Effects of Hypothetical Large Earthquakes on Repository Host Rock Fractures

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ABSTRACT

This report regards seismically induced fracture displacements at different distances from a large earthquake (>Mw 7). Applying a synthetic strike-slip stress field, which is in accordance with the notion of the crust being in frictional equilibrium, we propagate a rupture along a large steeply dipping, surface breaching joint plane and calculate secondary displacements on target fractures induced by both static and dynamic effects. The earthquake zone has its orientation according to site data for the BFZ214 Olkiluoto deformation zone, i.e. dip angle 77 degrees and dip direction N195E. However, its dimensions are 38 x 19 km (Length x Width), i.e. significantly larger than those given by the Olkiluoto site model. The model volume containing the fault and target fractures is discretized for proper propagation of stress waves with frequencies up to 1.5 Hz. The adequacy of this discretization is tested in a separate model.

The target fractures are located at five perpendicular distances from the earthquake fault plane, 300 m, 700 m, 1500 m, 2500 m and 4500 m. We consider six different fracture orientations: three steeply dipping sets and three gently dipping. All target fractures are circular with radius 150 m and they have their centres at 420 m depth. Two regions with identical target fracture configurations are included. Region #1 is located at the centre of the earthquake zone and Region #2 with an offset of about 12 km eastward along the zone strike.

We test six earthquake rupture scenarios. In Case 1 – Case 5, the hypocentre is located at about 9 km depth, close to the western end of the zone, whereas in Case 6, it is located at 2 km depth in the middle of the zone. In Case 1, the stress peak ahead of the rupture front is allowed to grow without any limit at long hypocentre distances, thus simulating a fault with a peak strength that depends on propagation distance. In Case 2 – Case 6, we assume a uniform fault peak strength with a peak friction coefficient μ = 0.6. In Case 3, we only let the western part of the fault rupture and the rupture is abruptly arrested at the centre of the fault, close to target fracture Region #1. This induces stress concentration effects close to the target fractures in this region. In Case 4 and 5, we test the effect of strong stress concentration effects further. We define a 1.5 km wide region at the centre of the fault that extends from the surface break to the bottom of the fault. In Case 4, fault slip in this region is limited by increasing the residual fault strength by a factor of 2.3. In Case 5 rupture and slip is completely suppressed in this region by assuming the strength to be infinite.

The earthquake scenarios all generate stress drops of the order of 10 MPa and slip velocities on par with those found in real large events as well as in theoretical models. The moment magnitudes are in the range Mw 7 – 7.3, which is above what is predicted by empirical regressions for earthquakes with similar rupture areas (360 – 720 km²). The ratio between the stress drop and the background stresses is about twice what is suggested for real events.

Regarding the target fracture displacements, the results can be summarized as follows:

- Out of all target fractures and all earthquake rupture scenarios tested here, no fractures at 700 m distances or more slip more than 40 mm.
For scenarios without sharp fault edges or unlimited stress peak growths, no fractures slip more than 40 mm, even at the shortest distance considered, 300 m. At 700 m distance the largest fracture displacement is about 25 mm.

There is a clear and systematic dependence of fracture orientation. With exception of fractures located at the smallest distances (300 m, 700 m) from sharp fault edges (Case 3, Case 5), none of the steeply dipping fractures slip more than 15 mm.

There is a clear and systematic dependence on position along the fault. With exception for Case 3 in which the fault does not extend beyond Region #1, and Case 5, which also has a sharp fault edge close to Region #1, fracture displacements are generally larger in Region #2. This fracture region is located with an offset relative to the fault centre, in a volume where the fault displacement generates increased horizontal stresses and consequential stability reductions of gently dipping fractures. Furthermore, in all scenarios, this fracture region is located farther away from the hypocentre than Region #1, and our results show that the dynamic effects tend to be more extensive at longer rupture propagation distances.

Considering the high moment magnitude-to-rupture area ratio of the synthetic earthquakes as well as the strong stress concentration effects around fault edges in Case 3 and Case 5, the results suggest that 90 mm can be regarded a robust upper bound estimate of seismically induced displacement on a 150 m radius fracture at 300 m distance from a large earthquake. Similarly, at 700 m distance 40 mm is a robust upper bound estimate. Additional results obtained from models modified to produce magnitudes that are more typical of the fault sizes considered here give induced displacements that are about half the upper bound results.

**Keywords:** Earthquake, modeling, fault, fractures, secondary displacement, 3DEC.
Suurten hypoteettisten maanjäristysten vaikutukset loppusijoitustilan kallioperän rakoihin

TIIVISTELMÄ

Tässä raportissa tarkastellaan seismisesti indusoituja siirrostumia raoissa, jotka sijaitsevat erityisesti suuresta maanjäristyksenä (>Mw 7). Analyysissa käytettiin synteeettistä strike-slip jännityskenttää, joka on yhtenevä olettamuksen kanssa että maan kuori on hertiovoin ja kitkavoiman suhteen tasapainoista, ja muodostettiin suureen pystyhköön, maan pinnalle puhkeavaan siirrosvyöhykkeeseen maanjäristys ja laskettiin siirroksen ympäriä sijaitsevissa kohdissa tapahtuvat sekundaariset siirrostumat. Sekundaariset siirrostumat aiheutuvat sekä maanjäristyksen staattisista että dynamisistä vaikutuksista. Mallinnetun siirrosvyöhykkeen kaade ja kaate suunta ovat samanlainisia kuin Olkiluodon geologisen paikkamallin vyöhykkeellä BFZ214, eli kaade 77 astetta ja kaate suunta 195 astetta. Työssä vyöhykke mallinettiin kuitenkin merkittävästi suuremmaksi kuin Olkiluodon paikkamallissa ja vyöhykkeen mitat olivat 38 x 19 km (Pituus x Leveys). Vyöhykkeen ja kohdoraot sisältävät mallitilavuus diskretisoitiin niin, että malli mahdollistaa jännitysaaltojen todennäköisen etenemisen aina 1.5 Hz taajuuksiin asti. Diskretisoinnin paikkansapitävyyttä testattiin erillisellä mallilla.

Kohdoraot sijaitsivat viidellä kohtisuoralla etäisyydellä maanjäristysvyöhykkeestä: 300 m, 700 m, 1500 m, 2500 m ja 4500 m. Työssä tarkasteltiin kuutta rakosuuntaa: kolmea pystyä suuntaa ja kolmea loivasti kaatavaa suuntaa. Kaikki kohdoraot olivat ympyrän muotoisia, säteeltään 150 m ja rakojen keskipisteet sijaitsivat 195 m vyöhykkeen keskkohdassa. Työssä käytettiin kahta aluetta joiden rakokokononnot olivat samanlaiset. Alue#1 sijaitsi maanjäristysvyöhykkeen kesikohdassa ja Alue#2 siirroksen kulun suunnassa itään päässä noin 12 km etäisyydellä Alueesta #1.

Työssä arvioitiin kuutta erillistä maanjäristystysskenaariota. Tapauksissa 1 - 5 maanjäristyksen hyposentrumi sijaitsi noin 9 km vyöhykkeen länsireunassa, kun taas tapauksessa 6 hyposentrumi sijaitsi 2 km vyöhykkeen keskiosassa. Tapauksessa 1 maanjäristyksen siirrostumarintaman edellä kulkevan jännitystilahuipun annettiin kasvavaksi jännityskentän alueella, jolloin tapaus vastaa siirrosvyöhykkettä, jonka huipullujuus riippuu siirrostuman etäisyydestä hyposentrumista. Tapauksissa 2 - 6 siirroksele oletettiin yhteenäköinen huipullujuus yhtenäisellä huipukitakakertoimella. Tapauksessa 3 ainoastaan siirrosvyöhykkeen lännessä annettiin siirrostua ja lisäksi siirrostuma pysäytettiin yhtenäisesti siirrosvyöhykkeen keskiosassa, lähellä Aluetta #1. Ääkilin siirrostuman pysäyttäminen aiheuttaa kannalta kohdarakojen sisältämän alueen lähellä. Tapauksissa 4 ja 5 testattiin edelleen voimakkaiden jännitystilahuipussa on esillä sekä siirrostuma pysäytettiin yhtenäisesti seuraavan alueen keskiosassa. Tapauksessa 4 siirrostuman määrää kyseisellä alueella rajoitettiin lisäämällä siirroksen jännöslujuutta alueella 2.3-kertaiseksi. Tapauksessa 5 sekä siirrostumarintaman eteneminen että siirrostuma kyseisellä alueella estettiin täysin asettamalla siirroksen lujuus alueella äärettömäksi.

Kaikki maanjäristystysskenaariot tuottivat jännitystilajan pudotuksia, jotka ovat luokkaj 10 MPa, sekä siirrosnopeuksia, jotka ovat yhteneväisiä niin todellisten suurten maanjäristysten kuin teoreettistenkin mallien kanssa. Momentimagnitudit ovat luokkaj 10
Mw 7 – 7.3, jotka ovat suurempia kuin empiristen regressiomallien avulla ennustetut magnitudit maanjäristykseille, joilla on saman suuruinen siirrostumapinta-ala kuin tässä työssä käytetyllä siirrosvyöhykkeelle (360 – 720 km²). Jännitystilan pudotuksen ja taustajännityksen suhde oli noin kaksinkertainen mitä on havaittu todellisilla tapahtumilla.

Kohderoissa tapahtuneiden sekundaarisirrostosten osalta analyysin tulokset voidaan koota yhteen seuraavasti:

- Kaikkien työssä testattujen kohderakojen ja maanjäristyskenaarioiden perusteella yksikään kohderako ei siirrostunut 40 mm enempää ≥700 m etäisyydellä maanjäristysvyöhykkeestä.

- Skenaarioille, joissa siirroksella ei ollut teräviä reunoja tai joissa huippujännityksen kasvu oli rajoitettu, yksikään kohderako ei siirrostunut 40 mm enempää, edes pienimmällä testatulla etäisyydellä maanjäristysvyöhykkeestä (300 m). 700 m etäisyydellä suurin siirrostuma kohderaossa oli n. 25 mm.

- Raon suunnalla oli selvä ja systemaattinen vaikutus kohderakoon siirrostumaan. Pois lukien raot, jotka sijaitsivat pienimmällä etäisyyksillä (300 m, 700 m) terävistä siirroksen reunoista (Tapaus 3, Tapaus 5), yksikään pystyrako ei siirrostunut enempää kuin 15 mm.

- Riippuen kohderaon sijainnista suhteessa siirrokseen nähdä, raon paikalla oli selvä ja systemaattinen vaikutus kohderaon siirrostuman suuruuteen. Tapauksia 3 (jossa maanjäristystiirros ei jatku Alueen #1 ohi) ja 5 (jossa siirroksella on terävä reuna Alueen #1 lähellä) lukuun ottamatta, kohderakojen siirrostumat olivat yleisesti ottaen suurempia Alueella #2. Kyseinen alue sijaitsi tietyllä etäisyydellä siirroksen keskiosasta, kalliotalavuudessa jossa vyöhykkeessä tapahtuva siirrostuma aiheuttaa horisontaalijännityksen kasvua ja vastaavasti stabilisuuden alenemaa loivasti kaatuvissa kohderoissa. Lisäksi kaikissa skenaarioissa Alue#2 sijaitsi kauempana hyposentrumista kuin Alue#1 ja tämän työn tulokset osoittavat, että maanjäristyksen dynaamiset vaikutukset ovat merkittävämpiä mitä kauemmaksi siirrostumamarinetan yhteen.

Ottaen huomioon sekä synteeettisen maanjäristyksen korkean momenttimagnitudin ja siirrostumapinta-alueen suhteen, että voimakkaat jännitystilan konseutraatiot siirroksen reunoilla tapaucksissa 3 ja 5, saadut tulokset osoittavat, että 90 mm voidaan pitää robustina ylärajana seismisesti indusoiduille siirrostumille halkaisijaltaan 150 m kohderaoireli, jotka sijaitsivat 300 m etäisyydellä suuresta maanjäristyksestä. Vastaavasti 700 m etäisyydellä 40 mm voidaan pitää robustina ylärajana kohderaon siirrostumalle. Kun analysoidaan malleja, jotka on muokattu tuottamaan nyt tutkitulle siirroskoolle tyypillisempiä magnitudia, saadaan tulokseksi puolestaan siirrostumia, jotka ovat suuruddeltaan noin puolet edellä mainituista ylärajoista.

**Avainsanat:** Maanjäristys, mallinnus, siirros, raot, sekundaarinen, siirrostuma, 3DEC.
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1. BACKGROUND AND INTRODUCTION

Posiva and SKB are planning for KBS-3 type nuclear waste repositories to be constructed in crystalline rock in Olkiluoto and Forsmark, respectively. The only effect of earthquakes that has been concluded to be of potential concern for these repositories is mechanical damage of waste canisters caused by shear displacements along intersecting fractures. At present, canisters sheared 50 mm or more count as damaged, regardless of the shear velocity and the fracture-canister intersection geometry. Displacements of these magnitudes require fractures of very considerable extension (cf. Fälth et al., 2010; Fälth and Hökmark, 2011; 2012). For risk assessment and for decisions on layout rules, it is essential to establish upper bound estimates of the shear displacements that could be induced by nearby earthquakes on differently oriented, sized and located host rock fractures.

Up to the present day, two studies that explicitly address the problem of seismically reactivated fractures at Olkiluoto have been reported (Fälth and Hökmark, 2011; 2012). In these studies, end-glacial earthquakes were simulated on four specifically selected, differently sized and differently oriented, Olkiluoto deformation zones (BFZ039, BFZ021, BFZ100, BFZ214). The host rock response, i.e., the secondary displacements induced on differently oriented rock fractures (target fractures) located at different distances from the slipping deformation zone (the primary fault) was recorded and analysed. The modelling approach can very briefly be summarized as follows:

- All earthquakes were assumed to be end-glacial earthquakes.
- The area of the deformation zones selected for the analyses ranged between 0.5 and 39 km².
- The earthquake rupture was initiated at a predefined hypocentre and programmed to propagate outwardly across the entire fault area until it was abruptly arrested at the fault edges, eventually resulting in an almost complete loss of the fault shear strength.
- Target fractures were perfectly planar and circular with 75 m radius.
- Target fractures were positioned at 100, 300 and 500 m distance from the plane of the primary fault, at different position along the strike of the fault and, for the smallest deformation zones, also at positions around the edges of the fault.
- A number of target fracture orientations, corresponding to orientations of fracture sets identified at the Olkiluoto site, were tried.

The results can, again very briefly, be summarized as follows:

- The earthquake moment magnitudes ranged between M_w 3.9 and M_w 5.9.
- The maximum induced, or secondary, target fracture slip, amounted to about 30 mm (recorded at the smallest fault-fracture distance, 100 m, in the BFZ100 model).
- For three out of the four Olkiluoto deformation zones analysed (BFZ021, BFZ039, BFZ214), no induced target fracture displacements exceeded 15 mm. For the fourth
(BFZ100), induced displacements exceeded 15 mm on 30% of the target fractures at the 100 m distance and on no fractures at the 300 and 600 m distances.

The modelling approach is characterized by deliberate schematic simplifications of complex mechanisms, most of them contributing to exaggerate the response of the target fractures, i.e., the induced secondary slip, considerably. For some mechanisms it should be possible to make realistic but still defensible modifications of the modelling approach and arrive at less exaggerated target fracture slip estimates.

The almost complete loss of fault shear strength associated with the rupture may be the most obvious example of a complicated, but schematically handled, process that will tend to give exaggerated secondary displacement on target fractures. The resulting moment magnitudes are well above typical magnitudes reported for real crustal earthquakes occurring over similar rupture areas.

The intensity of the stress concentrations around the fault edges is another example. The edge stress concentrations are results of an exaggeratedly abrupt transition from almost complete loss of strength on the fault plane to intact, ideally elastic rock at the fault edges. The largest induced target fracture shear displacements obtained in the analyses performed up to the present day can be shown to be caused by these stress concentrations and would, consequently, be less pronounced for more realistic descriptions of the arrest of the rupture.

There are, in conclusion, good prospects that it will be possible to argue that secondary displacements caused by seismic events on any of the local, moderately sized, Olkiluoto deformation zones in reality will be smaller, perhaps even significantly smaller, than those obtained in the numerical analyses performed so far. The issue of the present study is instead, however, to address the possibility that very large earthquakes, occurring on a large and more distant deformation zone could induce significant secondary displacements.

1.1 Scope

For this study the largest Olkiluoto deformation zone out of the twenty included in the work performed so far (numerical analyses: Fälth and Hökmark, 2011; Fälth and Hökmark, 2012 and stability assessment: Hökmark and Fälth, 2014), BFZ214, is used as example. The mapped length of this zone is not more than about 13 km (cf. Figure 1-1), and the estimated fault area not more than 28 km², but the zone orientation is taken to be reasonably representative of much larger zones in this part of Finland.

It should be pointed out that the mapped BFZ214 fault trace does not necessarily have to correspond to the surface rupture length of large individual single seismic events that occurred in the past on this deformation zone (or to surface rupture lengths of large future single events). Fault bends and intersections with other faults, such as those shown in Figure 1-1, are examples of geometrical properties and structural features that could define endpoints of single ruptures (Kneupfer, 1989), meaning that a number of smaller earthquakes could have ruptured segments of the total mapped fault trace. Even assuming the entire BFZ214 fault area to rupture in one single episode, that area (28 km² as given in the site description), is too small for a large earthquake (magnitude > 7)
to be possible. To address the large earthquake issue, hypothetical values of fault length and the fault width are used in this study. The results obtained here are therefore not necessarily relevant for BFZ214, but possibly for a hypothetical, much larger zone of similar orientation.

**Figure 1-1. BFZ214 trace. From Fält and Hökmark (2011).**

The following must be considered when attempting simulations of very large earthquakes:

- The model discretization. In dynamic modelling it is particularly essential that the model discretization is adequate. For stress waves to be transmitted properly, the resolution of the calculation grid should correspond to at least 8 elements per wavelength (Kuhlmeyer and Lysmer, 1973). In the models analysed in the previous studies the model volumes were very differently discretized with resolutions corresponding to frequencies up to 6.7 Hz for the inner, most critical regions containing the target fractures and the fault. For the large model volumes and the large fault-fracture distances required in this study, it will be necessary to allow for coarser grids because of computer run time and memory restrictions (even if the computational capacity at the time of this study is significantly increased in comparison to the capacity at the time of the previous reports).
The initial stress model. The stress models used to evaluate the stability of 20 local, differently oriented, Olkiluoto deformation zones (Hökmark and Fälth, 2014) were based on descriptions established for the Olkiluoto site. The site model stress field, extrapolated to depths that are relevant to large earthquakes, is not anisotropic enough that deep-extending deformation zones could be sufficiently close to failure over more than small fractions of their total fault areas for a large rupture to be a realistic possibility.

The residual strength. In the dynamic analyses of earthquakes on local Olkiluoto deformation zones reported so far, the residual strength was set as low as possible, just high enough to suppress obviously irrelevant post-rupture fault oscillations. As a consequence, these earthquakes released almost all the strain energy that theoretically could be released. Because of the modest stresses at the shallow depths relevant for the local zones, the resulting average stress drop did not exceed the maximum levels for typical intraplate earthquakes, i.e., around 15 MPa. To simulate realistic earthquakes on faults reaching depths of 10 km and below, i.e., where the driving shear stresses could amount to 50 MPa and more, it is necessary to calibrate the residual strength to obtain more realistic stress drops. There are, in addition, also empirically established bounds on the ratio between stress drop and shear stress (Scholz, 2002).

The model discretization issue is addressed in Chapter 2 and the stress model is presented and discussed in Chapter 3. In Chapter 0, we describe the numerical model and the results are presented in Chapter 5. Conclusions and discussion are found in Chapter 0.
2. MODEL DISCRETIZATION

2.1 Example model

The 3DEC approach used in previous dynamic modelling of Olkiluoto earthquakes and secondary slip on host rock fractures has been demonstrated, evaluated and benchmarked by Fälth et al. (2015). An example model with four square-shaped 0.36 km² target fractures at about 1 km distance from a square-shaped 16 km² fault, programmed to slip with a rupture mechanism similar to the mechanism used in the Olkiluoto earthquake simulations, was generated and analysed. An inner volume was discretized to transmit waves with frequencies up to 5.5 Hz and the remaining volume to transmit waves up to 3 Hz (Figure 2-1).

![Model outline showing a) the inner model volume, discretized for frequencies up to 5.5 Hz, embedded centrally in the larger outer volume, which is discretized for frequencies up to 3 Hz. The initial principal stress orientations are indicated. b) A close-up of the inner volume with primary fault and target fractures. c) Map view showing the horizontal projection of the primary fault and target fractures. The star indicates the hypocentre. From Fälth et al. (2015).](image-url)

The moment magnitude was $M_w$ 5.7. The slip induced on the target fractures ranged approximately between 9 and 14 mm which, for smaller fractures like the 150 m radius fractures considered in the Olkiluoto simulations, would correspond to 2-3 mm of slip. To check that the fault-fracture distance was large enough that the target fracture slip was caused exclusively by dynamic effects, i.e., that stress concentrations around the
fault edges did not contribute, the model was run also statically. As expected, no target fracture slip was observed in the static case, meaning that the 9-14 mm slip obtained in the dynamic case was a genuine effect of P and S waves reaching the target fractures. To increase the confidence in the dynamic effects, the modelling approach was benchmarked:

- First, the general relevance of the 3DEC logic to problems with a wave source within the interior of the model volume was checked by comparison between result obtained from a specifically designed 3DEC model and Stokes solution (see e.g. Lay and Wallace, 1995).
- Second, in the example model described above, 3DEC waveforms recorded at a number of surface receivers, positioned at about 5 km epicentral distance from the square-shaped rupturing fault, were compared with waveforms obtained in a corresponding, specifically designed Compsyn model. The efficiency of the 3DEC wave transmission to the receivers was determined by the coarsest discretization along the ray paths, i.e., the 3 Hz discretization used in the outer volume. The agreement between the 3DEC and Compsyn waveforms was found to be very satisfactory, in particular for the components with the most significant amplitudes.

The Compsyn code (Spudich and Xu, 2002) is an established tool for solving the wave equation in models with horizontal layers, and the waveform agreement strengthens the confidence in the relevance of the dynamic impact of the target fractures, i.e., the 9-14 mm slip range. Unfortunately, for the large-scale models attempted here with fault areas of about 700 km² and fault-fracture distances on the order of kilometres, it is not possible to discretize the relevant volumes for 5.5 or 3 Hz. This may, however not be necessary for the purpose of this study, which is to calculate relevant values of secondary slip induced on distant target fractures, not to produce correct waveforms at distant surface receivers. It is likely that high frequencies (> 1 Hz) do not contribute much to the induced slip:

Figure 2-2 shows the evolution of the rupture on the primary fault of the example model (a), the slip velocity at the four indicated fault positions (b) and the frequency spectra generated at the four points (c). Low frequencies (< 1 Hz) dominate.

Figure 2-3 shows the stability (Coulomb Failure Stress, CFS) variation (a) and the associated slip (b) on the four target fractures. The stability quantity, CFS, is defined as

\[ CFS = \tau - \mu (\sigma_n - P), \]

where \( \tau \), \( \mu \), \( \sigma_n \) and \( P \) are shear stress, coefficient of friction, normal stress and pore pressure, respectively.

The finite difference mesh is discretized for frequencies up to 5.5 Hz, (inner volume) and 3 Hz (outer volume), cf. Figure 2-1. However, as illustrated in the left part of the figure, the frequency of the CFS variation relevant for shear failure and slip is significantly lower (around 1 Hz).
Figure 2-2. a) Primary fault shear displacement contours at four time instances. The hypocentre is indicated by the star and the points A – D are recording points for slip velocity. b) Dip-slip slip velocity at points A – D low-pass filtered at 5 Hz. c) The unfiltered amplitude spectrum for the slip velocity functions in (b). The dash-dotted line to the lower left indicates the $\omega^{-2}$-slope. From Fälth et al. (2015).

Figure 2-3. Temporal development at the target fractures centres of a) Coulomb Failure Stress (CFS) and b) shear displacement. From Fälth et al. (2015). Here the model volume is discretized for 5.5 Hz (inner volume containing fault and target fractures, cf. Figure 2-1) and 3 Hz (outer volume with ground surface waveform receivers).
2.2 Coarsening the discretization - effects on secondary slip

For this study the example model described above was rerun with coarser grids. Figure 2-4 shows the CFS variation and the secondary slip on the four target fracture for two different discretizations: Inner - outer volumes discretized for 2 – 1 Hz and 1 – 1 Hz, respectively.

The main character of the CFS evolution does not appear to change much when the grid is coarsened from 5.5 Hz (Figure 2-3) to 1 Hz (Figure 2-4, bottom). At these fault-target distances, away from the stress concentrations around the fault edges, the net change in stability is a stability increase of around 2 MPa resulting from the overall stress relaxation associated with the strain energy release. The stabilization is accomplished within a little more than 2 seconds. It is clear from the figures that target fracture slip coincides with the short period (about 0.5 second) of stability loss associated with the dominating frequencies (of around 1 Hz), regardless of whether higher frequencies are transmitted or not. Logically, target fracture slip is basically the same in the finely (Figure 2-3 and coarsely (Figure 2-4) discretized models.

Figure 2-4. Results from modified discretization versions of the example model benchmarked by Fälth et al. (2015.)

Figure 2-4 also illustrates that, for target fractures that are initially close enough to the stability limit to slip in response to stress oscillations, oscillations of even lower frequency (with amplitudes as those of the 1 Hz waves) would extend the duration of
the target fracture slip pulses and potentially increase the resulting slip magnitudes. The large earthquake models analysed in this study will be discretized for frequencies of around 1.5 Hz in the region containing the fault and the fractures and are, based on the examples described here, concluded to capture effects of oscillations that are relevant to the process of secondary slip adequately.
3. STRESSES

Figure 3-1 (left) shows the stability (CFS) of BFZ214 as function of depth for a stress model based on the Olkiluoto site description, see the stability assessment made by Hökmark and Fälth (2014). The fault friction angle is set at 30 degrees. This stress field is qualitatively in agreement with the notion that the dominating style of deformation in the Finnish bedrock is slow creep at low stress threshold as suggested by Ojala et al. (2004). The stability increases with depth fast enough that it is not realistic to believe that a rupture originating at shallow depths, where the fault is reasonably close to the failure limit (CFS=0), could penetrate much deeper than one or two kilometres. For the extended steeply dipping BFZ214 deformation zone considered in this study to be close to the stability limit down to depths of about 20 km, a stress field similar to the synthetic strike-slip field discussed by Hökmark and Fälth (2014) must be assumed.

Figure 3-1 (right) shows the stability as function of depth for present-day conditions (P-D), glacial forebulge conditions (FB) and for end-glacial conditions when the edge of the retreating ice cover passes the site (EP). The stability is shown for two different synthetic stress fields, both with the major horizontal stress trending 90 degrees with respect to North and both resulting in frictional equilibrium for optimally oriented faults with a friction coefficient of 0.6. For the synthetic reverse stress field, the BFZ214 deformation zone is stably clamped by the high horizontal stresses at all depths (red curves), whereas it is reasonably close to the failure limit (CFS=0) for the synthetic strike-slip field (blue curves).

Figure 3-1 does also show that BFZ214 is stabilized compared to present-day conditions regardless of the stress model, at least down to depths of 3 - 4 km, during the end-glacial phase. The destabilisation (CFS EP > CFS PD) seen in Figure 3-1 (right) at larger depths is a consequence of the very uncertain and exaggerated extrapolation down to these depths of the model for residual pore overpressures made in these figures, cf. Hökmark and Fälth (2014). Because the BFZ214 is stabilized rather than destabilized during the end-glacial phase, even assuming the strike-slip background stress field it is not a credible potential postglacial fault.

Figure 3-2 (left) shows a Mohr circle representation of the synthetic strike-slip field at 10000 m depth. The white square shows the normal and shear stresses on a vertical plane with strike as BFZ214. The dashed Mohr circle shows corresponding stresses at the time of the forebulge, i.e., when the BFZ214 deformation zone is destabilized. For simplicity the glacial forebulge stresses have been aligned with background stresses such that the resulting stress anisotropy is maximized. The glacial stress additions are obtained from the UMISM ice reconstruction combined with the best fit Earth model, cf. Hökmark and Fälth (2014). Figure 3-2 (right) shows, for completeness, corresponding Mohr circles based on the synthetic reverse field. The white square shows the normal and shear stresses on an optimally oriented plane with dip as BFZ214. Again it is clear that a stress field allowing for rupture down to depths of 10 kilometres and more on BFZ214 should be something similar to the strike-slip field tried here. It is also clear that the impact of any forebulge stress additions would be very marginal.
Figure 3-1. Left: CFS as function of depth for base case Olkiluoto stress model considered in Olkiluoto stability analysis. FB = forebulge, EP = edge passing (i.e., at the time of ice retreat during the end-glacial phase), P-D = present day. Right: Same as left but for synthetic reverse and strike-slip stress fields. From Hökmark and Fälth (2014).

Figure 3-2. Mohr circle representation of synthetic stress fields at 10000 m depth. The white squares show the normal and shear stresses on a vertical plane with strike as BFZ214. The dashed circles show corresponding forebulge stresses assuming worst cases relative orientation of background stresses and glacial stresses.

For this study a large earthquake on BFZ214, hypothesized to extend about 40 km laterally and 20 km vertically, is assumed to be powered by the modified strike-slip field pictured in Figure 3-1 (right) and in Table 3-1 without consideration of any potential glacial disturbances. This stress field is in accord with the notion of the crust being in frictional equilibrium assuming hydrostatic pore pressure and a friction coefficient of 0.6 for optimally oriented vertical brittle deformation zones. The modification (addition of 12.45 MPa and 6.35 MPa, respectively, to the major and minor horizontal stresses of the idealized synthetic stress field) is made to give stresses at the repository level that are in approximate agreement with site model stresses. Figure 3-3 shows Mohr circle representations at 500 m depth of the modified synthetic strike-slip field (left) and the base case version of the site-specific field considered in the stability study (right), cf. Hökmark and Fälth (2014). At the repository level both stress fields are reverse faulting fields with optimally striking fractures being at (or very
close to) shear failure assuming a 0.6 friction coefficient. For the synthetic field also optimally striking, steeply dipping fracture are fairly close to the failure limit. The maximum shear stress is approximately the same for both fields. For the majority of the fault area the modification done to bring the synthetic field in agreement with site data at shallow depths are without importance.

Table 3-1. Modified synthetic strike-slip stress field. Depth $z$ is in metres.

<table>
<thead>
<tr>
<th>Synthetic strike-slip, modified</th>
<th>$\sigma_H$ (MPa)</th>
<th>$\sigma_H$ trend ($^\circ$)</th>
<th>$\sigma_h$ (MPa)</th>
<th>$\sigma_v$ (MPa)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.0351$z$+12.45</td>
<td>90</td>
<td>0.0179$z$+6.35</td>
<td>0.0265$z$</td>
<td></td>
</tr>
</tbody>
</table>

Figure 3-3. Left: Mohr circle representation of stress field assumed in this study at the level of target fractures. Right: Corresponding for site description-based base-case stress field (cf. Hökmark and Fälth, 2014).
4. **3DEC MODEL**

4.1 **Outline**

The model is analysed by use of 3DEC, version 5 (Itasca, 2013). The outline of the 3DEC model is shown in Figure 4-1. The model has the dimensions 100 x 100 x 50 km³ and the upper boundary represents the ground surface. The primary fault is located centrally in the model and is represented by one planar discontinuity that breaches the ground surface. It has dip angle 77° and dip direction N195E, in accordance with site data for BFZ214. This fault orientation gives a left lateral fault movement as shown in Figure 4-2. The model is intended to simulate an earthquake on BFZ214 that ruptures over an area that is significantly larger than what is indicated for BFZ214 by Posiva’s site model (cf. Fälth and Hökmark, 2011; Posiva, 2011). The length and width are set at 38 km and 19 km, respectively (Figure 4-1). There are two regions on the hanging wall side (south side of the fault) containing target fractures. The volume containing the fault and the target fractures is discretised such that it can transmit waves properly with frequencies up to 1.5 Hz. Outside this volume, the finite element grid is gradually made coarser towards the model boundaries (Figure 4-3). The rock mass between the discontinuities is modelled as a linear elastic, homogeneous and isotropic continuum. The material property parameter values are presented in Table 4-1. The parameter values of the rock mass correspond to S- and P-wave velocities of 3.1 km/s and 5.4 km/s, respectively.

![Figure 4-1. Outline of 3DEC model. Left: The fault (blue) is centrally located in the model box. The green boxes are the target fracture regions. Right: Close-up of the fault and target fractures.](image-url)
Figure 4-2. The fault is located at the centre of the model. The fault has dip angle 77° and dip direction N195E as indicated by the blue arrow. The trend of the major horizontal stress is N90E. The red arrows indicate the sense of fault displacement.

Figure 4-3. The inner regions containing the fault and target fractures (green and blue) are discretised for frequencies up to 1.5 Hz. In the remaining volume the discretisation is made gradually coarser towards the model boundaries.
4.2 Target fractures

The model includes two identical target fracture regions. Region #1 is located centrally along the fault trace and Region #2 with an offset of about 12 km (Figure 4-4). The intention is to examine the possible effect of the target fracture location relative to the hypocentre. In simulations made for the Forsmark site (Fälth, 2014), the secondary displacements increased when the hypocentre was located farther away from the target fracture region. According to observations by Fälth et al. (2015) the stress peak ahead of the rupture front tends to be stronger at longer hypocentre distances.

![Figure 4-4](image)

**Figure 4-4.** There are two identical target fracture regions. Region #1 is located centrally along the fault trace and #2 with an offset of about 12 km. Six different fracture sets are included at five perpendicular distances from the fault plane, see the inset. Their orientations are given by the legend. All fractures have radius 150 m and their centres are at 420 m depth.

The target fractures are located with their centres at 420 m depth. All fractures have radius 150 m, i.e. twice the fracture radius in the models analysed by Fälth and Hökmark (2011; 2012). The fractures are made larger here to admit a proper discretisation of the fracture surfaces, but still avoid very small finite difference zones. Small zones tend to give small time steps and thus very long computer run times for large models. Six fracture orientations are considered (Figure 4-4). Four of those are identical to those considered in previous Posiva studies (Fälth and Hökmark, 2012).
Then, there are two additional sets. One has the same orientation as the fault plane and the other is optimally oriented for failure, given the stress conditions at 420 m depth in the model. As illustrated by Figure 4-4, the fractures in each set are placed at different distances from the fault on a line perpendicular to the fault strike. Five perpendicular fault-fracture distances in the range 0.3–4.5 km are considered. This range spans over all distances from BFZ214 where canister will be located at Olkiluoto.

4.3 Calculation sequence

The calculation sequence comprises two phases, the static phase and the dynamic phase.

During the static phase, the primary fault is assigned a high cohesive strength in order to prevent slip, initial stresses are applied and the model is allowed to achieve static equilibrium under gravity. The target fractures are assigned their frictional properties according to Table 4-1. The bottom boundary is locked for displacements in the z-direction, the vertical boundaries are locked in the x- and y-directions and the top boundary is kept free. The static equilibrium state is the point of departure for the following dynamic phase.

During the dynamic phase, the top boundary is kept free and allows for surface reflections, while the other boundaries are redefined to be viscous, i.e. non-reflecting. Damping is accomplished only by means of the viscous boundaries. The method for rupture propagation is the same as that used in previous studies (e.g. Fälth and Hökmark, 2011). The rupture is initiated at the pre-defined hypocentre and the rupture front is programmed to move outwardly in the radial direction at a constant speed \( v_r \), here set to be 80% of the rock mass shear wave velocity. Prior to rupture front arrival, the shear strength is set either to match the current stress (i.e. practically infinite strength) or according to a finite frictional strength (see Section 4.4). At rupture front arrival, the shear strength is ramped down to a specified residual frictional strength over a specified amount of time \( \Delta t_{\text{red}} \), here set at 0.5 s. The residual friction coefficient is set at 0.35 (friction angle about 19 degrees), which gives a slip and corresponding seismic moment that can be regarded reasonably large relative to database regressions (Figure 5-1).
Table 4-1. Material property parameter values.

<table>
<thead>
<tr>
<th>Component</th>
<th>Parameter</th>
<th>Value</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rock mass</td>
<td>Density</td>
<td>2700</td>
<td>kg/m³</td>
</tr>
<tr>
<td></td>
<td>Young’s modulus</td>
<td>65</td>
<td>MPa</td>
</tr>
<tr>
<td></td>
<td>Poisson’s ratio</td>
<td>0.25</td>
<td></td>
</tr>
<tr>
<td>Primary fault</td>
<td>Residual friction angle</td>
<td>19</td>
<td>deg</td>
</tr>
<tr>
<td></td>
<td>Cohesion*</td>
<td>0</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Tensile strength</td>
<td>0</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Normal stiffness</td>
<td>30</td>
<td>GPa/m</td>
</tr>
<tr>
<td></td>
<td>Shear stiffness</td>
<td>30</td>
<td>GPa/m</td>
</tr>
<tr>
<td>Target fractures</td>
<td>Friction angle</td>
<td>30</td>
<td>deg</td>
</tr>
<tr>
<td></td>
<td>Dilation angle</td>
<td>5</td>
<td>deg</td>
</tr>
<tr>
<td></td>
<td>Cohesion</td>
<td>0.5</td>
<td>MPa</td>
</tr>
<tr>
<td></td>
<td>Tensile strength</td>
<td>0</td>
<td>MPa</td>
</tr>
<tr>
<td></td>
<td>Normal stiffness</td>
<td>10</td>
<td>GPa/m</td>
</tr>
<tr>
<td></td>
<td>Shear stiffness</td>
<td>1.5</td>
<td>GPa/m</td>
</tr>
</tbody>
</table>

* Final, after completed rupture

4.4 Simulation cases

Six simulation cases are tested (Table 4-2, Figure 4-5).

We test two different assumptions regarding the fault shear strength at locations not yet reached by the rupture front. In Case 1, the strength is assumed to be infinite. This assumption, which is the same as that made in earlier studies (e.g. Fälth and Hökmark, 2011), means that the stress peak ahead of the rupture front can grow large, in particular when the front has reached locations far from the hypocentre and the ruptured area has grown large. In earlier studies, when moderately sized events have been simulated, this effect was not very important. However, for the large event simulated here it becomes more pronounced.

In Case 2 – Case 6, the fault peak strength is set such that it corresponds to a frictional coefficient of 0.6. This frictional strength is in accordance with that assumed when the initial stress field was defined (cf. Chapter 0). With this assumption, the stress peaks do
not continue to grow as the ruptured area grows. Thus, the rupture is assumed to take place on a fault with no spatial variation of the peak strength.

In Case 3 – Case 5, different assumptions are made regarding the fault slip resistance of different regions on the fault. In Case 3, the eastern half of the fault plane is locked and only the western part ruptures. This means that there is a sharp fault edge that can cause stress concentrations close to target Region #1 (cf. Figure 4-4). In Case 4, the frictional strength is increased 2.3 times (residual friction coefficient 0.8) in a 1.5 km wide region from the surface to the bottom of the fault. This is done to schematically simulate possible effects of a fault bend and/or a region with higher shear strength. The same region is used in Case 5, but in this case the area is completely locked. Case 6 is similar to Case 2, but the hypocentre is located at 2 km depth close to the fault centre.

**Figure 4-5.** Schematic sketches indicating hypocentre locations and assumptions of fault shear resistance in the different cases. The fault plane is viewed from southwest, straight on along its normal.

**Table 4-2. Simulation cases.**

<table>
<thead>
<tr>
<th>Case</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Hypocentre located close to the west end of the fault, 14 km from fault centre along strike and at 8.9 km depth. Fault strength prior to rupture front arrival assumed to be infinite.</td>
</tr>
<tr>
<td>2</td>
<td>Same as Case 1, but fault peak strength limited to $\mu = 0.6$.</td>
</tr>
<tr>
<td>3</td>
<td>Same as Case 2, but only the western half of the fault ruptures. The eastern half is locked.</td>
</tr>
<tr>
<td>4</td>
<td>Same as Case 2, but the fault frictional strength is increased 2.3 times in a 1.5 km wide region close to fault centre.</td>
</tr>
<tr>
<td>5</td>
<td>Same as Case 4, but the 1.5 km wide region is completely locked.</td>
</tr>
<tr>
<td>6</td>
<td>Same as Case 2, but hypocentre is located at 2 km depth at fault centre.</td>
</tr>
</tbody>
</table>
5. RESULTS

5.1 Fault response

The simulated earthquakes generate an average stress drop of the order of 10 MPa and average fault shear displacements in the range 3.8 – 6.2 m (Table 5-1). The moment magnitude exceeds 7 for all cases except Case 3.

Figure 5-1 shows the relation between rupture area and moment magnitude for a number of synthetic 3DEC earthquakes, including the example model discussed in Chapter 2 and the one assumed here to occur on the hypothetically extended BFZ214. The results are plotted along with corresponding data compiled by Wells and Coppersmith (1994) for real crustal earthquakes. Earthquakes in stable continental regions (SCR), i.e., such as the Fennoscandian Shield, are specifically indicated. The regressions are based on Wells and Coppersmith (1994) data and on a similar study by Leonard (2010). The synthetic 3DEC earthquakes do all plot well above the regressions, even for BFZ214 cases where fault slip is systematically suppressed by a narrow strip of high shear strength (Cases 4 and 5). Specifically one should note that the magnitudes of all synthetic BFZ214 earthquakes are significantly higher than all corresponding SCR earthquakes in the Wells-Coppersmith catalogue.

Figure 5-2 shows moment magnitude vs maximum fault displacement for the synthetic 3DEC earthquakes along with corresponding data compiled for real crustal earthquakes by Wells and Coppersmith (1994). All 3DEC earthquakes plot well below the regression.

The comparisons in Figure 5-1 and Figure 5-2 between the 3DEC synthetic earthquakes and the database earthquakes indicate that the synthetic earthquakes are likely to be considerable overestimates as far as moment magnitude and fault slip are concerned. For the large hypothetical earthquake occurring on an extended version of the BFZ214 deformation zone this appears to be true for all of the cases listed in Table 5-1. The differences in the details of the fault response between the different cases are illustrated and discussed in the following.

### Table 5-1. Source parameters.

<table>
<thead>
<tr>
<th>Case</th>
<th>Rupture area (km²)</th>
<th>Average fault slip (m)</th>
<th>Peak slip (m)</th>
<th>Peak slip velocity* (m/s)</th>
<th>M₀ (Nm)</th>
<th>Mw</th>
<th>Average stress drop (MPa)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>720</td>
<td>5.8</td>
<td>8.5</td>
<td>8.4</td>
<td>1.1·10²⁰</td>
<td>7.3</td>
<td>11.0</td>
</tr>
<tr>
<td>2</td>
<td>720</td>
<td>5.8</td>
<td>8.5</td>
<td>8.4</td>
<td>1.1·10²⁰</td>
<td>7.3</td>
<td>11.0</td>
</tr>
<tr>
<td>3</td>
<td>360</td>
<td>3.8</td>
<td>6.3</td>
<td>6.5</td>
<td>3.6·10¹⁹</td>
<td>7.0</td>
<td>11.7</td>
</tr>
<tr>
<td>4</td>
<td>718</td>
<td>4.6</td>
<td>7.1</td>
<td>7.7</td>
<td>8.6·10¹⁹</td>
<td>7.2</td>
<td>9.3</td>
</tr>
<tr>
<td>5</td>
<td>691</td>
<td>4.0</td>
<td>6.6</td>
<td>7.6</td>
<td>7.3·10¹⁹</td>
<td>7.2</td>
<td>11.2</td>
</tr>
<tr>
<td>6</td>
<td>720</td>
<td>6.2</td>
<td>8.9</td>
<td>8.1</td>
<td>1.2·10²⁰</td>
<td>7.3</td>
<td>11.6</td>
</tr>
</tbody>
</table>

*The slip velocity at each point is calculated during analysis as a running temporal average over a 0.2 s time interval and the peak value is stored. The maximum value presented here is then determined from the peak values as running spatial average over a circular area with 600 m radius.
Figure 5-1. Moment magnitude vs rupture area for a number of 3DEC earthquakes (see text) and for database earthquakes (Wells and Coppersmith, 1994). Regression are from Wells and Coppersmith (1994), i.e., based on the catalogue data plotted here, and from Leonard (2010). The filled plot symbols represent events judged to have occurred in Stable Continental Regions (SCR).
Figure 5-2. Moment magnitude vs max displacement for a number of 3DEC earthquakes (see text) and for database earthquakes (Wells and Coppersmith, 1994).

Figure 5-3 shows fault shear displacements at different times after rupture initiation for all cases. The rupture algorithm is programmed to propagate the rupture in the radial direction. However, after about 4 seconds, when the rupture has reached the upper and lower edges of the fault (Case 1- Case 5), it becomes unilateral and propagates eastward, towards the target fracture regions. Since the stress drop increases with depth, the slip is also largest at depth (some 8-9 m for models with undisturbed rupture propagation). Still, the resulting total slip at the surface close to target fracture regions is considerable. For Cases 1, 2 and 6 (rupture propagates undisturbed over the entire fault area) it is between 5 and 7 m, systematically a bit larger in the central parts of the fault (close to target fracture Region #1) than towards the fault edges (close to target fracture Region #2). Figure 5-4 shows the temporal slip evolution at six points distributed along the strike of the fault at 500 m depth.

Figure 5-5 shows contours of fault slip velocity at different times for all model cases. As in Figure 5-3, it is clearly seen how the rupture propagates unilaterally towards east (Case 1- Case 5). The plots show that the slip velocity increases with depth due to the larger stress drop. It is also shown that the slip velocity increases as the ruptured area increases due to increasing stress effects around the rupture front. The increase of slip velocity with hypocentral distance is also shown by the peak slip velocity contours in Figure 5-6. This plot regards Case 2, i.e., the base case. The maximum peak slip velocity is about 8.4 m/s and is reached at large depth far from the hypocentre. The peak slip velocity is a quantity that cannot be easily and straightforwardly compared with and related to literature findings. The highest slip velocity ever instrumentally recorded for
real earthquakes, 4 m/s, is that of the 1999 Chi-Chi Mw 7.6 earthquake in Taiwan (Ma et al. 2001; Ma et al. 2003). The highest slip velocities of the Chi-Chi dip-slip earthquake were concluded to be located in the shallowest parts of the gently dipping deformation zone. For slip velocities on strike-slip faults at large depths there is, to our knowledge, no established empirical maximum value. Wald and Heaton (1994) estimated the maximum slip velocity of the Landers 1992 magnitude 7.3 strike-slip earthquake to be between 1.5 and 2.0 m/s. As far as the slip velocities estimated for the Chi-Chi and Landers earthquakes are concerned, the maximum slip velocity of the synthetic earthquake considered here appears to be, with some margin, high enough to count as an upper bound. The comparison is, however, not necessarily a fair one: the value will depend on over which areas and over which time intervals the background data and the resulting velocities have been averaged. The BFZ2014 velocities presented here are calculated as average values over a time interval of 0.2 s, while Wald and Heaton (1994) used a 1 s interval. Theoretical models of slip-weakening rupture processes occurring at sub-shear rupture velocities, such as for the synthetic BFZ214 event, give peak fault slip velocities of around 8 m/s, i.e., in agreement with the result shown in Figure 5-6 (Bizzarri, 2012). In principle it would be possible to explore the sensitivity of the results shown in Figure 5-5 and Figure 5-6 to the intervals used when processing the model raw data and then, perhaps, make more relevant comparisons between model results and estimates based on measurements.

Figure 5-7 shows the stresses on the fault plane before rupture initiation and Figure 5-8 the stress drop for the base case earthquake (Case 2). The initial shear stress, roughly averaged over the fault plane, is around 45 MPa and the average stress drop is 11 MPa (Table 5-1), meaning that the stress drop is around 20% of the initial stress.

Figure 5-9 shows contours of the seismic efficiency of the Case 2 earthquake. The seismic efficiency $\eta$ is defined as

$$\eta = \frac{\Delta \tau}{\tau_1 + \tau_2},$$

where $\tau_1$ and $\tau_2$ are the shear stresses before and after the earthquake, respectively, and $\Delta \tau$ is the stress drop ($\tau_2 - \tau_1$). At shallow depths with low initial stresses, the seismic efficiency as expressed above is sensitive even to modest stress drop variations, but on the majority of the BFZ214 fault area the seismic efficiency varies stably and smoothly between 0.10 and 0.20. A typical value of $\eta$ would be 0.06 (Scholz, 2002; McGarr, 1999), corresponding to a stress drop that is approximately 10% of the initial stresses. This is in agreement with Figure 5-1, where our BFZ214 model results plot above the regression lines.

Figure 5-10 shows the shear stress evolution at six points distributed along the strike of the fault at 500 m depth. At this depth the net stress drop ($\tau_2 - \tau_1$) is on the order of just a few MPa. The curves illustrate the way the shear stress increases ahead of the moving crack tip and drops as soon as the rupture front has passed. For Case 1, this moving stress peak grows without limits with the stress transfer from the expanding ruptured area, whereas a more realistic strength limit is applied in Cases 2 – 6.
Figure 5-3. Fault slip at different times after rupture initiation. The fault plane is viewed from southwest, straight on along its normal. The black dots indicate the hypocentre.
Figure 5-4. Evolution of fault slip at six points at 500 m depth.
Figure 5-5. Fault slip velocity at different time instances. The fault plane is viewed from southwest, straight on along its normal. The black dots indicate the hypocentre.
Figure 5-6. Contour plot of peak fault slip velocity in Case 2. The fault plane is viewed from southwest, straight on along its normal. The black dot indicates the hypocentre at 8.9 km depth. The target fracture regions are shown in red close to the upper fault edge.

Figure 5-7. Initial shear stress. The view is the same as in Figure 5-6. The target fracture regions are shown in red close to the upper fault edge.
Figure 5-8. Contour plot of stress drop in Case 2. Note that the colour scale saturates at 30 MPa even though there are locations at large depth where the stress drop exceeds that value. The view is the same as in Figure 5-6.

Figure 5-9. Seismic efficiency for Case 2. The view is the same as in Figure 5-6.
Case 1  Case 2  Case 3

Case 4  Case 5  Case 6

Figure 5-10. Shear stress evolution at 500 m depth at six points along the strike of the fault.
5.2 Off-fault stress effects

The sensitivity of the results presented on the fault response in the previous section seem to be logical and consistent with the difference input assumptions made in the different cases. In addition, the moment magnitude, the fault slip, the slip velocity and the stress drop obtained for the synthetic earthquakes on the hypothetically extended BFZ214 deformation zone do all appear to be upper bounds to estimates or measurements of corresponding quantities reported in the literature. In this section we look at the response of the surrounding rock.

Figure 5-11 shows the CFS (Eq. 1) evolution (Case 2) at the level of the repository for fracture planes dipping 30 degrees with dip direction 300 degrees with respect to north. This corresponds to the orientation of one of the fracture sets actually modelled in target fracture regions #1 and #2 (Figure 4-4). Note that CFS is here calculated from the stress tensors in the continuum, and not on explicitly modelled joint planes. The positions of the target fracture regions and the sense of fault shear are indicated. The contours show that, due to the compression of the rock mass generated by the fault movement, the stability after the earthquake is systematically reduced within and around Region #2 for this particular fracture set.

Figure 5-12 shows the temporal CFS evolution at the seven history points indicated in the last contour plot in Figure 5-11, bottom right panel. These points are all positioned in intact rock at about 1900 m distance from the fault. The CFS evolutions in Figure 5-12 are given for two hypothetical fracture plane orientations: the gently dipping set used in Figure 5-11 (left column), and a steeply dipping set (right column). Both orientations correspond to target fractures actually modelled in regions #1 and #2. The evolution is shown for the base (Case 2) and for three additional cases. As expected, the CFS evolutions of the gently dipping fracture set (left) are in agreement with the contours in Figure 5-11; points in the western end of the fault indicate increased stability due to stress relaxation while the eastern end is destabilized. For the steeply dipping set (right) the results are more difficult to interpret. However, we conclude that this set has a higher initial stability margin and that points #1 and #7 loose stability for all cases due to the proximity of the fault ends.

Figure 5-13, upper, shows examples of actual target fracture displacements in Region #1 induced by the Case 2 model. Figure 5-13, lower, shows the associated CFS variations recorded in the continuum a few hundred metres from the target and at the same distances from the fault. The results regard the target fracture orientations considered in Figure 5-12. Figure 5-14 shows corresponding results recorded in Region #2. There is a general agreement between the displacements and the CFS variations; fracture displacements tend to be larger at positions where the stability limit (CFS = 0) is exceeded by larger amounts or during longer periods of time. In Region #1, slip on the 300/30 fractures is induced by a temporary loss of stability whereas in Region #2, there is a permanent loss of stability that causes fracture slip. For the 195/77 set the stability limit is exceeded only during very short periods of time and the associated displacements become small.
Figure 5-11. CFS contours on a horizontal projection plane at 400 m depth in Case 2 assuming friction angle 30°, cohesion 0.5 MPa and stresses on a plane with dd/dip = 300/30. Blue and reddish colours indicate stable and unstable conditions, respectively. The black line indicates the fault plane and the small dots show the target fracture positions. The arrows in the upper left panel show the sense of fault shear movement. The square-shaped numbered dots in the lower right panel indicate history points (see Figure 5-12) located at approximately 1900 m perpendicular distance from the fault plane.
Figure 5-12. Temporal evolution of CFS at the points indicated in Figure 5-11 assuming friction angle $30^\circ$ and cohesion 0.5 MPa. The results are low-pass filtered at 2 Hz.
Figure 5-13. Temporal evolution of target fracture shear displacement at different distances from the fault plane in fracture Region #1 in Case 2 (upper). Results are shown for two fracture orientations (see Figure 4-4). CFS temporal evolution based on stress tensor recordings at points located in the continuum between the target fractures (lower).
Figure 5-14. Temporal evolution of target fracture shear displacement at different distances from the fault plane in fracture Region #2 in Case 2 (upper). Results are shown for two fracture orientations (see Figure 4-4). CFS temporal evolution based on stress tensor recordings at points located in the continuum between the target fractures (lower).

5.3 Target fracture displacements

The target fracture displacements for all cases are presented as cumulative plots in Figure 5-15 - Figure 5-20. It should be noted here that some of the cases are analysed mainly for the sake of general model understanding, i.e., to get perspectives on the importance of some of the schematic assumption made in previous analyses. These cases are (cf. Table 4-2):

- Case 1 - unlimited growth of stress peak travelling ahead of the moving rupture front (cf. Figure 5-10, upper left)
- Case 3 - abrupt arrest of rupture on fault edge close to target fracture region.
- Case 5 - abrupt arrest of rupture on narrow stability strip close to target fracture region.
The unlimited stress peak growth appears to have a small but clear and logical impact. Case 1 induced displacements are 1-2 mm larger than Case 2 (base case) displacements (cf. Figure 5-15 and Figure 5-16). There is a tendency that the differences are larger in Region #2, which is logical since this is where the rupture front has travelled the longest and the stress peak grown the most.

The effects of the stress concentration around the sharp fault edge close to Region #1 in cases 3 and 5 is clear from Figure 5-17 and Figure 5-19. Here the target fractures slip increase by about 60% and about 360% relative to the base case (Case 2) for Case 3 and Case 5, respectively, at the smallest distance (300 m). It is also worth noting that the largest displacements occur on steeply dipping fractures, while no steeply dipping fracture slips more than around 11 mm in models without sharp fault edges. This points to the importance of a realistic fault edge representation. In Case 4, for example, a similar stability strip as that in Case 5 is modelled, however with less sharp edges. Here the target fracture displacements are very similar to those obtained in the base case (Case 2).

In Figure 5-21 the slip induced in Region #1 for the six rupture cases at three fault-target distances (300 m, 1500 m and 4500 m) are compared. At the smallest distance, the stress concentrations associated with the abrupt rupture arrest at sharp fault edges determine the maximum displacements. At 1500 m distance there is only little influence of the stress concentrations and there is no target fracture displacement larger than 15 mm for any of the cases. The only effect of the fault irregularities in Case 3, Case 4, and Case 5, that remain at 4500 m appear to be the general reduction of earthquake magnitude, meaning that these cases produce less induced displacement.

In Figure 5-22, the maximum displacement for all six cases are plotted as function of fault-target distance, regardless of target fracture orientation and target fracture region. The following is observed:

- Again it is clear that there is a very systematic dependence on the distance to the primary fault. At distances larger than 700 m from the fault, no displacements on any of the 300 m diameter target fractures exceed 40 mm, regardless of target fracture orientation, target fracture region, hypocentre location, peak stress assumptions and occurrence of sharp fault irregularities.
- Again it is clear that the largest displacements occur in response to exaggerated stress peaks (Case 1) or in response to local stress concentrations around unrealistically sharp fault irregularities (Cases 3 and 5). These result are plotted with dashed lines.
- For cases without abrupt rupture arrest at sharp fault irregularities and with limits prescribed for the stress peak moving ahead of the rupture front (Cases 2, 4 and 6), there are no target fracture displacements in excess of 35 mm at any distance.
Figure 5-15. Cumulative plots of target fracture displacements in Case 1. The plot symbols indicate target fracture orientation. Note the different x-axis scales.

Figure 5-16. Cumulative plots of target fracture displacements in Case 2. The plot symbols indicate target fracture orientation.
Figure 5-17. Cumulative plots of target fracture displacements in Case 3. The plot symbols indicate target fracture orientation.

Figure 5-18. Cumulative plots of target fracture displacements in Case 4. The plot symbols indicate target fracture orientation.
Figure 5-19. Cumulative plots of target fracture displacements in Case 5. The plot symbols indicate target fracture orientation. Note the different x-axis scales.

Figure 5-20. Cumulative plots of target fracture displacements in Case 6. The plot symbols indicate target fracture orientation.
Figure 5-21. Comparison between cases at 300 m, 1500 m and 4500 m fault-target distances. Note that the scales differ between the three distances. Schematic illustrations of the six cases are shown in the right column.
Figure 5-22. Displacement as function of fault fracture distance for the six modelling cases. The curves show the maximum displacements obtained for the six cases, regardless of the fracture orientation and the target fracture region. The three cases considered to be less realistic because of the stress peak evolution (Case 1) and sharp edges close to fracture regions (Cases 3 and 5) are plotted with dashed lines.
6. CONCLUSIONS AND DISCUSSION

6.1 Summary of the results

In previous studies on earthquakes occurring on local, moderately sized, Olkiluoto deformation zones, the calculated induced displacements were modest, in particular at large distances. However, one might expect that very large earthquakes occurring on much larger deformation zones could have much larger ranges of influence and, possibly, induce critical shear displacements also at large distances. In this report we have addressed this question, i.e., of whether or not an earthquake occurring on a large fault that does not intersect the immediate surroundings of the repository could induce large displacement along repository host rock fractures.

We have picked the largest deformation zone considered in the Olkiluoto site description, the steeply dipping BFZ214, as example. The lateral extension and the area of BFZ214 as given in the site description are, however, not nearly sufficient to host an earthquake of magnitude 7 or larger. In the models described here we have simulated earthquakes on a significantly extended, hypothetical version of BFZ214.

In the previous studies, the synthetic Olkiluoto earthquakes were assumed to be triggered and powered by end-glacial stresses. This is in keeping with the notion that the time of ice retreat after the latest glaciation is the only period for which unambiguous evidence of large earthquakes in the Fennoscandian Shield exist. However, in contrast to the gently dipping Olkiluoto deformation zones, the steeply dipping BFZ214 appears to be stabilized at the time of ice retreat (Hökmark and Fälth, 2014). Therefore we don’t assume the earthquake to occur under end-glacial conditions, but under present-day conditions. Moreover, at depth, the BFZ214 deformation zone would be stable with very good margins also under present-day conditions if the local Olkiluoto stress field, as given in the site description, is straightforwardly extrapolated to depths relevant for a large earthquake occurring on the extended version of the zone. For the BFZ214 deformation zone to be reasonably close to failure on the majority of the hypothetically extended fault area, we have chosen here to assume a synthetic strike-slip stress field, i.e., a stress field in agreement with the notion of the crust being in frictional equilibrium determined by the strength of optimally oriented, steeply dipping, brittle deformation zones. The idealized conceptual synthetic stress field as defined by Lund and Schmidt (2011) was slightly modified in order to reproduce the site model stress field at the level of the repository. This was a necessary measure to establish relevant initial stability conditions for the target fractures, and has practically no impact on the fault response.

It should be noted that there are no indications that the synthetic strike-slip field, or any other stress field with similar anisotropy, should be a reality at the Olkiluoto site. Ojala et al (2004) for instance, suggest that the dominating style of deformation in the Finnish bedrock is slow creep at low stress thresholds. It is also worthwhile noting that the Wells and Coppersmith (1994) catalogue does not include any SCR (Stable Continental Region) earthquake of magnitudes larger than $M_w$ 6.7, (cf. Figure 5-1). Given all the above it is evident that an earthquake of magnitude larger than $M_w$ 7 close to the Olkiluoto site is an extremely unlikely event. It is beyond the scope of this study to attempt to set numbers to that small probability. What is presented here are the results of very hypothetical “what - if” scenarios.
The results can briefly be summarized as follows:

1. Regardless of assumptions made regarding the location of the hypocentre, the way stress peaks are allowed to grow ahead of the moving rupture front and occurrences of abrupt transitions from slipping portions of the fault area to infinitely stable portions, none of the 300 m diameter target fractures slipped more than 40 mm at distances of 700 m or more from the magnitude 7 – 7.3 earthquakes modelled here, cf. Figure 5-22.

2. For models without impacts of sharp fault edges or unlimited stress peak growths, none of the 300 m diameter fractures slipped more than 40 mm, even at the shortest (300 m) fault-target distance tried here (Cases 2, 4, and 6, cf. Figure 5-22). At 700 m distance the largest displacements are about 25 mm.

3. There is clear and systematic difference among the different sets of target fractures: with exception of fractures located at the smallest distances (300 and 700 m) from sharp fault edges (Case 3 and Case 5) none of the steeply dipping fractures (square plot symbols in Figure 5-15 - Figure 5-20) slipped more than 15 mm. This is a logical consequence of the high initial stability (low CFS values) of the steeply dipping fractures which are clamped by high horizontal stresses.

4. There is a clear and systematic difference between positions along the strike of the fault, i.e., among results recorded in regions #1 and #2, respectively, cf. Figure 5-21. With exception for Case 3 in which the fault does not extend beyond Region #1, and Case 5 with sharp fault edges close to Region #1, fracture displacements are generally larger in Region #2. There are logical explanations for this: 1) For gently dipping fractures, i.e., those with the smallest stability margins in the initial state, the CFS values have increased around the easternmost fault segments on the repository side of the fault after completed rupture (cf. $\Delta$CFS contours for a gently dipping plane in Figure 5-11, lower right panel), meaning loss of stability. 2) The duration of the stress peak travelling ahead of the rupture front seems to increase with the distance from the hypocentre (cf. Figure 5-10) and is, consequently, larger close to Region #2. That off-fault effects increase in this way with the distance from the hypocentre is a phenomenon observed also by others, see e.g. Andrews (2005).

The above statements, all regarding induced target fracture displacements which is the issue of direct concern for the repository, appear to be consistent with results presented for the fault behaviour and the general off-fault effects. All these results appear, in turn, to be qualitatively understandable. The results of the discretization check presented in section 2.2 indicate that dynamic effects are adequately accounted for. So, given the input assumptions regarding the initial stresses, the fault geometry (for the different cases), the residual fault strength, the shear strength and the perfectly planar geometry of the target fractures, we are confident that the target fracture slip results calculated here are relevant.

We argue that point #1 above effectively suggests that 40 mm should be considered a robust and safe upper bound estimate of the fracture displacements that could be induced on 300 m diameter target fractures by a large earthquake occurring at 700 m distance. It is a matter of judgement to decide how large the margins to corresponding results that potentially would be obtained from models based on realistic, best estimate assumptions of all input parameters need to be. For models without sharp edges and
exaggerated stress peaks, the maximum target fracture slip at 700 m distance was not more than around 25 mm.

One should note that also the scenarios without sharp fault edges and unlimited stress peaks analysed here give earthquakes that plot well above the regressions established by Wells and Coppersmith (1994) and Leonard (2010), cf. Figure 5-1. This is consistent with the high seismic efficiency found in all our synthetic earthquakes (cf. Figure 5-9 for an example); about twice the value suggested in the literature (McGarr, 1999; Scholz, 2002). With a value of the residual strength calibrated to give a more typical value of the seismic efficiency, the resulting earthquake would plot closer to the area-magnitude regressions. The target fracture displacements obtained from such models could not be used directly for layout decisions and risk analyses, but would give perspectives on the degree of conservativeness in the results presented so far.

The sharp fault edges are responsible for the largest target fracture shear displacements obtained here. The difference between Case 4 results (strip, or band, with high but finite stability) and Case 5 (same band with infinite stability) is very clear. At the smallest distance (300 m) the sharp edge model gives almost 90 mm of target fracture slip and the soft edge model a little less than 30 mm. For deformation zones located close to the repository, one issue that would need further attention, regardless of the assumptions made relating to seismic efficiency and stress drop, is to explore defendable ways to represent irregularities, e.g., fault bends, that potentially could have effects similar to those of the stability strips in Case 4 and Case 5. At large distances, 1500 m and more, the influence of local stress concentrations appear to be insignificant and the effect of different irregularities, such as those in Case 4 and Case 5, is rather to reduce the overall effect of the earthquake (Figure 5-22).

There are additional perspectives on the sharp edges and the stability strip in scenarios 4 and 5. In these cases the rupture front, as modelled for simplicity here, propagates without time-delay across the stability strip, meaning that the resulting stress waves and overall dynamic effects of the earthquake are those of one single episode of rupture. A more likely effect of the stability strip, which could represent, for instance, a fault bend, would be that a second episode of rupture is initiated at some low-stability point in the remaining fault segment.

The results obtained here were not influenced by the response of a realistic fracture network, meaning that no natural anelastic attenuation effects were accounted for. This may have contributed to give exaggerated displacement at the largest distances. A realistic network of slipping fractures is likely to redistribute stresses also close to the fault and possibly reduce the largest displacements on the most optimally oriented fracture and instead increase small displacements on others. The potential effects of a fracture network are difficult to quantify, but we judge that not accounting for it has contributed to increase the maximum target fracture displacements at all distances.

The results were not influenced by any variations in rock mass properties, i.e., we have approximated the site model values of the elastic parameters to be valid at all depths in these large models. The parameter of potential importance would be the wave velocities at larger depths. These velocities typically increase by about 20% from the repository horizon, i.e., the model data, down to the bottommost parts of the extended BFZ214 at
about 20 km depth, cf. Hyvönen et al. (2007). Since stress waves generated at fault segments reasonably close to the target fractures will dominate the response, we judge this modest variation to be of little importance.

For a large earthquake scenario involving a large hypothetical gently dipping deformation zone, such as for instance an extended version of BFZ021, the details of the results, but not the overall style of the results, could be different from those obtained here for the case of a steeply dipping zone. The effects of fault strength irregularities would dominate at small distances and die out a few kilometres from the fault. Similar to the case analysed here, it would be necessary to assume a stress field that is much more anisotropic than the extrapolated local site model stress field in order to bring the deeper parts of the zone reasonably close to failure. The differences between results obtained at the different positions along the strike of the fault would probably be less pronounced. Instead it is likely that the target fracture response on the hanging wall side and the footwall side would be different, cf. Fälth (2014).

6.2 Slip estimate margins

To be useful for layout decisions and risk analysis, estimates of seismically induced shear displacements along repository host rock fractures should be bounding estimates. The results shown in Figure 5-22 are all based on earthquake models with the highest magnitudes that, given the fault areas and the extensive amount of empirical data on crustal earthquakes compiled by Well and Coppersmith (1994), are realistically possible (c.f. Figure 5-1). In addition, for all 3DEC earthquakes the seismic efficiency, which is a measure of the relation between the stress drop and the pre-existing shear stresses, is about twice the value suggested for typical crustal earthquakes. As far as the impact of strain energy release is concerned, the results shown in Figure 5-22 are, therefore, judged to be upper bound estimates. At the smallest distances the impact of stress concentrations formed around sharp stability irregularities, such as in Cases 3 and 5, dominate. To get perspectives on the margins, i.e., the potential slip overestimates in Figure 5-22, results obtained from modified versions of two of the cases, Case 2 and Case 5, are presented below. In the revised versions the residual strength is increased to give typical, rather than exaggerated, seismic efficiencies.

The increased residual strength and the consequential stress drop and efficiency reduction does also give magnitude reductions. As seen in Figure 6-1, the revised versions still plot above the Wells and Coppersmith (1994) regression and close to the Leonard (2010) regression. Comparison between the results shown in Figure 5-22 and target fracture slip obtained from these models would therefore give useful perspectives on the slip estimate margins.
Figure 6-1. Close-up of relevant part of Area-Magnitude plot shown in Figure 5-1. Results of revised versions of Case 2 and Case 5 models are added (red triangles).

Figure 6-2 shows target fracture slip obtained in the revised version of the Case 2 model. Corresponding results from the unmodified model are included for comparison (dashed lines). Target fracture slip is generally reduced by about 50%. The stress disturbance generated in the revised model version is obviously not sufficient, even at the smallest distances, to push any of the steeply dipping target fractures over the stability threshold.

Figure 6-3 shows target fracture slip obtained in the revised version of the Case 5 model. Corresponding results from the unmodified model are included for comparison (dashed lines). Target fracture slip is generally reduced by about 50%. All displacements at all distances (300 m to 4500 m) are less than the 50 mm canister damage threshold for the 300 m diameter perfectly planar fractures considered here.
Figure 6-2. Cumulative plots of target fracture displacements obtained in revised version of Case 2 model (with increased residual strength and with seismic efficiency consequently reduced to value typical of real earthquakes). Dashed lines show, for comparison, corresponding results from the unmodified Case 2 model (i.e., same results as in Figure 5-16).

Figure 6-3. Cumulative plots of target fracture displacements obtained in revised version of Case 5 model (with increased residual strength and with seismic efficiency consequently reduced to value typical of real earthquakes). Dashed lines show, for comparison, corresponding results from the unmodified Case 5 model (i.e., same results as in Figure 5-19).
The case giving the largest induced displacements is Case 5 (cf. Figure 5-22) which includes both an exaggerated stress drop and maximized stress concentration effects of fault strength irregularities. It appears that it is sufficient to make more realistic assumptions regarding one of these issues to bring the maximum displacements below 50 mm at all distances:

- Reducing the seismic efficiency from an exaggerated to a typical value (by increasing the residual strength) while keeping the sharp edges of the stability band brings the maximum displacement from 89 mm to about 39 mm (Figure 6-3)
- Softening the stability band (as in Case 4) while keeping the residual strength (and the exaggerated seismic efficiency) brings the maximum displacement from 89 mm to about 25 mm (Figure 5-22). Admittedly, there is no established way to quantify effects of strength irregularities realistically at the present time, meaning that the magnitude of this type of slip reduction is uncertain.

We have not attempted to combine the two approaches to arrive at even smaller maximum slip values at the smallest distances (Table 6-1). It has yet been demonstrated in this section that the Case 5 results presented in Figure 5-22 should count as safe upper bound estimates.

**Table 6-1. Margin estimate overview.**

<table>
<thead>
<tr>
<th>Stability irregularity with sharp edges</th>
<th>Low residual strength, twice typical seismic efficiency. Above regressions in area-magnitude diagram</th>
<th>Increased residual strength, typical seismic efficiency. On regressions in area-magnitude diagram,</th>
</tr>
</thead>
<tbody>
<tr>
<td>Stability irregularity with softened edges</td>
<td>Case 5</td>
<td>Case 5 (re)</td>
</tr>
<tr>
<td>300 m distance</td>
<td>700 m distance</td>
<td>300 m distance</td>
</tr>
<tr>
<td>89 mm</td>
<td>36 mm</td>
<td>39 mm</td>
</tr>
<tr>
<td>Stability irregularity with softened edges</td>
<td>Case 4</td>
<td>Not modelled</td>
</tr>
<tr>
<td>300 m distance</td>
<td>700 m distance</td>
<td></td>
</tr>
<tr>
<td>25 mm</td>
<td>19 mm</td>
<td></td>
</tr>
</tbody>
</table>

6.3 **Final remarks**

In this study we have attempted to show that earthquakes, and even very large ones, occurring on deformation zones located away from the immediate vicinity of the rock volume designated for the nuclear waste repository, will induce only modest secondary slip on typical host rock fractures. Secondary slip has been shown to scale with fracture size (Fälth et al., 2010), meaning that the result shown in Figure 5-22 for 300 m diameter fractures can be used to find upper bound slip estimates for fractures of different sizes.
The upper bound estimates are based on results from models

- with moment magnitudes that, given the available fault area, are as high as reasonably possible. This is accomplished by setting the residual strength low
- with a maximized impact of fault strength irregularities.

The results of the additional modelling presented in the previous section show that it is sufficient to make typical, rather than very conservative, assumptions regarding one of these issues to arrive at significantly reduced estimates of the induced slip. This strengthens the confidence in the upper bound claims made here for the results presented in Figure 5-22.

In southern Fennoscandia, indications of paleoseismicity and postglacial faulting are fewer and weaker than in northern Fennoscandia where postglacial thrust faults are clear evidence of the bedrock response to glacial loading (Lagerbäck and Sundh, 2008). According to Hutri (2007), however, observations of possible seismic reactivation in old bedrock fractures zones in the sea areas around south-western Finland, indicate that the possibility for postglacial faulting in the Olkiluoto region may have been underestimated in the past. For the modelling of a seismic event on a large deformation zone undertaken in this study, however, we have not taken the potential impact of glacial stresses into account. This is because the steeply BFZ214 deformation zone used as example here would tend to be stabilized rather than destabilized under end-glacial conditions and would not be a credible typical postglacial thrust fault (cf. discussion in Chapter 0). It should also be noted that, on the majority of the deep-extending fault area, the glacial stresses are small compared to the existing background stresses, meaning that the results, as far as the fault behaviour is concerned, would differ only in details if glacial stresses were added in our BFZ214 models.

If end-glacial stresses were included in the modelling of a large earthquake on a large steeply dipping deformation zone, such as the modelling presented here, the main additional effect would probably be those of the change in the initial stability of the target fractures, i.e., gently dipping ones would be destabilized and steeply dipping ones stabilized. Given that the most probable large earthquake scenario for the Fennoscandian bedrock (and according to the Hutri (2007) not only in the northern parts of the shield), may be the postglacial faulting scenario, effects of end-glacial stresses should be examined, however rather in a model containing a credible postglacial fault than a steeply dipping fault plane. A suitable deformation zone would be the BFZ021 zone or the BFZ099 zone, possibly extended to depths large enough that the modelling results would capture also effects of hypothetical structures below the repository.
REFERENCES


