Post-seismic viscoelastic deformation and stress transfer after the 1960 $M_{9.5}$ Valdivia, Chile earthquake: effects on the 2010 $M_{8.8}$ Maule, Chile earthquake

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SUMMARY

After the 1960 $M_{9.5}$ Valdivia, Chile earthquake, three types of geodetic observations were made during four time periods at nearby locations. These post-seismic observations were previously explained by post-seismic afterslip on the downdip extension of the 1960 rupture plane. In this study, we demonstrate that the post-seismic observations can be explained alternatively by volumetric viscoelastic relaxation of the asthenosphere mantle. In searching for the best-fitting viscosity model, we invert for two variables, the thickness of the elastic lithosphere, $H_e$, and the effective Maxwell decay time of the asthenosphere mantle, $T_M$, assuming a 100-km-thick asthenosphere mantle. The best solutions to fit the observations in four sequential time periods, 1960–1964, 1960–1968, 1965–1973 and 1980–2010, each yield a similar $H_e$ value of about 65 km but significantly increasing $T_M$ values of 0.7, 6, 10 and 80 yr, respectively. We calculate the corresponding viscoelastic Coulomb stress increase since 1960 on the future rupture plane of the 2010 $M_{8.8}$ Maule, Chile earthquake. The calculated viscoelastic stress increase on the 2010 rupture plane varies gradually from 13.1 bars at the southern end to 0.1 bars at the northern end. In contrast, the stress increase caused by an afterslip model has a similar spatial distribution but slightly smaller values of 0.1–3.2 bars on the 2010 rupture plane.

Key words: Seismic cycle; Transient deformation; Seismicity and tectonics; Subduction zone processes; Dynamics: seismotectonics; South America.

1 INTRODUCTION

Post-seismic deformation is frequently observed following large earthquakes, which might be associated with post-seismic afterslip and/or volumetric viscoelastic relaxation (Freed 2005; Wang 2007). However, the relative importance of the different mechanisms is still an open issue, mainly because the spatial and temporal coverage of the post-seismic deformation observations is very limited. The world’s largest earthquake ever recorded, the 1960 $M_{9.5}$ Valdivia, Chile earthquake, provides a unique opportunity to investigate the post-seismic deformation and the processes that contribute to the deformation.

This study uses three types of geodetic observations collected after the 1960 Chile earthquake near the rupture zone (Fig. 1). Type 1 observations are land-level changes during 1957–1964 along a leveling line in the Central Valley (triangles in Fig. 1; Plafker & Savage 1970). Type 2 observations are land-level changes during 1960–1968 at 155 survey sites along the Chilean coastline and Corcovado Gulf (dots in Fig. 1; Plafker & Savage 1970). Type 3 observations are tidal records at Puerto Montt, which is only available within two time periods, 1965–1973 (Fig. 6a) and 1980–2010 (Fig. 6b), on the Corcovado Gulf coast.

Barrientos & Ward (1990) solved for a slip model that explains the Types 1 and 2 observations. Their inverted slip patches concentrate at two depth ranges, 0–50 km (solid lines in Fig. 1) and 70–150 km (dashed lines in Fig. 1). Co-seismic slip is thought to occur only within the seismogenic zone, whose lower bound has been estimated to be in the range of 48–53 km for the Chilean region (Tichelaar & Ruff 1991). Barrientos & Ward (1990) thus suggested that the upper patches (0–50 km) correspond to co-seismic slip, while the deeper patches (70–150 km) are associated with post-seismic ‘afterslip’. In this way, Barrientos & Ward (1990) proposed an afterslip model, that is the deeper patches of their slip model (70–150 km), to explain the post-seismic signals in Types 1 and 2 observations. For Type 3 observations, Barrientos et al. (1992) inverted for another afterslip model at the depth of 50–100 km.

In this study, we show that volumetric viscoelastic relaxation of the asthenosphere mantle, as opposed to the post-seismic afterslip models (Barrientos & Ward 1990; Barrientos et al. 1992), can explain the Types 1, 2 and 3 geodetic observations equally.
2 POST-1960 VISCOELASTIC DEFORMATION

2.1 Post-1960 observations

The Type 1 and 2 observations include both co-seismic and post-seismic components. We derive the post-seismic deformation by subtracting co-seismic deformation from the total observations. The co-seismic deformation is calculated as that caused by the co-seismic slip model of Barrientos & Ward (1990) (upper patches, 0–50 km) using the Coulomb 3.3 software (Retrieved on 4 October 2010 from http://earthquake.usgs.gov/research/modeling/coulomb/; Lin & Stein 2004; Toda et al. 2005). We then assume that the remaining signals in the Type 1 and 2 data are caused by volumetric viscoelastic relaxation after the 1960 earthquake. The regions of inferred positive post-seismic deformation are located about 100–300 km away from the Chile trench axis (Fig. 4b), in
patches of Barrientos & Ward (1990) for the 1960 earthquake (solid contours in Fig. 1). We define a three-layered Earth viscosity model that includes a low-viscosity asthenospheric mantle layer with a Maxwell rheology. The asthenospheric mantle layer is sandwiched between overlying elastic lithosphere and underlying lower mantle, both of which are of high viscosities (Table 1). The shear modulus and densities of the lithosphere and asthenosphere are based on the preliminary reference Earth model (Dziewonski & Anderson 1981), increasing with depth from $2.49 \times 10^{10}$ to $6.82 \times 10^{10}$ Pa and $2.6 \times 10^3$ to $3.44 \times 10^3$ kg m$^{-3}$, respectively. Other physical parameters in our models are listed in Table 1. We invert for two variables: the thickness of the elastic lithosphere, $H_e$, and the effective Maxwell decay time of the asthenospheric mantle, $T_M$. We hold the thickness of the asthenospheric mantle, $H_a$, fixed at 100 km. We define $T_M$ as $2\eta_a/\mu_a$, where $\eta_a$ and $\mu_a$ are the viscosity and shear modulus of the asthenospheric mantle, respectively (Table 1).

### 2.3 Best-fitting viscosity models

We search for viscosity models that minimize the rms misfits between viscoelastic relaxation models and post-seismic observations. Four best-fitting solutions for the post-seismic observations in Types 1, 2 and 3 (Periods a and b), are searched within the variable domain of $H_e$ between 50 and 80 km and $T_M$ between 0.3 and 640 yr. The best-fitting solution for the Type 1 post-seismic observations is $H_e = 65$ km and $T_M = 0.7$ yr with a rms misfit of 0.25 m (Solution 1, Fig. 3a). It is slightly smaller than the 0.35 m rms misfit that we calculate for the afterslip model of Barrientos & Ward (1990). The observation error for Type 1 data is 0.1 m (Plafker & Savage 1970). We define an acceptable range for $T_M$ values, 0.4 to 1.8 yr, with rms misfits of less than 0.4 m (horizontal error bar in Fig. 3a).

The best-fitting solution for the Type 2 post-seismic observations corresponds to $H_e = 67$ km and $T_M = 6$ yr with a rms misfit of 0.67 m (Solution 2, Fig. 3b), similar to the 0.68 m rms misfit for the afterslip model of Barrientos & Ward (1990). The observation error for Type 2 data is 0.34 m (Plafker & Savage 1970). An acceptable range for $T_M$ values defined by rms misfits of less than 0.7 m is between 2.9 and 14 yr (horizontal error bar in Fig. 3b). Figs 4(a) and (b) compare the best-fitting viscoelastic solutions, afterslip model and corresponding observations for Types 1 and 2 data, respectively.

Best-fitting solutions for the Type 3 tidal records yield larger best-fitting $T_M$ values. The best-fitting $T_M$ values are 10 yr for Period a

### Table 1. Parameters for post-1960 viscoelastic relaxation models.

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Description</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>$H_e$</td>
<td>Thickness of the elastic lithosphere (km)</td>
<td>50–80</td>
</tr>
<tr>
<td>$\rho_e$</td>
<td>Range of the density of the elastic lithosphere (kg m$^{-3}$)</td>
<td>2.6–3.38 $\times 10^3$</td>
</tr>
<tr>
<td>$\lambda_e$</td>
<td>Range of the Lamé’s first parameter of the elastic lithosphere (Pa)</td>
<td>2.93–7.86 $\times 10^{10}$</td>
</tr>
<tr>
<td>$\mu_e$</td>
<td>Range of the shear modulus of the elastic lithosphere (Pa)</td>
<td>2.49–6.82 $\times 10^{10}$</td>
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<tr>
<td>$\eta_e$</td>
<td>Viscosity of the elastic lithosphere (Pa s)</td>
<td>1 $\times 10^{29}$</td>
</tr>
<tr>
<td>$H_a$</td>
<td>Thickness of the asthenospheric mantle (km)</td>
<td>100</td>
</tr>
<tr>
<td>$\rho_a$</td>
<td>Range of the density of the asthenospheric mantle (kg m$^{-3}$)</td>
<td>3.38–3.44 $\times 10^3$</td>
</tr>
<tr>
<td>$\lambda_a$</td>
<td>Lamé’s first parameter of the asthenospheric mantle (Pa)</td>
<td>7.86 $\times 10^{10}$</td>
</tr>
<tr>
<td>$\mu_a$</td>
<td>Shear modulus of the asthenospheric mantle (Pa)</td>
<td>6.70 $\times 10^{10}$</td>
</tr>
<tr>
<td>$\eta_a$</td>
<td>Viscosity of the asthenospheric mantle (Pa s)</td>
<td>1 $\times 10^{21}$</td>
</tr>
<tr>
<td>$T_M$</td>
<td>Effective Maxwell decay time of the asthenospheric mantle defined as $2\eta_a/\mu_a$ (yr)</td>
<td>0.31–640</td>
</tr>
<tr>
<td>$\eta_{lower}$</td>
<td>Viscosity of the lower mantle (Pa s)</td>
<td>1 $\times 10^{22}$</td>
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<tr>
<td>$s$</td>
<td>Co-seismic slip of the 1960 rupture (m)</td>
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<tr>
<td>$\theta$</td>
<td>Strike of the 1960 fault plane</td>
<td>7°</td>
</tr>
<tr>
<td>$\delta$</td>
<td>Dip of the 1960 slip model</td>
<td>20°</td>
</tr>
<tr>
<td>$\lambda$</td>
<td>Rake of the 1960 slip model</td>
<td>105°</td>
</tr>
<tr>
<td>$H_s$</td>
<td>Maximum depth of the co-seismic slip plane (km)</td>
<td>60</td>
</tr>
</tbody>
</table>

**Figure 3.** The rms misfits between Type 1 (Panel a) and Type 2 (Panel b) post-seismic observations and volumetric viscoelastic relaxation models. Horizontal axis ($T_M$) is on a logarithmic scale. Best-fitting solutions are indicated by large black circles: Solution 1 (Panel a, $H_e = 65$ km, $T_M = 0.7$ yr); Solution 2 (Panel b, $H_e = 67$ km, $T_M = 6$ yr). White horizontal error bars indicate $T_M$ ranges that yield acceptable rms misfits.

(Solution 3, Fig. 5a) and 80 yr for Period b (Solution 4, Fig. 5b). We also define acceptable ranges of $T_M$ values, 1.8–40 yr (horizontal error bar in Fig. 5a) and 14–450 yr (horizontal error bar in Fig. 5b), for Periods a and b, respectively. However, Type 3 data do not provide a strong constraint for $H_e$ values (Figs 5a and b). Figs 6(a) and (b) compare the best-fitting solutions with the monthly tidal records for Periods a and b, respectively.

The variations in the best-fitting $T_M$ values can be described by an empirical double exponential function:

$$\log(T_M(t)/1\text{ yr}) = A[1 - e^{-(t - t_0)/\tau}]$$

(1)

where $t$ is the time since the 1960 earthquake. Regression analysis yields $A = 1.9$, $t_0 = 2.0$ yr and $\tau = 10.5$ yr (dashed line in Fig. 7). Similar regression processes for the acceptable $T_M$ ranges yield $A = 2.7$, $t_0 = 0.8$ yr and $\tau = 7.8$ yr for the upper bound, and $A = 1.3$, $t_0 = 5.9$ yr and $\tau = 14.8$ yr for the lower bound (within in the grey region in Fig. 7).
Figure 4. Comparison between observations (black dots and error bars) and best-fitting solutions (red dots): (a) Solution 1 for Type 1 post-seismic deformation and (b) Solution 2 for Type 2 deformation. The error bars indicate observation errors in the geodetic survey (Plafker & Savage 1970). Green dots are the corresponding deformation calculated for the afterslip model of Barrientos & Ward (1990). Red and green dashed lines in (b) show our modelling results near the trench axis for the best-fitting viscoelastic and the afterslip models, respectively.

Figure 5. The rms misfits between Type 3 tidal data and viscoelastic models: (a) Solution 3 for 1965–1973 (Period a); (b) Solution 4 for 1980–2010 (Period b). Horizontal axis ($T_M$) is on a logarithmic scale. The tidal data do not provide a strong constraint for $H_e$ values. Assuming $H_e = 65$ km, we invert for the best-fitting $T_M$ solutions: Solutions 3 (Panel a, $T_M = 10$ yr); Solution 4 (Panel b, $T_M = 80$ yr). White horizontal error bars indicate $T_M$ ranges that yield acceptable rms misfits.

The four best-fitting $T_M$ values correspond to viscosities of $2 \times 10^{18}$, $2 \times 10^{19}$, $3 \times 10^{19}$ and $2 \times 10^{20}$ Pa·s, assuming an asthenospheric shear modulus of $\mu_A = 6.7 \times 10^{10}$ Pa (Table 1). The range of our best-fitting $T_M$ values between 5 and 70 yr for the time period of 1965–1988 (dashed line in Fig. 7) is consistent with the result of Piersanti (1999). The lower bound of our acceptable range of $T_M$ values between 5 and 70 yr for the time period of 1965–1988 (dashed line in Fig. 7) is similar to the result based on GPS observations (Hu et al. 2004; Fig. 7). The time-dependent increase in the $T_M$ and viscosity values might reflect either a non-linear stress-dependent rheology (Freed & Bürgmann 2004; Freed et al. 2006) or a biviscous Burgers rheology of the asthenosphere (Pollitz 2003; Hetland & Hager 2006).

3 POST-1960 STRESS TRANSFER

3.1 Coulomb stress change

For the viscosity structure that best explains the post-seismic deformation, we calculate the Coulomb stress changes on the 1960 and 2010 rupture planes caused by the viscoelastic process. The Coulomb stress change on a receiver fault is defined as:

$$\Delta CFF = \Delta \tau + \mu \Delta \sigma,$$

where $\Delta \tau$ is the shear stress change resolved on the receiver fault (positive in the slip direction), $\Delta \sigma$ is the normal stress change on
the receiver fault (positive when the fault is unclamped) and $\mu$ is the apparent friction coefficient after accounting for pore pressure, which is assumed to be 0.4 (e.g. King et al. 1994). We calculate the stress changes caused by viscoelastic relaxation using the VISCO1D code. We assume the strike, dip and rake angles for the receiver fault of the 2010 earthquake to be 10°, 20° and 90° (Lorito et al. 2011), and for the 1960 earthquake to be 7°, 20° and 105° (Barrientos & Ward 1990).

3.2 Post-1960 stress transfer

We first calculate the viscoelastic stress changes on the fault planes of the 1960 and 2010 earthquakes. Viscoelastic deformation and stress changes scale with $T_M$ (Hetland & Hager 2006), so we calculate stress changes at various locations (i.e. locations 1, 2 and 3 in Fig. 8c) as a function of the scaled time, $t/T_M$, where $t$ is the time since the occurrence of the 1960 earthquake (Fig. 9a). Since the effective $T_M$ increases with time (Fig. 7), we define averaged $T_M$ values, $\bar{T}_M(t)$, by solving the equation $t/\bar{T}_M(t) = \int_0^t 1/T_M(t)\,dt$, where $T_M(t)$ is given by eq. (1). Using $\bar{T}_M(t)$ instead of $T_M$, we transform the stress evolution curves in Fig. 9(a) to those in Fig. 9(b).

The stress changes are calculated to increase rapidly at first and then level off towards constant values after less than 10 yr. Because the stress evolution depends on temporal $T_M$ functions (eq. 1), we calculate the stress changes for the best-fitting (black line), as well as the upper and lower bound $T_M$ curves (grey region in Fig. 9b).

On the rupture plane of the 1960 earthquake, the region of calculated viscoelastic stress increase is surrounded by areas of negative stress changes on the eastern and western margins (Fig. 8d).

Regions of greater viscoelastic stress increase are located at the depth of 20–50 km, coinciding with the largest co-seismic stress drop of the 1960 earthquake (Fig. 8b). For the best-fitting $T_M$ values and corresponding evolution curve (dashed line in Fig. 7), the maximum stress increase is about 40 bars at the depth of ∼30 km. The maximum stress increase corresponding to the lower and upper bounds of the temporal $T_M$ evolution curves (grey region in Fig. 7) is calculated to be about 30 and 45 bars, respectively. Thus, the viscoelastic stress changes tend to compensate for the co-seismic stress drop on the rupture plane of the 1960 earthquake. Meanwhile, the viscoelastic stress increase corresponding to the best-fitting $T_M$ curve (dashed line in Fig. 7) is calculated to be 13.1 bars at the southern end of the 2010 rupture plane (location 1 in Fig. 8c), gradually decreasing to 0.1 bars at the northern end (location 3). The hypocentre of the 2010 earthquake (location 2) is calculated to have experienced a viscoelastic stress increase of 0.5 bars. We also calculate the stress increases on the 2010 rupture plane for the lower and upper bounds of the temporal $T_M$ evolution curves (grey region in Fig. 7): Location 1 (9.2 and 16.4 bars), location 2 (0.3 and 0.6 bars) and location 3 (0.04 and 0.2 bars; grey regions in Fig. 9b).

We also calculate the stress changes caused by the afterslip model of Barrientos & Ward (1990) using the Coulomb 3.3 software. The afterslip also loads both the 1960 and 2010 rupture planes (Figs 8e and f). On the 1960 rupture plane, the region of greater stress increase due to afterslip is at the depth of about 60 km (Fig. 8f), which is deeper than that for the volumetric viscoelastic relaxation model (Fig. 8d). On the 2010 rupture plane, the afterslip stress increase has a similar along-strike distribution (Fig. 8e) as that
of volumetric viscoelastic relaxation (Fig. 8c), but with a smaller maximum stress increase (location 1) of 3.2 bars, compared with 13.1 bars for the viscoelastic relaxation. At the hypocentre of the 2010 earthquake (location 2), the stress increase caused by the afterslip and volumetric viscoelastic relaxation is similar: 0.4 bars (Fig. 8e) and 0.5 bars (Fig. 8c). In comparison, the calculated co-seismic stress increase at the hypocentre of the 2010 earthquake (location 2) caused by the 1960 earthquake is about 0.3 bars (Lin & Stein 2004; Fig. 8a), which is of the same order of magnitude as the post-seismic viscoelastic and afterslip stress increase.

4 DISCUSSION

4.1 Uncertainties in the Type 1 and 2 post-seismic deformation data

Barrientos & Ward (1990) and our study assume that Type 1 and 2 observations contain co-seismic and post-seismic deformation signals. Type 1 and 2 observations may contain other information as well. For example, Linde & Silver (1989) suggested that these observations include pre- and co-seismic signals. Moreno et al. (2009) attributed all the deformation to co-seismic slip by using a curved geometry rather than a plane fault. Despite alternative interpretations for the Type 1 and 2 observations, our study shows that the post-seismic deformation, which was previously interpreted as from the afterslip model of Barrientos & Ward (1990), could be equally well explained by the volumetric viscoelastic relaxation of the asthenosphere mantle.

4.2 Viscoelastic relaxation versus afterslip

Our results show that the volumetric viscoelastic relaxation and afterslip could equally explain the geodetic observations in the distance of about 70–370 km away from the Chile trench axis (Fig. 4b). Data that could possibly distinguish the two post-seismic mechanisms include: (1) near-trench seafloor observations (e.g. Newman 2011), (2) long-term geodetic observations or (3) high-density horizontal deformation observations. Our calculations show that the predicted deformation between the volumetric viscoelastic and afterslip models is similar at the distance of more than 70 km from the trench axis, but is different near the Chile trench axis (Fig. 4b). If future seafloor deformation observations are collected, the afterslip and volumetric viscoelastic relaxation models could be distinguished from each other. Another method of distinction is to use long-term geodetic observations: Post-seismic viscoelastic relaxation usually continues within tens of years (Hetland & Hager 2006), while afterslip occurs only within several years after a great earthquake (Helmltter & Shaw 2009). Finally, Wang et al. (2007) argued that the opposing horizontal deformation directions observed at coastal and inland sites of high-density GPS surveys could be explained by post-seismic viscoelastic relaxation, but not afterslip.

4.3 Stress transfer versus tectonic loading

The importance of stress transfer mechanisms to trigger an earthquake must be understood within the framework of the stress evolution in an entire earthquake cycle. The earthquake cycle of the 2010 M8.8 Maule, Chile earthquake could be considered to begin in 1835, when a M8.5 earthquake occurred in the same segment as the 2010 earthquake (Moreno et al. 2010; Beck et al. 1998). During its earthquake cycle, the 2010 rupture plane is loaded by plate convergence due to the coupling between the subducting Nazca Plate and the overriding South America Plate. We estimate this secular tectonic stress using a GPS-inverted plate-coupling model with variable coupling coefficient (Moreno et al. 2010), assuming a plate convergence rate of 6.6 cm yr$^{-1}$ (Angermann et al. 1999). Our calculations show that the tectonic stress increase and the 2010 co-seismic stress drop are of the similar ‘two-patch’ spatial patterns (Figs 10a and b). The 2010 earthquake is calculated to have released a similar magnitude of stress as the stress accumulated by the secular tectonic loading in 1835–2010.

The above similarities also suggest that the 2010 rupture plane was close to failure before the occurrence of the 2010 Maule earthquake. Therefore, a relatively small amount of additional stress increase, such as that due to co- and post-seismic processes caused by the nearby 1960 earthquake, could contribute to the triggering of the 2010 earthquake.

We also estimate the residual stress after the occurrence of the 2010 earthquake by adding the 1835–2010 tectonic and 2010 co-seismic stresses. The post-1835 tectonic stress accumulated on the slip patch to the north of the 2010 hypocentre seems to be fully released, while the southern patch is still associated with a certain...
amount of residual stresses (Fig. 10c), similar to the interpretations of Moreno et al. (2010) based on slip deficit calculation.

5 CONCLUSIONS

Our investigation of the post-seismic deformation of the 1960 M9.5 Valdivia, Chile earthquake and the associated stress transfer yields the following results:

1. Three types of post-1960 geodetic observations during four time periods could be explained by a volumetric viscoelastic relaxation model equally well as the previous afterslip models. The observations in four sequential time periods could be explained by a consistent elastic lithosphere thickness of about 65 km, but require a significant increase in the effective Maxwell decay times of the asthenospheric mantle.

2. The inverted effective Maxwell decay times of the asthenosphere mantle are 0.7, 6, 10 and 80 yr to best explain the observations during the time periods of 1960–1964, 1960–1968, 1965–1973 and 1980–2010, respectively.

3. The rupture plane of the 2010 Maule, Chile earthquake is calculated to have experienced stress increase of 0.1–13.1 bars due to post-1960 viscoelastic relaxation, in comparison to 0.1–3.2 bars due to afterslip. These post-1960 stress increases might have contributed to the triggering of the 2010 earthquake.

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REFERENCES


APPENDIX A: MODEL SENSITIVITY OF VISCOELASTIC DEFORMATION

The calculated post-seismic viscoelastic deformation depends on two key variables: the thickness of elastic lithosphere, $H_e$, and the effective Maxwell decay time of the asthenosphere mantle, $T_M$. We assume a 100-km-thick asthenosphere mantle.

1. Types 1 and 2 observations

With decreasing $H_e$ values, the calculated Type 1 deformation becomes more positive (Fig. A1a). Meanwhile, the distance of the region of the calculated maximum uplift from the trench axis decreases for Type 2 data (Fig. A1b). For both Type 1 (Fig. A2a) and Type 2 (Fig. A2b) data, a reduction in $T_M$ leads to more positive calculated deformation.

2. Type 3 observations

The calculated post-1960 deformation at the Puerto Montt station increases exponentially with time (Fig. A3). For both Period a (1965–1973, Fig. A3a) and Period b (1980–2010, Fig. A3b) data, all the $H_e$ values in the range of 50–80 km explain the observations well, assuming a constant $T_M$. Thus, Type 3 observations do not provide a strong constraint for $H_e$ values. For both Period a (Fig. A3c) and Period b (Fig. A3d), a reduction in $T_M$ leads to an increase in the calculated rate of deformation, assuming a constant $H_e$.

Figure A1. Calculated deformation for (a) Type 1 and (b) Type 2 observations assuming different lithospheric thickness: $H_e = 50$ km (red), 60 km (orange), 70 km (magenta), 80 km (blue). Green curves correspond to the afterslip model (Barrientos & Ward 1990). Black dots with error bars indicate the corresponding post-1960 land-level changes. Viscoelastic models assume constant $T_M$ values of 0.7 yr (Solution 1) in (a) and 6 yr (Solution 2) in (b).

Figure A2. Calculated Type 1 and 2 deformation for viscoelastic models with different $T_M$ values: best-fitting $T_M$ (red), as well as lower (blue) and upper (purple) bounds of the acceptable $T_M$ ranges. Viscoelastic models assume a constant $H_e$ value of 65 km. Green curves correspond to the afterslip model (Barrientos & Ward 1990). Black dots with error bars indicate the corresponding post-1960 land-level changes. (a) For Type 1 observations. (b) For Type 2 observations.

Figure A3. Calculated deformation at the tidal station Puerto Montt for two time intervals: (a, c) Period a of 1965–1973 and (b, d) Period b of 1980–2010. Black dots are the observed monthly time-series. Post-seismic viscoelastic models with different $H_e$ and $T_M$ values are shown by different colours. (a–b) Modelling results for different $H_e$ values, 50, 60, 70 and 80 km, assuming $T_M$ to be 10 yr in (a) and 80 yr in (b). (c–d) Modelling results for different $T_M$ values: best-fitting $T_M$ (red), as well as lower (blue) and upper (purple) bounds of the acceptable $T_M$ ranges, assuming $H_e = 65$ km.