Modelling of glacially-induced stress changes in Latvia, Lithuania and the Kaliningrad District of Russia

Holger Steffen, Rebekka Steffen and Lev Tarasov


Abstract. We model the change of Coulomb Failure Stress (δCFS) during the Weichselian glaciation up until today at 12 locations in Latvia, Lithuania and Russia that are characterised by soft-sediment deformation structures (SSDS). If interpreted as seismites, these SSDS may point to glacially-induced fault reactivation. The δCFS suggests a high potential of such reactivation when it reaches the instability zone. We show that δCFS at all 12 locations reached this zone several times in the last 120,000 years. Most notably, all locations exhibit the possibility of reactivation after ca. 15 ka BP until today. Another time of possible activity likely happened after the Saalian glaciation until ca. 96 ka BP. In addition, some models suggest unstable states after 96 ka BP until ca. 28 ka BP at selected locations but with much lower positive δCFS values than during the other two periods. For the Valmiera and Rakuti seismites in Latvia, we can suggest a glacially-induced origin, whereas we cannot exactly match the timing at Rakuti. Given the (preliminary) dating of SSDS at some locations, at Dyburiai and Ryadino our modelling supports the interpretation of glacially-induced fault reactivation, while at Slinkis, Kumečiai and Liciškėnai they likely exclude such a source. Overall, the mutual benefit of geological and modelling investigations is demonstrated. This helps in identifying glacially-induced fault reactivation at the south-eastern edge of the Weichselian glaciation and in improving models of glacial isostatic adjustment.

Keywords: Glacial Isostatic Adjustment; Earthquake; Coulomb Failure Stress; Finite element modelling; Baltic countries; Soft-sediment deformation structure; Seismites

INTRODUCTION

Repeated glaciations during the Pleistocene (2,588.0 to 11.7 thousand years before present (ka BP), Cohen et al. 2013 updated) have fundamentally shaped the surface of the Earth, especially due to the development and advance of large continental-scale ice sheets. The interaction of ice sheets with the solid Earth is summarised in the term glacial isostatic adjustment (GIA) and recognises several strongly tied processes (Whitehouse 2018). They include, among others, surface deformation as well as changes in the gravitational potential field, rotation, and stress field of the Earth (Steffen, Wu 2011). Ice sheets also removed water from the oceans so that mean sea level during the Last Glacial Maximum (LGM) was ca. 130–134 m lower than today (Lambeck et al. 2014; Peltier et al. 2015).

We focus herein on changes in the stress field and their consequences. The weight of an ice sheet bends the lithosphere and induces additional stresses to e.g., lithostatic pressure-induced and tectonic background stresses, in both the vertical and horizontal stress components. We use the term GIA stresses (Steffen et
al. 2014b) for these additional stresses. The vertical GIA stresses vanish when the ice melted completely but due to the visco-elastic nature of the Earth’s mantle the lithosphere is still deformed and readjusts only slowly towards isostatic equilibrium (Steffen et al. 2014a). Therefore, horizontal GIA stresses remain but can be released in earthquakes along pre-existing faults (Steffen et al. 2014a). This is evidenced, for example, in northern Fennoscandia in more than a dozen fault scarps of up to 30 m height (Mikko et al. 2015). Such faults are known as postglacial faults but are nowadays termed glacially-induced faults (GIF) (Lund 2015). Next to northern Fennoscandia, there is also evidence of such faults and traces of glacially-triggered earthquakes in Denmark, northern Germany, Poland and the United Kingdom (Sanderson, Jørgensen 2015; Brandes et al. 2018; Brandes et al. 2012; Grube 2019; Hoffmann, Reicherter 2012; Pisarska-Jamrožy et al. 2018; van Loon, Pisarska-Jamroży 2014; Stewart et al. 2001) as well as in North America (Fenton 1994). GIFs are found not only in the centre of former ice sheet locations but also at their margins and beyond (Brandes et al. 2012, 2015; Druzhinina et al. 2017; Sandersen, Jørgensen 2015; Stewart et al. 2000).

A few studies (e.g. Bitinas, Lazauskienė 2011; van Loon et al. 2016; Pisarska-Jamroży, Bitinas 2018; Pisarska-Jamroży et al. this issue) discuss GIFs or glacially-triggered seismicity in the Baltic area, which represents the south-eastern corner of the Scandinavian Ice Sheet (SIS) during the Weichselian glaciation (ca. 115.0 to 11.7 ka BP), the last of past north-European glaciations. Many tectonic faults have been identified in the crystalline basement of this area (Fig. 1). In view of the modelling results by Brandes et al. (2012, 2015, 2018) for central Europe (the south-western corner of the SIS), it is reasonable to assume that GIFs also developed in the countries of Latvia, Lithuania and Russia (Kaliningrad District), as has already been suggested by Bitinas, Lazauskienė (2011). This view is supported by prominent earthquakes in the East Baltic area in historic and recent times (Nikuļins 2011, 2019; Pačėsa, Šliaupa 2011) such as the 1616 earthquake in Latvia (Doss 1910) and the 2004 Kaliningrad earthquake (Uломов et al. 2008). These could be connected to the past glaciation in the same way as Brandes et al. (2015) and Brandes et al. (2019) suggested for historic and recent earthquakes in Germany.

In the Baltic countries and the Kaliningrad District of Russia, seismic activity is proposed for a dozen locations discussed in the recent literature (Fig. 1). These locations are mostly characterized by soft-sediment deformation structures (SSDS) (see Fig. 2, for example) that, if interpreted as seismites, may point to glacially-triggered seismicity in the near-field of potentially reactivated faults (Fig. 1, Table 1). At some locations the trigger remains unclear. For other locations, triggers such as collapse of river walls are more likely to have generated seismites-like SSDS (e.g. at the Liciškėnai outcrop, see Table 1).

The aims of our study are therefore: (i) to analyse modelled GIA stress changes in view of their poten-
Table 1 Overview of locations with known or assumed glacially-induced seismicity in Latvia, Lithuania and the Kaliningrad District of Russia. See Fig. 1 for geographical distribution. Lat. = Latitude, Long. = Longitude, SSDS = Soft-sediment deformation structures

<table>
<thead>
<tr>
<th>Location</th>
<th># in Fig. 1</th>
<th>Lat./Long.</th>
<th>Brief description</th>
<th>Timing information</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Särnate outcrop, Latvia</td>
<td>1</td>
<td>57.07 N 21.42 E</td>
<td>Lake or lagoon environment. SSDS in two units with hiatus, likely one single event, palaeoseismic event possible</td>
<td>Ice-free ca. 14.0 ka, age of organic deposits 7.73 ka and younger</td>
<td>Nartišs et al. (2018)</td>
</tr>
<tr>
<td>Baltmuža site, Latvia</td>
<td>2</td>
<td>56.93 N 21.26 E</td>
<td>Lacustrine sediments with 3 SSDS horizons, palaeoseismic event possible</td>
<td>Deposition 28.6–23.4 ka, ice-free ca. 14.0 ka</td>
<td>Belzyt et al. (2018a)</td>
</tr>
<tr>
<td>Valmiera section, Latvia</td>
<td>3</td>
<td>57.55 N 25.44 E</td>
<td>Glaciofluvial sediments, palaeoseismic event very likely</td>
<td>Deposition &lt;14.5 ka</td>
<td>Van Loon et al. (2016)</td>
</tr>
<tr>
<td>Rakuti section, Latvia</td>
<td>4</td>
<td>55.90 N 27.09 E</td>
<td>Glacial lacustrine sediments, palaeoseismic event very likely</td>
<td>Deposition 17.0–16.0 ka</td>
<td>Van Loon et al. (2016)</td>
</tr>
<tr>
<td>Girulai megaslides, Lithuania</td>
<td>5</td>
<td>55.75 N 21.09 E</td>
<td>360 m long mega-landslide, hypothetically triggered by earthquake</td>
<td>Happened 7.7 ka or any time thereafter</td>
<td>Damušytė, Bitinas (2018), Bitinas et al. (2016)</td>
</tr>
<tr>
<td>Juodkiai quarry, Lithuania</td>
<td>6</td>
<td>55.61 N 21.35 E</td>
<td>Glaciofluvial delta with SSDS, not further investigated</td>
<td>Not available (Late Weichselian)</td>
<td>Bitinas, Damušytė (2018)</td>
</tr>
<tr>
<td>Ventės Ragas outcrop, Lithuania</td>
<td>7</td>
<td>55.35 N 21.20 E</td>
<td>Sandy lacustrine and aeolian sediments with SSDS, not further investigated</td>
<td>Not available (Late Weichselian)</td>
<td>Bitinas, Damušytė (2018)</td>
</tr>
<tr>
<td>Dyburiai outcrop, Lithuania</td>
<td>8</td>
<td>55.94 N 21.60 E</td>
<td>Glaciofluvial inter-moraine sediments, palaeoseismic event very likely</td>
<td>Deposition 119.7–91.1 ka, ages subject to debate</td>
<td>Pisarska-Jamroży et al. (2018b)</td>
</tr>
<tr>
<td>Slinkis outcrop, Lithuania</td>
<td>9</td>
<td>55.09 N 23.45 E</td>
<td>Meandering fluvial system sediments with trapped SSDS, palaeoseismic event or glacial earthquake suggested</td>
<td>Deposition 24.0–21.2 ka</td>
<td>Belzyt et al. (2018b), Pisarska-Jamroży et al. (this issue)</td>
</tr>
<tr>
<td>Kumečiai outcrop, Lithuania</td>
<td>10</td>
<td>55.06 N 23.42 E</td>
<td>Fluvial meandering system sediments with several layers of SSDS, palaeoseismic event unlikely</td>
<td>Deposition 76.0–46.7 ka, ages subject to debate</td>
<td>Pisarska-Jamroży et al. (2018c)</td>
</tr>
<tr>
<td>Liciškėnai outcrop, Lithuania</td>
<td>11</td>
<td>54.60 N 24.21 E</td>
<td>Glaciofluvial sediments with SSDS, palaeoseismic event unlikely</td>
<td>Deposition 74.2–51.7 ka, ages subject to debate</td>
<td>Woronko et al. (2018)</td>
</tr>
<tr>
<td>Ryadino archaeological excavation, Russia</td>
<td>12</td>
<td>55.03 N 22.20 E</td>
<td>Glacial lacustrine sediments with SSDS, palaeoseismic event likely</td>
<td>Deposition 8.7–7.5 ka</td>
<td>Druzhinina et al. (2017)</td>
</tr>
</tbody>
</table>

In general, a GIA model consists of an earth model characterized by its specific rheology as well as an ice load history (Whitehouse 2018). Both parts affect the results and their fit to observations and thus there is still no consensus on the best GIA model for the whole of Fennoscandia (Steffen, Wu 2011). Many different ice-earth model combinations give reasonable fits to selected observational datasets and are thus advocated by respective groups (e.g., Lambeck et al. 2010; Pelletier et al. 2015). We test several of such combinations that are discussed for our investigation area to show the spread of possible results so that the reader can get a feeling for the uncertainty of the modelling.

We apply the FE software ABAQUS (ABAQUS 2018) to create a three-dimensional model of the lithosphere and mantle in Fennoscandia. We follow the flat-Earth approach outlined in Wu (1992, 2004) and Steffen et al. (2006), which has been shown to agree well with previous numerical solutions of GIA (Wu, Johnston 1998; Spada et al. 2011). The model consists of a centre of 4500 km × 4500 km size and a frame that extends the model horizontally to a size of 60,000 km × 60,000 km. This allows mantle material to flow outside the central area and minimize associated numerical errors (Steffen et al. 2006). The model reaches the core-mantle boundary at 2891 km depth (Table 2) which has been shown to be appropriate when working with continental-scale ice sheets (Steffen et al. 2015). The centre has 90 × 90 hexahedral elements of 50 km × 50 km extent in the horizontals, while the elements of the frame are variable in size, but side length increases from the centre to the edge. The sides of the model are fixed with rigid boundary conditions.

**MODELLING**

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The lithosphere is composed of model-dependent (due to lithospheric thickness) 8 or 9 element layers with the first 6 layers having 5 km thickness each. It represents an elastically behaving outer shell of the Earth whose thickness corresponds to loading processes such as GIA of time-scales of some 1000s to 100,000 years and should therefore not be confused with seismically or geologically derived lithospheric thicknesses (Eaton et al. 2009). The upper mantle down to 670 km has 4 or 5 element layers, and the lower mantle 5 element layers, see Table 2 for an overview. Material parameters density, shear modulus and Young’s modulus are volume-averaged values derived from the Preliminary Reference Earth Model (PREM; Dziewonski, Anderson 1981).

Previous studies (Brandes et al. 2012, 2015, 2018; Steffen et al. 2014b) have shown that the Earth model composition affects the timing of possible seismic activity. We therefore test a variety of Earth models (Table 3). The parameters represent commonly used Earth models on a global scale and dedicated models for northern Europe (Kierulf et al. 2014; Lambeck et al. 2010; Peltier et al. 2015; Steffen, Wu 2011). We thereby consider models with thinner (Lambeck et al. 2010; Peltier et al. 2015) and thicker (Kierulf et al. 2014) lithospheric thicknesses as well as softer (Peltier et al. 2015) and harder (Lambeck et al. 2010) lower mantle viscosity that are discussed in the literature. We also test models with lateral heterogeneities in either lithospheric thickness or mantle viscosity: Wang, Wu (2006) presented a global model of lithospheric thickness which we implement, see Brandes et al. (2018), for more information. Becker, Boschi (2002) provided the global seismic tomography model SMEAN, whereas we use the update SMEAN2 from 2016, which is converted to viscosity variations following Steffen et al. (2006) and Wu et al. (2013).

We test three different data-constrained ice chronologies as the load for the Earth models. The first is the commonly used global model ICE-6G by Peltier et al. (2015), version ICE-6G_C, at a spatial resolution of 0.5 × 0.5 degrees. The model has 500-year time steps from 26 ka BP to today. We use the North-European part of this model and add another sawtooth-type glaciation cycle from 216 to 126 (full load) to 116 ka BP (load-free) to increase result accuracy (Johnston, Lambeck 1999), and then crudely increase the load linearly until 26 ka BP. The second ice history model ANU-ICE is a combination of two regional ANU-ICE ice history models for the British Isles (Lambeck 1995; spatial resolution 0.125 × 0.25 degrees) and Fennoscandia together with the Barents and Kara seas (Lambeck et al. 2010; spatial resolution 0.25 × 0.5 degrees). This model spans two glaciation cycles from 240 ka BP to today. Time steps vary between 3 and 45 000 years. The third ice history model is the European part (Fennoscandia, the Barents/Kara seas and the British Isles) of GLAC (Tarasov et al. 2012; Tarasov 2013; Nordman et al. 2015), model number 71340, at a spatial resolution of 0.25 × 0.5 degrees. The ice model is implemented in the GIA model in 500 (during deglaciation) to 2000 years (during glaciation) time steps. For all three chronologies, corresponding ocean loading is not included, which according to Steffen et al.

Table 2 Vertical dimensions and subdivision of models with fixed lithospheric thickness

<table>
<thead>
<tr>
<th>Layer</th>
<th>Thickness [km]</th>
<th>Depth [km]</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>5</td>
<td>5</td>
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<td>2</td>
<td>5</td>
<td>10</td>
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<td>3</td>
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<td>7</td>
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<td>60</td>
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<tr>
<td>8</td>
<td>30</td>
<td>90</td>
</tr>
<tr>
<td>9</td>
<td>30/50/70</td>
<td>120/140/160</td>
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<tr>
<td>10</td>
<td>90/110/130</td>
<td>250</td>
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<tr>
<td>11</td>
<td>170</td>
<td>420</td>
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<tr>
<td>12</td>
<td>151</td>
<td>571</td>
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<tr>
<td>13</td>
<td>100</td>
<td>671</td>
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<tr>
<td>14</td>
<td>329</td>
<td>1000</td>
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<tr>
<td>15</td>
<td>330</td>
<td>1330</td>
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<tr>
<td>16</td>
<td>500</td>
<td>1830</td>
</tr>
<tr>
<td>17</td>
<td>500</td>
<td>2330</td>
</tr>
<tr>
<td>18</td>
<td>561</td>
<td>2891</td>
</tr>
</tbody>
</table>

(2006) can affect the quantities, i.e. near the coasts in Fennoscandia, by about one order of magnitude of the signal. This uncertainty is well accommodated by the signal spread of tested earth and ice models.

The potential of triggering earthquake activity at a certain location is tested by calculating and analysing δCFS (see Brandes et al., 2015, for details). The δCFS represents the minimum stress required to reach faulting in a certain stress regime (compressional, neutral (strike-slip) or extensional). A negative δCFS value thereby indicates that the fault is stable, while a positive value means that GIA stress can induce faulting, e.g. so that the accumulated stress is temporarily released by an earthquake. For better visualization we add the zero line in Figs. 3–7 to indicate the threshold where such onset of fault motion is possible. This separates each diagram into a zone of stability (below the zero line) and a zone of instability (above the zero line).

Although δCFS calculation is straightforward, there are many input parameter and assumptions that have to be made. For this study we assume the faults are optimally oriented. Thus, their strike and dip values promote faulting for a commonly used friction coefficient of 0.6 and faults are perpendicular to the maximum horizontal direction of the tectonic background stress. Inspection of the tectonic faults in Fig. 1 shows different fault orientations, and some are oriented in a nearly optimal sense, which supports our assumption. At the same time, the faults are assumed to be close to failure at the onset of glaciation which has been shown to be a reasonable assumption (Steffen et al., 2014a). This agrees with findings by Zoback, Townend (2001) and Zoback, Zoback (2015), who concluded that the intraplate crust is generally critically stressed and thus close to failure. Pore-fluid pressure is not investigated. The stress difference between maximum (S1) and minimum (S3) background stress is implemented in form of the R-value (stress ratio, Gephart, Forsyth 1984) and set to 0.5, which means that the average of minimum and maximum background stress is used for the intermediate stress S2. The δCFS is calculated for 12.5 km depth as our tests (not shown) yield only small depth-dependent differences that are within the spread of the Earth and ice models tested. The general δCFS behaviour (trends and turning points) is the same.

RESULTS

Figure 3 shows the δCFS at the 12 locations of Table 1 and Fig. 1 for the last 26,000 years if only ice model ICE-6G_C is used in combination with the 8 Earth models. At first, we consider a compressional background stress field which implies a thrust-faulting mechanism. A compressional background stress field is reasonable to assume based on the World Stress Map (Heidbach et al. 2018). During the LGM from ca. 26 to 19 ka BP (Clark et al. 2009), the δCFS is strongly negative, thus in the stability zone, at all locations. The δCFS spread between the different Earth models is ca. 1.0–1.5 MPa, whereas the models with thicker lithosphere of 120 km and larger are more negative. Between ca. 18 and 14 ka BP (location-dependent) δCFS strongly increases approaching or crossing the zero line, depending on the Earth model. In the Latvian locations 1–3, δCFS mainly crosses the zero line between 12 and 5 ka BP. At Rakuti (4) one model would reach the instability zone today. In the west-Lithuanian locations (5–8), the zero line is crossed earlier than in Latvia, at ca. 14 ka BP. After 5 ka BP, the δCFS of all Earth models is in the instability zone. The δCFS at the interior Lithuanian locations (9–11) as well as at Ryadino (12) in Russia reach instability earliest at ca. 15 ka BP. Choosing a different Earth model can move the zero-crossing close to today, e.g. at Liciškėnai (11). In general, Earth models with thinner lithosphere (90 km) reach instability earlier than those with thicker lithosphere or with lateral lithosphere thickness variations. The δCFS difference between all models today is less than 1 MPa.

As can be seen in Fig. 3, selecting Earth models L090_520_L221 and L160_SMEAN2 largely envelops the spread of tested models. Also, one can see that the δCFS behaviour in nearby locations is very similar, so we show the result for only one of the locations 1 and 2, 5–8, and 9–11, respectively. For clarity, we thus limit remaining analyses to those two Earth models, simply called 1D and 3D in the figures, and 6 different locations (1, 3, 4, 8, 10 and 12).

In Fig. 4, we also show δCFS results at those 6 locations for ice models ANU-ICE and GLAC. Clearly, both models yield different curves compared to ICE-6G_C. Maximum negative values during LGM reach more than 14 MPa and the increase thereafter is much steeper than for ICE-6G_C. Considering the 1D model, the zero line at Sārnate (1) is crossed with ANU-ICE 5000 years earlier than with ICE-6G_C. In turn, considering the 3D model, the zero line is crossed up to 2000 years later. Hence, possible onset of fault reactivation at Sārnate is suggested between 14–7 ka BP based on Figs. 3 and 4.

At Valmiera (3), the results with GLAC suggest a phase of possible GIA induced seismic activity 17–15 ka BP, then δCFS drops and may reach the instability threshold later again after deglaciation is completed. ANU-ICE 1D points to a possible short activity at 14 ka BP, then stable conditions until instability is reached again after deglaciation. Overall, fault reactivation at Valmiera could have been possible 17–15, 14, and after 9.5 ka BP based on our modelling results.
At Rakuti (4), some models reach the instability zone after ca. 11.5 ka BP, while others may only reach them in the future. At Dyburiai (8), activity could happen earlier (ca. 15.5 ka BP) than based on results with ICE-6G_C only, while at Kumečiai (10) and Ryadino (12) timing is like that of ICE-6G_C. As for Rakuti, some δCFS curves of 3D models for Kumečiai do not cross the zero line but suggest it for the future. The GIA model spread after 12 ka BP is less than 2 MPa at all locations.

Some sediments in the SSDS horizons of the locations of Table 1 were dated to get insight into their possible time of generation. At Dyburiai, Kumečiai and Liciškėnai times much older than Late Weichselian were retrieved. We therefore show in Figs. 5 and 6 δCFS at 6 locations over the last 120,000 years to cover those times of possible activity. We only plot results for ANU-ICE and GLAC as ICE-6G_C is not available before LGM. At all locations, positive values are found from 120 ka BP (thus during Late Saalian times) until ca. 96 ka BP. Then, depending on the ice model, the zero line is touched or crossed at many different times until the Late Weichselian. The positive values are low and the gradients are rather flat before the zero line is reached. Any activity between ca. 60 and 50 ka BP at any of the locations can be excluded based on the modelling results.

The results so far assume a compressional back-
ground stress regime. Fig. 7 shows example results for an extensional (normal faulting, dotted) and neutral (strike-slip, dashed) background stress regime for all three tested ice models in combination with the 1D Earth model. In Lithuania and Russia as well as at the coastal locations of Latvia, fault reactivation can be excluded after 22 ka BP under normal faulting or strike-slip conditions. In Valmiera and Rakuti in Latvia, the instability threshold is reached (dependent on ice model) during LGM and a short period thereafter, but then stable conditions remain after 13–11 ka BP. At all locations, fault reactivation due to an extensional or neutral background stress regime is only found for the ANU-ICE and GLAC ice models, while it can be excluded if ICE-6G_C is applied.

DISCUSSION

Considering the generally accepted compressional background stress regime for our area under investigation, our results support the findings of glacially-induced seismites for the locations of Valmiera and Rakuti in Latvia as was suggested by van Loon et al. (2016). They suggested a deposition after 14.5 ka BP for Valmiera (Table 1) and our modelling can match this point in time with ANU-ICE (Fig. 4). Using GLAC or ICE-6G_C times after 14.5 ka BP would also be possible, but mostly after 12 ka BP. GLAC also shows potential activity between ca. 17–15 ka BP. If an extensional background stress regime is assumed (Fig. 7), only GLAC would point to potential activity at exactly the time documented in van Loon et al. (2016). Hence, our different modelling results strongly support that SSDS at Valmiera are seismites due to glacially-induced seismicity. If a better dating of seismites would be achieved at Valmiera this may help in constraining our GIA models. Currently, ANU-ICE could be preferred but we note that we have made many assumptions whereas pore fluids can effectively alter the fault reactivation potential (Ranalli 1995), so that ICE-6G_C and GLAC model 71340 cannot be excluded yet.

Similarly, our models suggest activity at Rakuti after ca. 12 ka BP (Figs. 3 and 4). This does not match the findings of van Loon et al. (2016), who propose 17–16 ka BP. However, some models are very close to the zero line during that time, so we speculate that increased pore fluids, e.g. due to meltwater, could have acted as final trigger. Rakuti can therefore serve as a very crucial location to exclude some model configurations. Some models might be less preferable, if e.g. pore fluids are disregarded. For example, a model with thinner lithosphere of 90 km and either ice model ICE-6G_C or ANU-ICE would fit better than a 3D model with GLAC ice history. We note though that we have not tested other GLAC models which could potentially perform better.

Our results also support preliminary investigations of SSDS at Sārname in Latvia (Nartīs et al. 2018) and Ryadino in Russia (Druzhinina et al. 2017), which could be glacially-induced seismites. Deposits at those locations cover a period of roughly 9–7 ka BP, which is well covered by many of our models independent of the chosen ice and Earth model combination (Figs. 3 and 4).
At Baltmuža, Belzyt et al. (2018a) provide deposition ages of 28.6–23.4 ka. There is no model that shows glacially-induced activity during this time (Figs. 3, 4 and 7). Moreover, the δCFS of most models is strongly negative so that we would exclude a glacial source of these SSDS.

The Giruliai mega-landslide likely happened 7.7 ka BP or thereafter (Bitinas et al. 2016; Damušytė, Bitinas 2018). All our modelled δCFS are past the instability threshold (Figs. 3 and 4) and thus a nearby fault could have been reactivated due to GIA. This could have triggered the earthquake that led to the mega-landslide. However, this remains very speculative and more research needs to be undertaken to find further evidence such as the presence of nearby faults and date-able seismites.

SSDS from Juodikiai and Ventės Ragas have yet to be dated and it is still unclear if these can be interpreted as seismites (Bitinas, Damušytė 2018). Our modelling suggests unstable conditions after 15 ka BP for those locations (Figs. 3 and 4). Both sediments are Late Weichselian in age and thus glacially-induced activity cannot be excluded. However, SSDS should be dated and investigated if they can be regarded as seismites.

In Dyburiai, dating of sediments points to a very early SSDS development between 119.7–91.1 ka BP which is possibly supported by our modelling results for both ANU-ICE and GLAC (Fig. 6). However, our models cannot limit the time span nor suggest a more certain point of seismic activity. We must also note that the δCFS is only slightly positive and small model adjustments could move the curves to stable conditions. In this regard, sophisticated geological investigations and dating could help in constraining the GIA models and exclude certain model setups.

At the Lithuanian locations of Slinkis, Kumečiai and Liciškėnai, we suggest other mechanisms as potential sources of the SSDS, as was discussed by Belzyt et al. (2018b); Pisarska-Jamroży et al. (2018c) and Woronko et al. (2018), respectively. During the times specified in Table 1, our models exhibit mainly...
stable conditions (Figs. 3 and 6). For the location of Slinkis, this has interesting implications in view of the recent findings by Pisarska-Jamroży et al. (this issue), who suggest that “two well-defined layers with SSDS (dominated by water/sediment-escape structures and accompanying load structures) are linked to at least two phases of liquefaction processes, which may have been triggered by seismic shock derived from glacial isostatic earthquake or glacial earthquake”. With our current model setup, we would exclude an earthquake due to GIA. However, if a glacial earthquake could also be excluded at Slinkis, this location can, as Rakuti in Latvia, serve as a very crucial location to exclude some model configurations.

Independent of the background stress field, all curves exhibit low gradients after 15–13 ka, with many in the instability zone or very close to the threshold, which results in a long interval where activity could be initiated. For some locations, several models cross the zero line around present day. Hence, GIA cannot be excluded as trigger of historic and quite recent earthquakes in the Baltic countries and the Kaliningrad District of Russia. Such connection has also been suggested for Germany and Denmark (Brandes et al. 2015, 2019).

Of course, these results are still subject to our used GIA models and assumptions made in the δCFS calculation. They may change if, for example, non-optimal faults or pore-fluid pressure are considered.

CONCLUSION

We have calculated δCFS at 12 selected locations in Latvia, Lithuania and the Kaliningrad District of Russia by applying a commonly used FE model of GIA. The locations are characterized by SSDS, which, if interpreted as seismites, indicate seismic activity. Previous and new preliminary geological studies suggest for some locations a possible connection of the SSDS formation to earthquakes that were likely triggered by lithospheric stress field changes induced by the decay of the Weichselian ice sheet. Within the stated assumptions of our study, our results show that all locations at several points in time have reached a state of fault instability, independent of the background stress regime and the chosen GIA model setup. Based on the dating of SSDS horizons, we cannot confirm GIA as source for certain SSDS locations, such as Slinkis, Kumečiai and Liciškėnai. At Sarnate, Valmiera, Rakuti, Dyburiai and Ryadino such relation is likely. In addition, the behaviour of δCFS curves after 15 ka BP suggests that historic and recent earthquakes in the East Baltic area (see e.g., Ņikuļins 2011; Pačėsa, Šliaupa 2011) could be an aftermath of the last glaciation which has already been considered for Germany and Denmark (Brandes et al. 2015, 2019).

The SSDS of Rakuti and Slinkis may serve as crucial locations to improve our GIA models as glacially-induced fault reactivation has been suggested from geological analysis (van Loon et al. 2016; Pisarska-Jamroży et al. this issue), but the GIA modelling cannot yet support this. If more SSDS are found and categorised as seismites, thorough dating of the deposition horizon can help in GIA modelling by excluding or supporting certain GIA model configurations. Locations of glacially-induced faulting can therefore serve next to relative sea levels, global positioning and gravimetric observations as additional constraint on GIA modelling.
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