and a dashpot connected in parallel. When a stress is applied to a Kelvin body, the dashpot does not allow the spring to deform instantaneously: instead, the dashpot deforms slowly. If the stress is removed, the strain in the Kelvin body is fully recoverable. The Burgers model is *bi-viscous*, that is, one viscosity is associated with the Kelvin element and the other with the Maxwell element (Figure 1.4). This model will be further discussed in detail in the following sections.

FIGURE 1.4: A schematic of the Burgers model. a) The Maxwell element has a dashpot and a spring in series. The spring in the Maxwell element represents elastic strain, where μ_2 is rigidity, and the dashpot represents long-term viscosity. The Kelvin element has a spring in parallel with a dashpot, and represents transient viscosity.

1.2.1 Viscoelastic models of postseismic observations

Early studies used linear Maxwell models of the asthenosphere to explain the post-seismic geodetic data following large earthquakes at longer time scales (e.g., [Nur & Mavko, 1974](#); [Thatcher & Rundle, 1979](#)). However, [Pollitz \(2003\)](#) noticed that the early postseismic deformation following the 1999 Hector Mine earthquake could be explained using an effective mantle viscosity of $\sim 5 \times 10^{17}$ Pa s; after few months of the event the effective viscosity of the mantle needed explain the observations increased to $\sim 2 \times 10^{19}$ Pa s. Following this observation, he concluded that this could be due to either nonlinear or transient rheology of the mantle, and preferred a Burgers model, in which the Kelvin element represents the transient part.

1.2.2 Transient rheology of the lower crust/upper mantle

A linear Burgers rheology has frequently been used to explain postseismic observations following large earthquakes. For example, the geodetic observations following the 2002 Mw 7.9 Denali earthquake have been explained using a transient viscosity of 1.0×10^{17} Pa s and a steady-state viscosity of 2.8×10^{18} Pa s (Pollitz, 2005), and the postseismic deformation following the 2004 Sumatra earthquake and the 2010 Mw 8.8 Maule earthquake were explained using a similar transient mantle rheology (K. Wang et al., 2012). However, a Burgers rheology with linear relationship between stress and strain-rate is not adequate to represent laboratory observations, where a nonlinear relation between stress and strain-rate is common. Therefore, the Burgers rheology needs to be modified. We propose one solution to this problem in the next section.

1.3 Using insights from laboratory experiments in our modeling framework

As mentioned above, the approach taken in this thesis is to model postseismic deformation based on results of laboratory experiments. In order to do this, we need a flow law for olivine that captures both the transient and steady-state creep, and is consistent with laboratory observations. While several formulations include both transient and steady-state creep in different materials (Andrade, 1910; M. Ashby & Duval, 1985; Duval et al., 1983; S. Karato, 1998; J. Poirier, 1980), they lack the

essential properties of a rheological framework, such as time-frame invariance and reference-frame indifference. Hence, we propose a general formulation that includes both of these essential properties, and captures transient and steady-state creep consistently:

$$\dot{\varepsilon}^i = f(\sigma, \varepsilon^i, \theta), \quad (1.5)$$

where θ is a state variable representing either the evolution of microstructures of olivine or internal stress; σ is the applied/differential stress; ε^i and $\dot{\varepsilon}^i$ are the anelastic strain and strain rates, respectively; and f is a general function whose form will be established in the next section.

1.3.1 Constitutive models for the transient creep of olivine

We consider the Burgers model, which consists of the Maxwell and Kelvin elements connected in series (Figure 1.4). First, we decompose the total strain into elastic (ε^e) and anelastic strain components (ε^i):

$$\varepsilon = \varepsilon^e + \varepsilon^i. \quad (1.6)$$

In a Burgers model, the strain-rates in the Maxwell element and Kelvin element are additive:

$$\dot{\varepsilon}^i = \dot{\varepsilon}_K + \dot{\varepsilon}_M, \quad (1.7)$$

where the relation between stress and strain-rate in the Kelvin dashpot follows a law of the type

$$\dot{\epsilon}_K = f_K(\sigma_K), \quad (1.8)$$

with σ_K being the stress in the Kelvin system, and the strain rate in the Maxwell element follows

$$\dot{\epsilon}_M = f_M(\sigma), \quad (1.9)$$

where σ is the deviatoric stress in the Maxwell element. The total stress in the Kelvin element is the sum of the stress in the spring ($G_K \epsilon_K$) and the dashpot ($\sigma_K = f_K^{-1}(\dot{\epsilon}_K)$), and is balanced with the macroscopic stress as follows:

$$\sigma = G_K \epsilon_K + f_K^{-1}(\dot{\epsilon}_K). \quad (1.10)$$

Taking the function f_K on both sides, we get

$$\dot{\epsilon}_K = f_K(\sigma - G_K \epsilon_K). \quad (1.11)$$

Writing these together and with equation 1.6, we obtain constitutive equations in the form $\dot{\epsilon} = f(\sigma, \epsilon^i, \epsilon_K)$, namely

$$\begin{aligned} \dot{\sigma} &= -G(\dot{\epsilon}_M + \dot{\epsilon}_K) \\ \dot{\epsilon}^i &= f_M(\sigma) + f_K(\sigma - G_K \epsilon_K) \\ \dot{\epsilon}_K &= f_K(\sigma - G_K \epsilon_K), \end{aligned} \quad (1.12)$$

where in this context ϵ_K can be thought of as a state variable, and the equation for $\dot{\epsilon}_K$ represents the state evolution equation. f_M represents long-term steady-state creep. The functional form of steady-state creep and its associated flow law parameters are well known from laboratory experiments (discussed in section 1.1.1). The remaining task is to establish the functional form of f_K . As the exact form of f_K is unknown, we can test with a linear, nonlinear, or exponential form. For simplicity, we choose the functional form of the transient to be consistent with the corresponding steady-state creep.

1.3.2 Modified Burgers rheology for the diffusion creep of olivine

In the case of diffusion creep, the relation between stress and strain-rate is linear.

The functions f_K and f_M can be written as

$$\begin{aligned} f_K(\sigma - G_K \epsilon_K) &= A_K (\sigma - G_K \epsilon_K) d^{q_K} C_{OH}^{r_K} \exp \left(-\frac{Q_K + pV_K}{RT} \right) \\ f_M(\sigma) &= A_M \sigma d^{-q_M} C_{OH}^{r_M} \exp \left(-\frac{E_M + pV_M}{RT} \right), \end{aligned} \quad (1.13)$$

where the subscripts K and M refer to the Kelvin and Maxwell elements, respectively.

1.3.3 Modified Burgers rheology for the dislocation creep of olivine

In the case of dislocation creep, the relation between stress and strain-rate is non-linear. The functions f_K and f_M can be written as

$$\begin{aligned} f_K(\sigma - G_K \varepsilon_K^i) &= A_K (\sigma - G_K \varepsilon_K^i)^{n_K} C_{OH}^{r_K} \exp\left(-\frac{Q_K + pV_K}{RT}\right) \\ f_M(\sigma) &= A_M \sigma^{n_M} C_{OH}^{r_M} \exp\left(-\frac{Q_M + pV_M}{RT}\right), \end{aligned} \quad (1.14)$$

where the subscripts K and M refer to the Kelvin and Maxwell elements, respectively. The general constitutive relationship requires a tensor formulation and is written as

$$\begin{aligned} f_K(\sigma_{ij} - G_K(\varepsilon_K^i)_{ij}) &= A_K ||x||^{n_K-1} C_{OH}^{r_K} \exp\left(-\frac{Q_K + pV_K}{RT}\right) X_{ij} \\ f_M(\sigma_{ij}) &= A_M \sigma_{ij}^{n_M} C_{OH}^{r_M} \exp\left(-\frac{Q_M + pV_M}{RT}\right), \end{aligned} \quad (1.15)$$

where subscripts i and j are the tensor indices (Einstein's summation convention) and

$$X_{ij} = \sigma_{ij} - G_K(\varepsilon_K)_{ij}, \quad (1.16)$$

where X_{ij} is the deviatoric stress tensor. Substituting the functions $f_K(\sigma - G_K \varepsilon_K^i)$ and $f_M(\sigma)$ in 1.12 with the above equations, we get the coupled ordinary differential equations (ODE's), which we then solve numerically using the initial conditions of $\sigma = \sigma_0 = 1$ MPa, $\varepsilon_K^i = 0$, and $\varepsilon^i = 0$ (Figure 1.5).

FIGURE 1.5: Comparison of Maxwell steady-state creep with that of linear (diffusion) and nonlinear Burgers (dislocation) models with their corresponding transients. The anelastic strain is non-dimensionalized by σ_0/G and time by $t_M = \eta_M/G$, where η_M is the steady-state viscosity ($\sim 10^{18}$ Pa s).

1.4 Summary and thesis organization

In this chapter, we have discussed the deformation of the upper mantle from laboratory and modeling perspectives, and the approach taken in this thesis. We have also reviewed various observations supporting the existence of the transient rheology of olivine, noting that none of the existing models capture the transient and steady-state creep consistently. In order to address this issue, we have proposed a modified Burgers rheology to model the laboratory data of olivine deformation and geodetic observations following a large earthquake.

This thesis is organized in the following way. In the second chapter, we apply the above-proposed equations of the transient rheology of olivine to modeling of the postseismic observations following the 2012 Mw 8.6 Indian Ocean earthquake. We use the steady-state creep flow law parameters from [S. Karato & Jung \(2003\)](#).

However, due to a lack of values for the transient creep of olivine, we assume them to be the same as those of the steady-state creep. We estimate the water content of the Indian Ocean upper mantle, and show the importance of including transient creep in the postseismic deformation modeling studies. In the third chapter, we analyze laboratory data of single crystal olivine under hydrous conditions, where experiments were performed at different confining pressures and water contents. We determine the water content dependency and pressure dependency of the soft- and hard-slip systems of olivine. We propose a theoretical model of jog density to explain our results, and propose a physical model for the transient creep of olivine. As per this model, the transient creep of olivine is dominated by the slip on the soft-slip system, and the steady-state creep is dominated by the slip on the hard-slip system. In the fourth chapter, we use published data from [P. Chopra & Paterson \(1984\)](#), where deformation experiments were performed on dunites under anhydrous conditions, and we determine the transient flow law parameters of olivine. We get different temperature sensitivities of olivine deformation for the transient and steady-state creeps of olivine. We explain our results by extending the theoretical model proposed in the third chapter (i.e., the model of jog density, where the soft-slip system is saturated with jogs and the hard-slip system is undersaturated). Our results also suggest that the underlying physical mechanism of the transient creep of olivine is different from the physical mechanism controlling the steady-state creep.

Chapter 2

Upper mantle water stratification inferred from observations of the 2012 Mw 8.6 Indian Ocean earthquake*

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Abstract: Water, the most abundant volatile in Earth's interior, preserves the young face of our planet by catalyzing mantle convection, lubricating plate tectonics, and feeding forearc volcanism. Since planetary accretion, water has been exchanged between the hydrosphere and the geosphere, but its depth distribution in the mantle remains elusive. Here, we exploit the water-sensitivity of the strength of olivine, the weakest and most abundant mineral in the upper mantle, and the unique constraints brought by the exceptionally large 2012 Mw 8.6 Wharton Basin earthquake in the Indian Ocean to evidence stratification of water content in the upper mantle. We explain the transient deformation in the aftermath of the earthquake that was recorded by continuous geodetic stations along Sumatra by the wet dislocation creep of olivine, indicating a minimum water content of ~ 0.01 wt% or $1,600 \text{ H}/10^6 \text{ Si}$ in the asthenosphere, considering a wide range of temperature conditions and including the effect of transient creep of olivine. The earthquake ruptured conjugate faults throughout the entire oceanic lithosphere down to great depths, compatible with dry olivine in the thick, brittle, layer. We explain the steep rheological contrast by dehydration across the lithosphere-asthenosphere boundary, presumably by buoyant melt migration to form the oceanic crust.

Water is heterogeneously distributed in the Earth's interior among nominally anhydrous minerals (e.g., olivine and pyroxene), melts, and fluids. The principal source of water in the upper mantle is from its dissolution in olivine and other minerals as a point defect. Olivine, the most abundant upper mantle mineral, constitutes a water reservoir that may exceed the mass of the oceans ([Hirschmann, 2006](#)). Water drastically reduces the strength of olivine ([S. Karato et al., 1986](#)) and this effect

can be exploited to estimate the water content from the mechanical response of the asthenosphere to stress perturbations such as the ones following large earthquakes.

Since the 2012 Mw 8.6 Wharton Basin earthquake ruptured the entire oceanic lithosphere (Hill et al., 2015; Wei et al., 2013; Yue et al., 2012), the postseismic deformation (Fig. 2.1) following this event provides strong constraints on the rheological properties of the oceanic lithosphere-asthenosphere system. This is in contrast to most postseismic deformation, where earthquakes occur on the continental regions and hence the influence of lower crust needs to be corrected in order to infer the rheological properties of the upper mantle (Bürgmann & Dresen, 2008).

Several mechanisms may explain the postseismic deformation, including accelerated viscoelastic flow in the asthenosphere and triggered slip (afterslip) along the Wharton Basin faults or the Sunda megathrust. Vertical deformation is sensitive to the gradient of the horizontal displacements, providing useful constraints to identify the likely source of deformation (Bruhat et al., 2011; Rollins et al., 2015; Rousset et al., 2012). Afterslip around the coseismic slip area creates horizontal motion compatible with geodetic observations, but predicts subsidence of the fore-arc islands, opposite to what is observed (Figures 2.1, 2.5, & 2.9). Afterslip on the Sunda megathrust creates the right polarity of vertical motion but the thrust motion gives the reverse sense of horizontal displacements. Viscoelastic flow confined in a finite-thickness asthenosphere predicts the right polarity of vertical displacements and the right azimuth of horizontal displacements. Likely, both afterslip and

FIGURE 2.1: Postseismic deformation and aftershocks (red dots, NEIC catalogue) 1 year after the 2012 Mw 8.6 Wharton Basin earthquake. A) The stations of the Sumatran GPS Array north of the equator moved landwards while the others moved seawards (black vectors with uncertainties). The northern stations overall uplifted (colored circles, warm colors for uplift). Stations LEWK, BNON and BSIM on a forearc island produced the most uplift. The source mechanisms of the main shock and largest aftershock are from the Global Centroid Moment Tensor Catalog. The plate age is from Müller et al. (2008) that controls the temperature at depth is shown along the Sunda trench. The coseismic stress change driving the postseismic relaxation is shown in contours of 300 kPa and 30 kPa at 100 km depth. B) Cross section along the A-A' profile shown in A of the plate age, coseismic stress change and slab geometry. The red segments indicate faults in the brittle layer. No significant stress change occurs in the mantle wedge, making the experiment setup insensitive to rheology below the Sunda plate.

viscous flow in asthenosphere contribute to the displacements, and their mechanical coupling must be included in the analysis.

The plastic deformation of olivine is accommodated by a combination of diffusion creep and dislocation creep (e.g., [Bürgmann & Dresen, 2008](#); [G. Hirth & Kohlstedt, 2003](#); [S. Karato & Wu, 1993](#)), and the steady-state rheology (the constitutive stress-strain rate relationship) of olivine is a thermally activated flow law of the form (e.g., [G. Hirth & Kohlstedt, 2003](#); [S. Karato & Jung, 2003](#))

$$\dot{\epsilon} = A \sigma^n d^{-m} (C_{\text{OH}})^r \exp \left(-\frac{Q + pV}{RT} \right) \quad (2.1)$$

where $\dot{\epsilon}$ is the strain rate, A is a pre-exponential factor, σ is the deviatoric stress, n is the stress exponent, Q is the activation energy, R is the universal gas constant, T is the temperature, V is the activation volume, p is confining pressure, d is grain size, m is the grain size exponent, and C_{OH} and r are the water concentration and its exponent. Diffusion creep is for $n = 1$ and $m = 2 - 3$, and dislocation creep for $n = 3 - 5$ and $m = 0$. The temperature weakening and pressure hardening in the Arrhenius law define the top and bottom boundaries of the mechanical asthenosphere, respectively. The strain rates from the two mechanisms add up, but one of them dominates depending on stress, temperature, pressure, and grain size (e.g., [S. Karato et al., 1986](#); [S. Karato & Wu, 1993](#)). Ubiquitous observations of seismic anisotropy in the upper mantle (e.g., [Gaherty et al., 1996](#); [Montagner & Tanimoto, 1990](#)) suggest that dislocation creep is the dominant deformation mechanism at these depths, which occurs for grain sizes of $d = 5$ mm and above, compatible with observations of grain size ranging from 5 to 20 mm in mantle xenoliths ([S. Karato & Wu, 1993](#)). Because of the competition between grain size reduction by dislocation creep and grain growth during diffusion creep the two creep mechanisms may

coexist at steady state (S. Karato & Wu, 1993), but the dominance of dislocation creep is enhanced during the postseismic period because of the power-law dependence on stress and the large stress perturbation from the mainshock (Figure 2.2B, 2.7).

Laboratory experiments show that olivine creep exhibits a rapid transient before reaching steady-state creep (P. N. Chopra, 1997; Hanson & Spetzler, 1994; S. Karato, 1998) and previous studies have suggested that postseismic signals cannot be explained with realistic Earth properties without including the transient effect (e.g., Freed et al., 2010; Pollitz, 2003; K. Wang et al., 2012). The underlying mechanism for the transient behavior is not fully understood but some experiments suggest that it may be caused by the transition in the mechanism of strain accommodation in dislocation creep (e.g., S. Karato, 1998) or the evolution of internal stress distribution in diffusion creep (e.g., Mackwell et al., 1985; Raj, 1975). Here, we propose an addition to the rheology of olivine that includes these transient effects in the diffusion creep or dislocation creep regime (see Methods).

$$\dot{\varepsilon}_K = A_K [\sigma - G_K \varepsilon_K]^n (C_{OH})^r d^{-m} \exp \left(-\frac{Q + pV}{RT} \right) \quad (2.2)$$

where ε_K is the transient strain and G_K is a work hardening coefficient. The formulation is compatible with the linear Burgers rheology that combines a Kelvin and a Maxwell element in series for $n = 1$, which corresponds to the transient of diffusion creep. (Here, the subscript K corresponds to the constitutive properties

FIGURE 2.2: Probability density of water content in the asthenosphere from modeling of postseismic deformation. A) Marginal (solid profile) and conditional (dashed profile) probability density of water content constrained by geodetic observations. The colors indicate different mantle temperatures (T_m). These models include the transient creep of olivine except for the blue one. The prior density of water content from laboratory experiments (solid black profile) is a log-normal distribution with a mean value of $600 \text{ H}/10^6 \text{ Si}$ and a standard deviation of 1. Geochemical estimates (vertical dashed profiles) of water content in the mid-ocean ridge basalt source provide independent lower and upper bounds. B) Relaxation times and effective viscosity for the dislocation (solid profiles) and diffusion creep (dashed profiles) regimes. We assume a background strain rate of $\dot{\epsilon} = 10^{-15} \text{ s}^{-1}$ and a mantle temperature of 1380°C ; other parameters are from [S. Karato & Jung \(2003\)](#). The strength of dislocation creep is dynamically reduced due to stress changes from the mainshock (black solid profile). C) Observed (black arrows and colored circles, same as in Fig. 1) and modeled (white arrows for horizontals and background color for verticals) displacements after 1 year assuming afterslip on the Wharton Basin faults and viscoelastic flow in the asthenosphere. The slab contours are from the USGS Slab1.0. D) Time series of GPS (blue squares with $\pm 1\sigma$ error bars) and modeled (red profile) displacements following the 2012 Mw 8.6 Wharton Basin earthquake including the transient creep of olivine.

of the Kelvin element.). For larger power exponents and $m = 0$, the formulation represents the transient of dislocation creep; both are thermally activated (parameters such as A , r , Q and V for transient creep may be different from those for

steady state, but for simplicity, we assume that these are the same between transient and steady-state creep except for the pre-exponential factors A and A_K). The transient and steady-state strain rates add up to relax the same stress, leading to a more rapid initial deformation than with steady-state creep alone (Figure 2.6). At steady state the internal stress is relaxed ($\sigma - G_K \epsilon_K = 0$) so the transient flow does not affect the long-term strength of the upper mantle.

We propose a physical model for the geodetic observations where viscoelastic relaxation in the asthenosphere and afterslip on the Wharton Basin faults work in concert to relax the coseismic stress perturbation from the mainshock and the largest aftershock (Barbot & Fialko, 2010b). We build a three dimensional model with a brittle lithosphere that subducts below the Sunda trench (Figure 2.2C) and rides above a low-viscosity asthenosphere (see Methods). The strength of the upper mantle is controlled by friction in the lithosphere and the temperature and pressure dependence of olivine flow in the asthenosphere (we use a cooling half-space and the oceanic plate age model of Müller et al. (2008)). We choose the effective friction coefficient (Figure 2.3A) to obtain a brittle-ductile transition depth of 45 km, based on the deep rupture of the mainshock (Hill et al., 2015; Wei et al., 2013). We model afterslip with rate-strengthening friction (Barbot et al., 2009; Marone et al., 1991) down to 70 km depth, below and around fault areas with negative coseismic stress change.

We use a Bayesian approach to explore the likelihood of the physical parameters

FIGURE 2.3: Strength of the lithosphere and earthquake rupture. A) Moment release from the mainshock from Wei et al. (2013) (red profile), average stress drop (dotted-dashed profile) and lithosphere strength assuming dry olivine (dashed profile), wet olivine (green profile) and stratified water content (black dashed profile). The brittle strength assumes a low effective friction coefficient based on the near-orthogonal conjugate faults that ruptured during the earthquake (inset). B) Effective viscosity for dry olivine, wet olivine and stratified water content (same as in A) using the constitutive equation of S. Karato & Jung (2003). The water content (red profile) changes across the lithosphere-asthenosphere boundary (LAB) between about 45 and 60 km depth.

that control olivine flow and afterslip by assimilating the constraints from our geodetic observations. We test a range of water content from 0.0003 to 0.04 wt% (50 to 6,000 H/ 10^6 Si), mantle temperatures in the cooling half space from 1350 to 1400°C with the adiabatic temperature gradient of 0.4 °C/km (Fig. 2b, see methods), and reference velocities for the rate of afterslip in the range $V_0 = 0-2.75$ $\mu\text{m/s}$. We first evaluate the bivariate probability density and then integrate out the afterslip parameter to produce a single-variate probability density for water content in the Wharton Basin asthenosphere (Figure 2.2A). The result incorporates the uncertainties originating from both sources of deformation, i.e., postseismic slip and viscous flow after the earthquake. We extend the parameter search manually as necessary to investigate the role of the background strain rate, the mantle temperature, the activation volume and transient parameters of olivine (see methods).

We can explain the geodetic observations with the laboratory-derived constitutive properties of olivine and realistic in situ conditions with a water content ranging from 0.01 wt% (1,600 H/10⁶ Si) for a mantle temperature of 1,400 °C to 0.027 wt% (4,200 H/10⁶ Si) for a mantle temperature of 1,350 °C (Figures 2.2, 2.8 & 2.11). The best-fit model for $T_m = 1380^\circ\text{C}$ has $A = 10^{0.56}$, $Q = 418.5 \text{ kJ/mol}$, $V = 11 \text{ cm}^3/\text{mol}$, $r = 1.2$, $m = 0$ for dislocation creep, and $A = 10^6$, $Q = 335 \text{ kJ/mol}$, $V = 4 \text{ cm}^3/\text{mol}$, $r = 1$, $m = 3$, $d = 10 \text{ mm}$ for diffusion creep, and water content of $C_{\text{OH}} = 2,240 \text{ H}/10^6 \text{ Si}$ in Eq. 1 and the same parameters except $A_K = A/2$ and $G_K = G$ in Eq. 2. There is little tradeoff between water content and afterslip speed (Figure 2.8) and the maximum likelihood value of the afterslip parameter falls in the range found in other tectonic settings (Barbot et al., 2009). Our estimated water content is in good agreement with geochemical estimates for the mid-ocean ridge basalt source (Hirschmann, 2006) in the range 0.005-0.02 wt% (800-3,200 H/10⁶ Si). Ignoring the transient creep of olivine introduces bias in estimates of water content. With mantle temperatures of 1,350 °C and below, our estimates of water content are above full water saturation at 100 km depth (Figure 2.2A).

The geodetic data are also sensitive to the thickness of the asthenosphere, which is controlled by the pressure hardening of olivine through the activation volume or the transition to a high-pressure phase at 410 km. Models with an activation volume lower than 5 cm³/mol for dislocation creep fail to explain the observations,

indicating a finite thickness of the mechanical asthenosphere during the postseismic transient. As diffusion creep has a low activation volume (G. Hirth & Kohlstedt, 2003; S. Karato & Jung, 2003), the result is a further indication that dislocation creep is the dominant deformation mechanism in the asthenosphere during the early postseismic transient. The geodetic time series are better modeled assuming a low background strain rate for dislocation creep, indicating that diffusion creep is faster than dislocation creep at steady-state (see methods), compatible with grain sizes of about 5 mm (Figure 2.2B).

The 2012 Mw 8.6 Wharton Basin earthquake propagated into the lithosphere-asthenosphere boundary with an average stress drop of 17 MPa (Hill et al., 2015), and a slip distribution tapering off near the brittle-ductile transition, from 45 to 65 km depth (Wei et al., 2013) (Figure 2.3A). The rupture took place on near-perpendicular conjugate faults (Hill et al., 2015; Wei et al., 2013; Yue et al., 2012), indicating a low effective coefficient of friction (see methods). The thick lithosphere is compatible with the strength of dry olivine (Figure 2.3). These observations imply that water content is stratified with a steep gradient at the lithosphere-asthenosphere boundary (Figure 2.3B). Dehydration of the lithosphere may happen by buoyant partial melt to form the oceanic crust, as the solubility of volatiles is higher in the liquid phase (G. Hirth & Kohlstedt, 1996; S. Karato et al., 1986).

Partial melting is often invoked to explain weak rheology and high electrical conductivity (S. Karato et al., 1986; Naif et al., 2013). However, the deformation

FIGURE 2.4: Schematic of water content distribution across a subduction zone. The oceanic brittle lithosphere is dry (0.001-0.005 wt%) and the oceanic asthenosphere is wet (~ 0.015 wt%) with a water content given by the probability density (inset) assuming the mantle temperature $T_m = 1,380^\circ\text{C}$. The water content in the continental lithosphere and the mantle wedge are from [S. Karato \(2010\)](#).

occurring during and after the 2012 Mw 8.6 Wharton Basin earthquake can be explained by a steep rheological gradient across the lithosphere-asthenosphere boundary controlled by temperature and water stratification without weakening caused by partial melting (Figure 2.4). Application of the proposed methodology to other great and giant earthquakes can refine our estimates of water content in the Sumatran mantle wedge and other tectonic regions.

2.1 Supplementary Information

2.1.1 GPS data processing

We process the SuGAR raw data using the GPS-Inferred Positioning System and Orbit Analysis Simulation Software (GIPSY-OASIS) software (hosted at <https://gipsy-oasis.jpl.nasa.gov/>). To extract the postseismic transient due to the 2012 M_w 8.6 Indian Ocean earthquake, we identify and remove the signals from other earthquakes in the entire available time series for each component of each station. In this procedure, we simultaneously estimate the linear long-term rates, coseismic, postseismic, and seasonal signals by nonlinear leastsquares fitting. A complete description of these steps can be found in [Hill et al. \(2015\)](#) and [Feng et al. \(2015\)](#). We isolate the postseismic deformation from the 2012 Indian Ocean earthquake sequence by removing the contribution of all other identified sources. The resulting time series is shown in [Figure 2.5](#).

2.1.2 Formulation of the transient rheology of olivine

The effect of the transient creep of rocks, whether for diffusion creep or dislocation creep, is well known in the laboratory, and some studies suggest that it is important to incorporate in models of postseismic deformation ([Freed et al., 2010](#); [Pollitz, 2003](#); [K. Wang et al., 2012](#)) or postglacial rebound (e.g., [Ranalli, 1980](#)). Several attempts have been made to formulate a rheological law for transient creep (e.g., [Andrade, 1910](#); [Chinh et al., 2004](#); [Hart, 1970](#); [J. Poirier, 1980](#); [Voce, 1948](#)), but

FIGURE 2.5: Time series of GPS displacements at the Sumatra GPS Array stations where the effect of previous earthquakes, their postseismic relaxation, and other spurious effects have been modeled and removed from the data. The resulting time series isolate the postseismic transient following the 2012 Mw 8.6 Wharton Basin earthquake. The arbitrary colors indicate the GPS station. The vertical lines mark the onset of the postseismic transient. The methodology for data preparation is described in [Hill et al. \(2015\)](#) and [Feng et al. \(2015\)](#).

they have important shortcomings. For example, laws such as

$$\sigma = A\epsilon^q \dot{\epsilon}^r \quad (2.3)$$

where A , q and r are positive constitutive parameters ([J. Poirier, 1980](#); [Voce, 1948](#)), or

$$\sigma = B \left[1 - \exp \left(- \left(\frac{\varepsilon}{\varepsilon_0} \right)^s \right) \right] \dot{\varepsilon}^r \quad (2.4)$$

where B , ε_0 , r and s are positive constants ([Chinh et al., 2004](#); [Voce, 1948](#)), can only create a single transient creep episode, as plastic strain systematically increases, and therefore they do not allow transient creep for repeated earthquake stress perturbations. Some other laws, such as the so-called Andrade creep

$$\dot{\varepsilon} = \frac{ABnt^{n-1}}{1 + At^n} + \dot{\varepsilon}_{ss} \quad (2.5)$$

where $\dot{\varepsilon}_{ss}$ is the steady state strain rate and A and B are positive parameters ([Andrade, 1910](#)), are singular ([S. Karato, 2008](#)) at $t = 0$, for $n > 1$. More fundamentally, laws such as (Eq. 2.5) fail to satisfy fundamental principles such as time frame invariance (the same flow should be predicted for any clock).

There are several ways to formulate a rheology for the transient creep of olivine by invoking work hardening. For simplicity, we propose a generalization of the Burgers rheology. The Burgers rheology may be readily appropriate to represent transient creep in the diffusion creep regime, and we adapt it to be compatible with the power-law steady state of dislocation creep.

In a Burgers material, where the Kelvin element and the Maxwell element are in series, the anelastic strain rate can be written

$$\dot{\epsilon}^i = \dot{\epsilon}_K + \dot{\epsilon}_M, \quad (2.6)$$

where $\dot{\epsilon}_K$ is the anelastic strain rate in the Kelvin element and $\dot{\epsilon}_M$ is the anelastic strain rate in the Maxwell element. The strain-rate of the Maxwell element is given by Eq. 2.1 and strain-rate of the Kelvin element is given by

$$\dot{\epsilon}_K = A_K [\sigma - G_K \epsilon_K]^n (C_{OH})^r d^{-m} \exp \left(-\frac{Q + pV}{RT} \right) \quad (2.7)$$

where ϵ_K is the transient strain and G_K is a work hardening coefficient. The formulation is compatible with the linear Burgers rheology that combines a Kelvin and a Maxwell element in series for $n = 1$, which corresponds to the transient of diffusion creep. For larger power exponents and $m = 0$, the formulation represents the transient of dislocation creep; both are thermally activated. The transient and steady-state strain rates add up to relax the same stress, leading to a more rapid initial deformation than with steady-state creep alone (Figure 2.4). At steady state the internal stress is relaxed ($\sigma - G_K \epsilon_K = 0$) so the transient flow does not affect the long-term strength of the upper mantle.

In the absence of detailed laboratory data on wet olivine, we assume that the thermodynamic parameters for transient and steady-state creep are the same (for dry olivine, [P. N. Chopra \(1997\)](#) and [R. Post \(1977\)](#) show that flow law parameters

are similar between transient and steady-state creep except for the pre-exponential factor). However, laboratory experiments indicate that if transient creep is due to the transition of slip from the soft slip system to the hard slip system of olivine, transient creep should be less sensitive to the water content (Mackwell et al., 1985).

At the background stress σ_0 , if the flow is at steady-state (that is, $\dot{\varepsilon}_K = 0$), we have $\sigma_0 - G_K \varepsilon_K^0 = 0$, where ε_K^0 is the cumulative strain at the background stress. After a stress perturbation from an earthquake, the stress changes from σ_0 to $\sigma + \Delta\sigma$ and the rate of transient creep, unlike steady-state creep, is not sensitive to the background stress during the postseismic transient, unless it was not at steady-state. Multiple stress perturbations also lead to multiple transient creep episodes.

We explore the predictions of the proposed rheology in a spring-slider system. Together with the constitutive relationship, conservation of momentum leads to the system of coupled ordinary differential equations

$$\begin{aligned}\dot{\sigma} &= -G(\dot{\varepsilon}_M + \dot{\varepsilon}_K) \\ \dot{\varepsilon}_M &= A\sigma^n (C_{OH})^r d^{-m} \exp\left(-\frac{Q+pV}{RT}\right) \\ \dot{\varepsilon}_K &= A_K [\sigma - G_K \varepsilon_K]^n (C_{OH})^r d^{-m} \exp\left(-\frac{Q+pV}{RT}\right)\end{aligned}\tag{2.8}$$

We solve these equations numerically with unit stress perturbations and typical laboratory-derived constitutive properties (Figure 2.6). The transient creep accelerates the immediate relaxation that follows the stress perturbation and the

subsequent relaxation is slower, compared to when operating at steady state only. The hardening coefficient G_K controls the amount of strain that is relaxed by transient creep.

We generalize the constitutive relationship for transient creep for three-dimensional deformation with isotropic rheology. The constitutive relationship becomes

$$(\dot{\varepsilon}_K)_{ij} = A_K q^{n-1} (C_{OH})^r d^{-m} \exp\left(-\frac{Q + pV}{RT}\right) Q_{ij}, \quad (2.9)$$

where subscripts i and j are the tensor indices and

$$Q_{ij} = \sigma_{ij} - G_K (\varepsilon_K)_{ij} \quad (2.10)$$

is the internal deviatoric stress tensor and $q^2 = Q_{kl}Q_{kl}$ is the norm of the internal deviatoric stress tensor (we use Einstein's summation convention). We implement these constitutive relationships in the community code Relax, such that we can simulate three-dimensional models of postseismic deformation that includes after-slip, viscoelastic flow and the transient creep of olivine in a self-consistent manner. Figure 2.6B and 2.6C show the predicted time series at a few SuGAR stations for various values of G_K .

FIGURE 2.6: Effect of transient creep of olivine on postseismic transients. A) Creep tests. We compare relaxation tests that include the transient creep of olivine (solid red profile) with the response at steady-state for power-law (dashed blue profile) and linear (dashed black profile) viscoelastic materials. The stress perturbation and relaxation times are chosen to highlight typical behaviors. The resulting strain, stress, and time are non-dimensionalized. The transient creep accelerates the initial response, but slows down the subsequent relaxation. B) Effect of transient creep on models of postseismic relaxation following the Wharton Basin earthquake. Inclusion of transient creep accelerates the deformation at stations BNON and BSIM, other parameters being the same. C) Role of the hardening coefficient G_K on models of postseismic deformation with transient creep at GPS station BSIM. The hardening coefficient controls how much stress is relaxed by transient creep and how long the effect is sustained. The hardening coefficient G_K in our models for the Wharton Basin is equal to the background rigidity.

2.1.3 Bounds on pre-stress

The dynamics of postseismic deformation is sensitive to the stress preceding the coseismic perturbation in the asthenosphere because of the power-law dependence in the stress-strain rate relationship. Our mechanical model for the power-law flow of olivine includes (1) the background strain rate due to the shear arising from the vertical gradient of horizontal flow, which is associated with the long-term oblique subduction of the oceanic lithosphere below Sumatra, and (2) the pure shear associated with internal deformation by conjugate strike-slip faulting of the Wharton basin along the diffuse boundary between the Indian and Australian plates. We convert the background strain rate to pre-stress, accounting for the water content, temperature and the other physical parameters of olivine rheology. We assume a homogeneous velocity gradient from the surface (plate velocity of 5.6 cm yr^{-1} oriented 10° N ; Figure 2.1) to a relative fixed transition zone. The orientation of the conjugate faults in the Wharton basin suggests that the deformation in the lithosphere is horizontal pure shear [Delescluse & Chamot-Rooke \(2007\)](#) with principal stress orientation between -30° N and -25° N for the compressive component and between 60° N and 65° N for the extensive component (Figure 2.3).

The long-term strain rate is the sum of the linear (diffusion creep) and nonlinear (dislocation creep) strain rate contributions

$$\dot{\epsilon} = \dot{\epsilon}_{\text{diffusion}} + \dot{\epsilon}_{\text{dislocation}} \quad \text{or} \quad \dot{\epsilon} = \tau \left(\frac{1}{\eta_{\text{diffusion}}} + \frac{1}{\eta_{\text{dislocation}}} \right) \quad (2.11)$$

where $\eta_{\text{diffusion}}$ and $\eta_{\text{dislocation}}$ are the effective viscosities for diffusion and dislocation creep, respectively. While the linear viscosity due to diffusion creep has little effect on the initial postseismic deformation and cannot be directly estimated (Figure 2.2B), it does affect the background stress

$$\tau = \frac{\eta_{\text{dislocation}}}{1 + \frac{\eta_{\text{dislocation}}}{\eta_{\text{diffusion}}}} \dot{\epsilon} \quad (2.12)$$

So a low-viscosity diffusion creep reduces the background stress and the associated strain rate for dislocation creep.

We consider two end-member models for the background strain rate $\dot{\epsilon}$. In a first model, plate motion is accommodated in the mantle across a 100-km-thick region, leading to a background strain rate of $2 \times 10^{-14} \text{ s}^{-1}$. In a second model, the region is 400km thick, leading to a background strain rate of $5 \times 10^{-15} \text{ s}^{-1}$. Considering these models one at a time, we investigate a range of strain rates for dislocation creep ranging from full background strain rate (corresponding to vanishing diffusion creep in the model) to $1 \times 10^{-17} \text{ s}^{-1}$ (corresponding to dominant diffusion creep at steady state). We find that our geodetic data are best explained with a dislocation strain rate of the order of 10^{-17} s^{-1} to 10^{-16} s^{-1} , corresponding to grain size between 6mm and 10mm depending on the background strain rate. This indicates that diffusion creep and dislocation creep have similar strengths in the asthenosphere at steady state. The strength of diffusion creep is lower than the one for dislocation creep for grain sizes lower than 6mm (Figure 2.2B).

FIGURE 2.7: Comparison between the stress changes impacted by the 2012 Mw 8.6 Wharton Basin (Wei et al., 2013) earthquake and the 2004 Mw 9.2 Aceh-Andaman (Chieh et al., 2007) and the 2005 Mw 8.6 Nias (Konca et al., 2007) earthquakes (coloured beach balls) at 100 km depth. The stress change to the giant and great Sunda megathrust earthquakes is one to two orders of magnitude lower than the one due to the 2012 event in the Wharton Basin asthenosphere. The negligible effect of the megathrust events is due to their stress changes concentrating below the megathrust and in the mantle wedge. In addition, some of this stress has been relaxed in the many years that separate the events.

The recent great and giant earthquakes along the Sunda megathrust affected the background stress preceding the 2012 earthquake sequence (Delescluse et al., 2012). To estimate this effect, we evaluate the deviatoric stress change in the asthenosphere at the time of the 2012 event caused by the nearby 2004 M_w 9.2 Aceh-Andaman and the 2005 M_w 8.6 Nias earthquakes (Figure 2.7). The deviatoric stress near the rupture area of the 2012 event at 100 km depth due to these earthquakes is about two orders of magnitude smaller than the coseismic stress change.

For simplicity we ignore the stress caused by the Sunda megathrust earthquakes in our simulations.

2.1.4 Lithosphere strength

The strength of the brittle lithosphere can be evaluated from the orientation of the conjugate faults using Byerlee's law ($\tau = \mu' \sigma_n$). The effective coefficient of friction is given by

$$\mu' = \frac{1}{\tan 2\theta}, \quad (2.13)$$

where θ is the angle between the fault and the principal stress. The direction of the principal stress should bisect the conjugate faults. In the Indian Ocean earthquake rupture, the main conjugate faults are nearly orthogonal (Figure 2.1 and 2.3), providing an estimate of the effective coefficient of friction in the range 0.1-0.2. The exact orientation of the rupture fault is still subject to debate (Hill et al., 2015), affecting our estimate of the effective coefficient of friction. The inferred small friction coefficient in the Wharton Basin reduces the strength of the brittle lithosphere (Figure 2.3A).

The thickness of the lithosphere can be independently estimated from the depth of the coseismic rupture of the 2012 Mw 8.6 Indian Ocean earthquake as substantial coseismic slip occurred down to 60 km depth (Wei et al., 2013). The reason for the great depth extent of the rupture is unclear, but strong dynamic weakening from frictional melting may have allowed the rupture to penetrate below the seismogenic

zone. The depth range from 45km to 60km where coseismic slip tapers from its maximum value may correspond to the lithosphere-asthenosphere boundary (Figure 2.3A).

2.1.5 Forward models of postseismic deformation

We build a three-dimensional model of the lithosphere and asthenosphere in which the flow parameters depend on depth, except in the subducted slab where viscosity is infinite, resulting in a three-dimensional rheological model. The background depth-dependent viscosity is thermally activated following a cooling half-space model

$$T(z, t) = T_0 + T_m \operatorname{erf} \left(\frac{z}{\sqrt{4Kt}} \right) + \alpha(z - z_0)H(z - z_0) \quad (2.14)$$

with $T_0 = 0^\circ\text{C}$, where T_m is the mantle temperature (a free parameter), the plate age (t) varies spatially following the model of Müller et al. (2008), the thermal diffusivity is $K = k/(\rho_m C_p)$ with conductivity $k = 3.138 \text{ Wm}^{-1}\text{K}^{-1}$, the specific heat is $C_p = 1.171 \text{ J kg}^{-1} \text{ K}^{-1}$, the density is $\rho_m = 3.330 \text{ kg m}^3$, $\alpha = 0.4^\circ\text{C km}^{-1}$ is the adiabatic temperature gradient, $H(z)$ is the Heaviside function, and $z_0 = 100 \text{ km}$ is the depth of thermal boundary layer for a 60-million-year old oceanic plate (Figure 2.2B). In our models, most of the postseismic deformation occurs around 100 km depth, where the coseismic stress change is high, so the adiabatic temperature gradient has almost no effect on our water content estimates. The elastic slab is constructed using the Slab 1.0 model of Hayes et al. (2012) with a uniform thickness of 80 km. Because the accelerated flow is concentrated in areas of high

coseismic stress change, immediately below the mainshock, the effect of the slab does not greatly affect the fit to the geodetic data.

We create models of postseismic relaxation where the coseismic stresses from the mainshock and the largest aftershock are potentially relaxed by both afterslip in the lithosphere and viscoelastic flow in the asthenosphere, depending on the rheological parameters. We use the coseismic slip model of [Wei et al. \(2013\)](#) to produce the stress perturbation and to define the geometry of faults for afterslip. In our relaxation models, both afterslip and viscoelastic deformation are driven by stress and the coupling between the two mechanisms is taken into account. Afterslip is allowed only on the faults that ruptured coseismically, around the areas of negative coseismic stress change. We assume uniform friction properties outside the rupture area, as only near-field data would motivate finer-tuned models. We model afterslip with rate-strengthening friction, a simplification of the rate-and-state friction law that is adequate to represent triggered aseismic slip at steady state (e.g., [Bruhat et al., 2011](#); [Rollins et al., 2015](#); [Rousset et al., 2012](#)). In the rate-strengthening approximation the afterslip velocity V is given by

$$V = 2V_0 \sinh \left[\frac{\Delta\tau}{(a-b)\sigma} \right] \quad (2.15)$$

where $\Delta\tau$ is the stress change after the mainshock, $(a-b)$ is the steady-state friction parameter and σ is the effective normal stress on the fault. We assume a uniform $(a-b) = 6$ MPa in our simulations and we explore values of V_0 that

best explain the geodetic observations in combination with other mechanisms of deformation (Figure 2.8).

FIGURE 2.8: Probability density of water content in the Wharton Basin asthenosphere from assimilation of the SuGAR time series of postseismic deformation and prior information from geochemical estimates. A) The bivariate probability density of water content and the afterslip parameter. There are little tradeoffs between the two parameters. B) Prior (black profile), marginal (blue profile) and conditional (dashed red profile) probability densities of the afterslip parameter. The prior distribution is $1/V_0$, i.e., the limit no prior information for Jeffrey's parameters (*Tarantola, 2004*). C) Prior information, marginal and conditional probability densities for water content in the Wharton Basin asthenosphere. The prior density is a log-normal distribution with a mean value of $600 \text{ H}/10^6 \text{ Si}$ and a standard deviation of 1.

The calculations are performed using the Relax software, which employs a spectral method to evaluate the quasi-static deformation caused by stress perturbations.

The method incorporates an equivalent body-force representation of dislocations that allows us to include detailed finite slip distributions. The equivalent body-force method is advantageous compared to other approaches such as finite-element methods because meshing of the domain around three-dimensional faults is automatic. The numerical approach has been validated using comparisons with analytic solutions for fault slip and comparisons with finite-element solutions in simple geometric settings (Barbot & Fialko, 2010a,b; Barbot et al., 2009). Another advantage of the Relax software and the equivalent body-force method is the ability to simulate several physical mechanisms of deformation simultaneously and to include nonlinear rheologies for plastic flow (Bruhat et al., 2011; Rollins et al., 2015; Rousset et al., 2012).

2.1.6 Limits of single-mechanism deformation models

Afterslip as a single mechanism of postseismic deformation predicts horizontal deformation compatible with the observation. However, the vertical displacement predicted by afterslip in the nearest stations is opposite to observations, suggesting another active process of postseismic deformation, such as viscoelastic relaxation in the asthenosphere (Figure 2.9).

Viscoelastic flow in the upper mantle can occur by diffusion creep, dislocation creep, or a combination of both (Bürgmann & Dresen, 2008). The physical properties of dislocation creep in olivine are well known from a wealth of laboratory

FIGURE 2.9: Simulation of the surface postseismic displacements after 1 year due to stress-driven afterslip on the Wharton Basin faults. The predicted horizontal displacements (white arrows) are aligned with the GPS observations (black arrows), but the modeled vertical displacements (background colours) is opposite to measured by GPS (coloured circles).

experiments documenting the effect of temperature, pressure, and water content (e.g., [G. Hirth & Kohlstedt, 2003](#); [S. Karato & Jung, 2003](#); [S. Karato et al., 1986](#); [Mei & Kohlstedt, 2000b](#)). The properties of diffusion creep of olivine are less well known because of its great sensitivity to grain size. Several field samples show that both diffusion creep and dislocation creep act in tandem to deform mylonite shear zones below the brittle-to-ductile transition ([Jin et al., 1998](#)). Experimental work suggests that dislocation creep of olivine is the dominant mechanism of

deformation in the upper-mantle conditions (e.g., Behr & Platt, 2014; S. Karato & Wu, 1993). But competition between grain growth during diffusion creep and grain size reduction by dislocation creep can promote comparable strain rate at these depths at equilibrium. In this case, dislocation creep should dominate post-seismic deformation following a large stress perturbation because of the power-law stress-strain rate dependence. As the stress is reduced by postseismic relaxation the contributions of both mechanisms should become similar again.

Following these considerations we first tested the potential of dislocation creep to explain the postseismic transient. Initially, we ignored the transient creep of olivine. We find that nonlinear viscoelastic deformation produces vertical displacements compatible with observations and horizontal displacements in the correct azimuth. However, the amplitude of horizontal displacements is lower than observed. We conclude that steady-state dislocation creep cannot explain the entire set of observations.

2.1.7 Inverse models of postseismic deformation

We use a Bayesian approach to estimate the water content in olivine, assimilating prior information from geochemical estimates and additional constraints from

geodetic observations. The posterior probability density is

$$\sigma_M(\mathbf{m}) = \nu \exp \left[-\frac{1}{2}(\mathbf{d}_{\text{obs}} - \mathbf{g}(\mathbf{m}))^T \mathbf{C}_D^{-1}(\mathbf{d}_{\text{obs}} - \mathbf{g}(\mathbf{m})) \right] \rho_M(\mathbf{m}), \quad (2.16)$$

where ν is a constant, $\mathbf{d}_{\text{obs}}$ is the GPS data vector, $\mathbf{g}(\mathbf{m})$ is the forward model based on parameters $\mathbf{m}$, $\mathbf{C}_D$ is the data covariance matrix and ρ_M is the prior information. Geochemistry provides limits on the water content in the mid-ocean ridge basalt source. We include these constraints on water content using a log-normal distribution (Figure 2.2A). For the afterslip parameter V_0 , we assume a uniform prior $1/V_0$, which corresponds to the limit of a log-normal distribution for infinite variance (Figure 2.8A). We assume that the geodetic data are independently and normally distributed. In principle, the joint probability density of all physical parameters can be estimated, but we limit our exploration to two physical parameters at a time to reduce the computational burden. In a first step, we jointly estimate the water content in olivine and the rate-strengthening parameters of afterslip, and assume all other parameters to be fixed. In a second step, we change the *in situ* parameters manually to a range of conditions.

We sample the posterior probability density using the Neighbourhood algorithm, a derivative free Markov Chain Monte Carlo method (Sambridge, 1999). The inversion tool, dubbed Relax-Miracle, uses Relax for the forward models. We benchmark the approach using a synthetic data set using the GPS network configuration of the Sumatra GPS Array. We create a forward model with a known water content

and afterslip parameter and we use this data set as an input data to our inversion scheme. The posterior probability density function not only recovers the target parameters but also informs us of inherent tradeoffs between the two relaxation mechanisms (Figure 2.10).

FIGURE 2.10: Benchmark of the Bayesian inversion method for water content in the Wharton Basin asthenosphere. We produce a synthetic data set of GPS time series at the SuGAR network (inset in A and black profile with uncertainties in B) assuming the afterslip parameter $V_0 = 3.0 \times 10^{-6}$ m/s, a water content of $C_{OH} = 1000 \text{ H}/10^6 \text{ Si}$, and no transient creep. C) The bivariate probability density for water content and the afterslip parameter has a maximum at the target value of the model parameters and documents the tradeoffs between model parameters. A) Conditional probability density along the red horizontal profile in C and marginal probability density of water content in the asthenosphere. D) Conditional probability density along the red vertical profile in C and marginal probability density of the afterslip parameter. In principle, the spatial distribution of the SuGAR GPS stations with 1 year of data provides an unbiased estimate of the water content in the asthenosphere.

We explore uniform water content C_{OH} in the asthenosphere in the range 0.0003 to 0.04 wt% (i.e., 50 to 6,000 H atoms per million Si atoms), mantle temperatures T_m from 1,350° C to 1,400° C, and reference velocities for afterslip in the range

$V_0 = 0 - 2.75 \mu\text{m}^{-1}$. We explore different values of the transient creep parameters A_K from $A_K = 0$ to $A_K = 3A$, and G_K from $G/2$ to $3G$, where G is the rigidity modulus. We assume all other physical and *in situ* parameters same for transient creep and steady-state creep. We assume that the laboratory-derived values for the constitutive parameters, which were carried out at much higher strain rate than the typical geological conditions, scale to natural conditions. Geodesy cannot independently constrain the water sensitivity of transient creep and the parameters A_K , so the product ($A_K C_{\text{OH}}^r$) for transient creep should be considered as a lump parameter. We also investigate a range of background strain rates for dislocation creep from 10^{-17} s^{-1} to $2 \times 10^{-14} \text{ s}^{-1}$. Our best fitting model in Figure 2.2 has $V_0 = 1.75 \times 10^{-6} \text{ ms}^{-1}$, $A_K = A/2$, and $G_K = G$. Incorporating transient creep affects the estimate of water content required in the mantle, as previously inferred in other studies (Freed et al., 2010; Pollitz, 2015), by lowering the water content required to fit the data (Figure 2.2). The inferred afterslip parameter V_0 is similar to what is found in other tectonic settings (Barbot et al., 2009; Wei et al., 2015).

We obtain the probability density of the water content in the asthenosphere by either marginalizing out the afterslip parameter

$$\sigma_{\text{marginal}}(m_1) = \int_0^\infty \sigma_M(\mathbf{m}) dm_2, \quad (2.17)$$

where m_1 is the water content in olivine and m_2 is the afterslip parameter; or taking the conditional probability density around the most likely value of the bivariate

distribution

$$\sigma_{\text{conditional}}(m_1) = \frac{\sigma_M(m_1, m_2)}{\int_0^\infty \sigma_M(m_1, m_2) dm_2}, \quad (2.18)$$

where m_2 is the most likely value of the afterslip parameter. The posterior probability distribution for water content and afterslip in the Wharton basin is shown in Figures 2.2 and 2.8.

The best-fit forward model is shown in Figure 2.2 and 2.11. The small misfit in the GPS time series may be due to our simplifying modeling assumptions, which ignore reactivation of the megathrust or the Sumatran fault and internal deformation of the accretionary prism.

FIGURE 2.11: Observed (blue squares with uncertainties) and modeled (red profiles) GPS time series of postseismic displacements after the 2012 Mw 8.6 Wharton Basin earthquake. The figure shows all the SuGAR time series available and considered in the study. Left) Horizontal displacements. Right) Vertical displacements. The model reproduces the subtle uplift of the forearc island stations in the northern hemisphere and the temporal evolution of displacements at most stations. The small misfit at some stations (Fig. 2d) can be attributed to the model simplifying assumptions such as no re-activation of the Sunda megathrust or the Sumatran fault and no internal deformation in the accretionary prism.

Chapter 3

Anisotropic high-temperature creep in hydrous olivine single crystals and its geodynamic implications*

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3.1 Abstract

We compile the experimental data on deformation of hydrated olivine single crystals to determine the influence of water and pressure. We analyze the data from low pressure ($P=0.1$ MPa and 0.1-0.3 GPa) high-resolution experiments as well as the results from higher pressures (2 to 6 GPa) containing larger uncertainties, using a flow law of the form $\dot{\epsilon}_i = \dot{\epsilon}_i^{dry} + \dot{\epsilon}_i^{wet}$ with (i :slip system). The data are normalized to a common stress (150 MPa) and temperature (1573 K) and we determine the parameters water content exponent (r) and activation volume (V^*). We find largely different values of r and V^* for different orientations (slip systems), implying that the influence of water and pressure is highly anisotropic, and provides largely different values of r and V^* . For the $[110]_c$ orientation where the $[100](010)$ slip system is activated, we obtain $r_{[100]}=0.35 \pm 0.08$, and $V_{[100]}^*=11.0 \pm 3.0$ cm³/mol, while for the $[011]_c$ orientation where the $[001](010)$ slip system is activated, $r_{[001]}=1.3 \pm 0.30$, and $V_{[001]}^*=5.6 \pm 3.6$ cm³/mol (for a fixed water content). The highly anisotropic effects of water and pressure suggest that creep in olivine is not controlled solely by diffusion but also controlled by the density of jogs (or kinks) on dislocations that is controlled by water and pressure. We find that the easier slip system changes from a-slip (slip along the $[100]$ direction) at low water content and low pressure to c-slip (slip along the $[001]$ direction) at high water content and high pressure, suggesting that c-slip is softer than a-slip in most of the asthenosphere, whereas the opposite is true in most of the lithosphere. Implications of these results are discussed using a model of deformation of a polycrystalline

aggregate made of anisotropic crystals. It is suggested that in the asthenosphere where most of plastic deformation occurs, short-term time-dependent small strain deformation associated with post-seismic and post-glacial deformation is dominated by c-slip and strongly dependent on the water content, whereas the long-term large strain deformation associated with mantle convection is dominated by a-slip that is only weakly dependent on water content.

3.2 Introduction

The effect of water (hydrogen) on the plastic deformation of olivine has been well appreciated (e.g., [Blacic, 1972](#); [S. Karato & Jung, 2003](#); [S. Karato et al., 1986](#); [Mackwell et al., 1985](#); [Mei & Kohlstedt, 2000a,b](#)). In the simplest model of high-temperature creep, the rate of deformation is directly controlled by diffusion of the slowest moving species (e.g., [S. Karato, 2008](#); [Weertman, 1968](#)). [D. L. Kohlstedt \(2006\)](#) adopted this model to explain water weakening in olivine. However, [Mackwell et al. \(1985\)](#) reported a high degree of anisotropy in plastic flow of hydrous olivine. These findings contradict the simplest version of a diffusion-controlled model because diffusion of Si and O in olivine is nearly isotropic ([Costa & Chakraborty, 2008](#)). A report by [Fei et al. \(2013\)](#) on the influence of water on Si diffusion in Mg_2SiO_4 is in marked contrast with other studies (e.g., [S. Karato & Jung, 2003](#); [S. Karato et al., 1986](#); [Mackwell et al., 1985](#); [Mei & Kohlstedt, 2000a,b](#)), possibly due to the anisotropic effect of water on olivine deformation, which is caused by factors other than diffusion.

Plastic anisotropy in hydrous olivine has never been characterized quantitatively. This is partly due to difficulties in the experimental study of the influence of water under high-pressure conditions (e.g., [S. Karato & Weidner, 2008](#)), but also due to the complications caused by the combined effects of water and pressure ([S. Karato & Jung, 2003](#)). In this paper, we focus on characterizing the anisotropic effect of water and pressure based largely on the experimental results on single crystals under hydrous conditions. In the following sections, we review the available laboratory data at low and high pressure that constrain plastic anisotropy of hydrous olivine. We characterize the potential errors associated with bringing all available results to a single standard of temperature and pressure conditions. Using these data we identify the flow law parameters individually for the two dominant slip systems. Finally, we discuss the implications for fabric transitions and the rheological properties of polycrystalline olivine aggregates.

3.3 Errors in the experimental study on plastic deformation

In order to characterize the water and pressure effect, it is important to analyze data that cover a broad range of water content and pressure conditions. As any single apparatus cannot cover the range of pressure conditions found in the lithosphere-asthenosphere system, results obtained using different apparatus and methods must be compared. In order to combine these data, we evaluate the uncertainties associated with different apparatus and sample preparation methods.

The characterization of these uncertainties will allow us to describe the activation volume and power exponent of water content for a broad range of pressure and water content, while mitigating potential biases from using data gathered in different laboratory settings.

3.3.1 Errors in stress, strain rate and temperature measurement

In this section, we summarize the uncertainties in each method with a focus on the measurements of macroscopic quantities (stress, strain-rate and temperature). Errors associated with characterizing the chemical environment (water content, oxygen fugacity, oxide activity) will be discussed in the next section.

The plastic deformation of single crystal minerals can be studied under room pressure ($P=0.1$ MPa), high temperature conditions (e.g., [D. L. Kohlstedt & Goetze, 1974](#)). The influence of water on plastic deformation can be studied at room pressure if a gas mixture with different water fugacities is used ([S. Karato, 1989](#); [Poumellec & Jaoul, 1984](#)). Under these conditions, the water fugacity is small and hence the water effect is small. However, because both stress and strain rate can be measured precisely under these conditions (errors in stress measurements are less than 0.1 MPa, errors in strain-rate measurements are less than 10%), these studies provide an important data set to understand the influence of water at low water fugacities (~ 10 kPa corresponding to the water content of 5×10^{-4} ppm wt).

The temperature in these experiments is estimated with an error of ± 5 K.

Results for higher water content can be obtained using a high-pressure deformation apparatus. Up to ~ 0.3 GPa pressure, a high-resolution gas-medium deformation apparatus (e.g., [Paterson, 1990](#)) can be used. Under these conditions, substantially larger effects of water can be observed because the water fugacity increases with pressure (at $P=0.3$ GPa, water fugacity is ~ 100 - 150 MPa at 1573 K for Ni-NiO oxygen fugacity buffer). With the Paterson apparatus where an internal load cell is used to measure the stress, one can obtain results with small errors (~ 1 MPa) and small errors in strain rate ($< 10\%$) ([Paterson, 1990](#)). Temperature in these experiments is estimated with an error of ± 10 K. Indeed, it is the study by [Mackwell et al. \(1985\)](#) on olivine single crystals using this apparatus in which a markedly different anisotropic water weakening effect was reported (see also [S. Karato, 1989](#)). Although the mechanical results by such a study are of high quality, the range of water content and pressure is so narrow that it is difficult to characterize the anisotropic flow laws applicable to a broad range of upper mantle condition from their data alone.

The Griggs-type solid medium apparatus is adequate at pressures from 0.5 to 3 GPa. With this type of apparatus, a conventional method of stress measurements using an external load cell has large errors ([Gleason & Tullis, 1993](#)), but with an improved method of stress measurements such as the use of dislocation density, one could minimize the errors in stress measurements to $\pm 10\%$ ($\pm \sim 20$ - 30 MPa)

(S. Karato & Jung, 2003). Blacic (1972) conducted experiments on single and polycrystalline olivine samples using a Griggs apparatus at high pressures (1.5 GPa). However these experiments were at low temperatures (< 1300 K) and corrections for friction were not applied. Therefore, the results by Blacic (1972) are not used in the present study. No studies have been conducted on olivine single crystals using a Griggs apparatus at high temperatures (> 1500 K).

Beyond ~ 3 GPa, those conventional deformation apparatus (a gas-medium deformation apparatus and the Griggs apparatus) cannot be operated. New types of deformation apparatus such as the rotational Drickamer apparatus RDA (Yamazaki & Karato, 2001) or deformation DIA (D-DIA) (Y. Wang et al., 2003) must be used (see also W. B. Durham et al., 2002). At these high pressures, much higher water fugacity can be realized, and the influence of pressure can be determined. In these experiments, stress is measured using X-ray diffraction and the errors in stress estimate are relatively large due to the complex relationship between the microscopic stress on each grains (that is measured by X-ray diffraction) and the macroscopic stress acting on the sample (Chen et al., 2006; S. Karato, 2009; Singh, 1993). The data we use are from Girard (2011) and Girard et al. (2013), in which the strength of a single crystal olivine was estimated from X-ray diffraction on polycrystalline Al_2O_3 using several diffraction planes. Each diffraction plane provides an estimate of strength, and the average strength is taken as an arithmetic average and the error is one standard deviation (Table 1). Temperature measurements in these experiments have large uncertainties caused by the temperature gradient; in

these cases, we estimate an error of ± 60 K.

3.3.2 Errors in characterizing the chemical environment

[Poumellec & Jaoul \(1984\)](#) conducted deformation experiments at atmospheric pressure using single crystal olivine on two different orientations: $[110]_c$ and $[101]_c$. They controlled the water fugacity and oxygen fugacity using different gas mixtures ($\text{H}_2/\text{H}_2\text{O}$ and CO/CO_2). The water effect was determined by comparing the results for the same temperature and oxygen fugacity but at different water fugacity. However, [Poumellec & Jaoul \(1984\)](#) did not measure the water content of their samples. Indeed, the water contents in their samples would be so small that one would not be able to measure them with the currently available methods. We estimate the water content in their samples based on the thermo-chemical environment in their experiments using the experimental results of water solubility by [Zhao et al. \(2004\)](#). In addition, the oxide activity was not controlled in their experiments. However, we can infer the oxide activity from their experimental procedure. In order to avoid chemical reaction, they inserted a platinum (Pt) foil between their samples (San Carlos olivine single crystals) and an Al_2O_3 piston. Platinum reacts with olivine to remove FeO, and therefore we assume that their sample was under silica-rich conditions equivalent (or close) to the opx buffer.

In an experimental study with a gas apparatus, the oxygen fugacity is controlled

by metal-metal oxide buffer (Fe-FeO, Ni-NiO). Oxide activity is opx buffered. Regarding the water content, we used the reported values of water content of samples determined by infrared absorption spectroscopy and also saturation values at 300 MPa (i.e., ~ 16 ppm wt). There is some evidence for water escape in some experiments (e.g., [Mackwell et al. \(1985\)](#)). However, using saturation values or the reported values does not affect the results significantly and we report the results of using the saturation values.

Other important data sets are those from [Girard \(2011\)](#) and [Girard et al. \(2013\)](#). These are deformation experiments on olivine single crystals at pressures of 4.7-6.1 GPa using a D-DIA apparatus. [Girard et al. \(2013\)](#) published the results of high pressure deformation experiments on olivine single crystal for the $[110]_c$ orientation. Results for the $[101]_c$ and $[011]_c$ orientations are also available from [Girard \(2011\)](#). These are important data because the pressure range is substantially higher than in other olivine single crystal studies.

However, there are several limitations on these data. First, water content during an experiment is not well constrained. They added water in the early stage of an experiment through the dehydration of talc. After dehydration at ~ 1100 K, the temperature was increased to 1473 K and deformation experiments were performed. The water content was measured only in the samples after the experiment. The water contents reported in their papers are less than the saturation limits under

those conditions. Consequently, some water loss occurred during these experiments. Therefore, we use the reported water content and the stress measurements at the later stage of deformation experiments in our analysis. In addition, the measured stress values in their experiments did not change by more than 20% with time (strain), which is much less than the stress uncertainties in these experiments. Therefore, this assumption will not have a large effect on the results of our analysis.

The oxygen fugacity is not well constrained in these high pressure experiments (4.7-6.1 GPa). They used a rhenium (Re) foil to separate the sample from alumina, but they also used a W-Re thermocouple close to the sample. Oxygen fugacity is likely between W-WO_{x/2} (x=4-6; [Cottrell et al. \(2009\)](#)) and Re-ReO₂ buffer. The oxygen fugacity corresponding to the Fe-FeO is in between them, and we consider that the uncertainty in oxygen fugacity is $\sim\pm 2$ orders of magnitude. Assuming $q\sim 0.2$ ([Keefner et al., 2011](#)), this leads to an uncertainty in strain-rate of a factor of $\sim\pm 2.5$. As for the oxide activity, their samples are likely controlled by the opx buffer because the samples were in contact with talc that produces pyroxene upon dehydration.

3.4 Method of analysis

In studying the deformation of olivine single crystals, three different orientations are used to activate different slip systems: $[110]_c$, $[101]_c$ and $[011]_c$ ($[110]_c$ means the orientation where normal stress is applied in a direction bisecting the $[100]$ and

the $[010]$ orientations). For the $[110]_c$ orientation, the $[100](010)$ slip system is activated, for the $[011]_c$ orientation, the $[001](010)$ slip system is activated, and for the $[101]_c$ orientation, both the $[100](001)$ and the $[001](100)$ slip systems are activated.

One of the most important factors to characterize a dislocation in a crystal is the Burgers vector, a displacement vector by which the periodicity of the underlying lattice structure is preserved. As the energy density of a dislocation line is proportional to s^2 , (s is the total slip), the length of the Burgers vector b controls the energetics of deformation (J. Hirth & Lothe, 1982, Ch.1). The length of the Burgers vector is different between $\mathbf{b}=[100]$ (0.48 nm) and $\mathbf{b}=[001]$, (0.63 nm). This leads to $\sim 50\%$ difference in dislocation energy. Therefore, it is important to compare the creep behaviour of different slip systems with different Burgers vectors. Because $[110]_c$ and $[011]_c$ orientations activate the slip systems whose slip plane is common ((010) plane) but slip directions are different ($[100]$ versus $[001]$), a comparison of creep properties of samples with $[110]_c$ and $[011]_c$ orientations provide insight into the plastic anisotropy and such anisotropic effects have also been observed in the dislocation recovery experiments (Yan, 1992). The $[101]_c$ orientation is expected to give the results in between because for this orientation, two slip systems involving $\mathbf{b}=[100]$ or $\mathbf{b}=[001]$ are activated (we call them a-slip and c-slip respectively).

We analyse these data assuming a flow law of the form

$$\dot{\epsilon}_i = A_i \sigma^{n_i} C_W^{r_i} f_{O_2}^{q_i} \exp\left(-\frac{E^* + pV_i^*}{RT}\right) + \dot{\epsilon}_i^{dry} \quad (3.1)$$

for each orientation, where $\dot{\epsilon}_i$ is the strain rate for the i -th orientation (such as $[110]_c$), $\dot{\epsilon}_i^{dry}$ is the strain rate for the i -th orientation under dry condition, A_i is the pre-exponential constant, σ is the stress, n_i is the stress exponent, E^* is the activation energy, R is the universal gas constant, T is the temperature, p is confining pressure, V^* is the activation volume, C_W is the water content, r_i is the water content exponent, f_{O_2} is the oxygen fugacity and q_i is the oxygen fugacity exponent. The correction for $\dot{\epsilon}_i^{dry}$ is important only for the room pressure experiments by [Poumellec & Jaoul \(1984\)](#) where the influence of water is small (this correction was made based on the data presented by [Poumellec & Jaoul \(1984\)](#)).

The pre-factor A_i includes the effect of oxide activity, but we consider only the opx buffered case, and treat this as a constant independent of thermodynamic conditions. The influence of oxygen fugacity can be large because for a given oxygen fugacity buffer, corresponding oxygen fugacity changes with pressure (and temperature) (e.g., [S. Karato, 2008](#), Ch.2). The influence of pressure (and temperature) on oxygen fugacity is included in our analysis.

Table 3.1 summarizes the experimental data used in this study. The table includes all the data including those from high-resolution modestly high-pressure results ([Mackwell et al., 1985](#); [Tielke et al., 2017](#)), room pressure ([Poumellec & Jaoul, 1984](#)), and high-pressure (exceeding 4 GPa) data ([Girard, 2011](#); [Girard et al., 2013](#)).

TABLE 3.1: Experimental conditions and results of previous studies used in our study. The influence of oxygen fugacity is corrected using the pressure dependence of oxygen fugacity. The results shown correspond to the oxygen fugacity buffered by Fe-FeO. The water content reported in this table is based on the (Paterson, 1982) calibration.

Source	Orientation	P ¹ (GPa)	Strain rate ² (10 ⁻⁵ s ⁻¹)	Stress (MPa)	T (K)	Water content ³ (ppm wt)
Poumellec & Jaoul (1984)	[110] _c	10 ⁻⁴	0.017	30 ±0.1	1564	5x10 ⁻⁴
Poumellec & Jaoul (1984)	[101] _c	10 ⁻⁴	0.018	43 ±0.1	1564	5x10 ⁻⁴
Mackwell et al. [1985]	[110] _c	0.3	1	58 ±1	1573	6.87
Mackwell et al. [1985]	[110] _c	0.3	10	141 ±1	1573	6.87
Mackwell et al. [1985]	[011] _c	0.3	1	135 ±1	1573	6.87
Mackwell et al. [1985]	[101] _c	0.3	1	48 ±1	1573	6.87
Girard et al. [2013]	[110] _c	4.7	3.06	541 ±100	1473	18.3 ±4.5
Girard et al. [2013]	[110] _c	4.8	1.28	390 ±140	1473	21.7 ±5.4
Girard et al. [2013]	[110] _c	6.1	2.4	280 ±100	1473	37.5 ±25
Girard et al. [2013]	[011] _c	4.7	1.74	541 ±100	1473	18.3 ±4.5
Girard et al. [2013]	[011] _c	6.1	6.1	280 ±130	1473	37.5 ±25
Girard [2011]	[101] _c	4.9	3.9	249 ±110	1473	18.1 ±6.25
Girard [2011]	[101] _c	4.9	2.42	292 ±80	1473	18.1 ±6.25
Tielke et al. [2017]	[101] _c	0.3	2.3	100 ±1	1473	12.5
Tielke et al. [2017]	[101] _c	0.3	6.2	122 ±1	1473	12.5
Tielke et al. [2017]	[101] _c	0.3	1.2	150 ±1	1473	12.5
Tielke et al. [2017]	[101] _c	0.3	2.3	204 ±1	1473	12.5
Tielke et al. [2017]	[101] _c	0.3	1.3	86 ±1	1473	12.5
Tielke et al. [2017]	[101] _c	0.3	5.6	133 ±1	1473	12.5
Tielke et al. [2017]	[101] _c	0.3	2.8	206 ±1	1473	12.5
Tielke et al. [2017]	[101] _c	0.3	4.4	250 ±1	1473	12.5
Mei & Kohlstedt (2000a)	All	0.3	6.7	120 ±1	1473	16.25
Mei & Kohlstedt (2000a)	All	0.1	2.74	90 ±1	1473	6.87

¹ For the gas-medium apparatus, the uncertainties in the pressure are assumed to be 0.5 MPa, and for high pressure (> 4 GPa) they are assumed to be 0.5 GPa.

² Calculation of errors in strain rate, including the influence of error propagation is given in the supplementary material.

³ Uncertainties in measurements such as water content are assumed to be 10% (when not given).

In order to determine water content sensitivity (“r”) and pressure sensitivity (“V*”), we normalize all the data to common temperature (1573 K), common oxygen fugacity (corresponding to that of the Fe-FeO buffer), common oxide activity (opx buffer) and a common stress (150 MPa). In doing this normalization, $E^*=470$ kJ/mol, $n=3$ and $q=0.2$ are assumed (e.g., Bai et al., 1991; Keefner et al., 2011; Mei & Kohlstedt, 2000b; Poumellec & Jaoul, 1984).

One limitation of the currently available data set is that there are only two direct deformation experiments for the $[011]_c$ orientation (Girard, 2011; Mackwell et al., 1985). Consequently, we add two constraints based on the experimental observations of polycrystalline olivine. First, we use the observations of fabric transitions (e.g., Jung & Karato, 2001; S. Karato, 2008) at $P=0.5\text{-}2$ GPa. These studies showed that fabric transitions from $[100]$ slip (A-/E-type fabric) to $[001]$ slip (C-type fabric) at a water content of ~ 80 ppm wt at $P \sim 2$ GPa ($T \sim 1500$ K). This implies that the creep strength for the a-slip ($[110]_c$) and the c-slip ($[011]_c$) should agree at this condition (e.g., Kaminski, 2002; Lister, 1979).

Second, we also use the results from Mei & Kohlstedt (2000b), who performed experiments on polycrystalline olivine at pressures from 0.1 to 0.45 GPa. In a polycrystalline olivine deformation experiment at steady-state, more than one slip system is active at any given time. Based on the results suggesting that the c-slip is the rate-controlling slip system at low pressures with a modest water content, we assume that their results are controlled by the c-slip. In order to use these data, we need to determine the proportionality constant between single crystal and polycrystal strength using the single crystal experimental data carried out at similar thermodynamic conditions (such as confining pressure). In this context we use single crystal experiments from Mackwell et al. (1985), which were under similar conditions to those of Mei & Kohlstedt (2000b), to determine the proportionality constant, and then use this constant to determine the data point corresponding to a confining pressure of 0.1 GPa. Using these additional constraints, we are able to

determine all the necessary parameters for the flow law for this orientation.

3.5 Data Analysis

Because the magnitude of errors is considerably different among different studies, we analysed the data using the weighted least squares method, where the weights are calculated as the reciprocal of the variance of each data point. The key in this step is the error estimate that includes the estimate of error in each data set as well as the errors caused by the normalization to a reference temperature and stress.

The errors in each measurement are discussed above. Now we estimate the errors associated with normalization. We normalize the results of various studies to a common condition of $T=1573$ K and stress=150 MPa (and oxygen fugacity of Fe-FeO buffer). In doing so, we assume a single stress exponent and activation energy. In reality, these parameters have some uncertainties. For example, $n=3\pm0.5$ and $E^*=470\pm50$ kJ/mol (Mackwell et al., 1985; Tielke et al., 2017). We take these errors in the parameters n and E^* and propagate the uncertainties during normalization (see Supplementary Material).

As all of the experimentally derived strain-rate, water content, and pressure have uncertainties, we estimate the water content stress exponent r and the activation volume V^* using a generalized least-squares method that account for errors in

TABLE 3.2: Flow laws for three different orientations corresponding to the Fe-FeO buffer. Water content sensitivity and activation volume estimated using the single crystal experimental data for different slip systems using weighted least squares (errors are for one standard deviation).

Orientation	$[110]_c$	$[101]_c$	$[011]_c$
Pre-exponential factor (A in $\text{s}^{-1} \text{MPa}^{-n} \text{ppm wt}^{-1}$)	$10^{4.88 \pm 0.03}$	$10^{5.54 \pm 0.01}$	$10^{2.63 \pm 0.24}$
Water content exponent (r)	0.35 ± 0.08	0.63 ± 0.06	1.32 ± 0.3
Activation volume (V^* in cm^3/mol)	11.0 ± 3.0	14.2 ± 3.9	5.63 ± 3.6

multiple axes (Ghilani & Wolf, 2006) and the weighted least squares method. The details and comparison of these methods are given in the Supplementary Materials. As can be seen in **Table 4.3**, different methods give essentially the same results. The difference between the mean values of water content exponent and activation volume is not too high when comparing the weighted least-square to that of generalized least-squares. However, the uncertainties are slightly higher in the case of generalized least-squares than the weighted least-squares, as expected. We conclude that the uncertainties in water content and pressure of fabric transition data point mentioned above is not too large and hence we report/discuss values from the weighted least-squares approach in the main paper. The parameters A , r , and V^* determined for $[110]_c$, $[101]_c$, $[011]_c$ are shown in **Table 3.2**, and the data and model predictions are shown in **Fig. 3.1** and **Fig. 3.2**.

3.6 Results

A marked difference in the values of flow law parameters among different orientations can be noted. Particularly important is a large difference in the water content exponent, r . For the $[110]_c$ orientation, $r=0.35 \pm 0.08$, whereas that for the $[011]_c$

FIGURE 3.1: Strain-rate versus water content relationships for three different orientations. Strain-rates are normalized to $T = 1573$ K and $P = 1$ GPa. Pressure effects are corrected using the results of the weighted least square fit to the data. Color of the symbol and the profile indicate different slip systems. Different symbols correspond to the different studies.

orientation, $r = 1.32 \pm 0.3$. For the $[101]_c$, $r = 0.63 \pm 0.06$. Since $\mathbf{b} = [100]$ dislocations are involved for the $[110]_c$ orientation, and $\mathbf{b} = [001]$ dislocations are involved for the $[011]_c$ orientation, we interpret that the creep involving $\mathbf{b} = [100]$ and creep involving $\mathbf{b} = [001]$ have different sensitivities to water content. The $[101]_c$ orientation involves both $\mathbf{b} = [100]$ and $\mathbf{b} = [001]$, and the estimated value of r for this orientation

FIGURE 3.2: Strain-rate versus pressure relationships for three different orientations. Strain-rates are normalized to $T = 1573$ K and $C_W = 1000$ ppm H/Si or 62.5 ppm wt. Water effects are corrected using the results of the weighted least square fit to the data. Color of the symbol and the profile indicate different slip systems. Different symbols correspond to the different studies.

is in between those for the $[110]_c$ and $[011]_c$ orientations. Similarly, the estimated activation volumes are different for creep involving different Burgers vectors. However, the activation volume for $[101]_c$ is not in between $[110]_c$ and $[011]_c$ as expected.

Based on these results, we propose that flow law of olivine single crystals can

be classified into two classes: one involving $\mathbf{b}=[100]$ dislocations, and another involving $\mathbf{b}=[001]$ dislocations. The results from the $[110]_c$ correspond to $\mathbf{b}=[100]$ dislocations, and the results from the $[011]_c$ to $\mathbf{b}=[001]$ dislocations, while the results from the $[101]_c$ are expected to be in between $[110]_c$ and $[011]_c$. The relative ease of deformation by these two classes of slip systems changes with various conditions, particularly with pressure and water content (**Fig. 3.2**). In the following, we will use this simplified model and discuss the relative contributions of these slip systems to plastic flow in Earth's upper mantle.

3.7 Discussions

3.7.1 Interpretation of the anisotropic water and pressure effects

The marked difference in the water content dependence for different slip systems supports the notion that high-temperature creep of olivine might not be controlled solely by diffusion, as some of the simple models suggest (e.g., [D. L. Kohlstedt, 2006](#); [Weertman, 1968, 1999](#)). [Yan \(1992\)](#) reported similar results, namely, anisotropy in dislocation recovery in olivine (stronger dependence of dislocation recovery on the water content for the c-slip than for the a-slip). In contrast, diffusion of Si (and O) in olivine is remarkably isotropic ([Costa & Chakraborty, 2008](#)).

Consequently, we consider an alternative model where dislocation creep is controlled by recovery with the recovery rate depending on the jog density as well as on the diffusion coefficient. In this model, when jog formation is difficult (due to high Peierls stress, for example), then the dislocation climb rate depends not only on the diffusion coefficient but also on the jog density viz., (e.g., [J. Hirth & Lothe, 1982](#); [S. Karato, 2008](#), Ch.17 & 9, respectively),

$$\dot{\epsilon} \propto D(T, P, C_W) \cdot c_j(T, P, C_W) \quad (3.2)$$

where D is the relevant diffusion coefficient and c_j is the jog density. Not only diffusion is water content dependent but also the jog density could be water content dependent if a jog is hydrated ([S. Karato, 2008](#), Ch.10). The energy of a jog and hence the water content dependence of jog density depends on the crystallographic nature of a dislocation (e.g., [J. Hirth & Lothe, 1982](#)), hence, the influence of water on jog density likely varies among different slip systems, leading to anisotropic plastic deformation.

The inferred small water content dependence for the $[110]_c$ orientation ($\mathbf{b}=[100]$ slip), similar to Si diffusion, suggests that creep by this slip system is directly controlled by diffusion. This implies that for this type of dislocation, jog formation is easy, and hence the dislocation line is saturated with jogs ([J. Hirth & Lothe, 1982](#)). In contrast, for slip systems involving $\mathbf{b}=[001]$ dislocations, jog formation energy is larger, and hence the dislocation line is not saturated with jogs. This

contrast in the water content dependency between the two slip systems is likely due to different jog densities between the two slip systems.

The jog density at thermal equilibrium is given by $c_j \propto \exp\left(-\frac{H_j^*}{RT}\right)$ where H_j^* is the jog formation energy, given by (e.g., [J. Hirth & Lothe, 1982](#), Ch.8)

$$H_j^* = \beta \cdot \frac{\mu}{1 - \nu} \cdot b^2 a \quad (3.3)$$

where b is the length of the Burgers vector, a is the length of the atomic distance in the direction normal to the glide plane, β is a non-dimensional constant that may contain a variable that depends on the property of a dislocation, μ is shear modulus and ν is Poisson's ratio. For the $[100](010)$ and $[001](010)$ slip systems, a is same but b is different, and the formation energy of a jog H_j^* is different by $\sim 50\%$: c_j is higher for the $[100](010)$ dislocations than for the $[001](010)$ dislocations.

The first-order conclusion from relations (3.2) and (3.3) is that the length of the Burgers vector has an important influence on the jog density. Given the difference in the Burgers vector between $\mathbf{b}=[100]$ (0.48 nm) and $\mathbf{b}=[001]$ (0.63 nm), we expect that the jog density for the $\mathbf{b}=[100]$ dislocations is much higher than that for the $\mathbf{b}=[001]$ dislocations. In such a case, it is possible that the $\mathbf{b}=[100]$ dislocations are saturated with jogs, whereas the $\mathbf{b}=[001]$ dislocations are under-saturated with jogs. This provides one possible explanation for the anisotropic effect of water and

pressure; there could be other models.

Jogs of a dislocation may react with hydrogen (e.g., [S. Karato, 2008](#); [S. Karato & Jung, 2003](#); [Mei & Kohlstedt, 2000b](#)). This will introduce additional water sensitivity such that the water fugacity sensitivity of creep will be given by ([S. Karato, 2008](#), Ch.10)

$$\dot{\epsilon} \propto f_{H_2O}^{r_{\text{diff}}+r_{\text{jog}}} \quad (3.4)$$

where f_{H_2O} is water fugacity and r_{diff} and r_{jog} are the water fugacity exponents for the diffusion coefficient and jog density, respectively. Using the empirical relation $C_W \propto f_{H_2O}$ (e.g., [D. Kohlstedt et al., 1996](#)), this relation can be translated to a relation between strain-rate and water content ($\dot{\epsilon} \propto C_W^{r_{\text{diff}}+r_{\text{jog}}}$). For the slip system involving $\mathbf{b}=[100]$ dislocations, $r_{\text{jog}} = 0$ if it is saturated with the jogs. But for the slip system involving $\mathbf{b}=[001]$ dislocations, dislocations are not saturated with jogs, and if jogs are hydrated, then $r_{\text{jogs}} > 0$. This explains the stronger sensitivity of the c-slip system to water content.

A similar explanation may be presented for the anisotropic pressure effect. From equation (2), one has

$$V_{\text{creep}}^* = V_{\text{diff}}^* + V_{\text{jog}}^*, \quad (3.5)$$

where V_{creep}^* , V_{diff}^* and V_{jog}^* are the activation volumes for creep, diffusion and jog formation respectively. For the slip system involving $\mathbf{b}=[100]$ dislocations, the pressure dependence of creep rate is the same as that of diffusion, $V_{\text{creep}}^* = V_{\text{diff}}^*$. In

contrast, for the slip system involving $\mathbf{b}=[001]$ dislocations, pressure dependence includes that of jog density, consequently, $V_{\text{creep}}^* = V_{\text{diff}}^* + V_{\text{jog}}^*$. There is no study on V_{jog}^* , but if we used the results of theoretical modeling on the pressure dependence of dislocation mobility (e.g., [Skelton, 2018](#)) as a guide, we may speculate that $V_{\text{jog}}^* < 0$ for the $\mathbf{b}=[001]$ dislocations, which explains our observations. However, we admit that this explanation is highly speculative because there is no direct calculation for jog formation energy in olivine as a function of pressure.

Anisotropy in kink formation (and migration) energy is an alternative model, but based on the marked anisotropy in dislocation recovery reported by [Yan \(1992\)](#), we prefer a model involving jogs.

3.7.2 Creep in olivine for different slip systems

Using the flow laws for the $[001](010)$ and the $[100](010)$ slip system(s) (a- and c-slip system), we calculate the viscosity-depth profiles for two slip systems (**Fig. 3.3**). We use the oceanic geotherm corresponding to the age of 60 Myrs where a cooling half-space model for temperature is assumed. All profiles show a marked minimum in viscosity at ~ 120 km. From this plot, we can see that the relative strengths of these two slip systems change with depth.

FIGURE 3.3: A plot of effective viscosity (for a fixed stress (10 kPa)) for for all three orientations. An oceanic geotherm corresponding to 60 Ma old oceanic plate with cooling half-space model for temperature is assumed (red profile)

To illustrate the transition in the softer slip system with depth more clearly, we calculated conditions where the strain rates by two slip systems ($\dot{\epsilon}_{[110]}$ (the a-slip) and $\dot{\epsilon}_{[011]}$ (the c-slip)) agree (**Fig. 3.4**). The results are shown in the pressure-water content space for a few assumed temperatures. We find that under low pressure and/or low water content, the $[100](010)$ slip system with Burgers vector $\mathbf{b}=[100]$ is softer (higher strain rate for a given stress) than the $[001](010)$ slip system with Burgers vector $\mathbf{b}=[001]$, whereas under high pressure and/or high water content, the $[001](010)$ slip system with Burgers vector $\mathbf{b}=[001]$ is softer (higher strain rate

for a given stress) than the $[100](010)$ slip system with Burgers vector $\mathbf{b}=[100]$. This is consistent with the results on the deformation of polycrystalline olivine based on the results from $P=0.1$ to 2 GPa showing that the water content exponent is ~ 1.2 , close to the water content exponent for the $[011]_c$ orientation ($\mathbf{b}=[001]$) (S. Karato & Jung, 2003). Similar results were also obtained using low pressure data by G. Hirth & Kohlstedt (2003). However, in the deeper regions of the upper mantle with high water content (e.g., the asthenosphere), the strong slip system likely changes to the one involving $\mathbf{b}=[100]$ (Fig. 3.3).

FIGURE 3.4: Diagram showing the dominant slip system as a function of water content and pressure with 95% confidence intervals at temperatures of 1273 K and 1673 K. The a-slip is the slip system that is activated for the $[110]_c$ orientation (with $\mathbf{b}=[100]$, (010) slip plane) and the c-slip is the slip system that is activated for the $[011]_c$ orientation (with $\mathbf{b}=[001]$, (010) slip plane).

3.7.3 Creep of a polycrystalline olivine and some geophysical implications

When there are multiple slip systems with largely different strengths, plastic deformation and deformation microstructures of a polycrystalline aggregate are connected to different slip systems in different ways for different properties. There have been a few experimental studies on the transient creep ([P. N. Chopra, 1997](#); [Hanson & Spetzler, 1994](#); [Thieme et al., 2018](#)) and steady-state creep of olivine aggregates (e.g., [S. Karato & Jung, 2003](#); [Mei & Kohlstedt, 2000b](#)), and some models have been proposed. However, here we discuss a model based on our results and observation.

Creep by the softest slip system and creep by the hardest slip system provide the upper and the lower limits of real strain-rate for an aggregate that contains crystals oriented in various directions; the actual strain-rate is in between these two limits. The creep behaviour of such a polycrystal evolves with strain caused by the evolving interaction between the two slip systems.

[M. Ashby & Duval \(1985\)](#) provided a theoretical model of deformation of a polycrystalline material that is made of highly anisotropic single crystals. Assuming homogeneous strain that includes elastic strain, they derived a flow law equation for a polycrystal in terms of the flow laws for the soft and hard slip systems that contains internal stress caused by the interaction between the soft- and hard-slip

systems. Because internal stress evolves with strain, deformation behaviour also evolves with strain. At low strain (comparable to or lower than the elastic strain), internal stress is small and deformation of a polycrystal is dominated by the soft slip system, whereas at high strain, internal stress becomes large and deformation is dominated by the hard slip system. Experimental studies on ice (highly anisotropic crystal) support these notions (e.g., [Duval et al., 1983](#); [Kocks, 1970](#)).

In a geophysical context, low strain deformation is relevant to most time-dependent deformation that is used to infer mantle viscosity, e.g., post-glacial rebound ([Peltier et al., 1980](#)) and post-seismic deformation ([Freed et al., 2010](#); [Masuti et al., 2016](#); [Qiu et al., 2018](#); [Sun et al., 2014](#)), where strain magnitude is comparable to or less than the elastic strain ([S. Karato, 1998](#)). In contrast, deformation associated with mantle convection is large. Consequently, in the former, the soft slip system dominates, whereas the hard slip system dominates in the latter (for more details of transient creep, refer to [S. Karato \(1998\)](#)).

Fig. 3.4 shows that in most of deep, water-rich regions, the a-slip system is stronger than the c-slip system. This change in A-type to C-type fabric through E-type can be further supported by seismological observation of the change in strength of radial anisotropy ([Visser et al., 2008](#)) (see also [S. Karato \(2008\)](#)). In such a case, steady-state creep is dominated by the a-slip system. For such a slip system, water content dependence is weak, and therefore steady-state creep strength of these regions is likely not very sensitive to water content. In contrast,

transient creep in these regions is highly sensitive to water content.

FIGURE 3.5: Change in viscosity upon dehydration. Dehydration due to partial melting will strengthen a rock (G. Hirth & Kohlstedt, 1996; S. Karato, 1986). A dehydration depth of 65 km is assumed where $T = 1473$ K corresponding to dehydration beneath the mid-ocean ridge. Initial water content is assumed to be 200 ppm wt. The slip system controlling the steady-state rheology is the one involving $\mathbf{b}=[100]$ at the beginning. Consequently, the initial stage of hardening due to dehydration is limited. However, as the water content is reduced, the dominant slip system changes to the one involving $\mathbf{b}=[001]$, and more substantial weakening will occur.

Note that, even though steady-state creep in the asthenosphere may not be sensitive to water content, the dehydration that occurs beneath the mid-ocean ridge will strengthen a rock substantially. This is because as one removes water, the rate-controlling slip system for the steady-state creep will change from the a-slip system

to the c-slip system, and the net change in the strength will be substantially higher than a case where the a-slip system alone controls the creep strength (**Fig. 3.5**).

3.7.4 Implications for the fabric transitions in the deep upper mantle

Deformation fabrics such as the lattice-preferred orientation (LPO) of minerals is controlled by the weakest slip system (e.g., [S. Karato, 2008](#); [Lister, 1979](#)). Some studies have shown that the LPO changing with change water content under low pressure conditions ([Jung & Karato, 2001](#)), while others have suggested that it is pressure induced under dry conditions at the pressure above 3 GPa ([Jung et al., 2009](#); [Raterron et al., 2009](#)). However, from our results we suggest that the weakest slip system in olivine changes with water content and pressure. Therefore, the deformation fabric (LPO) of olivine will change with both water content and pressure.

Up to ~ 3 GPa (~ 100 km depth), pressure effects are minor and water effects dominate. The present results are consistent with the fabric transitions in olivine reported by [S. Karato et al. \(2008\)](#) summarizing the experimental results below $P=2$ GPa. Using our new results, we further examine the likely LPO in the deeper regions down to ~ 200 km ($P=6$ GPa). An inspection of **Fig. 3.3** shows that down to these pressure conditions, high water content corresponds to the olivine LPO controlled by c-slip (e.g., the C-type or the B-type fabric), whereas low water content corresponds to a-slip (e.g., the A-type or the E-type fabric). The reported

trend from the A-type to the C-type fabric transition through the E-type fabric at relatively low stresses is consistent with the present results. We conclude that the results summarized by [S. Karato et al. \(2008\)](#) are valid to $P \sim 6$ GPa (~ 200 km).

Applicability of our results to higher pressures is not clear. Similar to our results, there are reports showing that the activation volume for a-slip is larger than for c-slip under dry (or water-poor) conditions ([Girard, 2011](#); [Raterron et al., 2007](#)). These results suggest that, at low water content, c-slip becomes easier than a-slip at pressures higher than ~ 6 -8 GPa at the laboratory strain rate. If one were to apply our results showing larger water content sensitivity for c-slip than for a-slip, then one would expect that at pressures of 7-11 GPa and at high water content, c-slip is easier than a-slip. This inference does not agree with the results of [Ohuchi & Irifune \(2013\)](#), who reported that the A-type fabric (due to a-slip) dominates under deep upper mantle conditions ($P=7$ -11 GPa) with high water content. We do not know how to explain this discrepancy. Further experimental studies are needed to solve this puzzle.

3.8 Summary and concluding remarks

We analysed the experimental data on plastic deformation of olivine single crystals with the presence of water (hydrogen). The data include those obtained at pressures of 0.1 MPa to 6 GPa. Assuming that the stress exponent and activation energy are the same for all slip systems, we determined the water content and

pressure dependence of flow laws for three different orientations assuming the flow law of the form, $\dot{\varepsilon}_i^{wet} \propto C_W^{r_i} \exp\left(-\frac{PV_i^*}{RT}\right)$. We show a marked difference in these parameters (r_i , V_i^*) among different orientations involving different slip systems. The slip system involving a-slip has a weak water content dependence ($r=0.35$) and strong pressure dependence ($V^*=11$ cm³/mol), while the slip system involving c-slip has a strong water content dependence ($r=1.3$) and weak pressure dependence ($V^*=5.6$ cm³/mol).

The water content exponent that we determine is consistent with the lab results at relatively low pressures showing $r \sim 1.2$ (e.g., [S. Karato & Jung, 2003](#)) where the [001](010) slip controls the rate of steady-state creep. Similarly, our conclusion is consistent with those obtained from low pressure data, suggesting $r \sim 0.2-0.3$ (e.g., [Poumellec & Jaoul, 1984](#)) in single crystals where the a-slip system dominates. However, the estimates of activation volume contain large errors. This is due to the large experimental errors in high-pressure studies that are currently available.

We must emphasize that our analysis contains a few assumptions. One of them is the assumption that the sensitivity of deformation to temperature, stress exponent and oxygen fugacity is the same for all slip systems. This is an over-simplification. If deformation with the c-slip system involves jog formation but that of the a-slip system does not, then the activation energy for the former will be somewhat higher than the latter. However, this will cause no major change as far as we consider deformation at high temperatures close to the normalized temperature ($T=1573$

K) that we used. Similarly, the influence of the different oxygen fugacity is not large because the influence of oxygen fugacity is modest (exponent is ~ 0.2). Potentially important is the possible difference in the stress exponent. There are some observations suggesting that the stress exponent for the $\mathbf{b}=[001]$ slip system is larger than that for the $\mathbf{b}=[100]$ slip system (e.g., [Goetze, 1978](#)). Since we need to extrapolate stress (to a lower level) to apply the lab data to Earth, this has a large effect. Indeed, based on the results summarized by [Goetze & Poirier \(1978\)](#) and [S. Karato et al. \(2008\)](#), the stress exponent for the a-slip system should be smaller than that of the c-slip system. If this is the case, a comparison of the relative strength of these slip systems at a geological stress level needs to be made with great care.

Implications of our results for deformation of a polycrystalline aggregate of olivine are discussed based on a theoretical model developed by [M. Ashby & Duval \(1985\)](#). This model predicts evolving flow behaviour: small strain deformation dominated by the soft slip system but large strain deformation by the hard slip system. This is a general behaviour of deformation of a polycrystalline aggregate as far as homogeneous strain is assumed, but it is confirmed only for ice where anisotropy is large. Similar tests for olivine and other mantle materials will be important.

Our results predict that water content sensitivity will be different between steady-state creep and transient creep, and will change with depth (under the assumption that the weakest slip system controls the transient creep and strongest controls

the steady-state creep). Our results also explain the apparent discrepancy in the water sensitivity between creep and diffusion. However, the applicability of our results to pressures exceeding 6 GPa remains to be tested. Also important is the possible difference in stress exponent and the activation enthalpy among different slip systems. High-resolution deformation experiments under high pressures with well-constrained chemical environment are needed to address these issues.

3.9 Supplementary Information

Here, we give additional details related to numerical method used and jog density. Dislocation creep of olivine aggregates is controlled by dislocation climb. The dislocation climb rate depends on the diffusion coefficient (D) and jog velocity (v_j). Jog and climb of dislocation are shown in Fig. 3.6.

FIGURE 3.6: A simple cartoon showing jog, glide plane, and climb direction.

We perform the data analysis in two steps. (1) We normalize the data to a common stress and a common temperature. This normalization propagates errors. (2) After normalization, we invert the data to determine the most likely values of water content exponent (r) and activation volume (V^*). In this second step, we impose different errors for the different data sets. The errors in the data used in inferring r and V^* include (i) errors in individual measurements and (ii) errors caused by normalization. We use “weighted least squares” and “generalized least squares” techniques to perform this inversion. Some details on these two steps are explained in this Supplementary Information.

3.9.1 Error propagation

Here we describe the preparation of errors for least-square analysis. In the next section, the analysis of data using different methods is explained. We normalize the data to a common stress (150 MPa) and temperature (1573 K) assuming a stress exponent (3 ± 0.5) and activation energy (470 ± 50 kJ/mol) that have quoted uncertainties. The uncertainties in the stress measurement and other sources of errors need to be propagated, and in order to do that we use the general error propagation equation for uncorrelated variables

$$\sigma_y^2 = \sigma_u^2 \left(\frac{\partial y}{\partial u} \right)^2 + \sigma_v^2 \left(\frac{\partial y}{\partial v} \right)^2 + \dots \quad (3.6)$$

where σ^2 is the variance, y is a function of u and v , and u and v are independent variables. We apply the above formula to propagate the errors in stress measurement, stress exponent, temperature measurement, and activation energy to strain rate.

3.9.1.1 The error propagation caused by the errors in stress measurement and stress normalization

The relation between stress and strain-rate is (in the dislocation creep regime)

$$\dot{\epsilon} \propto \tau^n, \quad (3.7)$$

where $\dot{\varepsilon}$ is the strain-rate, τ is the stress and n is the stress exponent (note that in order to distinguish stress from the variance, we use τ for stress). The errors in strain-rate are caused by errors in stress measurement (σ_τ) and the stress exponent (σ_n). Differentiating the above equation with respect to τ and applying error propagation,

$$\frac{\sigma_{\dot{\varepsilon}}^2}{\dot{\varepsilon}^2} = \frac{n^2 \sigma_\tau^2}{\tau^2}. \quad (3.8)$$

Similarly, for the stress exponent

$$\dot{\varepsilon}_2 = \dot{\varepsilon}_1 \left(\frac{\tau_2}{\tau_1} \right)^n. \quad (3.9)$$

Taking the logarithm, differentiating and using the error propagation equation, we obtain,

$$\frac{\sigma_{\dot{\varepsilon}_2}^2}{\dot{\varepsilon}_2^2} = \sigma_n^2 [\ln(\tau_2) - \ln(\tau_1)]^2. \quad (3.10)$$

The net error in strain rate due to errors in stress measurement and assumed stress exponent is the sum of (3.8) and (3.10).

3.9.1.2 The error propagation caused by the errors in temperature measurement and temperature normalization

The relation between temperature and the strain rate is (in the dislocation creep regime)

$$\dot{\varepsilon} \propto \exp \left(-\frac{E^* + pV^*}{RT} \right), \quad (3.11)$$

where E^* is the activation energy, V^* is the activation volume, p is the pressure, and R is the universal gas constant. Differentiating the above equation with respect to T and applying the error propagation,

$$\frac{\sigma_{\dot{\varepsilon}}^2}{\dot{\varepsilon}^2} = \frac{\sigma_T^2}{T^4} \left(\frac{E^* + pV^*}{R} \right)^2 \quad (3.12)$$

The strain-rate is normalized for a temperature of 1573 K assuming an activation energy of 470 kJ/mol. However, uncertainty in the activation energy (assumed 50 kJ/mol) introduces errors in the data where the experiments were not conducted at 1573 K. To correct this error,

$$\dot{\varepsilon}_2 = \dot{\varepsilon}_1 \frac{\exp(-E^* + pV^*/RT_2)}{\exp(-E^* + pV^*/RT_1)}. \quad (3.13)$$

Using the error propagation, we have

$$\frac{\sigma_{\dot{\varepsilon}_2}^2}{\dot{\varepsilon}_2^2} = \left(\frac{\sigma_E}{R} \right)^2 \left(\frac{1}{T_1} - \frac{1}{T_2} \right)^2 \quad (3.14)$$

So, the total error in strain-rate is the sum of these errors (i.e., equations (3.8), (3.10), (3.12) and (3.14)). The data after normalizing and applying error propagation is shown in Table 3.6. Since the activation volume (V^*) is a model parameter, we do not know its value. We assume an initial activation volume of 10 cm³/mol, calculate the error in strain-rate, and perform the weighted least squares analysis. In order to get the best fit value, we iterate the above steps (i.e., calculation of

strain rate and perform data analysis) until the weighted mean square error becomes a constant. The algorithm converges within 2-3 iterations. However, the inclusion of the activation volume in the above error propagation and iterating has a small effect on the final value of the activation volume and its uncertainty, compared to not including it in the analysis.

3.9.2 Data Analysis

We assume the following equation for the strain-rate versus T (temperature), p (pressure), and water content (C_W),

$$\dot{\varepsilon} = A' C_W^r \exp\left(-\frac{pV^*}{RT}\right) \quad (3.15)$$

where $A' = A \tau^n \exp\left(\frac{-E^*}{RT}\right)$, $\dot{\varepsilon}$ is the strain-rate, V^* is the activation volume, A is the pre-exponential factor, τ is the differential stress, n is the stress exponent, E^* is the activation energy, R is the universal gas constant, and T is the temperature. Linearizing the above equation,

$$\ln(\dot{\varepsilon}) = \ln(A') + r \ln(C_W) - \frac{pV^*}{RT} \quad (3.16)$$

where $\dot{\varepsilon}$, C_W , p , and T are measured in the experiment, and other variables A' , r , V^* are the unknowns. We analyze the data using two different methods viz., weighted least squares and generalized least squares.

3.9.2.1 Weighted least squares

In the weighted least squares case, only errors in the dependent variable are considered, i.e., the weight matrix is a diagonal matrix with reciprocal of variance as their element $w_i = 1/\sigma_i^2$,

$$\mathbf{W} = \begin{pmatrix} w_1 & 0 & \dots & 0 \\ 0 & w_2 & \dots & 0 \\ \vdots & \vdots & \ddots & \vdots \\ 0 & 0 & \dots & w_n \end{pmatrix}. \quad (3.17)$$

The least squares solution is,

$$\begin{aligned} \mathbf{m} &= [\mathbf{G}^T \mathbf{W} \mathbf{G}]^{-1} \mathbf{G}^T \mathbf{W} \mathbf{d}_{\text{obs}} \\ \tilde{\mathbf{C}}_{\mathbf{M}} &= \sigma^2 [\mathbf{G}^T \mathbf{W} \mathbf{G}]^{-1} \end{aligned} \quad (3.18)$$

where $\mathbf{m}$ is a vector of model parameters, $\mathbf{G}$ is the design matrix that contains the partial derivatives with respect to model parameters, $\mathbf{d}_{\text{obs}}$ is a vector of the observed data (i.e., strain rate), and $\tilde{\mathbf{C}}_{\mathbf{M}}$ is a posterior covariance matrix where diagonal values are the variances of the model parameters. Table 3.3 contains the correlation matrix for all the slip systems using weighted least squares. The correlation coefficient between A and r is very high suggesting that within the standard deviation there are other values which are equally likely to give the same fit to the data.

TABLE 3.3: Correlation matrix for all slip systems using weighted least squares

Slip system	$[110]_c$			$[101]_c$			$[011]_c$		
	A	r	V^*	A	r	V^*	A	r	V^*
A	1	-0.56	0.27	1	-0.66	0.37	1	-0.96	-0.61
r	-0.56	1	0.16	-0.66	1	0.17	-0.96	1	0.73
V	0.27	0.16	1	0.37	0.17	1	-0.61	0.73	1

3.9.2.2 Generalized least squares

For generalized least squares, errors in both the dependent and the independent variables are considered. The multiplication of the unknown coefficient with the error in the independent variable makes this a nonlinear problem that has to be solved using an iterative algorithm. The general form of the equation we are talking about is,

$$(y + e_y) = a(x + e_x) + b(z + e_z) + c \quad (3.19)$$

where y is the dependent variable, x and z are independent variables with uncertainties e_y , e_x , and e_z , respectively, and a, b , and c are the unknown coefficients. The solution to such a problem is obtained by first order Taylor expansion and is described in Chapter 22 of Ghilani & Wolf (2006). For our problem, instead of y we have strain rate ($\dot{\epsilon}$), water content (C_W) instead of x , and pressure (p) instead of z , and their measurement uncertainties. The results of using both these methods are shown in Table 3.4. Correlation coefficients using this method are shown in Table 3.5.

TABLE 3.4: Comparison of flow law parameters and their standard deviations estimated from weighted least squares and generalized least squares. Units of A are the same as Table 3.2.

Orientation	Weighted least squares			Generalized least squares		
	A	r	V*	A	r	V*
$[110]_c$	$10^{4.90 \pm 0.03}$	0.35 ± 0.08	11.0 ± 3.0	$10^{4.9 \pm 0.1}$	0.35 ± 0.08	11.0 ± 3.1
$[101]_c$	$10^{5.50 \pm 0.03}$	0.63 ± 0.06	14.0 ± 3.9	$10^{5.5 \pm 0.08}$	0.64 ± 0.06	14.2 ± 4.1
$[011]_c$	$10^{2.65 \pm 0.24}$	1.32 ± 0.3	5.63 ± 3.6	$10^{2.53 \pm 0.37}$	1.43 ± 0.35	5.8 ± 3.8

TABLE 3.5: Correlation matrix for all slip systems using the generalized least squares.

Slip system	$[110]_c$			$[101]_c$			$[011]_c$		
	A	r	V*	A	r	V*	A	r	V*
A	1	-0.56	0.27	1	-0.64	0.38	1	-0.96	-0.56
r	-0.51	1	0.16	-0.64	1	0.17	-0.96	1	0.67
V*	0.27	0.16	1	0.17	0.38	1	-0.56	0.67	1

3.9.2.3 Dominant slip system

To determine which slip system might dominate under different water content and pressure, we calculate the conditions under which the strain rates of the two slip systems are equal (Fig. 3.4). In a low pressure and low water content case, a-slip dominates, whereas under high pressure and high water content conditions, c-slip dominates. We use the flow law parameters shown in Table 3.2 and different pressures under which the experiments were carried out. We have,

$$\dot{\epsilon}_{[110]} = \dot{\epsilon}_{[011]}$$

$$A_{[110]} (\sigma_{[110]})^n C_W^{r_{[110]}} \exp\left(-\frac{E^* + pV_{[110]}^*}{RT}\right) = A_{[011]} (\sigma_{[011]})^n C_W^{r_{[011]}} \exp\left(-\frac{E^* + pV_{[011]}^*}{RT}\right) \quad (3.20)$$

Rerranging and solving for C_W gives,

$$C_W = \left[\left(\frac{A_{[011]}}{A_{[110]}} \right) \left(\frac{\sigma_{[011]}}{\sigma_{[110]}} \right)^n \exp\left(\frac{p\{V_{[110]}^* - V_{[011]}^*\}}{RT}\right) \right]^{\left(\frac{1}{r_{[110]} - r_{[011]}}\right)} \quad (3.21)$$

TABLE 3.6: Strain rate normalized to a common stress of 150 MPa, temperature of 1573 K, with errors from different experiments. The errors in the last column are computed as the sum of equations (3.8), (3.10), (3.12) and (3.14)

Source	Orientation	P (GPa)	Normalized $\dot{\epsilon}$	Error in $\dot{\epsilon}$
(Poumellec and Jaoul 1984)	[110]c	10^{-4}	4.50E-06	3.50E-06
(Poumellec and Jaoul 1984)	[101]c	10^{-4}	2.18E-06	1.36E-06
(Mackwell et al., 1985)	[110]c	0.3	2.70E-04	1.61E-04
(Mackwell et al., 1985)	[110]c	0.3	1.20E-04	2.80E-05
(Mackwell et al., 1985)	[011]c	0.3	1.37E-05	3.25E-06
(Mackwell et al., 1985)	[101]c	0.3	3.05E-04	1.88E-04
(Girard et al., 2013)	[110]c	4.7	1.19E-06	2.13E-06
(Girard et al., 2013)	[110]c	4.8	1.38E-06	2.73E-06
(Girard et al., 2013)	[110]c	6.1	6.71E-06	1.30E-05
(Girard et al., 2013)	[011]c	4.7	6.74E-07	1.21E-06
(Girard et al., 2013)	[011]c	6.1	1.70E-05	3.31E-05
(Girard et al., 2013)	[101]c	4.9	1.55E-05	3.23E-05
(Girard et al., 2013)	[101]c	4.9	5.96E-06	1.08E-05
(Tielke et al., 2017)	[101]c	0.3	8.90E-04	2.98E-04
(Tielke et al., 2017)	[101]c	0.3	1.32E-03	3.78E-04
(Tielke et al., 2017)	[101]c	0.3	1.38E-03	3.67E-04
(Tielke et al., 2017)	[101]c	0.3	1.05E-03	3.23E-04
(Tielke et al., 2017)	[101]c	0.3	7.91E-04	3.05E-04
(Tielke et al., 2017)	[101]c	0.3	9.21E-04	2.52E-04
(Tielke et al., 2017)	[101]c	0.3	1.24E-03	3.85E-04
(Tielke et al., 2017)	[101]c	0.3	1.09E-03	4.03E-04
(Mei & Kohlstedt., 2000)	All	0.3	1.09E-04	1.62E-05
(Mei & Kohlstedt., 2000)	All	0.1	4.11E-06	1.19E-06

The above equation can be used to calculate the water content for the pressures at which the experiments were conducted; this data is plotted in Figure 3.4 as black circles. The above equation can be linearized and regressed to fit the line; here a-slip and c-slip are equally dominant (blue and red lines in Figure 3.4).

Chapter 4

A constitutive law for the transient creep of dry olivine aggregates

Abstract: The rheology of the upper mantle impacts a variety of geodynamic processes, including postseismic deformation following great earthquakes and post-glacial rebound. The deformation of upper mantle rocks is controlled by the rheology of olivine, which has well constrained steady-state properties. However, the physical mechanism underlying transient creep, an evolutionary, hardening phase converging asymptotically to steady-state, is still poorly understood. Here, we use mechanical data from laboratory experiments performed on dunites under anhydrous conditions, a nonlinear Burgers model, and a Markov Chain Monte Carlo method to constrain the transient creep parameters. We find different properties for the transient and steady-state components of dunite, with activation energies of $E_K = 430 \pm 20 \text{ kJ/mol}$ and $E_M = 535 \pm 15 \text{ kJ/mol}$, and stress exponents $n_K = 2.0 \pm 0.1$ and $n_M = 3.6 \pm 0.1$, respectively. These differences suggest that the physical mechanism controlling transient creep differs from steady-state creep

of olivine aggregates. The lower activation energy of transient creep could be due to a higher jog density of the corresponding soft-slip system.

4.1 Introduction

Since the start of the 21st century, multiple faults around the world have ruptured in great earthquakes, for example, the 2004 M_w 9.1 Aceh-Andaman earthquake, the 2005 M_w 8.6 Nias-Simeulue earthquake, the 2010 M_w 8.8 Chile earthquake, the 2011 M_w 9.0 Tohoku-Oki earthquake, and the 2012 M_w 8.6 Indian Ocean earthquake. These earthquakes and their postseismic observations were captured by dense geodetic networks, providing an opportunity to study the underlying physical mechanisms. Most postseismic deformation studies of these earthquakes have shown that including viscoelastic relaxation of the upper mantle is important to explain the observations (Klein et al., 2016; Pollitz et al., 2006; Sun et al., 2014). While these studies included only steady-state creep into their models, recently, the steady-state of the upper mantle has been shown to be preceded by transient creep (e.g., Freed et al., 2012; Masuti et al., 2016; Qiu et al., 2018; Tang et al., 2019; Weiss et al., 2019).

Understanding the transient rheology of the upper mantle is essential to accurately model geodynamic processes such as postglacial rebound (e.g., S. Karato, 1998; Peltier et al., 1980) and postseismic deformation (e.g., Freed et al., 2012; Masuti et al., 2016; Qiu et al., 2018; Tang et al., 2019; Weiss et al., 2019). However, our current understanding of the rheology of the upper mantle comes mainly from rock

deformation studies in the steady-state creep regime (e.g., [Carter & Ave'Lallemant, 1970](#); [P. Chopra & Paterson, 1981, 1984](#); [S. Karato et al., 1986](#); [S. Karato & Wu, 1993](#); [Mei & Kohlstedt, 2000a](#)), whose flow law parameters are well defined (e.g., [G. Hirth & Kohlstedt, 2003](#); [S. Karato & Jung, 2003](#)).

In contrast, only a handful of studies have analyzed data in the transient creep regime ([P. N. Chopra, 1997](#); [Hanson & Spetzler, 1994](#); [R. Post, 1977](#); [Thieme et al., 2018](#)). [R. Post \(1977\)](#) and [P. N. Chopra \(1997\)](#) performed experiments on dunites under constant stress conditions using a solid-medium and a gas-medium apparatus, respectively, and concluded that transient creep and steady-state creep are controlled by similar physical mechanisms. [P. N. Chopra \(1997\)](#) analyzed the stress vs. time data using a linear Burgers rheology, and suggested that the ratio r of transient to steady-state viscosity is 0.17 to 0.67, noting that it was slightly higher than the value used in geophysical modeling of ~ 0.1 (e.g., [Sabadini et al., 1985](#); [Yuen et al., 1986](#)). Since then, different values of r ranging from 0.01 to 0.4 have been used in postseismic studies (e.g., [Hoechner et al., 2011](#); [Pollitz, 2005](#); [K. Wang et al., 2012](#)).

[Hanson & Spetzler \(1994\)](#) conducted experiments on olivine single crystals and argued that transient creep could be anisotropic, as they observed strain softening instead of strain hardening in the case of the $[110]_c$ orientation. [Hanson & Spetzler \(1994\)](#) and [P. N. Chopra \(1997\)](#) found that the dislocation density evolves during the transition from transient creep to steady-state creep, in agreement with theoretical models that assume an evolving internal microstructure (e.g., [S. Karato, 2008](#), Ch. 3). [Thieme et al. \(2018\)](#) conducted a detailed study of the transient creep

of olivine aggregates under anhydrous conditions at a high temperature (1473 K) and 300 MPa of confining pressure using a gas-medium apparatus. Dislocation microstructures were analyzed at different cumulative strain in the transient regime, but evidence for a systematic evolution of the microstructure was inconclusive.

Microstructure evolution during transient creep can be modeled with a flow law that involves the evolution of a state parameter (e.g., [Masuti et al., 2016](#); [Sherburn et al., 2011](#)). The formulation proposed by [Masuti et al. \(2016\)](#) uses the concept of internal stress and considers the anelastic strain (i.e., thermodynamically irreversible) from the steady-state and transient creep to be additive. [Sherburn et al. \(2011\)](#) employ a formulation derived from viscoplasticity theory ([Nemat-Nasser, 2004](#); [Nemat-Nasser & Hori, 1999](#)), using isotropic and kinematic hardening. Since transient creep corresponds to an initial rapid displacement followed by a slower response, work hardening is appropriate. However, the formulation of [Sherburn et al. \(2011\)](#) contains a large number of free parameters that cannot be fully constrained with available laboratory data. We therefore consider a simpler formulation based on the Burgers rheology, but modified with a nonlinear stress/strain-rate relationship in the Kelvin and Maxwell elements. We use mechanical data from laboratory experiments to determine the flow law parameters of dunite, a rock assembly made of more than 94% olivine, providing a consistent treatment of transient creep and steady-state creep.

FIGURE 4.1: Schematic of the Burgers model and ideal stress vs. strain curve for olivine aggregates. a) The Maxwell element has a dashpot and the spring in series. The spring in the Maxwell element represents the elastic strain, where G_M is the Young's modulus, and the dashpot represents the long-term inelastic strain. The Kelvin element has the spring in parallel with the dashpot, and represents the transient inelastic strain, where G_K is the hardening coefficient. b) Under the constant strain rate assumption, the ideal response of olivine aggregates. It will have an instantaneous elastic strain, followed by transient and steady-state creep. When the strain rate is lowered to new constant value, similar response follows. c) Strain rates during the constant strain-rate experiments.

4.2 Experimental data

To determine the flow law parameters of the transient creep, we can use data from experiments conducted either under constant stress or constant strain-rate conditions. Due to limited availability of high-temperature and high-resolution

data under constant stress conditions, we use data mainly from constant strain-rate experiments, specifically from [P. Chopra & Paterson \(1984\)](#) as summarized in Table 4.3. In their study, experiments were carried out on two types of dunites (Åheim, of 900 μm grain size and Anita Bay, of 100 μm grain size) under dry and wet conditions at different temperatures (1273 K to 1573 K) using a gas-medium apparatus. In this paper, we use data from experiments conducted on both Åheim and Anita Bay dunites under dry conditions and at high-temperatures (> 1473 K).

Since no measurement uncertainty was reported by [P. Chopra & Paterson \(1984\)](#), we assume a uniform uncertainty of 2 MPa in the stress measurement. Additional errors in the stress and strain measurements may come from apparatus distortion. To correct for this type of error, [P. N. Chopra \(1997\)](#) conducted an experiment with polycrystalline alumina. Unfortunately, these experiments are not discussed in sufficient detail to correct the dry dunite data used in our study. However, the calibration graph reveals that apparatus distortion is not temperature dependent. Consequently, ignoring apparatus distortion will have no effect on our estimation of the activation energy of the transient or steady-state creep.

Similar to [P. Chopra & Paterson \(1984\)](#), dry and wet experiments using a solid-medium apparatus were also conducted on dunites by [R. Post \(1977\)](#) under constant stress conditions. This data set is not included in our study, because the uncertainties in the stress measurements are large compared to a gas-medium apparatus. Similarly, we have excluded data from Chopra's (1997) transient creep study conducted under wet conditions, as our work is limited to dry conditions.

4.3 Flow law parameters for transient creep

Several attempts have been made to formulate a rheological law for transient creep, in particular for metals (e.g., [Andrade, 1910](#); [Chinh et al., 2004](#); [Hart, 1970](#); [J. Poirier, 1980](#); [Voce, 1948](#)), but they typically cannot be extrapolated to other physical conditions. We seek a general formulation that satisfies fundamental principles for a rheological framework, such as time frame invariance and reference frame indifference. The same deformation should be predicted for any clock and any coordinate system. We propose a constitutive relationship that consistently captures transient creep and steady-state creep. We consider the Burgers model, which consists of a Maxwell element in series with a Kelvin element (Figure 4.1), but modified with a nonlinear stress/strain-rate relationship for the dashpots ([Masuti et al., 2016](#)). The spring in the Kelvin element represents the internal stress within the representative volume element. In this idealized model, the Maxwell and Kelvin elements are in series, and the corresponding strain-rates are additive. Hence, the total anelastic strain-rate is

$$\dot{\epsilon}^i = \dot{\epsilon}_K + \dot{\epsilon}_M , \quad (4.1)$$

where $\dot{\epsilon}_K$ is the strain-rate in the Kelvin element, $\dot{\epsilon}_M$ is the strain-rate in the Maxwell element, and $\dot{\epsilon}^i$ is the total anelastic strain rate.

The strain-rate in the Maxwell element is given by a nonlinear stress/strain-rate relationship compatible with previous findings (e.g., [G. Hirth & Kohlstedt, 2003](#);

S. Karato & Jung, 2003)

$$\dot{\epsilon}_M = A_M \sigma^{n_M} d^{-m_M} (C_{OH})^{r_M} \exp \left(-\frac{E_M + pV_M}{RT} \right), \quad (4.2)$$

where A_M is the pre-exponential factor, σ is the deviatoric stress, n_M is the stress exponent, E_M is the activation energy, R is the universal gas constant, T is the temperature, V_M is the activation volume, p is the confining pressure, d is the grain size, m is the grain size exponent, C_{OH} is the water concentration, and r_M is the water content exponent. The formulation captures the behavior of various microphysical mechanisms of deformation at steady-state depending on the choice of parameters (e.g., Bürgmann & Dresen, 2008). For example, diffusion creep corresponds to $n_M = 1$, and dislocation creep to $m_M = 0$ and $n_M > 1$.

In general, the strain-rate in the Kelvin element is a function of the effective stress

$$q_K = \sigma - G_K \epsilon_K, \quad (4.3)$$

where ϵ_K is the cumulative strain in the Kelvin element, and G_K is a work-hardening coefficient. The effective stress drives the flow in the Kelvin element and captures the equilibrium of the microstructure under an ambient macroscopic stress. The effective stress is modulated by the internal stress $G_K \epsilon_K$, representing the internal stress concentration within the representative volume element. The strain-rate in the Kelvin element can then be written $\dot{\epsilon}_K = f_K(q_K)$, where f_K is a zero-crossing at the origin, monotonically increasing function. The exact form of f_K is unknown, with a possibly linear, nonlinear, or exponential form. For consistency with the

functional form of stress/strain-rate at steady-state, we adopt the following non-linear, thermally activated formulation

$$\dot{\epsilon}_K = A_K q_K \|q_K\|^{n_K-1} (C_{OH})^{r_K} d^{-m_K} \exp\left(-\frac{E_K + pV_K}{RT}\right), \quad (4.4)$$

where the form $x \|x\|^{n-1}$ is to allow for possible retrograde deformation in the Kelvin element under some conditions, for example, rapid stress reduction. When the effective stress is negative, expressions of the form x^n would give rise to complex numbers and are therefore inadequate. The parameters A_K , n_K , m_K , r_K , E_K and V_K play the same roles as in equation (4.2), but for the Kelvin element. [Masuti et al. \(2016\)](#) describe a tensor equivalent of equation (4.4), but a scalar formulation is sufficient to model laboratory data.

The formulation is appropriate to consistently represent the transient creep and steady-state creep. For example, steady-state implies $\dot{\epsilon}_K = 0$, such that the strain-rate in the Maxwell element represents steady-state plastic flow. During transient creep, both the Kelvin and Maxwell components contribute to the total anelastic strain. Therefore, rapid deformation of the Kelvin element is associated with the phase of transient creep, even though the Maxwell component also contributes strain during that time.

The olivine flow law parameters and their uncertainties at steady-state are well known, with the possible exception of the activation volume V_M , which is still debated (e.g., [Bürgmann & Dresen, 2008](#); [Korenaga & Karato, 2008](#)). However, the flow law parameters and their uncertainties for transient creep are unknown. To

constrain the parameters r_K and V_K in equation 4.4, experimental data conducted at different water contents and confining pressures are necessary. Similarly, to constrain the parameter m_K , experiments in the diffusion creep regime with different grain sizes are needed, but these are not available. In this study, we focus on dry dislocation creep experimental data conducted at confining pressure of 300 MPa and known steady-state creep parameters (see Table 2 of [P. Chopra & Paterson, 1984](#)), and we restrict ourselves to the determination of the transient creep parameters A_K , G_K , n_K , and E_K . The effective activation energy therefore includes the pressure effect at 300 MPa.

Unlike for the steady-state creep parameters, which can be determined using any of the least-squares methods that solve a linear inverse problem, determining the transient creep parameters requires modeling of the time series of strain or stress under constant stress or constant strain-rate conditions, respectively, and the resulting inverse problem is nonlinear. We model the stress versus strain evolution with the Burgers model described above as follows. We consider a spring-dashpot model (Figure 4.1), where the total strain-rate is the sum of the elastic and anelastic components ([Andrews, 1978](#); [Barbot & Fialko, 2010b](#))

$$\dot{\epsilon} = \dot{\epsilon}^e + \dot{\epsilon}^i, \quad (4.5)$$

where $\dot{\epsilon}^e$ and $\dot{\epsilon}^i$ are the elastic and anelastic strain-rates, respectively. Substituting Hooke's law $\dot{\epsilon}_e = \dot{\sigma}/G_M$, where G_M is the Young's modulus, we obtain

$$\dot{\sigma} = G_M(\dot{\epsilon} - \dot{\epsilon}^i). \quad (4.6)$$

Equations (4.2), (4.3), (4.4), and (4.6) form a closed set of governing equations. We solve the resulting set of coupled ordinary differential equations numerically using the Runge-Kutta method (Press et al., 1992). For each assumed set of rheological parameters, we obtain time series of stress and strain components and the associated parametric curves of stress versus total anelastic strain-rate. The remaining step is to identify the parameters that best explain the available experimental data and their uncertainties.

4.4 Deformation data analysis

We use a Markov Chain Monte Carlo method with the Gibbs sampling to explore the model space, as Gibbs sampling provides a simple and efficient technique to search for best-fit parameters and their uncertainties, even for nonlinear problems (e.g., Casella & George, 1992; Geman & Geman, 1984). For a given set of physical parameters, we simulate the mechanical response for all of the experimental conditions. We then compare the predictions to all measurements at once (i.e., 5 single strain rate experiments and 7 strain rate stepping experiments). The procedure therefore represents a global optimization of the model parameters. Gibbs

sampling requires only a few simple tuning parameters, i.e., the number of samples to be generated in computing the conditional distribution for each parameter, and the number of burn-in and step samples used to reduce the effect of artificial correlation and initial guess. We generate 10^5 samples with conditional evaluations of 1000 samples in each dimension. We discard the initial 10^4 samples and we retain every tenth sample from the remaining samples.

FIGURE 4.2: Joint and marginal distribution of the flow law parameters generated from the Gibbs sampling (10^5 samples with burn-in= 10^4 and step=10) for the constitutive parameters.

With the selected random samples, we generate marginal and joint distributions of the corresponding model parameters (Figure 4.2), and we discuss the mean values and their standard deviation. We find a strain-hardening coefficient $G_K = 33 \pm$

0.3 GPa, approximately an order of magnitude smaller than Young's modulus, but with the same order of magnitude as the rigidity (shear modulus) of the same rock. We find an effective activation energy of $E_K = 430 \pm 20$ kJ/mol and stress exponent of $n_K = 2.0 \pm 0.1$, lower than those of the steady-state creep of $E_M = 535 \pm 15$ kJ/mol and $n = 3.6 \pm 0.1$, respectively.

Using the mean of the probability distributions as input parameters, we simulate the deformation and compare it to the stress/strain-rate experimental curves (Figures 4.3 and 4.6). The data at fixed strain-rate are explained well. The misfit is due to both tradeoffs that naturally occur when all data are optimized simultaneously as well as the simplifying assumptions of the model. The model also explains the strain-rate stepping experiments, but the fit to the second part, at smaller strain rate, is poorer than the first part, at higher strain rate. This could be due to change in internal stress of the mineral, or permanent changes in the microstructure due to large strain reached, affecting the transient more than the steady-state behavior. This can also be seen in another model with a hardening coefficient set to $G_K = 10$ GPa in the second phase of the experiment that explains the data slightly better (dashed red line in Figures 4.3 and 4.6).

We also explore the potential of the internal state variable (ISV) constitutive framework of Sherburn et al. (2011). This constitutive law involves more than 10 parameters, including isotropic and anisotropic hardening coefficients. This rheology may provide superior predicting capability, but no data is currently available to constrain all of the parameters individually. Sherburn et al. (2011) used a subset of the data of P. Chopra & Paterson (1984) considered here. We find that the ISV's

TABLE 4.1: Flow law parameters of dry dunites and San Carlos olivine aggregates determined from the constant-strain rate experimental data of [P. Chopra & Paterson \(1984\)](#). The bounds column contains the upper and lower bounds used for the Gibbs sampling. Steady-state parameters for dunites and San Carlos olivine aggregates are from [P. Chopra & Paterson \(1984\)](#) and [S. Karato et al. \(1986\)](#), respectively (errors are for one standard deviation).

Parameter	Symbol	Value		Bounds	Unit
		Dunite	Olivine		
Pre-exponential factor	A_K	$10^{5.5 \pm 0.6}$	10^6	10^{-2} to 10^9	$s^{-1} MPa^{-n_K}$
Hardening coefficient	G_K	33 ± 0.3	15	0.1 to 100	GPa
Stress exponent	n_K	2.0 ± 0.1	2.2	0.1 to 10	
Activation energy	E_K	430 ± 20	410	100-800	kJ/mol
Young's modulus	G_M	92 ± 2	92	10-200	GPa
Pre-exponential factor	A_M	$10^{4.46 \pm 0.09}$	$10^{5.38}$	-	$s^{-1} MPa^{-n_M}$
Stress exponent	n_M	3.6 ± 0.1	3.5	-	
Activation energy	E_M	535 ± 15	540	-	kJ/mol

best fit model performs poorly against the data that were not used to estimate their best fit model (Figure 4.5). In contrast, the nonlinear Burgers model is fully constrained from the data available and the fit is obtained from a smaller set of physical properties.

4.5 Discussion

The analysis of mechanical data for dry dunites reveals an activation energy of $E_K = 430 \pm 20$ kJ/mol and a stress exponent $n_K = 2.0 \pm 0.1$ for transient creep. Since dunites contain more than 94% of olivine, we tested how our model performs against the data of dry San Carlos olivine aggregates from [S. Karato et al. \(1986\)](#). Using the transient creep values from this study and the steady-state creep values from [S. Karato et al. \(1986\)](#), we explain the data of dry San Carlos olivine aggregates (Fig. 4.4). We manually adjust the transient creep parameters for olivine; however,

FIGURE 4.3: Stress vs. strain series data (black dots) for dunites with best-fit models. The red line represents the best fit model with nonlinear Burgers. The dashed blue line indicates the nonlinear Burgers with an activation energy of 500 kJ/mol, with all other parameters remaining the same, and similarly, the dashed red line indicates the nonlinear Burgers model with $G_K = 10$ GPa.

these values are within the uncertainties of values obtained using dunites data, except for the hardening coefficient. The transient creep values of dunites equally represent the transient creep of dry San Carlos olivine aggregates (Table. 4.1). Hence, we discuss the implication of this finding for olivine aggregates. The lower

activation energy and stress exponent of the transient creep mean that the physical mechanism behind the transient creep of olivine aggregates could be different from the mechanism behind steady-state creep, in contrast to the findings of [R. Post \(1977\)](#) and [P. N. Chopra \(1997\)](#).

FIGURE 4.4: Stress vs. strain series data for San Carlos olivine aggregates with best-fit models. Using the values from Table. 4.1 for San Carlos olivine aggregates, we explain the laboratory data of dry San Carlos olivine aggregates from [S. Karato et al. \(1986\)](#). These experiments were conducted under anhydrous conditions at 300 MPa, similar to [P. Chopra & Paterson \(1984\)](#).

4.5.1 Interpretation of the lower activation energy of the transient creep

Based on the similarity between the activation energy of the diffusion of Si in olivine (e.g., [Costa & Chakraborty, 2008](#)) and the activation energy of climb-controlled dislocation creep (i.e., power-law dislocation creep), diffusion of the slowest species has been proposed as the rate-controlling mechanism, if the dislocations are fully saturated with jogs ([D. L. Kohlstedt, 2006](#)). However, this diffusion-based theory does not explain the plastic anisotropy observed in the dislocation creep rate of olivine aggregates, as the degree of anisotropy observed in the diffusion coefficients is small ([S. Karato, 2010](#)).

An alternative model of anisotropy in jog density (number of jogs per unit length) has thus been proposed ([S. Karato, 2010](#)). In this model, the dislocation climb velocity (v_c) depends not only on the diffusion coefficient but also on the jog density (c_j):

$$v_c \propto D c_j , \quad (4.7)$$

where D is the relevant diffusion coefficient (see [S. Karato \(2008\)](#) and/or [J.-P. Poirier \(1985\)](#)). Hence, we can say that the activation energy of climb-controlled dislocation creep (E_{creep}) is the sum of the activation energy of diffusion (E_D) and the activation energy of jog formation (E_{c_j}):

$$E_{\text{creep}} = E_D + E_{c_j} . \quad (4.8)$$

Recently, a high jog density of the soft-slip system, which controls the transient creep, and a low jog density of the hard-slip system, which controls the steady-state creep, has been used to explain the anisotropic water effect (Masuti et al., 2019). If we assume this model is valid under dry conditions, that is, the transient creep is controlled by the soft-slip system, then the soft-slip system has high jog density, and we do not need to create further jogs for deformation to occur (i.e., jog formation energy is low). Hence, the activation energy of the transient creep is equal to the activation energy of diffusion alone (i.e., $E_{\text{creep}} \approx E_D$; in the case of olivine, it is the diffusion of Si). In contrast, steady-state creep is controlled by the hard-slip system, where jog density is low, and further jogs need to be created for deformation to occur (i.e., jog formation energy is high). Hence, the activation energy of transient creep is smaller than that of steady-state creep.

A recent diffusion study by Fei et al. (2013) also supports the theoretical model of jog density discussed above. They concluded that the activation energy of the diffusion of Si in forsterite and olivine in both dry and wet conditions is ~ 410 to 450 kJ/mol. However, one of the outstanding unknowns is the activation energy of the soft-slip system. Bai et al. (1991) did experiments on olivine single crystals under anhydrous conditions at different orientations including the orientations $[110]_c$ and $[011]_c$, which activate the slip on the soft and hard slip systems, respectively, and found the activation energy of the hard slip system to be ~ 540 kJ/mol, in agreement with the steady-state creep of dunites and San Carlos olivine aggregates. However, in their study, the activation energy of the soft slip system (~ 230 to 1000 kJ/mol) is not well resolved.

Additionally, the lower stress exponent of transient creep compared to steady-state creep suggests that transient creep is controlled by the soft-slip system and is in agreement with previous studies (e.g., [Goetze & Poirier, 1978](#); [S. Karato et al., 2008](#)).

4.5.2 Outstanding problems and future directions

The physical mechanism behind the transient creep of olivine aggregates is currently unknown. A number of physical models have been proposed to explain the transient creep of olivine aggregates, viz., soft-slip to hard-slip transition ([S. Karato, 1998](#); [Masuti et al., 2016](#)), microstructure such as dislocation density and/or grain size evolution ([S. Karato, 2008](#)), isotropic and kinematic hardening ([Sherburn et al., 2011](#)) in the case of dislocation creep, and stress/strain re-distribution at the grain boundary in the case of diffusion creep ([S. Karato, 1998](#)). However, how robustly these models explain laboratory data recorded under various thermodynamic conditions is yet to be validated, except the soft-slip to hard-slip transition discussed in this paper. The ISV model of isotropic and kinematic hardening has been partially validated but performs poorly compared to our nonlinear Burgers model with the same data set (see Supporting material).

Another unknown is the *in situ* mechanics of deformation during postseismic deformation ([Freed et al., 2010](#); [Muto et al., 2019](#); [Rollins et al., 2015](#); [Rousset et al., 2012](#)). The upper mantle transient viscosity during the postseismic deformation following great and giant earthquakes along the Sunda arc can be explained with

a linear Burgers rheology (Qiu et al., 2018), implying deformation by diffusion creep. The lower-crustal postseismic deformation following the 1999 Mw=7.8 Chi-Chi, Taiwan earthquake can be explained using our proposed nonlinear Burgers rheology (Tang et al., 2019), implying deformation by dislocation creep. Similarly, dislocation creep has also been used to explain the postseismic deformation following the 2011 Tohoku earthquake (Agata et al., 2019; Muto et al., 2019). A linear Burgers rheology has so far been the most widely used to explain postseismic observations (Barbot et al., 2008; Biggs et al., 2009; DeVries et al., 2017; Hu et al., 2004; Pollitz, 2003; Pollitz et al., 2006, 2001; Sun et al., 2014; K. Wang et al., 2012). More experimental work on different minerals (e.g., enstatite, clinopyroxene, and orthopyroxene) and rock assemblies (e.g., peridotite and dunite) under hydrous conditions and confining pressure that is representative of asthenosphere are required to better constrain the transient rheological behavior of the lithosphere-asthenosphere system.

4.6 Conclusion

We explain the experimental deformation data derived from dry dunites and dry olivine aggregates using a nonlinear Burgers model. The proposed rheological framework consistently captures the transient creep and steady-state creep of olivine. The model can in principle apply to deformation in the upper mantle, except that the effect of water on transient creep is still unconstrained. The activation energy and the stress exponent for transient creep are lower than those for steady-state

creep. This suggests that the physical mechanism behind the transient creep of olivine aggregates may be different from that associated with steady-state creep, possibly due to a higher jog density in the soft slip system. Consequently, the transient creep of olivine aggregates may be controlled by the soft-slip system, which is saturated with jogs, and steady-state creep may be controlled by the hard-slip system, which is undersaturated with jogs. The constitutive framework and the associated physical parameters provide a reference for modeling deformation in the lithosphere and upper mantle at the time scales of the seismic cycle and post-glacial rebound.

4.7 Supplementary Information

In these supplementary materials, we provide additional information related to

- ISV model,
- Gibbs sampling benchmark,
- Gibbs sampling convergence,
- Covariance and correlation matrices,
- Code availability.

Internal state variable model (ISV)

Sherburn et al. (2011) conducted a similar study using their ISV model on a subset of the dataset used in this paper. Their flow rule uses two state variable, that is, the kinematic hardening coefficient α and the isotropic hardening coefficient R . The anelastic strain-rate in their ISV model is

$$\dot{\epsilon}^i = f(T) \sinh \left[\frac{(\sigma - \alpha) - R - Y(T)}{V(T)} \right] \frac{\sigma - \alpha}{\|\sigma - \alpha\|}, \quad (4.9)$$

where σ is the deviatoric stress, T is the temperature, and $f(T)$, $Y(T)$, and $V(T)$ are Arrhenius-type temperature-dependent functions, given by

$$\begin{aligned} V(T) &= C_1 \exp\left(-\frac{C_2}{T}\right), \\ Y(T) &= 0.5C_3 \exp\left(\frac{C_4}{T}\right) (1 + \tanh[C_{19}C_{20} - T]), \end{aligned}$$

and

$$f(T) = C_5 \exp\left(-\frac{C_6}{T}\right),$$

where the coefficients C_i are constants. Kinematic hardening represents the geometrically necessary dislocations. Isotropic hardening represents the randomly inherited dislocations. Assuming no kinematic hardening, Sherburn et al. (2011) modeled the stress vs. strain curves using the above flow law with isotropic hardening, associated with the following evolution law,

$$\dot{R} = \sqrt{\frac{2}{3}} H(T) \dot{\epsilon}^i - \left[\sqrt{\frac{2}{3}} R_d(T) \dot{\epsilon}^i + R_s(T) \right] R^2, \quad (4.10)$$

where $R_d(T)$, $H(T)$, and $R_s(T)$ are given by

$$R_d(T) = C_{13} \exp\left(-\frac{C_{14}}{T}\right),$$

$$H(T) = C_{15} - C_{16}T,$$

and

$$R_s(T) = C_{17} \exp\left(-\frac{C_{18}}{T}\right).$$

Using the best fit model parameters for the constants C_i , we plot the ISV model against the data used (Figure 4.5). Sherburn et al. (2011) used only part of the data from Chopra & Paterson (1984), and their best fit model does not explain the remaining data, for example, runs 4667, 4506, and 4461 (Figure 4.5).

In order to compare the nonlinear Burgers model against the ISV model, we optimize the nonlinear Burgers model parameters using a subset of the data. Using the best fit values from a subset of the data, both the nonlinear Burgers model and the ISV model perform poorly against the remaining data. Yet, despite its simplicity, the nonlinear Burgers model performs incrementally better than the ISV model (Figure 4.5).

Gibbs sampling convergence

In Gibbs sampling, as all the proposed samples are accepted, successive samples can be highly correlated. One way to check whether the samples are independent of each other is to compute the autocorrelation between samples. The autocorrelation reaches zero by about the 800th lag (Figure 4.7).

Covariance and correlation matrices

The covariance and correlation matrices are given in Table 4.2. Combining the information from Tables 4.1 and 4.2, the joint and marginal distributions can be reproduced.

FIGURE 4.5: Stress vs. strain data and the best fit curves from the ISV and the nonlinear Burgers models. **a-c)** The model fits to the data that were used by Sherburn et al. (2011). All the constants ($C_1 - C_{20}$) are from Table. 1 in Sherburn et al. (2011), except $C_{13} = 0.83$, which could be a typo in their paper. Also, the value of temperature for run 4580 is incorrectly used as 1573 K. **d-f)** Neither the ISV nor the nonlinear Burgers model fits the supplementary data from Chopra & Paterson (1984) that were not used in the optimisation process, however, the nonlinear Burgers model, which uses fewer parameters compared to the ISV model, still performs better.

FIGURE 4.6: Stress vs. strain series data (black dots) with best-fit models not shown in the main paper. The nonlinear Burgers model with a reduced hardening coefficient ($G_K=10 \text{ GPa}$) fits the relaxation data slightly better. Note that the gap between the hardening and relaxation parts is due to the fact that each segment is solved independently with appropriate initial conditions.

FIGURE 4.7: The autocorrelation function (ACF) for the samples generated using Gibbs sampling for the nonlinear Burgers model.

Code availability

All the MATLAB code necessary to reproduce our results will be shared on the *GitHub* page (<https://github.com/smasuti>).

TABLE 4.2: Covariance and correlation matrices for the selected samples from Gibbs sampling.

	covariance					correlation				
	A_K	G_K	n_K	E_K	G_M	A_K	G_K	n_K	E_K	G_M
A_K	3.88E-01	9.79E-04	1.68E-03	1.18E-02	1.46E-03	1.00E+00	3.88E-01	9.50E-02	9.09E-01	2.05E-01
G_K	9.79E-04	1.64E-05	-3.37E-05	1.84E-05	-2.47E-06	3.88E-01	1.00E+00	-2.93E-01	2.18E-01	-5.33E-02
n_K	1.68E-03	-3.37E-05	8.05E-04	2.93E-04	5.90E-05	9.50E-02	-2.93E-01	1.00E+00	4.97E-01	1.82E-01
E_K	1.18E-02	1.84E-05	2.93E-04	4.32E-04	5.00E-05	9.09E-01	2.18E-01	4.97E-01	1.00E+00	2.10E-01
G_M	1.46E-03	-2.47E-06	5.90E-05	5.00E-05	1.31E-04	2.05E-01	-5.33E-02	1.82E-01	2.10E-01	1.00E+00

TABLE 4.3: Data included from Chopra & Paterson (1984) in our study, where the high-temperature experiments were conducted under dry conditions at a confining pressure of 300 MPa.

No	Run Number	Temperature (K)	Strain rate ($\dot{\epsilon}$ in s^{-1})
1	4580	1673	1.00×10^{-4}
2	4580	1673	3.2×10^{-5}
3	4580	1673	1.0×10^{-5}
4	4469	1573	1.09×10^{-4}
5	4469	1573	1.05×10^{-5}
6	4469	1573	1.1×10^{-4}
7	4526	1483	5.26×10^{-5}
8	4667	1483	3.09×10^{-5}
9	4667	1483	3.43×10^{-6}
10	4461	1473	1.13×10^{-5}
11	4668	1483	9.72×10^{-6}
12	4668	1483	1.14×10^{-6}
13	4581	1573	1.07×10^{-4}
14	4581	1573	3.25×10^{-5}
15	4581	1573	1.1×10^{-5}
16	4576	1573	1.08×10^{-4}
17	4462	1473	1.25×10^{-5}
18	4666	1483	3.58×10^{-5}
20	4666	1483	3.74×10^{-6}
21	4675	1483	1.18×10^{-5}
22	4675	1483	1.32×10^{-6}
23	4506	1553	1.17×10^{-4}

FIGURE 4.8: Gibbs sampling benchmark. We generate synthetic data using similar thermodynamic conditions to those of the experimental data using $\log_{10}(A_K) = 5$, $\log_{10}(G_K) = 4.602$, $\log_{10}(n_K) = 0.477$, and $\log_{10}(E_K) = 5.6$. The histograms are plotted from the samples generated by Gibbs sampling (10^5 samples with burn-in= 10^4 and step=10 samples). The mean of the marginal distribution matches the true value from which the synthetic data were generated.

Chapter 5

Discussion and Conclusions

In this thesis, I have investigated the rheology of the upper mantle by modelling geodetic and experimental observations. First, I used data from the postseismic deformation following the 2012 M_w 8.6 Indian Ocean earthquake, the largest strike slip event ever recorded. The coseismic rupture reached depths of up to 60 km, and hence could have triggered significant flow in the asthenosphere, where olivine is the major mantle mineral. Generally, to explain the postseismic observations following large earthquakes, olivine steady-state creep has been invoked. As olivine exhibits accelerated transient creep before reaching steady-state creep, some studies have shown that it is necessary to include transient creep. Yet, while numerous constitutive relations for the transient creep have been proposed, most do not satisfy time frame invariance and coordinate indifference properties. Hence, I proposed a constitutive relation by modifying the Burgers model for the rheology of olivine that captures both the transient and steady-state creep. Using this nonlinear Burgers constitutive relation for the viscoelastic flow of olivine in the asthenosphere and the

rate-strengthening model of afterslip from the rate-and-state framework, the geodetic observations following the 2012 event are explained with high water content, indicating that the asthenosphere is wet. Since the earthquake ruptured conjugate faults down to great depths, compatible with dry olivine in the oceanic lithosphere, the water content in the upper mantle is likely stratified, with a steep gradient at the lithosphere-asthenosphere boundary. The steep rheological contrast could be due to dehydration across the lithosphere-asthenosphere boundary, presumably by buoyant melt migration to form the oceanic crust.

Second, due to lack of flow law parameters for the transient creep of olivine, the transient creep parameters in the above study were assumed to be same as those of steady-state creep. However, such an assumption is not valid, if the physical mechanism behind the transient and steady-states is different. In order to understand the physical mechanism behind transient creep, I have determined the transient creep flow law parameters using mechanical data from laboratory experiments performed on natural dunites containing at least 94% olivine under anhydrous conditions, a nonlinear Burgers model, and a Markov Chain Monte Carlo method. I have found different properties for the transient and steady-state components of dunites, with activation energies of $E_K = 430 \pm 20$ kJ/mol and $E_M = 535 \pm 15$ kJ/mol, and stress exponents $n_K = 2.0 \pm 0.1$ and $n_M = 3.6 \pm 0.1$, respectively. These differences suggest the physical mechanism controlling transient creep differs from the mechanism behind the steady-state creep of olivine aggregates, possibly due to a higher jog density in the soft-slip system. Consequently, the transient creep of olivine aggregates may be controlled by the soft-slip system, which is saturated with jogs,

and the steady-state creep may be controlled by the hard-slip system, which is undersaturated with jogs.

Third, the water content sensitivities of the transient and steady-state creeps of olivine were assumed to be same in the first study. In order to quantify the effect of water on the transient creep of olivine, I collected experimental data on deformation of hydrated single-crystal olivine. Using data from low pressure (~ 0.1 MPa) to high pressure (i.e., up to 6 GPa) with different water content, I found that the water effect on the soft-slip system is small, while on the hard-slip system it is large. If transient creep is controlled by the soft-slip system and steady-state creep is controlled by the hard-slip system, then the short-term time-dependent small-strain deformation associated with the post-seismic period is strongly dependent on the water content.

I have demonstrated the application of the nonlinear Burgers model to both geodetic data and laboratory data. However, three outstanding issues need to be discussed. First, while I have determined the stress exponent and activation energy in the case of dry olivine aggregate, the asthenosphere is wet and the effect of water on transient creep is unknown. Similarly, the effect of pressure (i.e., activation volume) on the transient creep is unknown. More rigorous experiments on olivine aggregates with varying water content and pressure are needed to accurately quantify these effects on the transient creep of olivine aggregates.

Second, while the flow laws of the soft-slip and hard-slip systems under wet conditions have been quantified ([Masuti et al., 2019](#)), due to limited availability of data,

the effects of temperature (i.e., activation energy) and stress (i.e., stress exponent) remain unknown. More experimental data at varying temperatures and stresses on different orientations (i.e., $[110]_c$, $[101]_c$, and $[011]_c$) are needed to fully constrain the flow laws of single-crystal olivine.

Last, once the effects of water and pressure on transient creep are accurately determined, postseismic deformation following large earthquakes needs to be re-analyzed to accurately quantify the effect of transient creep.

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