A secondary zone of uplift caused by megathrust earthquakes:
Insights from seismo-thermo-mechanical models and geodetic data

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Study Figure 4 which shows a section of the Earth and the mechanics of plate tectonics.

(i) Annotate Figure 4, to show the following: (Mark on diagram)
1. an ocean trench;
2. a zone of deep-focus earthquakes.

(ii) Account for the origin of each of these features: (Explain)
1. ocean trench:
   On a destructive margin, one plate is subducted under the other. When a oceanic plate is subducted under
   a continental plate, on one side there are folded mountains, on the other side there is a trench.
   Why?
2. Central Ridge:
   Two plates diverge, and out of the ridge come
magma which can build up a island arch.
   Iceland is an island which has formed in this way.

In a geography class a few years ago, I had to take this exam on plate tectonics. With a 4.25 I wasn’t very successful, but this didn’t stop me from studying earth science. Thank you, Nick Hampton, for Fascinating me for the geosciences!
Acknowledgements

First of all, I would like to thank Ylona van Dinther and Taras Gerya for the competent and motivating supervision. I also thank my co-examiners, Prof. Guy G. Drijkoningen and Dr. Auke Barnhoorn. My additional thanks go to my L\TeX expert Karol Kozioł, to Harold Tobin, to my awesome office mates, to the friendly librarian at the NO library, to Adrian and Patrizia, to Hanspeter and Franziska, to Sarah and Finn, to Andrea, to Toni, Mischa, Erzi and Private Benjamin Urben, to Cédéric, Bart De Wever, and the memorable train conductor from Norwich to Sheringham, to the notorious AT gang with Pascal, Simon, Fabi, Elias and Flo, to Claudio and Dr. in spe Martin W. Frischknecht, to the German taxpayers and to Wintershall for the scholarships, and to Total for giving me pens made from plastic, not oil. Finally, special thanks go to a special person.
Abstract

This thesis is intended as a manuscript for publication in a scientific journal. Additionally, the following abstract was submitted to American Geophysical Union for consideration in the 2014 Fall meeting (ID# 25819).

The 1960 M9.5 Valdivia and 1964 M9.2 Alaska megathrust earthquakes caused a zone of secondary coseismic uplift a few hundred kilometers landward of the trench. The secondary uplift exceeds 1m in Valdivia and 0.3m in Alaska. Its universal occurrence during other earthquakes as well as the physical mechanisms governing this process remain unclear, although several mechanisms have been proposed in the last century.

I combined survey data from these megathrust earthquakes in the last century with recent high-quality GPS data recorded during the 2010 M8.8 Maule and 2011 M9.0 Tohoku-Oki megathrust earthquakes. Based on these natural observations, I confirm the existence of a zone of secondary uplift in all earthquakes studied and discuss the similarities and differences between them. Dip and curvature of the slab and overriding plate structure seem to influence the amplitude of this phenomenon as well as the location of the transition from subsidence to secondary uplift. This transition ranges from 200 km distance to the trench up to 500 km. The amplitude of the secondary zone of uplift scales with the size of the earthquakes.

Using numerical seismo-thermo-mechanical modeling, I identified the low-viscosity lower crust as a crucial factor. A weak viscous lower crust does not accumulate significant stresses and allows for the bending and elastic buckling of a thin and strong elastic upper crust. The amplitude and relative position of the secondary zone of uplift also depends on the interplate decoupling depth as well as the shear modulus and thickness of the upper crust.

In conclusion, a secondary zone of coseismic uplift due to megathrust earthquakes exists and it is mainly controlled by the difference between a rheologically strong elastic upper and weak ductile lower crust. Since a weaker lower crust is observed to affect surface displacements distinctly, ignoring this feature will have a profound impact on the estimated slip on the interface. Therefore I suggest including this unequivocal feature in geodetic-based source inversions, especially for settings with a warm, continental overriding plate.
# Table of Contents

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Acknowledgements</td>
<td>ii</td>
</tr>
<tr>
<td>Abstract</td>
<td>iii</td>
</tr>
<tr>
<td>1 Introduction</td>
<td>1</td>
</tr>
<tr>
<td>2 Natural Data: Does a second zone of uplift exist?</td>
<td>4</td>
</tr>
<tr>
<td>3 Seismo-thermo-mechanical models: What are the governing parameters?</td>
<td>9</td>
</tr>
<tr>
<td>4 Laboratory-scale models: What geometrical parameters are crucial?</td>
<td>12</td>
</tr>
<tr>
<td>5 Discussion</td>
<td>17</td>
</tr>
<tr>
<td>6 Conclusions</td>
<td>20</td>
</tr>
<tr>
<td>A Numerical method</td>
<td>22</td>
</tr>
<tr>
<td>B Calculation of wavelength</td>
<td>25</td>
</tr>
</tbody>
</table>
Chapter 1

Introduction

The recent great earthquakes, such as the 2011 M9.0 Tohoku-Oki earthquake, occurred at subduction zones (e.g., Lay, 2012). Those megathrust earthquakes ($M > 8.5$) are still unpredictable and can cause widespread destruction and give rise to tsunamis. Many aspects regarding the physics governing those earthquakes remain unclear despite decades of research. A key limitation in research is the limited data available for analysis in both space and time. Only a handful of megathrust earthquakes have been observed and measured with some accuracy. Numerical modeling allows to overcome this gap and to study specific phenomena in the required detail.

The quasi-static (i.e., non-seismic) displacement field of earthquakes is routinely measured and published. The tectonic movements shed light on the physical processes operating on a local- to lithosphere-scale and are key to understand the generation of tsunamis. Additionally, geodetic data can be used for source inversions to calculate the slip distribution on the fault (Vigny et al., 2011), but it also provides a way to assess numerical models (van Dinther et al., 2013b).

During an earthquake, the surface above the hypocenter subsides, whereas the area closer to the trench is strongly uplifted. This is a reversal of the displacements accumulated during the interseismic period (Moreno et al., 2010) and explained by the elastic rebound theory Reid (1910). In the largest earthquakes, a secondary zone of uplift is measurable landward of the hypocenter. It was first observed by Plafker (1969) for the 1964 M9.2 Alaska earthquake. Subsequently, it was also identified for the strongest earthquake recorded so far, the 1960 M9.5 Valdivia (Plafker and Savage, 1970). Despite the fact that this signal is also present for the two most recent megathrust earthquakes (Chapter 2) it has not been observed as such. Due to measurement uncertainties and an absence of physical understanding of why it should occur, researchers have not — according to my knowledge — discussed it.
In standard (visco-)elastic models, a zone of secondary uplift is not observed (Wang, 2007). It was however observed in models which reconstruct specific events. The 'zone of minor uplift' could be reproduced for the Valdivia earthquake by introducing curvature (approximated by straight segments of different angles) in the fault (Plafker and Savage, 1970). They, however, noted: "[T]here is no difficulty in finding models which reproduce the observed distribution of major uplift, subsidence, and slight uplift, although none of these models can closely duplicate both the amplitudes and widths of the deformed zones. [...] The available data do not permit firm conclusions regarding the origin of the slight uplift, and it is entirely possible that the uplift may result from some other factor rather than fault curvature." (Plafker and Savage, 1970, p. 1024–1025)

Linde and Silver (1989) reanalyzed the same dataset of the Valdivia earthquake and concluded that aseismic afterslip below the seismogenic zone has occurred. They base it on the fact that their slip needed to go below the seismogenic zone to match the observed deformation pattern. This idea was also discussed by Plafker and Savage (1970), who were undecided whether aseismic creep happened before or after the Valdivia earthquake. However, to explain the secondary zone of uplift, the model of Linde and Silver (1989) also required a kinked fault with a more steeply dipping section landward of the second hinge point.

For the Alaska earthquake, Plafker (1969) noted the zone of slight uplift under ‘major unresolved problems’. He speculated that its cause might be a sudden increase in horizontal compressional strain and termed it 'Poisson bulge'. Vita-Finzi and Mann (1994) explained the deformation pattern in Valdivia by elastic buckling due to compression and extension of the overriding plate, "by analogy with the continuum behaviour of an elastic beam".

Wdowinski et al. (1989) present a ”thin viscous sheet model”, which — in a transient version — could explain the zone of secondary uplift by a visco-elastic response in the mantle wedge. They postulate a corner flow in the asthenospheric mantle bounded by a rigid subducting plate and a deformable overlying plate. Their predictions of deformation depend on the angle of subduction, the lithosphere thickness of the overriding plate, and the ratio of asthenospheric to lithospheric viscosities.

Although several mechanisms have been proposed in the last century, there is no consensus on what physical process is responsible for this observation. There is not even consensus yet that this zone of secondary uplift is a true physical phenomenon, and the typical (visco-)elastic models do not show such a feature (Wang, 2007). Recent observations, including the high-quality GPS data recorded during the 2010 M8.8 Maule and 2011 M9.0 Tohoku-Oki megathrust earthquakes, allow new insights into the enigma of the zone of secondary uplift. To understand the physical mechanisms governing the secondary zone of uplift, I combined new GPS data with the older survey data and determined the controlling parameters using numerical models.

In this manuscript, I will first compare the data obtained during four megathrust earthquakes to confirm its existence (Chapter 2). Subsequently, I test the role of the mechanical parameters involved in a large scale seismo-thermo-mechanical numerical model (van
Dinthar et al., 2013b) in which this phenomenon was observed (Chapter 3). The position and size of the zone of secondary uplift is fairly stable for varying mechanical parameters but identical geometries. To distill the effect of geometry, I need to use a simplified laboratory-scale numerical model (Chapter 4). Finally, by integrating these three data types, I discuss that the presence of a strong upper and a weak lower crust provides a new explanation of the secondary zone of uplift caused by megathrust earthquakes (Chapter 5).
Chapter 2

Natural Data:
Does a second zone of uplift exist?

A secondary zone of uplift in nature can be detected by surveying land elevations before and after a megathrust earthquake. Decades ago relative sea level changes were mapped using local markers such as high-tide lines or growth limits of mussels or vegetation with resulting measurement uncertainties in the decimeter range. Nowadays, continuous GPS data is the method of choice and it is usually precise within a few centimeters. Only certain megathrust earthquakes are suitable to study the coseismic displacement at distances, where the zone of secondary uplift is expected: the absence of a coastline (in the case of relative sea level changes measurements) or land (in the case of GPS measurements) in the region of interest makes it impossible to provide suitable measurements.

Figure 1 shows the vertical surface displacements of megathrust earthquakes in a cross section perpendicular to the trench. It is a collection of published data from 1960 M9.5 Valdivia (Plafker and Savage, 1970), 1964 M9.2 Alaska (Plafker, 1969), 2010 M8.8 Maule (Vigny et al., 2011) and the 2011 M9.0 Tohoku-Oki (Ozawa et al., 2011; Sato et al., 2011) earthquake. Table 1 states important aspects on the origin of these datasets and parameters of the corresponding subduction zones. The acquisition method indicates the uncertainty in the data. The time interval between the earthquake and the survey ($\Delta t$) shows the amount of postseismic deformation that is potentially included in the data.

The M9.5 Valdivia earthquake has a distinct zone of secondary uplift with a transition from subsidence to secondary uplift (second hinge point) at around 200 km from the trench and a magnitude of $\sim$1 m (Figure 1a). The data points are more scattered than in the other earthquakes and the measurement errors are rather large (in the range of 20 cm to 60 cm). This is partly due to the large lateral extent with different tectonic
Natural Data: Does a second zone of uplift exist?

Figure 1: Cross section showing the coseismic subsidence and uplift for the a) 1960 M9.5 Valdivia, b) 2010 M8.8 Maule, c) 2011 M9.0 Tohoku-Oki and d) 1964 M9.2 Alaska earthquakes. The gray dots are data points, whereas the colored lines are means of bins of 15 km width. The solid black line is zero vertical displacement, and the dotted lines indicate the two hinge points $HP_1$ and $HP_2$. Note the different vertical scales. All earthquakes show a secondary zone of uplift, but its amplitude and the position of the second hinge point varies. The distances to the trench were calculated using the geolib package for Matlab (Karney, 2013).
settings of the of area affected, and also due to the difference of eight years between earthquake and acquisition (Table 1). The smaller M8.8 Maule earthquake in roughly the same subduction zone also shows a distinct secondary uplift with roughly 10 times smaller amplitude and marginally shifted away from the trench (second hinge point at 240 km, Figure 1b). The M9.0 Tohoku-Oki earthquake produced a slightly different pattern of surface displacements and the zone of secondary uplift is not distinctly visible (Figure 1c). The map view however confirms its widespread, small magnitude existence with the second hinge point located at roughly 335 km from the trench (Figure 2). The M9.2 Alaska earthquake caused a maximum of 0.4 m of secondary uplift with a contrast between the eastern profile (recorded along the Richardson Highway) and the western profile (Alaska railroad, Figure 1d). The western hinge point is at ~500 km compared to ~350 km for the east profile (Figure 2). This difference might arise from the sharp bend of the slab and the trench in Alaska, whereas the trenches in Chile and in Tohoku are rather straight. Due to the bend, the dip of the subducting slab changes, and it is generally more steeply dipping in the west than in the east (Abers, 2008), but in shallow depths it is steeper in the east (Doser et al., 2008). Another difference is the amount of slip, which was larger in the west than in the east (Holdahl and Sauber, 1994).

Summarizing suggestions from all four earthquakes, the amount of secondary uplift is not clearly associated with peak slip of the earthquake (Table 1). This is contradicting the general understanding of more displacement at the interface (slip) resulting in larger surface displacements. The distance of the second hinge point to the trench (Figure 1) seems to be related to the rupture width of the corresponding event (Table 1). Dip of the interface also plays a role for the coseismic horizontal surface displacements. Alaska is the flattest subduction zone and shows the most horizontally stretched pattern. At shallow depths, the subducting plate in Japan dips also at a low angle, whereas in Chile, the slab is initially steeper than in Japan. In the area of the Valdivia earthquake, it is steeper than at the area of the Maule earthquake (Hayes et al., 2012), and it could potentially explain that the displacement pattern is the most compact of all. Besides, the subduction interfaces are curved and not straight and the degree of curvature might also have an influence on the size and position of a secondary zone of uplift.

To conclude, all four studied megathrust earthquakes show a similar displacement pattern including a zone of secondary uplift. Differences in amplitude and hinge point position are considerable and are likely due to different subduction zone geometries and potentially other aspects. To facilitate our physical understanding of why this secondary zone of uplift exists and to find out what causes these differences, numerical models of various complexities are necessary.
Natural Data: Does a second zone of uplift exist?

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<td>Growth limits, leveling survey&lt;sup&gt;2&lt;/sup&gt;</td>
<td>GPS&lt;sup&gt;3&lt;/sup&gt;</td>
<td>GPS&lt;sup&gt;4&lt;/sup&gt;, seafloor geodesy&lt;sup&gt;5&lt;/sup&gt;</td>
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<td>2–20 days</td>
<td>4h (GPS), 1–4 months (sea)</td>
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<td>30 cm</td>
<td>1–10 cm</td>
<td>2 cm (GPS), 20–60 cm (sea)</td>
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<td>$S_{max}$</td>
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<td>22–30 m&lt;sup&gt;6,7&lt;/sup&gt;</td>
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<td>50–60 m&lt;sup&gt;8&lt;/sup&gt;</td>
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<tr>
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<td>$U_{2,max}$</td>
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<td>0.3 m</td>
<td>0.12 m</td>
<td>0.04 m</td>
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Table 1: An overview over the differences between the earthquakes studied and the data acquisition methods. ∆t refers to the time span between earthquake and survey and indicates the amount of postseismic data included, $S_{max}$ are estimates of peak slip, $W_r$ is downdip rupture width (defined as width of the zone with slip $>$5 m), and θ is the average dip of the interface. $H P_1$ is the transition from primary uplift to subsidence (hinge point), whereas $H P_2$ is the second hinge point (the transition from subsidence to secondary uplift). Data sources: <sup>1</sup> Plafker and Savage (1970), <sup>2</sup> Plafker (1969), <sup>3</sup> Vigny et al. (2011), <sup>4</sup> Ozawa et al. (2011), <sup>5</sup> Sato et al. (2011), <sup>6</sup> Johnson et al. (1996), <sup>7</sup> Holdahl and Sauber (1994), <sup>8</sup> Yue and Lay (2013), <sup>9</sup> Moreno et al. (2009), <sup>10</sup> Holdahl and Sauber (1994), <sup>11</sup> Moreno et al. (2014), <sup>12</sup> Hayes et al. (2012)

Figure 2: Map view of the coseismic surface displacements in a) Southern Chile due to the M9.5 Valdivia earthquake, b) Central Chile due to the 2010 M8.8 Maule earthquake, c) north-east Japan due to the 2011 M9.0 Tohoku-Oki earthquake, and d) due to the 1964 M9.2 Alaska earthquake. Sources are given in the "Data type" row of Table 1. Uplift is red, while subsidence is blue. In c), triangles are data points not in used in this study. Green dots denote epicenters. Thick black lines are the trenches (Coffin et al., 1998) and thin black lines are parallels to the trenches at multiples of 100 km. In all earthquakes studied, there is a zone of secondary uplift, but the second hinge point is at different distances from the trench.
Natural Data: Does a second zone of uplift exist?
Chapter 3

Seismo-thermo-mechanical models: What are the governing parameters?

The seismo-thermo-mechanical numerical models of van Dinther et al. (2013b) were used to further investigate the coseismic displacements (Appendix A). The model is based on the continuum visco-elastic-plastic thermo-mechanical code I2ELVIS (Gerya and Yuen, 2007). It was extended from geological time scale processes to seismic cycle time scales by (among other adaptations) including slip-rate dependent friction to generate analog earthquakes (van Dinther et al., 2013a). The 2-D large scale model in this chapter represents a 1500 km × 200 km cross section perpendicular to the Southern Chile subduction zone (Figure A1 a) in Appendix).

The coseismic surface displacements of the reference event of van Dinther et al. (2013b, Figure 5 therein) show a distinct secondary zone of uplift beyond ∼220 km from the trench with a peak around 275 km. In this study, an additional 31 events of this model were analyzed (black lines in Figure 3). The coseismic displacements and hinge points agree well with the reference event of van Dinther et al. (2013b, Figure 5 therein). A parameter study with varying shear modulus ($G$) and viscosity ($\nu$) of the mantle allowed me to assess the impact of those material parameters and the mantle on the presence and shape of the zone of secondary uplift (Figure 3). The position of the secondary hinge point is rather stable for all models. However, some models show a slightly shorter wavelength at the secondary zone of uplift. Those models are characterized by either double (dark red in Figure 3) or half the mantle viscosity (light red) compared to the reference model. Therefore, it is unlikely that the mantle viscosity has a significant impact on wavelength, but it is rather the variability between individual events (including the magnitude) that causes the apparent difference in wavelength.
Seismo-thermo-mechanical models: What are the governing parameters?

Figure 3: Cross section showing the coseismic subsidence and uplift for different seismo-thermo-mechanical models. Black line (mean) and grey shading ($\pm$1σ) is the reference model with dashed lines showing the maximum and minimum of all events in the reference model. Red, blue, and green lines are generated using different mechanical parameters for the mantle of the subducting and overriding plate given in the legend.

If the zone of secondary uplift was a purely elastic feature (as suggested by Vita-Finzi and Mann (1994)), its wavelength can be predicted by an analytical expression (Turcotte and Schubert, 2002 and Appendix B). This equation, an analytical solution for an infinitely long plate under compression, predicts an increase by $\sim$19% for double the shear modulus and a decrease by $\sim$16% for half the shear modulus. Since I observe increases by $\sim$3.7% ($G^*2$) and $\sim$0.4% ($G/2$), the secondary zone of uplift does not seem to be a purely elastic effect.

As mentioned earlier, the inter-event variability is high. An analysis of 148 events of all the models with estimated magnitudes between M8 and M9 shows that the position of the second hinge point is weakly correlated to the rupture width (Figure 4) as suggested
by the natural data. The numerical models also show a qualitative positive correlation between the amplitude of the secondary uplift and the rupture width of each event (Figure 4). This is in contrast to the natural data (Table 1). However, rupture width in the models is a proxy for event size as there is no third (along-trench) dimension. In nature, earthquake magnitude scales with rupture area. The M9.5 Valdivia earthquake has a large magnitude despite a relatively small rupture width due to its rupture length of about 1000 km (Moreno et al., 2009). Taking rupture width in the models as an indicator for magnitude, I can conclude that larger events generate larger amplitudes in the secondary zone of uplift in both nature and numerical models.

To summarize, the amplitude of the secondary zone of uplift scales with the size of the event. The position of the secondary hinge point in the models appears to be robust as it is not systematically influenced by the mechanical parameters of the mantle and only weakly by the rupture width. In the natural observations a larger rupture width more clearly corresponds to the second hinge point farther landward, although statistically not very robust due to the limited number of observations. However, the natural data was acquired at different subduction zones with different geometries, whereas the geometry of the numerical models is the same for all events. To assess the influence of geometry, I use a simplified laboratory-scale subduction model.

![Figure 4](image_url)

**Figure 4:** The influence of the rupture width on a) the position of the second hinge point and b) on the amplitude of the secondary zone of uplift. The color code denotes the downdip rupture limit (in km to the trench), with lighter colors being closer to the trench. The reference model is marked with crosses.
Chapter 4

Laboratory-scale models: What geometrical parameters are crucial?

The laboratory setup consists of a visco-elastic gelatin wedge, which is analogous to the overriding plate and a flat slab subducting at a constant 10° angle below it (Corbi et al., 2013, and Figure A1b). The subducting plate has a seismogenic zone made of velocity-weakening sandpaper. It produces analog earthquakes, which qualitatively agree with observations from natural subduction earthquake cycles (Corbi et al., 2013). This laboratory experiment was numerically modeled by van Dinther et al. (2013a) using the code I2ELVIS (see section 3 and Appendix A), which will be used in this study.

In the basic configuration (Figure A1) the resulting coseismic displacements resolve the primary zone of uplift closest to the trench and the zone of subsidence, but no secondary zone of uplift (Figure 5b). However, the magnitude of the subsidence is too large compared to the magnitude of primary uplift. If the laboratory-scale model is tuned to depict a more realistic viscosity and strength structure of the earth (resulting from temperature dependence of viscosity, Figure 3a in van Dinther et al. (2013b)), the resulting displacement is more similar to the natural one (Figure 5).

A deep low-viscosity wedge representing the asthenospheric mantle is not sufficient in my models to generate a secondary zone of uplift as could be expected from the analog model of Rosenau et al. (2009). They observe a secondary zone of uplift (M. Rosenau, pers. comm.), but their model consists only of an elastoplastic lithosphere and a viscoelastic asthenosphere. In my models, a low viscosity mantle wedge generally leads to smaller surface displacements, which improves the comparison to nature. Additionally, the mantle wedge can influence the position of the secondary hinge point (Figure 6). The peak of the secondary zone of uplift correlates with the interplate decoupling depth, while the primary zone of uplift remains unaffected in both size and location. The angular shape of those asthenospheric wedges is hardly realistic. The sharp corner might facilitate the generation of a secondary zone of uplift similar to the kinked interface in the models of (Linde and Silver, 1989).
Laboratory-scale models: What geometrical parameters are crucial?  

Figure 5: a) The model setup with upper crust, lower crust, lithospheric mantle, and asthenospheric mantle wedge. Upper crust and lithospheric mantle have viscosities of $5 \times 10^{23} \text{ Pa s}$ (scaled to nature), whereas the lower crust and asthenospheric mantle have viscosities of $1 \times 10^{18} \text{ Pa s}$. The seismogenic zone is depicted in dark gray. b) The resulting coseismic displacements scaled to natural values. The basic model does not show a secondary zone of uplift. The zone of secondary uplift is produced by including a lower crust of low viscosity. An asthenospheric mantle wedge of low viscosity is not sufficient to produce a secondary zone of uplift. The surface displacements are an order of magnitude too large due to too much slip as a result of longer recurrence intervals (van Dinther et al., 2013a).
The bar-shaped low-viscosity zone depicting the lower crust is essential in reproducing the secondary zone of uplift. This weak lower crust acts as a reduced stress zone and allows the elastic compression of a distinct rigid upper crust. I use again the equation of Turcotte and Schubert (2002), in which shear modulus and thickness of a plate are important parameters, to predict the wavelength (Chapter 3, Appendix B). Increasing the shear modulus $G$ of the upper crust only slightly reduces the wavelength (Figure 8), whereas the equation predicts an increase of $\sim 19\%$ for $G^*2$ instead. For altering the thickness of the upper crust, distinct changes in wavelength are expected: $+68\%$ for double the plate thickness and $-41\%$ for half the plate thickness. My models behave in a different way; a thicker upper crust leads to a significant reduction in wavelength (as well as amplitude), although it moves the hinge points away from the trench (Figure 7). Based on these observations, I can therefore say that my system behaves different to the equation of Turcotte and Schubert (2002), since their system is a lot simpler than mine. They calculate the steady-state deformation of a uniform infinite beam under orthogonal compression. Instead, I am looking at snapshots of a finite beam compressed by a time-dependent loading mechanism under an angle. Therefore, I am probably not dealing with a single sinusoidal phenomenon as predicted by the equation, but by a superposition of several wavelengths that are sensitive to these parameters in different ways. Therefore, the equation to predict the wavelength (Appendix B) fails for this case but it correctly predicts the importance of the thickness of the plate and the shear modulus. However, my model behaves in a consistent way: A thicker upper plate leads to smaller amplitudes and moves the displacement pattern away from the trench (Figure 7). These effects are likely due to the changing coupling of the seismogenic zone and the low-viscosity lower crust, and due to the larger amount of energy needed to bend a thicker plate. Moreover, a larger shear modulus leads to similar amplitudes of primary and secondary zone of uplift due to higher stiffness.
Figure 6: Different sizes of low-viscosity asthenospheric mantle wedges result in a shift in secondary hinge point. The interplate decoupling depth (marked by a circle) correlates with the peak of the secondary zone of uplift.
**Figure 7:** The surface displacements of a model with a weak viscous lower crust (10 km thick for all) and an upper crust of variable thickness. Blue is the same model as in Figure 5. A thicker upper crust moves the general displacement landward and results in smaller amplitudes.

**Figure 8:** A larger shear modulus of the upper crust results in more equal amplitudes of primary and secondary zones of uplift. Lower and upper crust are each 10 km thick. The blue model is again the same as in Figure 5.
Figure 9 combines results from my natural data and numerical model studies. The latter are scaled according to the recurrence interval. The amplitudes of the primary and secondary zones of uplift and the zone of subsidence are comparable to natural megathrust earthquakes. The second hinge points of the Valdivia and Maule earthquakes are also in agreement with the large scale reference model, which was designed to resemble the Southern Chilean margin. The spatial variations are likely due to different subduction zone geometries such as variable dips and curvature of the interface (Chapter 2). Since the subducting slab in Alaska is gently dipping, this could explain why the second hinge point of Alaska is that far away from the trench (Figure 10). My modelling results suggest to consider the viscosity structure of the overriding plate as well. Japan is an island arc with a cold and strong lower crust, whereas Alaska and Chile are continental subduction zones with a hot and weak lower crust. This might explain the relatively small amplitudes of the secondary zone of uplift in Japan. The primary cause of the secondary zone of uplift is a weak viscous lower crust beneath a rigid elastic upper crust. The lower crust does not accumulate significant stresses and its viscous rheology allows bending of the thin upper crust. This bending results from elastic buckling due to compression of the forearc and the thin, rigid upper crust within the interseismic period and release of these elastically built-up displacements during megathrust earthquakes. The elastic buckling hypotheses has been posed by Vita-Finzi and Mann (1994) as well, although they discard the simultaneous need of a weaker viscous lower crust to facilitate bending of a thinner beam. The thin viscous sheet model of Wdowinski et al. (1989), who ascribed the zone of secondary uplift to an asthenospheric corner flow, is unlikely the sole explanation given that a low-viscosity mantle wedge alone does not produce a secondary zone of uplift. It could potentially play a minor role, especially considering the relation between the interplate decoupling depth and the second hinge point observed in both laboratory-scale (Figure 6) and large scale model (van Dinther et al., 2013b). Wdowinski et al. (1989) are however correct when they predict that deformation depends on the angle of
Figure 9: Cross section showing the coseismic subsidence and uplift for the earthquakes analyzed compared to the numerical models. Dots are the second hinge points. The natural data are means of bins of 15 km width (Figure 1). The black line is the mean of the scaled large scale reference model (Figure 3). A scaling factor of $1/3.5$ ($\approx 0.3$) was applied to the reference model to compensate for the 985 years average recurrence interval compared to 385 years in the Valdivia (Cisternas et al., 2005) and 175 years in the Maule earthquake (Madariaga et al., 2010), since the model is set up to represent Southern Chile. Cyan is the scaled laboratory-scale model, which has a lower crust and a mantle wedge of low viscosity (Figure 5). It was scaled by a factor of 0.05 due the 5518 years average recurrence interval.

subduction, the lithosphere thickness of the overriding plate, and the difference between asthenospheric and lithospheric viscosity. Deep aseismic slip on a kinked interface, which was used as an explanation for the secondary zone of uplift by Linde and Silver (1989), is unlikely as a unifying explanation. Aseismic creep is common, but a kinked interface
is not present at the second hinge point location in all subduction zones where this phenomenon occurs (Figure 10).

An open question is the amount of postseismic uplift present within both the observational data and the numerical simulations. In the numerical model it is possible that some postseismic displacements are included in the coseismic phase, since the coseismic duration is several orders of magnitudes higher than in nature and a separation of the two phases is rather arbitrary (van Dinther et al., 2013b). In the Valdivia and Alaska measurements, there is certainly an amount of postseismic displacement included due to the large time span between earthquake and survey (Table 1). In the region of secondary uplift of the Maule earthquake, the only continuously measuring station shows an uplift of a few millimeter one day after the earthquake and 10 cm after 2 weeks. This indicates a potentially postseismic origin or enhancement of the secondary zone of uplift. Conversely, the secondary zone of uplift in Tohoku shows general subsidence within 2 weeks after the earthquake (Ozawa et al., 2011). Vita-Finzi and Mann (1994) also note a postseismic recovery at several sites. Due to these opposing observations, additional analysis of the postseismic response is necessary.

**Implications**

The importance of a weak lower crust has a significant impact on geodetic-based source inversions. A weak visco-elastic lower crust should therefore be included between the mainly elastic upper crust and the lithospheric mantle in the forward model of these slip inversions. A different forward model will influence the synthetic data predicted by the model and therefore also effect the end result, the estimated slip on the interface.
Conclusions

I have analyzed natural data of four megathrust earthquakes to confirm the existence of a secondary zone of coseismic uplift. Hinge point locations as well as amplitudes are varying between these earthquakes, amongst others due to different dip and curvature of the interface. The amplitude of the surface displacements and the secondary zone of uplift scales with the size of the earthquakes as verified by seismo-thermo-mechanical modeling. Further numerical models revealed the necessity to include a low-viscosity lower crust to create a secondary zone of uplift, since the more viscous lower crust does not accumulate significant stresses and facilitates the bending of a thin upper crust. In the models, the amplitude and hinge point of the secondary zone of uplift depend on the interplate decoupling depth as well as shear modulus and thickness of the upper crust. A more natural coseismic vertical displacement pattern can therefore be achieved in simplified subduction models by adding a low-viscosity lower crust. Additionally, slip inversions can be improved by introducing a low viscosity lower crust in the calculations. Further research is needed to assess the behavior of the region of secondary uplift in the postseismic phase and the role of postseismic creep. Additionally, the role of geometry has to be studied more in simplified and analog models, and tied more closely to natural observations.

Outlook

A few issues in this thesis have been addressed, but not all are fully answered due to time constraints. To resolve the questions raised and to be able to submit this manuscript for publication, I suggest:

- Verification of the role of the weak lower crust by testing a large-scale model without a lower crust. In a preliminary attempt to do this, the stresses in the lower crust did not evolve enough to change the displacements pattern.

- Further examination of the role of kinked and curved interfaces. It was not possible
to make laboratory models with a curved interface to run stably within the limited time of this study.

- Exploring analytical expressions for the deformation of finite plates with complex geometries.

- Collection of other geometrical data of the subduction zones, including interface geometries, thicknesses of the upper and lower crust of the overriding plates, seismogenic zone widths, volcanic arc distances, interplate decoupling depths, and also heat flow. Tomographic images are a potential source for such data. Figure 10 is a preliminary attempt to show slab depths at various distances using hypocenter locations.

- An analysis of postseismic data in the region of secondary uplift of the Maule earthquake. The postseismic data will be provided by C. Vigny and will help to reach statistical significance of the postseismic response in Southern Chile.

- An investigation of the postseismic displacements of the laboratory-scale model to resolve the influence of the different geometrical elements.

- Another attempt to collect data from the 2004 M9.2 Sumatra-Andaman earthquake. Vertical GPS data is probably not available for the (potential) zone of secondary uplift, but geodetic measurements, satellite imagery and possibly InSAR data might be retrievable.

![Figure 10: Preliminary cross section of slab depths from hypocenter locations. Especially the horizontal axis is not well constrained. Additionally, the Alaska depth sections are not parallel to the profiles in Figure 1. Data sources: Hayes et al. (2012), Abers (2008), Bohm et al. (2002), Gutscher (2002)](image-url)
Appendix A

Numerical method

I2ELVIS is a 2-D implicit, conservative finite difference continuum visco-elastic-plastic thermo-mechanical code (Gerya and Yuen, 2007). It works on a fully staggered Eulerian grid combined with a Lagrangian marker-in-cell technique. This enables the properties (lithology, stress histories) to be advected along with the ‘particles’ they were attached to. The code solves for the pressure as well as horizontal and vertical velocity on each respective node assuming conservation of mass (equation A-1) and momentum (equations A-2 and A-3).

\[
\frac{\partial v_x}{\partial x} + \frac{\partial v_z}{\partial z} = 0 \quad (A-1)
\]

\[
\frac{\partial \sigma'_{xx}}{\partial x} + \frac{\partial \sigma'_{xz}}{\partial z} - \frac{\partial P}{\partial z} = \rho \frac{Dv_x}{Dt} \quad (A-2)
\]

\[
\frac{\partial \sigma'_{zx}}{\partial x} + \frac{\partial \sigma'_{zz}}{\partial z} - \frac{\partial P}{\partial z} = \rho \frac{Dv_z}{Dt} - \rho g \quad (A-3)
\]

The reological model is based on

\[
\dot{\varepsilon}'_{ij} = \frac{1}{2\eta} \sigma'_{ij} + \frac{1}{2G} \frac{D\sigma'_{ij}}{Dt} + \left\{ \begin{array}{ll}
0 & \text{for } \sigma'_{ij} < \sigma_{yield} \\
\chi \frac{\partial g_{plastic}}{\partial \sigma_{ij}} & \text{for } \sigma'_{ij} = \sigma_{yield} \\
\dot{\varepsilon}'_{ij(plastic)} & \text{for } \sigma'_{ij} > \sigma_{yield}
\end{array} \right. \quad (A-4)
\]

which relates deviatoric stresses \( \sigma'_{ij} \) and strain rates \( \dot{\varepsilon}'_{ij} \) in a non-linear visco-elastic-plastic manner. The code is typically used for long-term geodynamic processes, such as subduction dynamics and deformation (Gerya, 2011). It was adapted for seismic cycle simulations by (among other changes) incorporating rate-dependent friction (van Dinther et al., 2013a). The effective friction coefficient \( \mu_{eff} \) is calculated as

\[
\mu_{eff} = \mu_s (1 - \gamma) + \frac{\mu_s \gamma}{1 + \frac{\gamma}{\psi}} \quad (A-5)
\]
where $\mu_s$ is the static friction coefficient, $V_c$ is the characteristic velocity, a velocity at which half of the friction change has occurred, and $\gamma$ represents the amount of slip velocity-induced weakening as

$$\gamma = 1 - \frac{\mu_d}{\mu_s}$$

(A-6)

where $\mu_d$ is the dynamic friction coefficient.

Different to conventional seismological models, gravity is included but material compression as well as inertial dynamics in terms of pressure wave propagation are neglected. The main advantages of this tool are spontaneously developing faults (van Dinther et al., 2014), as well as a self-consistently evolving absolute stress distribution. For the large scale model, the heat equation, a viscous flow law (depending on temperature, pressure and stress), fluid flow, the treatment of (de)hydration as well as erosion on the surface are taken into account (van Dinther et al., 2013b).
Figure A1: a) The setup of the seismo-thermo-mechanical model (van Dinther et al., 2013b). This subduction zone resembles the natural setting in Southern Chile. b) The setup of the laboratory-scale model and its corresponding numerical model.
Appendix B

Calculation of wavelength

An infinitely long elastic plate will buckle into sinusoidal shape under horizontal compression. Its wavelength can be predicted using equation (3-124) from Turcotte and Schubert (2002):

\[
\lambda_c = 2\pi \left( \frac{E h^3}{12(1-\nu^2)(\rho_m - \rho_w)g} \right)^{1/4}
\]

where \( \lambda_c \) is the wavelength of the buckling at the critical stress (critical horizontal force necessary to initiate buckling), \( E \) is Young’s modulus, \( h \) is the thickness of the elastic plate, \( \nu \) is Poisson’s ratio, \( \rho_m \) and \( \rho_w \) are density of mantle and water and \( g \) is acceleration due to gravity. Using

\[
G = \frac{E}{2(1+\nu)}
\]

and knowing that \( \nu = 0.5 \) in my models, I get two proportionalities:

\[
\lambda_c \sim G^{1/4}
\]

\[
\lambda_c \sim h^{3/4}.
\]

Equation B-3 predicts an increase of the wavelength by \( \sim 19\% \) for double the shear modulus, and a decrease by \( \sim 16\% \) for half the shear modulus. Equation B-4 predicts an increase of the wavelength by \( \sim 68\% \) for double the plate thickness, and a decrease by \( \sim 41\% \) for half the plate thickness.
Bibliography


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