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Seismicity and Sources of Stress in Fennoscandia

SØREN GREGERSEN, CONRAD LINDHOLM, ANNAKAISA KORJA, BJÖRN LUND, MARJA USKI, KATI OINONEN, PETER H. VOSS AND MARIE KEIDING

ABSTRACT

The stress field in the Earth's crust and lithosphere is caused by several geological and geophysical factors. This chapter investigates the Fennoscandian area of uplift since the latest Ice Age and addresses the question of whether glacial isostatic adjustment may influence current seismicity. The region is far from plate tectonic boundaries, so investigation occurs in an intraplate area, with stresses caused by the lithospheric relative plate motions, as have been investigated by many authors over the years. Discussions on whether uplift and plate tectonics are the only causes of stress have been going on for many years in the scientific community. We present the earthquake distribution, the uplift pattern, the coast lines and the large postglacial faults, in a geographical overview. This is compared with the geological zones and zone boundaries. This review takes into account the improved sensitivity of the seismograph networks, and at the same time attempts to omit manmade explosions and mining events in the pattern, in order to present the best possible earthquake pattern.

From the earthquake data, focal mechanisms are derived that give indications of the present-day stress orientations. Supplemented by other stress measurements, it is possible to evaluate the stress orientations and their connection to the uplift pattern and known tectonics. Besides plate motion and uplift, one finds that some regions are affected stresswise by differences in geographical sediment loading as well as by topography variations. The stress release in the present-day earthquakes shows a pattern that deviates from that of the time right after the Ice Age. This chapter treats the stress pattern generalized for Fennoscandia and guides the interested reader to more details in the following chapters of this book.

10.1 Introduction and Geological Setting

Fennoscandia is, in a global perspective, a low seismicity region in the NW part of the Eurasian lithospheric plate. It has nevertheless exhibited some of the largest earthquakes in continental Western Europe north of the Alps (latitudes above 38°N) over the past few hundred years (FENCAT, 2020). The current tectonic regimes comprise continental intra-plate regions in the east and the Caledonian mountain range and the passive continental

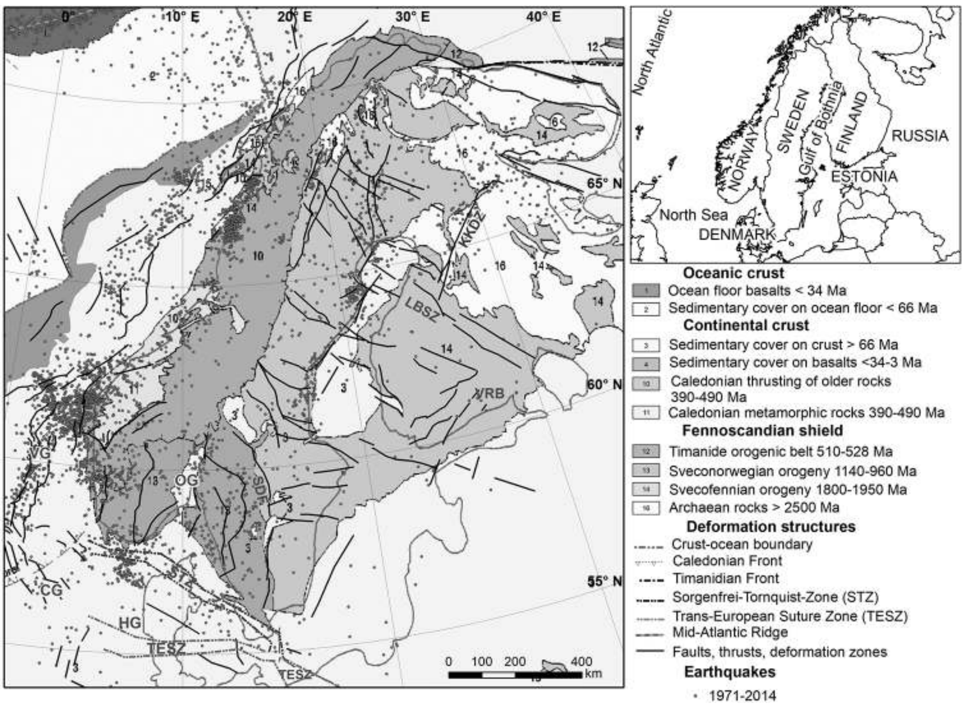

Figure 10.1 Major tectonic units, deformation zones and recent seismicity of Fennoscandia. Tectonic units and simplified deformation zones after Sigmond (2002). Instrumental seismicity, M $\geq$ 2 earthquakes in 1971–2014 from FENCAT (2020). Deformation zones and geological units: KKDZ – Kuusamo-Kandalaksha, LBSZ – Ladoga-Bothnian Bay, SDF – Sveconorwegian Deformation Front, STZ – Sorgenfrei-Tornquist Zone, TESZ – Trans-European Suture Zone and VRB – Vyborg rapakivi granite batholith. Grabens: CG – Central, HG – Horn, OG – Oslo, VG – Viking. (A black and white version of this figure will appear in some formats. For the colour version, please refer to the plate section.)

margin in the west. The continental part consists of the East European Craton (EEC) flanked by the Caledonides in the west to south-west. The craton is partly overlain by Mesoproterozoic to Phanerozoic platform sedimentary cover (Figure 10.1). In the south and south-west, the craton is separated from the Phanerozoic Europe by several deformation zones through Denmark defining the craton boundary via thickness changes in (1) lithosphere, (2) crust and (3) sedimentary cover (see Sandersen et al., Chapter 15). The latest major tectonic event that affected Fennoscandia was the Cenozoic opening and the spreading of the North Atlantic Ocean initiating 60 Ma ago (e.g. Ramberg et al., 2013) with ridge push force affecting all of Northern Europe and with the creation of major faults on- and off-shore western Norway in addition to the major oceanic transform faults.

The complex compressional and extensional evolution left the region with a multitude of deformation zones, faults and fractures that are and have been potential locations of geological reactivation. During the Pleistocene the region was subjected to repeated glaciations with varying duration and ice thickness. The latest/Weichselian glacial retreat started some 19,000 years before present (19 ka BP), and abrupt warming some 11.5 ka BP accelerated the ice retreat (Ramberg et al., 2013). In Fennoscandia, the glacial rebound is still ongoing, with a maximum uplift rate of about 1 cm/year on the Swedish north-east coast.

This chapter provides a regional overview of the seismicity of Fennoscandia and the stresses that drive that seismicity.

10.2 Present-Day Seismicity

Seismicity and sources of seismicity in Fennoscandia have been studied by numerous authors (e.g. Kolderup, 1905; Kjellén, 1912; Renqvist, 1930, and later Stephansson et al., 1986; Slunga, 1989; Bungum et al., 1991; Gregersen, 1992; Fejerskov & Lindholm, 2000; Byrkjeland et al., 2000; Muir Wood, 2000; Fjeldskaar et al., 2000; Pascal & Cloetingh, 2009; Gregersen & Voss, 2009; and many others). The older studies are based primarily on macroseismic observations, whereas the later ones are based primarily on instrumental recordings.

Fennoscandian earthquake observations are quite heterogeneous with respect to the level of magnitude of completeness and uncertainties of source parameters. Although the first seismograph stations in Fennoscandia were already installed in the early 1900s, density of the seismic stations remained for a long time rather sparse, and spatial coverage has been heterogeneous during most of the century. Most earthquakes have taken place on blind faults at significant focal depths. Accuracy of routine earthquake location usually has not been sufficient for identifying an individual fault as an earthquake causative fault. The situation prevails today with a few notable exceptions.

A marked expansion of the national seismic networks has taken place since year 2000, significantly decreasing magnitude detection thresholds and improving event location accuracy. Today, earthquake observations are based on continuous online seismic monitoring by the national seismological networks in Denmark (GEUS), Finland (University of Helsinki and University of Oulu), Norway (University of Bergen and NORSAR) and Sweden (University of Uppsala). Parametric earthquake data are kept in national databases and are also compiled into the Fennoscandian Earthquake Catalogue (FENCAT, 2020). In the following we are using two subsets of the FENCAT data to illustrate regional seismicity patterns in Fennoscandia. The first subset covers historical earthquakes with magnitude $M \geq 4.0$ in 1400–1970 and the second subset $M \geq 2$ instrumentally recorded earthquakes in 1971–2014. The data have been filtered for human-induced events (explosions, rock bursts, collapses, etc.) as well as events of questionable seismic origin (cf. Korja et al., 2016). We recognize that some explosions may still remain in the database, but these are so few that they do not bias the overall spatial distribution. Figure 10.2 includes both datasets and illustrates the spatial variations as well as magnitude variations of the data. Although some of the locations and magnitudes of the larger historical earthquakes are disputed, Figure 10.2 provides a fair overview of the spatial distribution of the earthquake activity as well as the locations of the largest earthquakes.

In a global perspective, Northern Europe is tectonically and seismically quiet, but western Scandinavia is still the most earthquake-active region north of the Alpine mountain

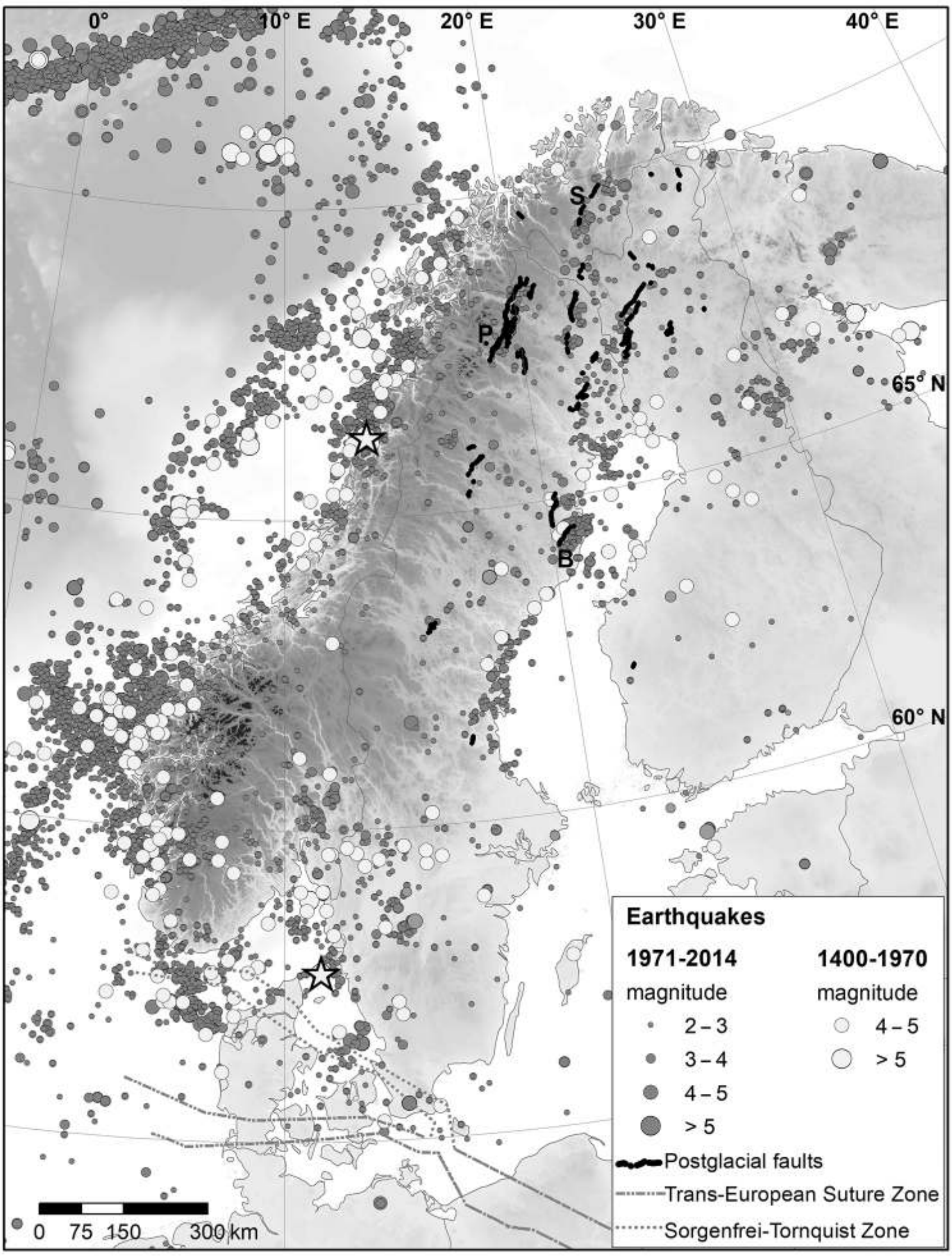

Figure 10.2 Seismicity in Fennoscandia. Earthquakes, magnitude M $\geq$ 4 historical events in 1400–1970 and M $\geq$ 2 instrumental events in 1971–2014 from FENCAT (2020). See legend for more information. The 1759 Kattegat and 1819 Lurøy earthquakes are marked with asterisks and the Burträsk, Pärvie and Stuoragurra faults with letters B, P and S. Postglacial faults from Munier et al. (2020), topographic data from GLOBE Task Team (1999) and bathymetric data from NOAA (Amante & Eakins, 2009).

chain. Figure 10.2 demonstrates the spatial relationship between earthquake activity and topography, with less activity in the eastern lowlands and the central mountain chain and higher activity in the westernmost offshore and coastal regions.

Seismicity in Fennoscandia is classified as intraplate within the Fennoscandian Shield, seismicity of various kinds at the craton boundary in Denmark, failed rift seismicity in the North Sea, Skagerrak and Oslo Graben and passive margin seismicity on- and off-shore the Norwegian coast (Figure 10.2). The largest earthquakes are localized along the failed rifts (Oslo and Viking grabens) and along the passive margin off-shore the Norwegian coast. Norway recognized in 2019 the '200-year anniversary' for the largest earthquake in continental Western Europe north of the Alps: the 1819 Lurøy earthquake (Figure 10.2). Although rare, earthquakes with magnitude M > 5 are found both in the historical and the instrumental datasets (e.g. Hansen et al., 1989; Pirli et al., 2010).

As can be seen in Figure 10.2, the seismicity in the Archean and Palaeoproterozoic part of Fennoscandia is to some extent diffusely distributed and, in some areas, more well-defined. In large parts of Fennoscandia, earthquakes and mapped faults are only marginally overlapping (Gregersen & Voss, 2010, 2014), whereas for other regions, e.g. Norway shelf regions, earthquake and fault density correlate well. Some of the large well-known post-glacial faults (PGFs) are still very active (e.g. Lindblom et al., 2015): the Burträsk Fault (Figure 10.2) is the most seismically active area in Sweden. Clusters of seismic activity are found along the north-east coast of Sweden and the NE-SW-trending Kuusamo–Kandalaksha deformation zone in Finland and north-west Russia (Figure 10.1). Recently, very shallow microearthquake clusters have been observed along minor faults in the Vyborg rapakivi granite batholith in south-eastern Finland and adjacent Russia (Uski et al., 2006; Smedberg et al., 2012; Assinovskaya et al., 2019). The Neoproterozoic Sveconorwegian Vänern region in south-western Sweden has also an elevated level of seismicity, associated with the Sveconorwegian Deformation Front.

The Sorgenfrei–Tornquist Zone (STZ) is the southern boundary of the Fennoscandian Shield. The STZ is a major deformation zone, across which the crustal and lithosphere thickness change abruptly (Berthelsen, 1998; Wilde-Piórko et al., 2002; Gregersen et al., 2006) and which has been suggested to be an old plate boundary (e.g. Mazur et al., 2015). Together with the Trans-European Suture Zone (TESZ), STZ outlines a wedge-shaped block in the south-western most corner of the craton. The block is highly deformed and has a thick Phanerozoic sedimentary cover (Gregersen et al., 2008). Seismicity is broadly associated with the STZ and its south-eastern extensions. South-west of the Danish activity there are, essentially, no earthquakes.

The failed Permian rifts, i.e. Oslo Graben in southern Norway and in the Viking Graben and Central Graben in the North Sea west of Norway and Denmark, seem to exhibit enhanced seismicity compared to surrounding regions. These regions have thinned crust and large normal faults, and the observation of enhanced seismicity in these regions is corroborated by global observations from stable continental regions (SCR), e.g. Landgraf et al. (2017). An outlier in this respect is the large 1759 Kattegat earthquake (e.g. Muir Wood, 1989), which is not easily associated with a graben structure, albeit by a steep crustal thinning in the region of the earthquake. In the sea between Sweden, Denmark and

Norway, very small earthquakes take place; their tectonic significance has been energetically debated (Hansen, 1986; Gregersen & Voss, 2010; Mörner, 2003).

The most intense earthquake activity in Fennoscandia (both in terms of regularity and the largest magnitudes) is found (i) offshore western Norway, (ii) in the Norwegian coastal region between steep mountains and offshore sedimentary basins and (iii) along the western shelf edge west of mid Norway. The passive margin is greatly extended, and it comprises seismically active zones along the coast and shelf edge.

Information on the focal depth distribution is not optimal due to the combination of sparse station density and large lateral variations in the crustal structure. Routine source depth estimates may contain significant uncertainties, and fixed depth estimates are frequently used by some seismic observatories.

Lindholm et al. (2000) made a detailed study for the west coast of Norway where the focal depth was somewhat shallower (10–15 km) near shore and onshore than what was found offshore (>20 km). NORSAR and NGI (1998) furthermore investigated Norwegian focal mechanisms and found that reverse faulting earthquakes had a median depth of 20 km, whereas the normal faulting events were shallower at 15 km median depth. Based on a subset of the most recent intraplate earthquakes in the Fennoscandian Shield, Korja and Kosonen (2015) and Korja et al. (2016) suggested that shallow seismicity (down to ∼15 km in depth) dominates in most of Finland and northern Sweden. A trend of westward deepening of seismic sources was observed in the Gulf of Bothnia and onshore Sweden, but only a few events occurred below ∼35 km in depth. They further suggested that a detachment zone controlling the depth extent of local fault zones is also controlling the depth distribution of seismicity. Deeper earthquakes, near the bottom of the crust or just below the crust, seem to be associated with major/crustal scale deformation zones, like the STZ in Denmark.

The seismic activity along the PGFs in northern Norway, Sweden and Finland is distinct from the regional background seismicity in terms of location, number of events and magnitude (Figures 10.2 and 10.3). There is a remarkable correlation between activity of low-magnitude seismicity and the mapped PGFs (Lagerbäck, 1978; Arvidsson, 1996, Lindblom et al., 2015). In the map view, the seismicity mainly clusters south-east of the main fault scarps, in accordance with the south-easterly dip of the faults. A study based on earthquake data from permanent stations and a local seismic network around the Pärvie Fault concluded that 71 per cent of the observed earthquakes north of 66° N in Sweden locate within 30 km to the south-east and 10 km to the north-west of the PGFs (Lindblom et al., 2015). Reflection seismic surveys have shown that the reverse Pärvie Fault dips steeply 50–65° to the south-east (Juhlin et al., 2010; Ahmadi et al., 2015), suggesting that the main event reactivated an old weakness zone in the crust. The seismicity near the Pärvie Fault does not correlate with a simple fault plane, but rather occurs in a zone dipping 30–60° to the south-east (Lindblom et al., 2015), that is, seismicity mainly locates in the crustal volume above the fault plane. Well-constrained earthquakes near the Pärvie Fault locate down to a depth of 35 km, indicating that the crust is seismogenic to at least 35 km depth (Lindblom et al., 2015). The focal mechanisms around the PGFs are mainly oblique reverse to strike-slip (see Figure 10.3).

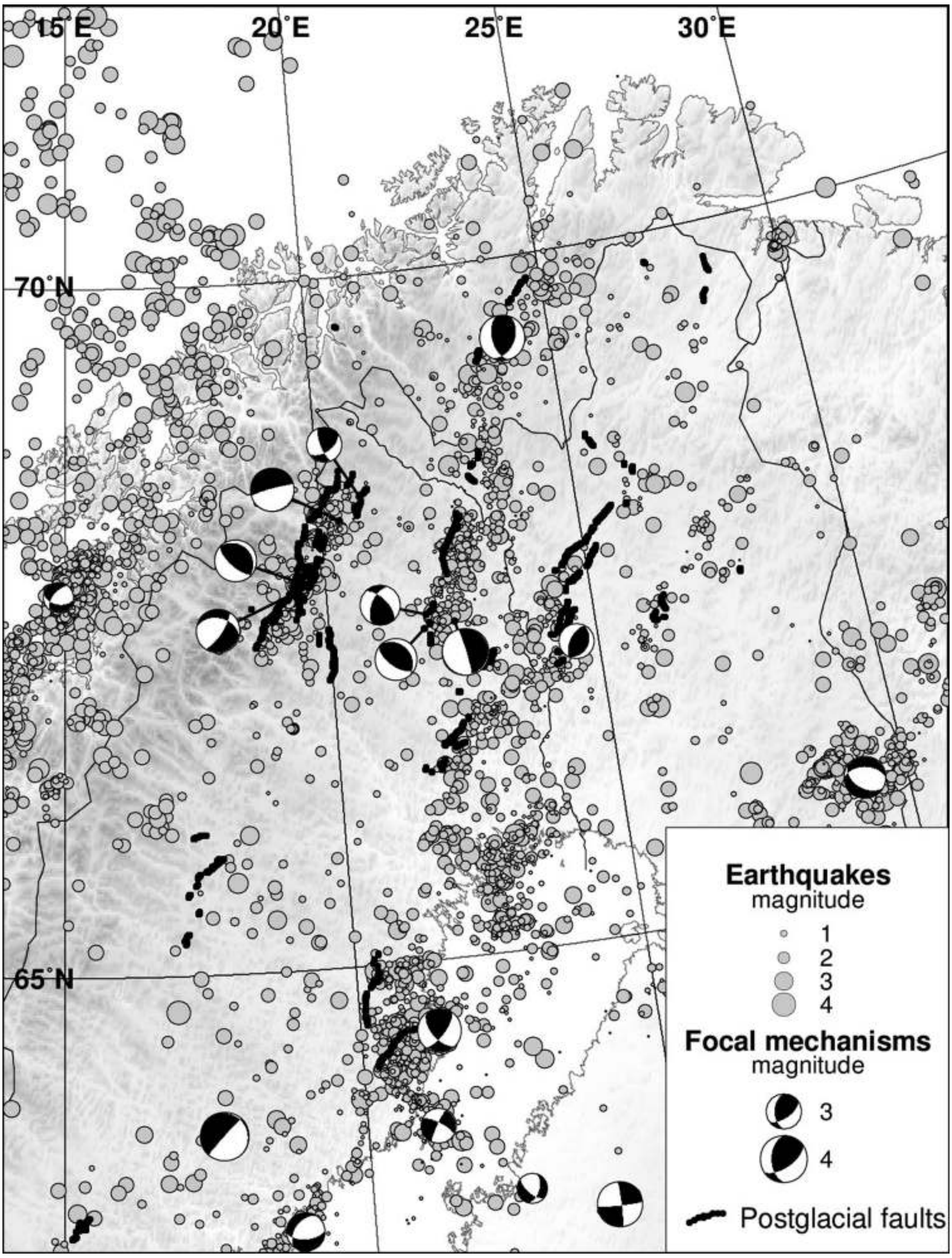

Figure 10.3 Seismicity near the postglacial faults in northern Fennoscandia. Earthquakes, all recorded events during 1971–2014 from FENCAT (2020). Earthquake focal mechanisms, M $\geq$ 2.5 events from a compilation in Keiding et al. (2015). Postglacial faults from Munier et al. (2020).

Interestingly, only the PGFs north of 64° N correlate with elevated seismicity. The proposed PGFs in southern Sweden, Denmark and Germany are not associated with current seismicity. The correlation between PGFs and seismicity is strongest in northern Sweden, corresponding to the region with the highest modelled glacial isostatic adjustment (GIA) stresses (Lund et al., 2009). The cause of the elevated seismicity is not fully understood. Lindblom (2011) showed that focal mechanisms on the Pärvie Fault imply a mainly tectonic stress regime. An alternative suggestion is that the elevated seismicity is caused by the static stress change induced by the large earthquakes that created the PGFs (Ronald Arvidsson, personal communication, 2016). Such a long-term effect would indicate a very long lithospheric relaxation time, or a very low tectonic stressing rate, as Stein and Liu (2009) argued for very long intraplate aftershock sequences. Yet another possibility is that the seismicity occurs as a result of the remaining GIA stresses or possibly a combination of all three processes.

10.3 The Regional Stress Direction in Fennoscandia

Sources of information on crustal stress are diverse and reflect the different stress conditions at different depths. Earthquake focal mechanisms reflect conditions at the hypocentre, and other methods such as overcoring and borehole breakouts reflect stresses closer to the surface. In the following, we will put more weight on the regionally significant stress derived from earthquake focal mechanisms.

An earthquake is the result of shear stresses on a specific structure. If it is possible to estimate reliably, then the earthquake source parameters provide important information on the crustal stress and the rupture process. For large earthquakes (M = 6+) a highly detailed model of the rupture and of the causative stress is possible (like what is routinely done by many international agencies today). However, for many smaller earthquakes, frequent in Northern Europe, focal mechanism qualities remain uncertain. In the study area, each country has a national database of focal mechanisms where magnitudes, quality and determination methods vary greatly.

From Norwegian regions some 200 focal mechanisms have been established, mostly from small and local earthquakes on- and offshore, with large uncertainty in source mechanisms. However, even if the individual mechanisms are very uncertain, the general pattern seems to be consistent. The focal mechanisms suggest a regionally consistent NW-SE compression with significant regional deviations. In Norway, reverse faulting dominates with pockets of normal faulting locally (Fjeldskaar et al., 2000; Hicks et al., 2000; Hicks & Ottemøller, 2001; Janutyte & Lindholm, 2017). In Sweden and Finland, there are some focal mechanisms from small earthquakes (Slunga, 1991; Lindblom, 2011; Uski et al., 2003; Uski et al., 2006). These point to a NW-SE to WNW-ESE direction for the maximum horizontal compressional stress release. Oblique strike-slip to reverse mechanisms with some local variations dominate in Finland, Estonia and north-west Russia, whereas in and around the Gulf of Bothnia, strike-slip is the dominant component of motion, although normal faulting also occurs. Recent focal mechanisms for small mid-crustal earthquakes along the eastern coast of Bothnian Bay suggest a transtensional setting where strike-slip faults have a small to significant component of extension. These events are spatially associated with the major Ladoga–Bothnian Bay shear zone. Reverse mechanisms occur more frequently in southern Finland and north of the Bothnian Bay area (cf. Lindblom et al., 2015; Korja & Kosonen, 2015; Korja et al., 2016 and references therein).

In south-central Sweden, earthquake focal mechanisms are generally of strike-slip type, with a clear NW-SE direction of compression (Slunga, 1991). In Denmark only few focal mechanisms are available. They show large uncertainty but are judged to be of the same class as those of southern Norway and southern Sweden.

In addition to plate-tectonic stress sources, the isostatic uplift (Figure 10.4) since the last glaciation, which ended about 9,000 years ago, is expected to contribute to the observed crustal stress in Fennoscandia. Discussions on whether uplift and plate tectonics are the only causes of stress and seismicity have been going on for many years in the scientific community (e.g. Gregersen & Basham, 1989; Arvidsson et al., 1991; NORSAR & NGI, 1998; Hicks et al., 2000; Gregersen, 2002; Lund, 2015; Pascal et al., 2010; Bungum et al., 2010; Redfield & Osmundsen, 2015; Korja & Kosonen, 2015; Brandes et al., 2015; Keiding et al., 2015) and are also to some extent discussed in Chapters 11–18 of this book.

Earthquake focal mechanisms reflect the stress field present at depth in the crust. Other methods, such as overcoring and borehole breakouts, reflect stresses closer to the surface. Stress magnitudes are significantly more difficult to estimate than stress directions. Therefore, most techniques only give the directions of the principal stresses, sometimes rotated into the maximum σ_H and minimum σ_h horizontal stresses. Observations on regional stress field indicators have been collected and standardized in the global World Stress Map Project (www.world-stress-map.org/). We are using data from Northern Europe (Figure 10.5; Heidbach et al., 2016) comprising earthquake focal mechanisms, borehole breakouts and in situ stress measurements (overcoring, hydraulic fracturing, borehole slotter). Heidbach et al. (2016) classified the data based on their reliability and precision into four classes A, B, C and D, with standard deviation of $\sigma_H \pm 15°$, $\pm 15-20°$, $\pm 20-25°$ and $\pm 25-40°$, respectively. Northern European observations in classes A–C are shown in Figure 10.5. The data are largely based on earthquake focal mechanisms but are supplemented by three other datasets (mainly offshore borehole breakouts). It is furthermore important to recognize that much more stress information exists in each Nordic country that has not yet been added to the WSM global database. Nevertheless, the maximum horizontal stress field indicator data indicate that NW-SE compression prevails throughout Northern Europe even at shallow depths. This NW-SE compression is most often attributed to the ongoing spreading process at the Mid-Atlantic Ridge, which exhibits a continuous ridge-push tectonic stress field with maximum horizontal compression oriented approximately NW-SE.

In evaluating the stress directions in Figure 10.5 it is, however, important to recognize that stresses are deduced from quite different methods, crustal depths, topographic terrains and rock conditions, which all have their influence on the derived stresses. In SCR like Fennoscandia the relative stress magnitudes are small, and the magnitude-wise similarity between σ_H and σ_h consequently results in a small deviatoric stress regime. This was lately

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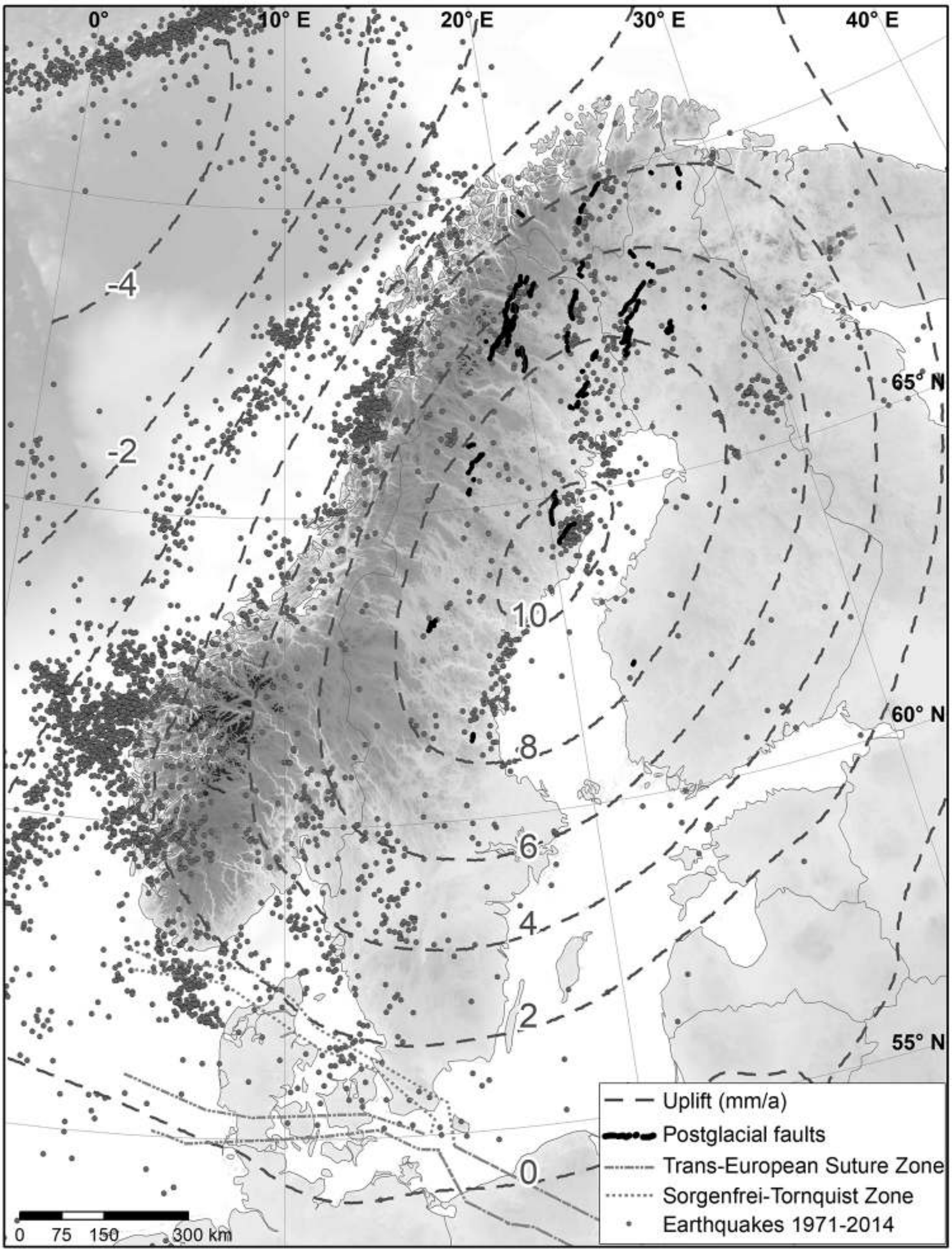

Figure 10.4 Recent seismicity and postglacial uplift in Fennoscandia. Seismicity data, M $\geq$ 2 earthquakes in 1971–2014 from FENCAT (2020). The absolute land uplift velocity (mm/a) model for Fennoscandia NKG2016LU_abs of the Nordic Geodetic Commission (Vestøl et al., 2019). Postglacial faults from Munier et al. (2020), topographic data from GLOBE Task Team (1999) and bathymetric data from NOAA (Amante & Eakins 2009).

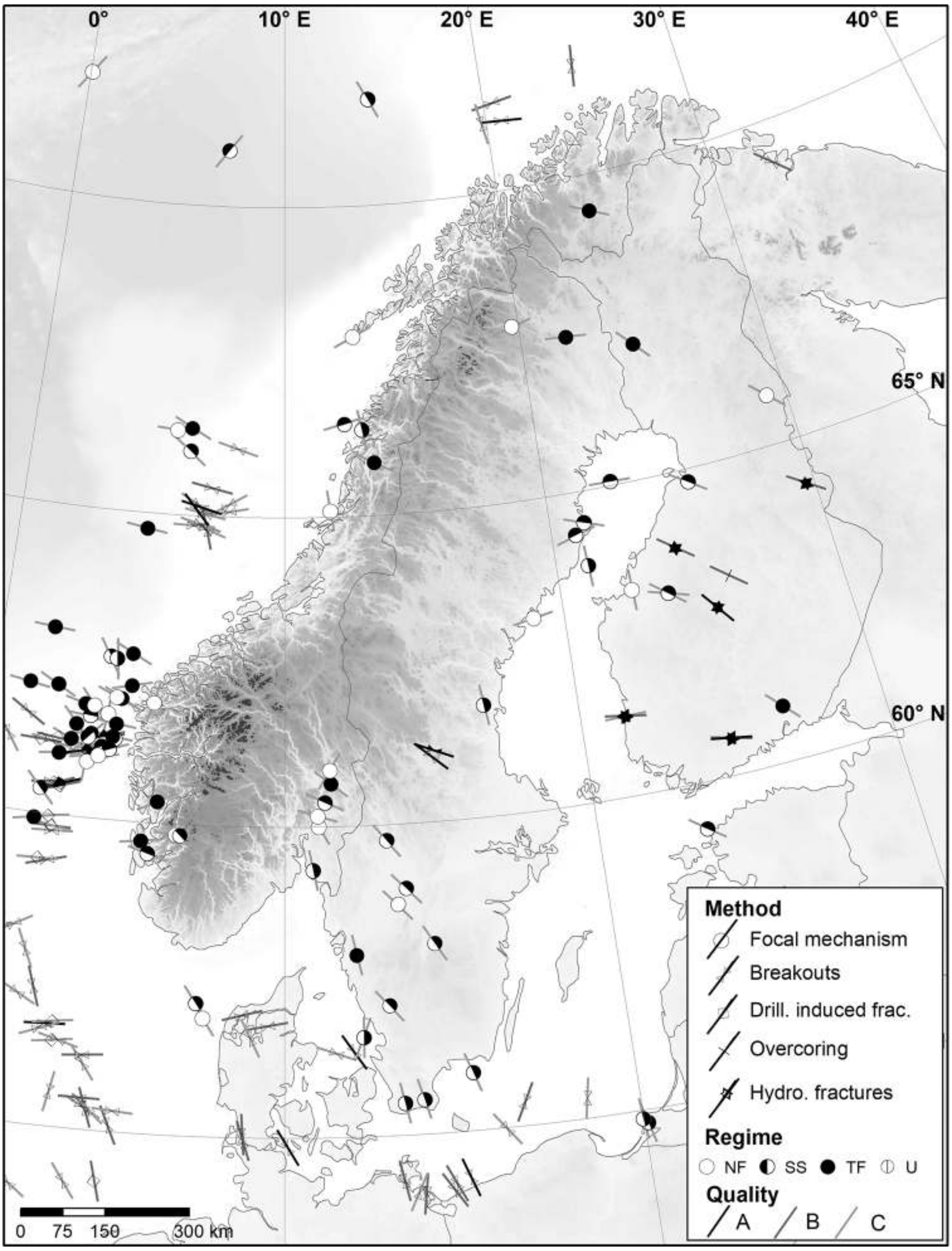

Figure 10.5 Contemporary stress field in Fennoscandia. The map displays quality A–C records from the 2016 World Stress Map database, complemented with some quality C earthquake focal mechanisms in Sweden and Finland. Shallow measurements (depth < 500 m) are excluded. Lines show the orientation of maximum horizontal stress $S_{H\max}$ and line shade represents quality ranking. Different stress regimes (NF – normal faulting, SS – strike-slip faulting, TF – thrust faulting, U – unknown) and stress indicators are marked with different symbols, see legend for more information. Topographic data from GLOBE Task Team (1999) and bathymetric data from NOAA (Amante & Eakins, 2009).

confirmed for northern Norway in the comprehensive NEONOR 2 project, where also regions of 90° stress rotation are identified (Olesen et al., 2018; Janutyte & Lindholm, 2017).

The other primary source of deformation relates to the Pleistocene glaciations. Northern Europe has been subjected to repeated glacial cycles and associated loading and unloading events resulting in reshaping of the geomorphology (Donner, 1995; Fredén, 2002). Loading and unloading of the lithosphere with an ice sheet produces an isostatic imbalance which is compensated by GIA, where the viscous mantle flows to accommodate variations in ice load and thereby produces subsidence or uplift. Today, the remaining isostatic imbalance is causing slow land uplift centred in north-eastern Sweden and the Bay of Bothnia, and according to a recent absolute land uplift model by the Nordic Geodetic Commission (NKG) NKG2016LU (Vestøl et al., 2019), the maximum rate of absolute uplift is 10 mm/year (Figure 10.4). The model demonstrates that GIA is still taking place (e.g. Janutyte & Lindholm, 2017; Keiding et al., 2015; Kierulf et al., 2014; Fjeldskaar et al., 2000; Fejerskov & Lindholm, 2000), causing an additional stress contribution in Fennoscandia (see also Steffen et al., Chapter 2).

10.4 Discussion on Earthquake Causes

Already Lagerbäck (1978) and Muir Wood (1989) suggested that landslides, faulting and associated enhanced earthquake activity in Lapland is related to the deglaciation process. This has been confirmed by Lagerbäck and Sundh (2008), Ojala et al. (2018) and Sutinen et al. (2019). In Figure 10.4 it is observed how the northern large postglacial faults are associated with present-day microseismic activity and how the pattern of seismic activity has poor correlation with the uplift curves in general. A good correlation between observed seismic activity and the uplift curves gradient is only found in the Norwegian coastal areas, where the uplift curves align with the coastlines and where the crust bending is expected to be at its maximum. Recently Lindholm (2019) observed, how the present-day earthquake activity is spatially much better correlated with the deglaciation isochrons than with the uplift model (and with present-day glaciated regions).

A key question raised early on was the source of the stress causing deformations in plate interiors. Pioneering work on global distribution of crustal earthquake-driving stresses were published from the mid-1970s by the seminal contributions of Solomon et al. (1975), Richardson et al. (1979), Harper (1989) and Zoback et al. (1989), and for Fennoscandia by Stephansson et al. (1986), Slunga (1989), Bott (1991) and Bungum et al. (1991) and many others. The relative importance of various crustal stress generating mechanisms in Fennoscandia, such as ridge-push, GIA, sedimentation and crustal density contrasts, have been discussed and modelled (e.g. Gregersen & Basham, 1989; Arvidsson et al., 1991; Hicks et al., 2000; Fejerskov & Lindholm, 2000; Muir Wood, 2000; Gregersen, 2002; Pascal & Cloetingh, 2009; Lund et al., 2009; Pascal et al., 2010; Lund, 2015; Redfield & Osmundsen, 2015; Brandes et al., 2015; Korja & Kosonen, 2015; Olesen et al., 2018). On a regional scale it has been difficult to conclude on the relative importance of these four stress sources. Recently, however, modelling has shown that GIA and ridge push may act constructively and increase crustal stresses in regions like middle Norway (Muir Wood, 2000; Fejerskov & Lindholm, 2000; Gregersen & Voss, 2009 and others).

The seismicity in Northern Europe can be classified as intraplate with preferred earthquake sources along (a) rifted passive margins, (b) palaeosutures and (c) failed rifts. The seismicity distribution is largely consistent with conclusions from global studies of so-called SCR (Johnston et al., 1994; Schulte & Mooney, 2005; Landgraf et al., 2017), maintaining that rifted passive margins and failed rifts are the two main types of host structures responsible for the largest earthquakes in such areas. A new concept that challenges the 'ridge-push concept' as the primary earthquake cause for the observed earthquake activity in western Fennoscandia has been proposed by Redfield and Osmundsen (2015). The concept takes its outset from the geological development perspective (not a stress perspective) and suggests that the first order patterns of Fennoscandian seismicity reflect the domain boundaries of the Mesozoic rifted margin. Redfield and Osmundsen (2015) identify three distinct belts of earthquakes striking subparallel to the generalized line of breakup, and its originality is that it refers to large crustal weakness zones and boundaries as a first order explanation to the observed seismicity (rather than stress). We will in this chapter not pursue these ideas further, but the reader should be aware of this alternative/complementary perspective that emphasizes pre-existing weakness zones in the crust as important for the earthquake generation.

The crustal stress generated by the GIA uplift has been modelled by many groups (e.g. Fjeldskaar, 2000, and a comprehensive review by Steffen & Wu, 2011), and they have demonstrated that major compressional stresses in the lower crust are large in the western coastal region of Norway (lately in Olesen et al., 2018). See more detailed discussion in this book.

A fundamental observation is that the exposed bedrock areas of Fennoscandia (Figures 10.2 and 10.4) exhibit more earthquakes than the surrounding sediment-covered platform areas of Denmark, Germany, Poland, Baltic States and Russia. This could be caused by crust/lithosphere differences or by the uplift/downwelling. But already in the discussion of Figures 10.1 and 10.5, we have concluded that the spatial distribution of earthquakes does not correlate well with the overall uplift pattern. So even if the postglacial uplift contributes to crustal stress, it cannot alone drive the present-day seismicity. It can, however, trigger seismicity.

Zoback et al. (1989) confirmed the earlier models of Richardson et al. (1979), where stress in the plate interiors is attributed to the relative plate motions at the plate boundaries. For Fennoscandia, the ridge-push has been interpreted as the dominating source of stress (Stephansson et al., 1986; Fejerskov & Lindholm, 2000; Korja & Kosonen, 2015; Gregersen & Voss, 2014). While the regionally observed stress directions largely support the plate boundary interpretation, the detailed analysis of focal mechanisms clearly indicates that other stress-generating mechanisms are also at work in some regions (Hicks et al., 2000; Janutyte & Lindholm, 2017; Olesen et al., 2018; Lindholm, 2019). A sometimes overlooked stress-generating factor is the gravitational effect caused by lateral changes in lithosphere thickness, as e.g. at the transition between oceanic and continental crust

(Fejerskov & Lindholm, 2000; Pascal & Cloetingh, 2009). Specifically, the relief of the western Scandinavian mountain range as well as the sharp crustal thickness variations play a vital role, as already pointed out by Fejerskov and Lindholm (2000). The Figure 10.4 uplift contours show a poor correlation with the pattern of the earthquake activity; however, when uplift isochrons are plotted with earthquake distribution some correlation is striking (Lindholm & Bungum, 2019; Lindholm, 2019). This indicates that postglacial uplift is influential in a different way: as shown by Lindholm (2019), the regions with the highest gradient in the deglaciation isochrons are clearly showing more present-day earthquakes than the regions of highest uplift gradients show. The mentioned curves on the maps are dependent on mathematical interpolation methods, so a discussion will have to be sorted out via comparison of mathematical interpolation methods.

While the existence of crustal stress is a natural prerequisite for earthquake generation, the existence of 'lubricated' faults that are favourably oriented in the stress field is another prerequisite (Copley, 2017; Landgraf et al., 2017). The existence of lubricated faults is in many cases more important than regional stress because large structures may themselves alter the surrounding stress field (e.g. Fjeldskaar et al., 2000) and observed local σ_H may deviate from the regional stress field.

An enigma in terms of stress and earthquake activity is presented by the large northern PGFs. The structures are all NE-SW striking, and they presumably ruptured during large earthquakes (less than 10,000 years ago) in postglacial times. Very recent (2018–2019) new trenching analyses from the Stuoragurra Fault, in Norway, surprisingly revealed rupture activities as late as 4,000–600 years ago (Olsen et al., 2020). These findings are so new that the full implication of these results is yet to come. Today, microseismicity is observed and clearly associated with these faults. Large earthquakes such as the ones that ruptured the PGFs are not expected in the current stress field.

We have above concentrated our short discussions of the PGFs to the main structures in the north and refrained from discussing the more recent claims of PGFs in central and southern Fennoscandia (and a few others in the north). These claims (among them, Mörner, 2003; Olesen et al., 2013; Mikko et al., 2015; Brandes et al., 2015) are important indicators that more surface deformations of postglacial origin may exist, and these are also analyzed and discussed more in this book.

The estimated tectonic strain rates for Fennoscandia are very small, and measuring these (attempted by Scherneck et al., 2001; Keiding et al., 2015) is further complicated by the fact that the surface velocity field is dominated by GIA, in both vertical and horizontal directions (e.g. Lidberg et al., 2010; Kierulf et al., 2014). Any estimate of the tectonic deformation signal in the Fennoscandian strain rate field must therefore attempt to quantify the GIA effect, which is very difficult within the uncertainty limits of the Global Positioning System (GPS) signal (e.g. Scherneck et al., 2001; Keiding et al., 2015). Previous geodetic studies concluded that glacial rebound drives earthquake activity in Fennoscandia (e.g. Gudmundsson, 1999), but mounting evidence from focal mechanisms and stress measurements as well as more detailed GIA modelling indicate that the remaining glacially induced stresses are small in magnitude (e.g. Wu et al., 1999; Lund et al., 2009). This implies that even though GIA may act as a stress contributor to the seismicity the lithospheric plate motions or other stress sources such as sediment loading or topographical loading are probably the main driving forces of Fennoscandian seismicity.

10.5 Summary

The Fennoscandian intraplate area can, in terms of seismicity, be divided between the more active mountainous region to the west (Norway with Caledonian overthrust belt) and the low-active region of the Fennoscandian Shield to the east (Sweden and Finland). The seismically most active zones appear to be optimally oriented for reverse and strike-slip faulting driven by the opening of the Atlantic.

The current strain rates in Fennoscandia are low but high enough to reactivate old structures, joints and extension fractures. As is well known, the reactivation of pre-existing deformation zones and faults depends on the stress field and the fault frictional state. What is often not sufficiently recognized is that major structures also alter the crustal stresses so that nearby structures may experience enhanced or decreased shear stress and with a direction different from the regional trends.

Finally, we may highlight the many-sided observations on the locations of modern Fennoscandian earthquakes. Many earthquakes are concentrated within and along mapped fault zones, and the larger earthquakes seem to favour aborted rift zones with thinned crust. Many earthquakes occur along the passive continental margin where the crustal thickness is subject to significant lateral variations, and along parts of the Norwegian coast. In other regions several of the scattered earthquakes occur unexpectedly on unknown or disregarded deformation zones. The Fennoscandian earthquake activity takes place in the middle to lower crust in the shelf regions and at shallower depths in the coastal regions, and in the upper to middle crust in the shield area (Finland and Sweden). The northern PGFs are associated mostly with shallow upper crust earthquakes.

The glacial isostatic adjustment contribution cannot be ignored, but the contribution is, on the regional scale, considered less than the plate tectonic contribution. In certain regions the GIA effect may, however, be important.

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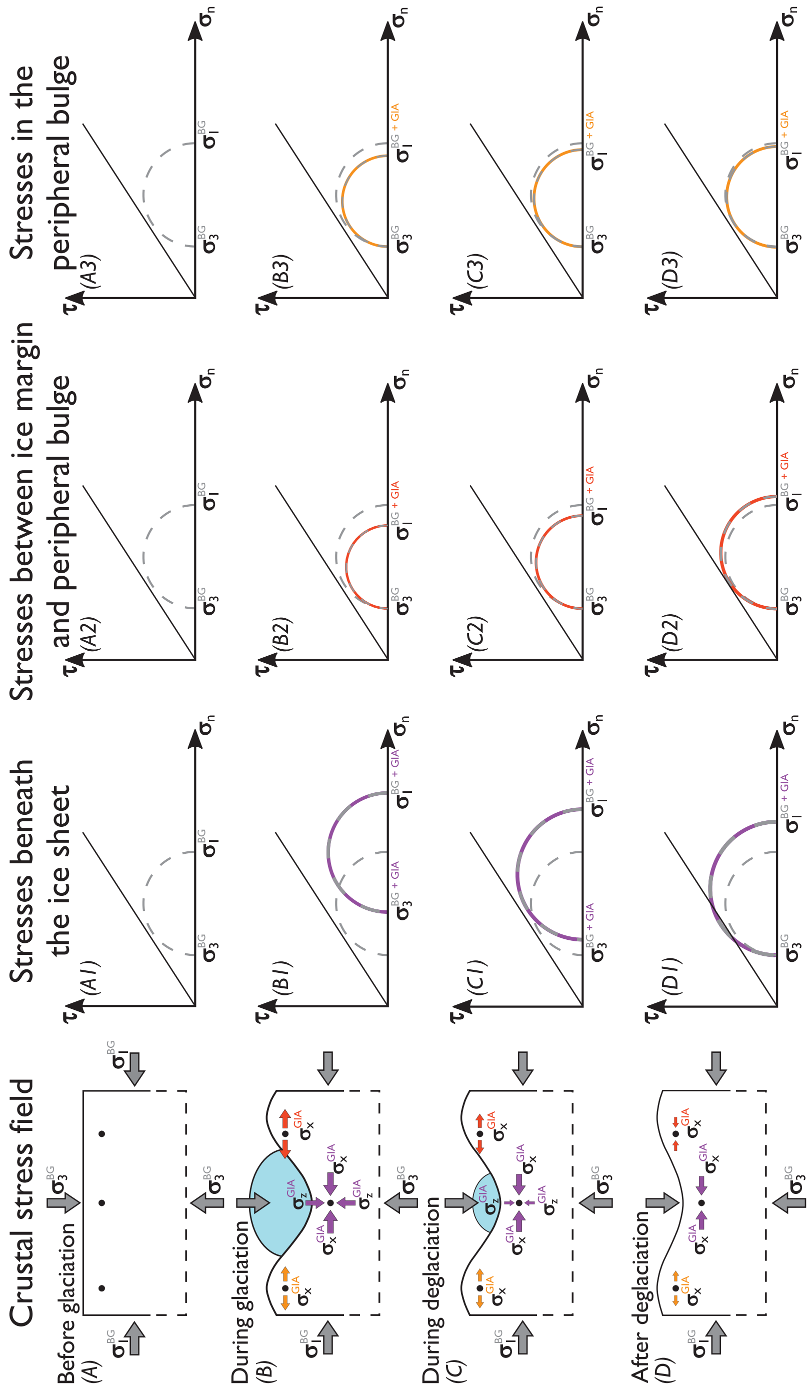

Figure 2.4

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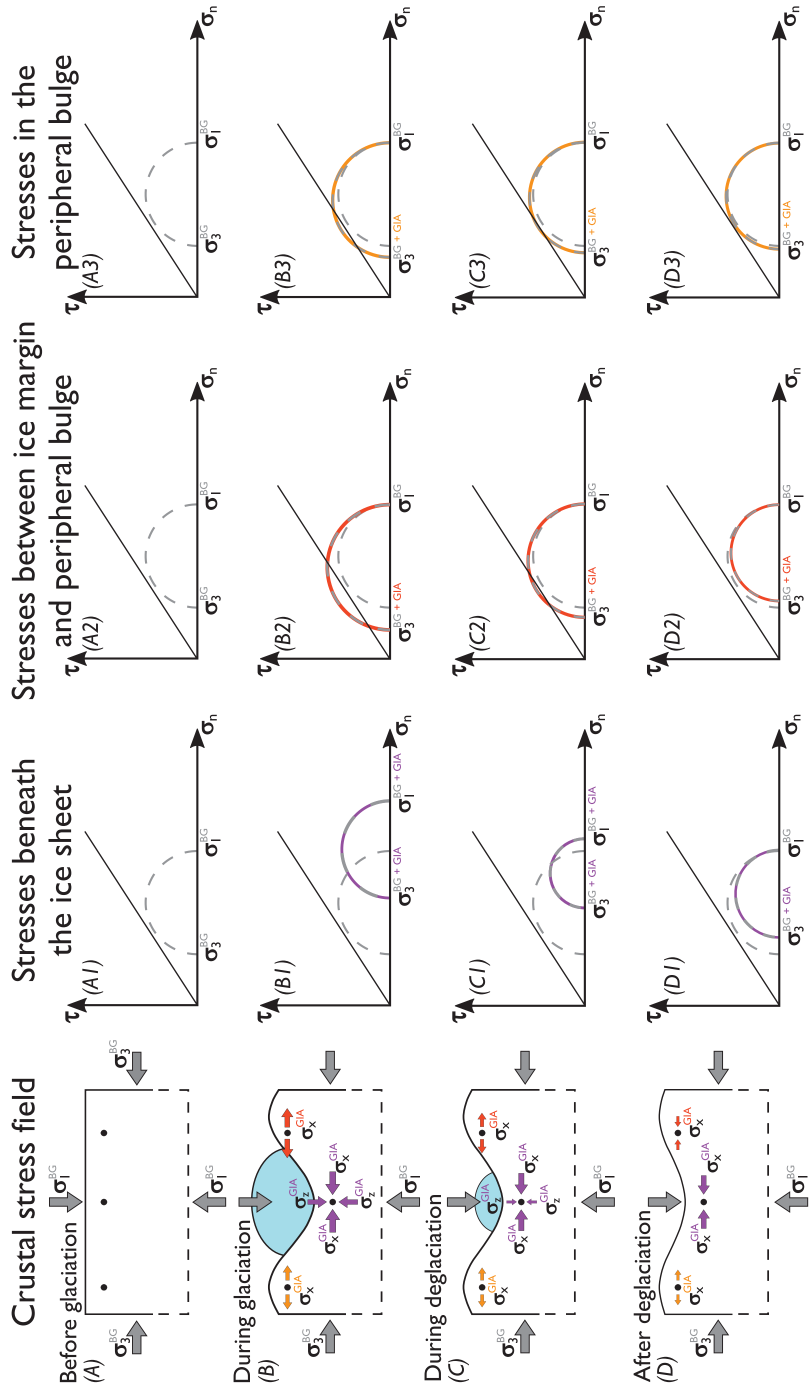

Figure 2.6

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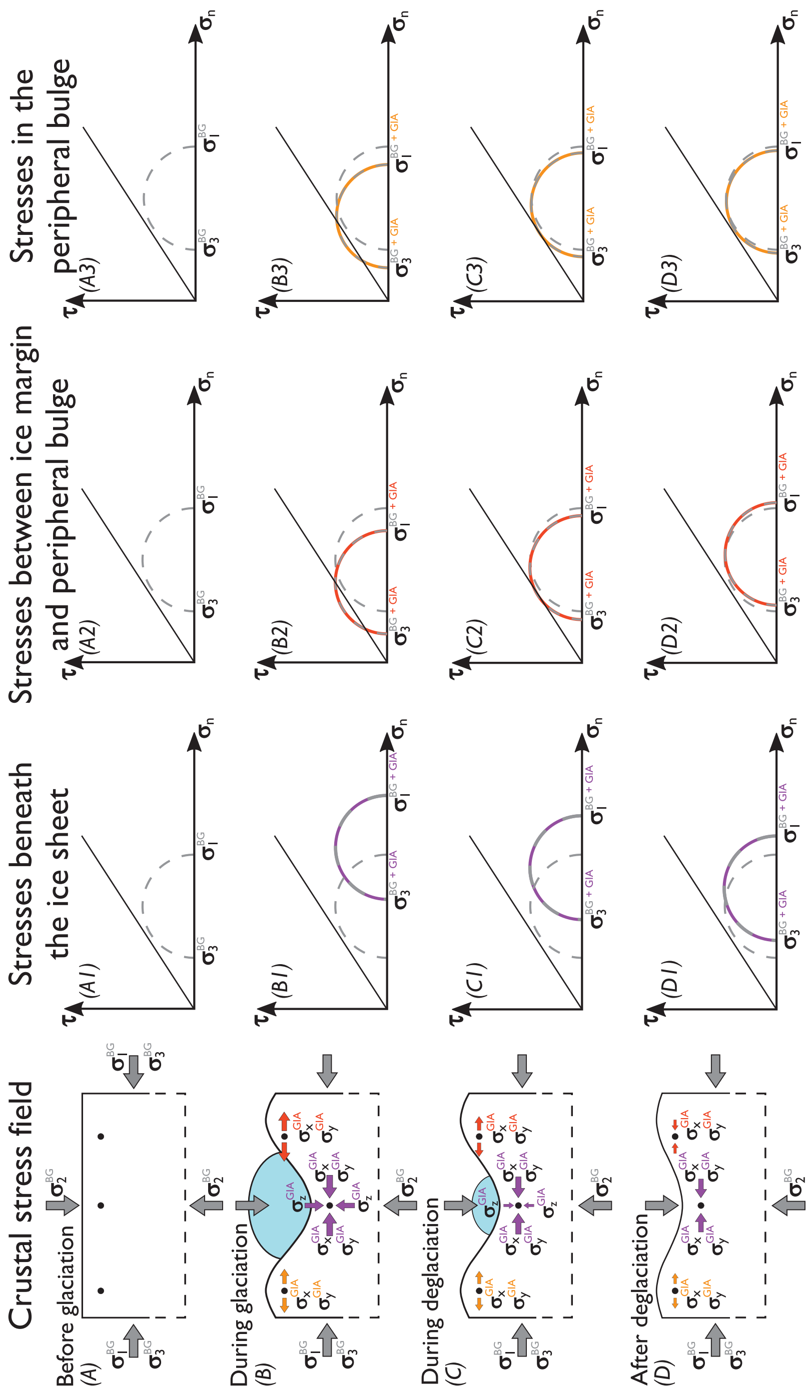

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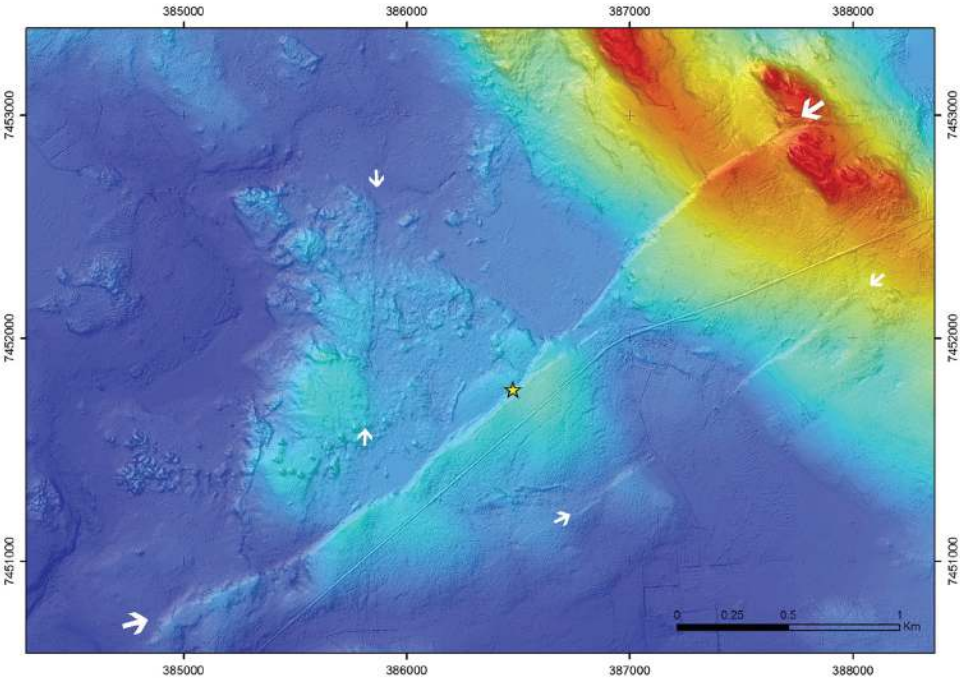

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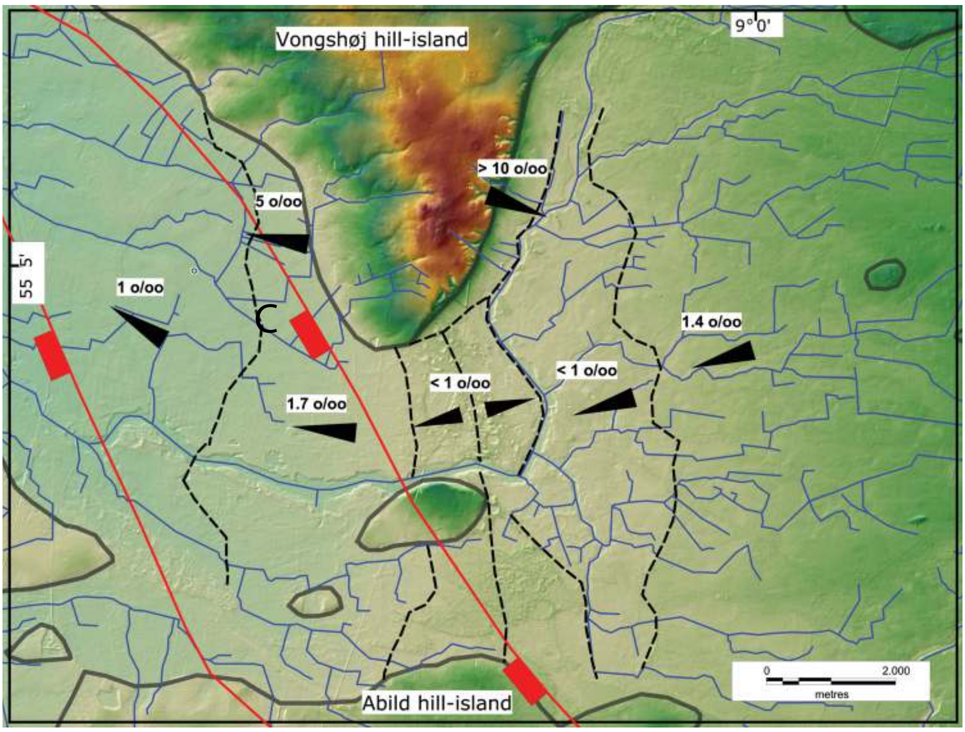

Figure 3.4

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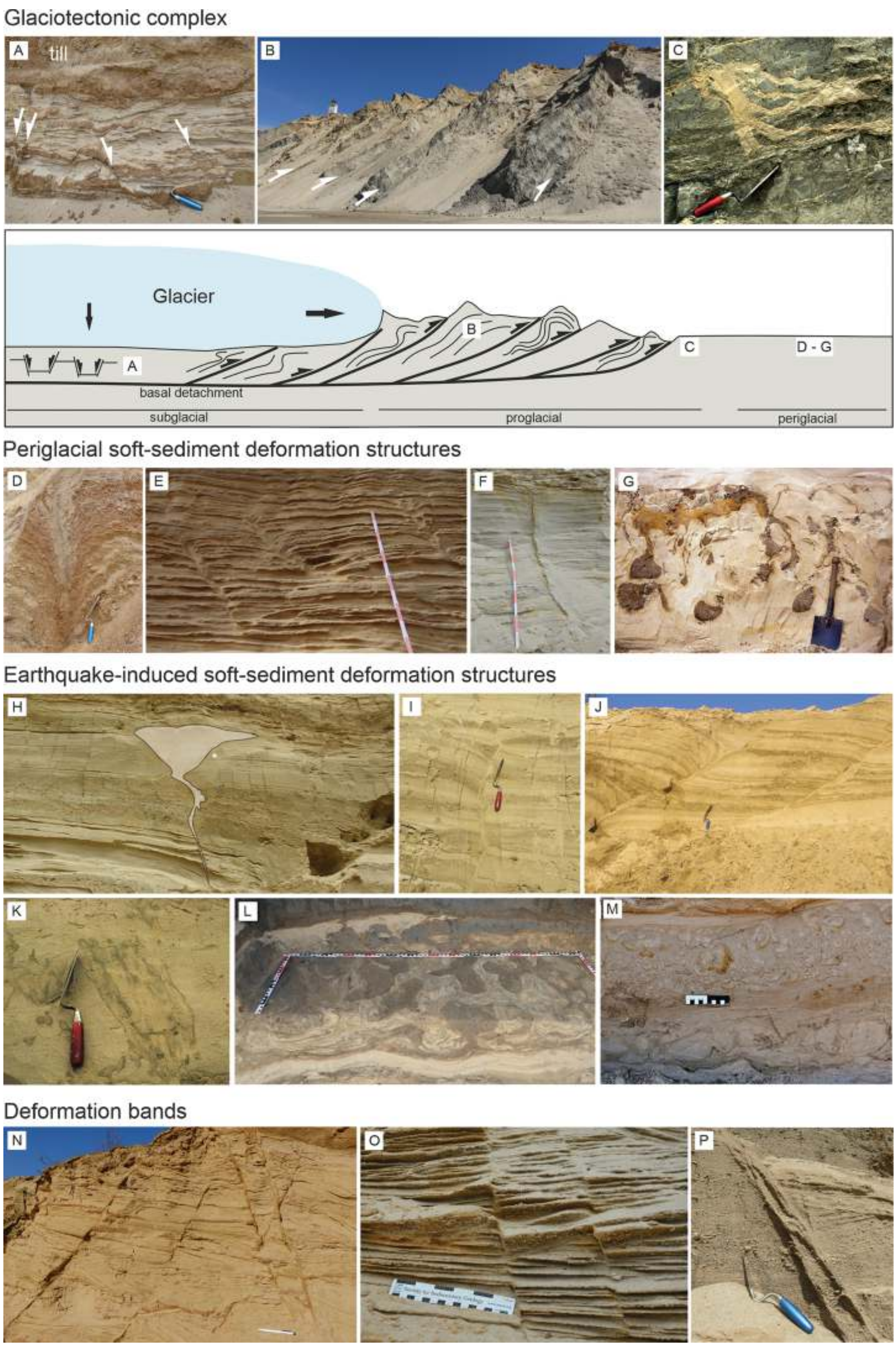

Figure 4.3

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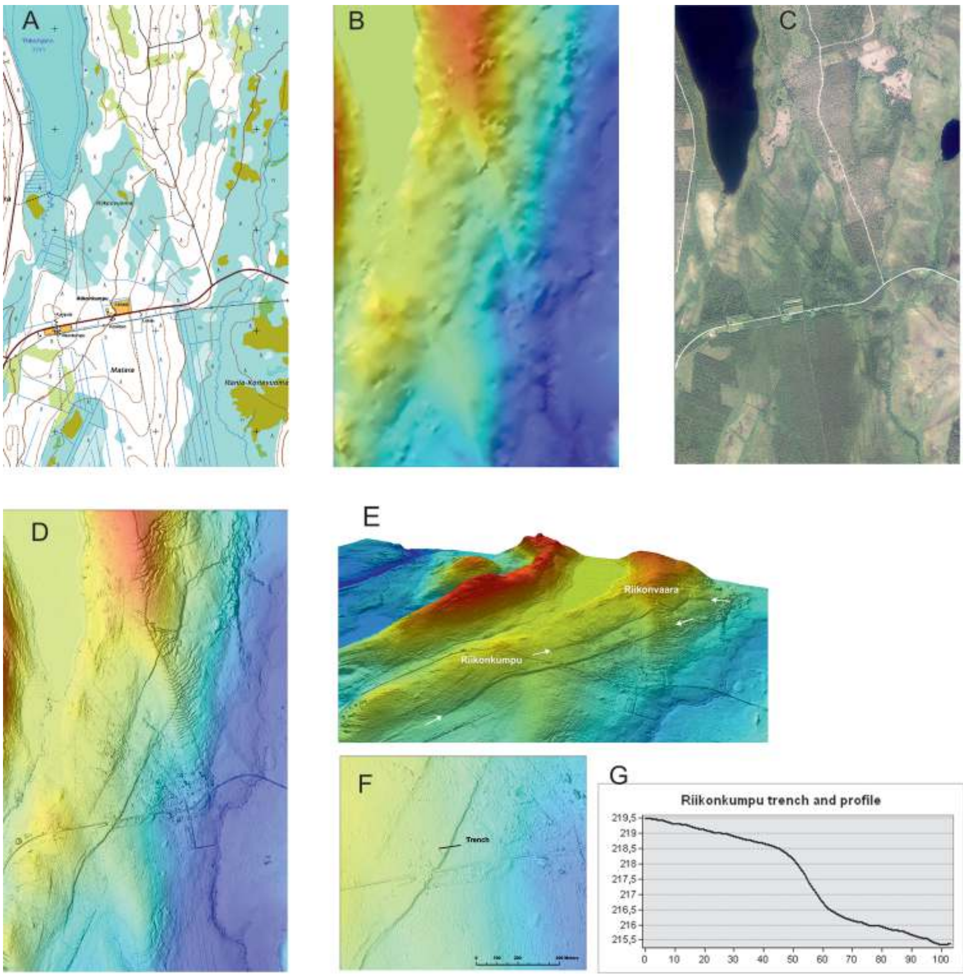

Figure 5.2

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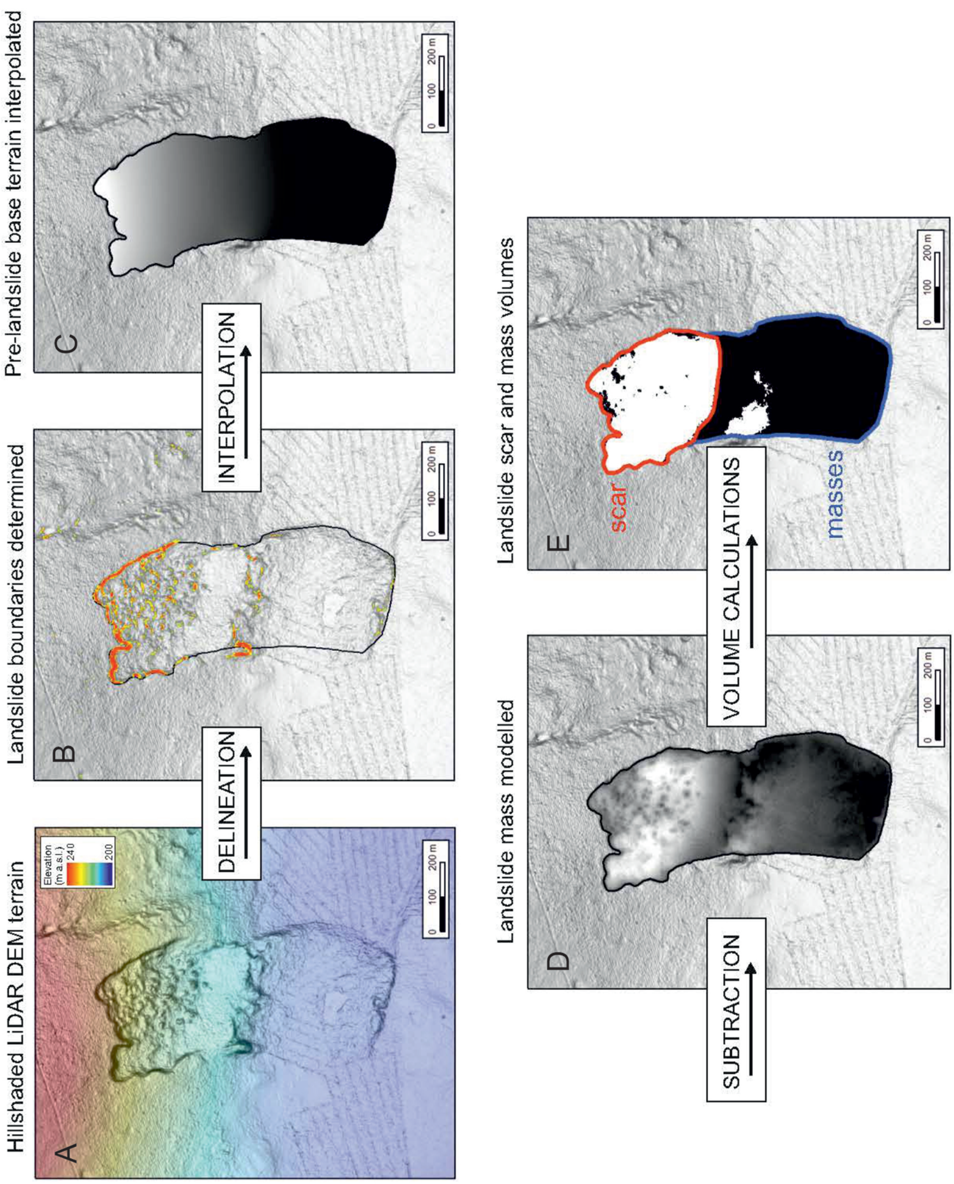

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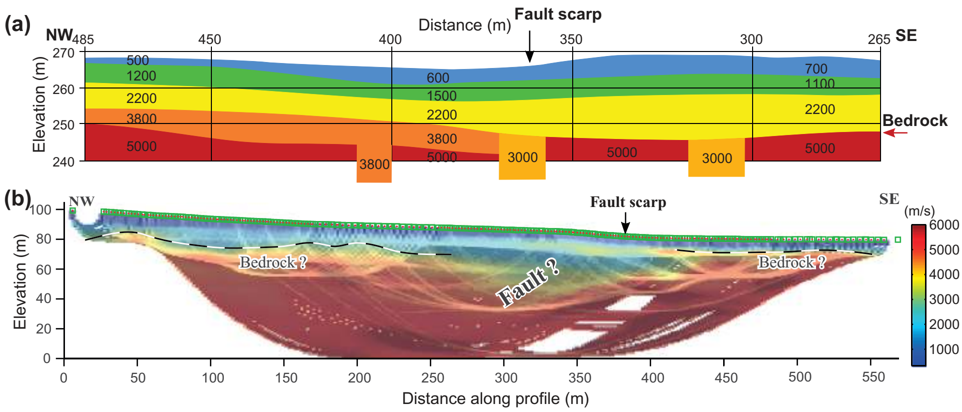

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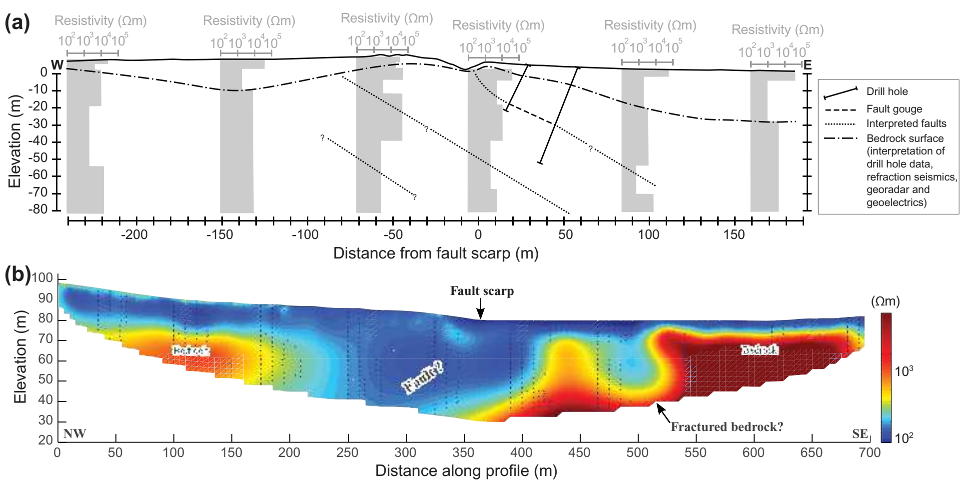

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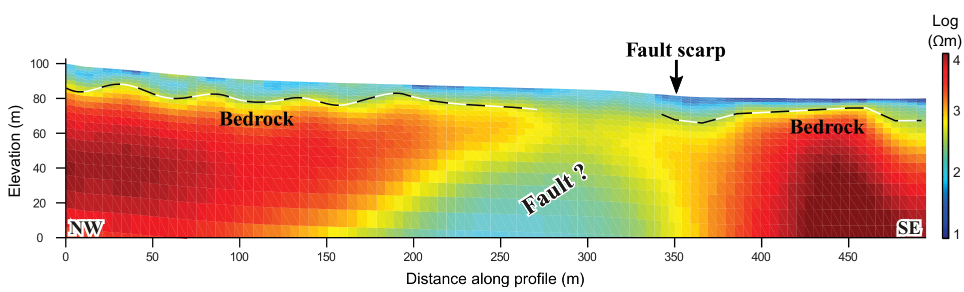

Figure 7.6

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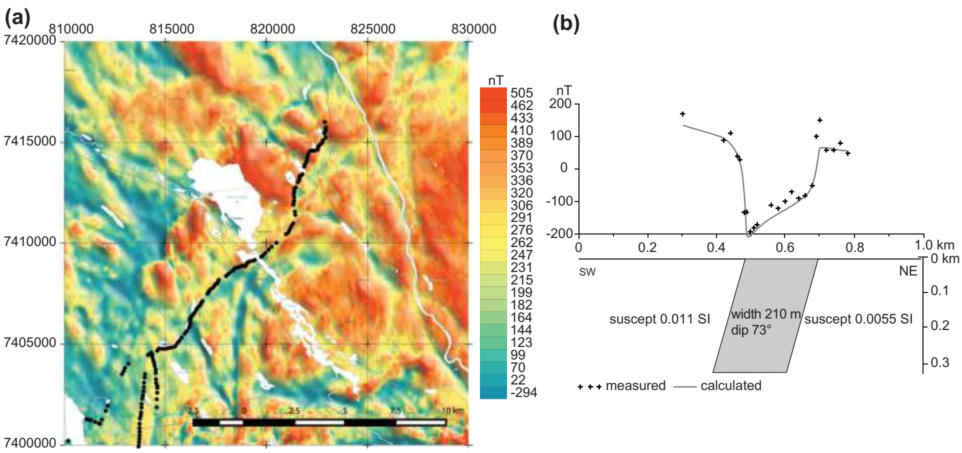

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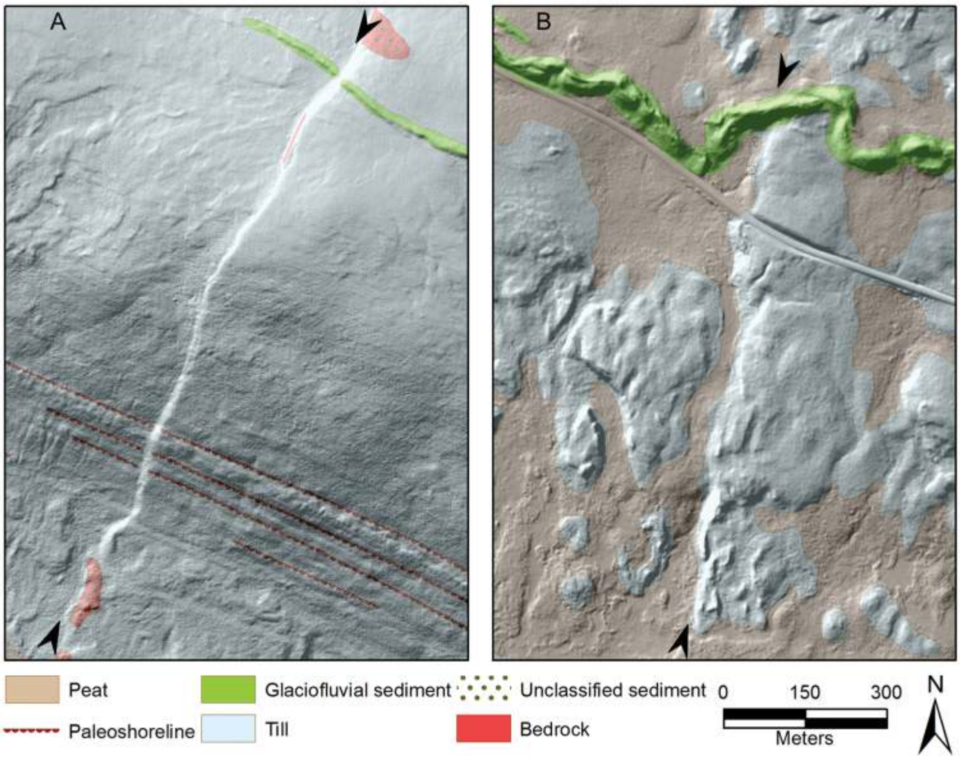

Figure 8.2

Søren Gregersen

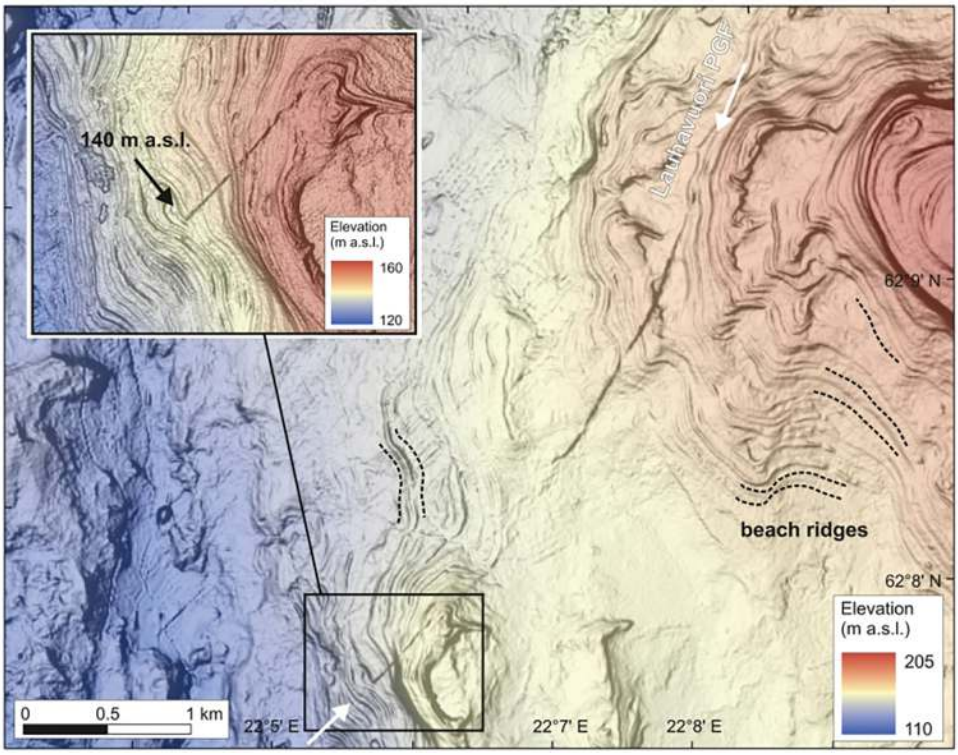
140 m a.s.l.
Elevation
(m a.s.l.)
160
120
Lauhavuori PGF
62°9' N
beach ridges
62°8' N
Elevation
(m a.s.l.)
205
110
0
0.5
1 km
22°5' E
22°7' E
22°8' E

Figure 8.4

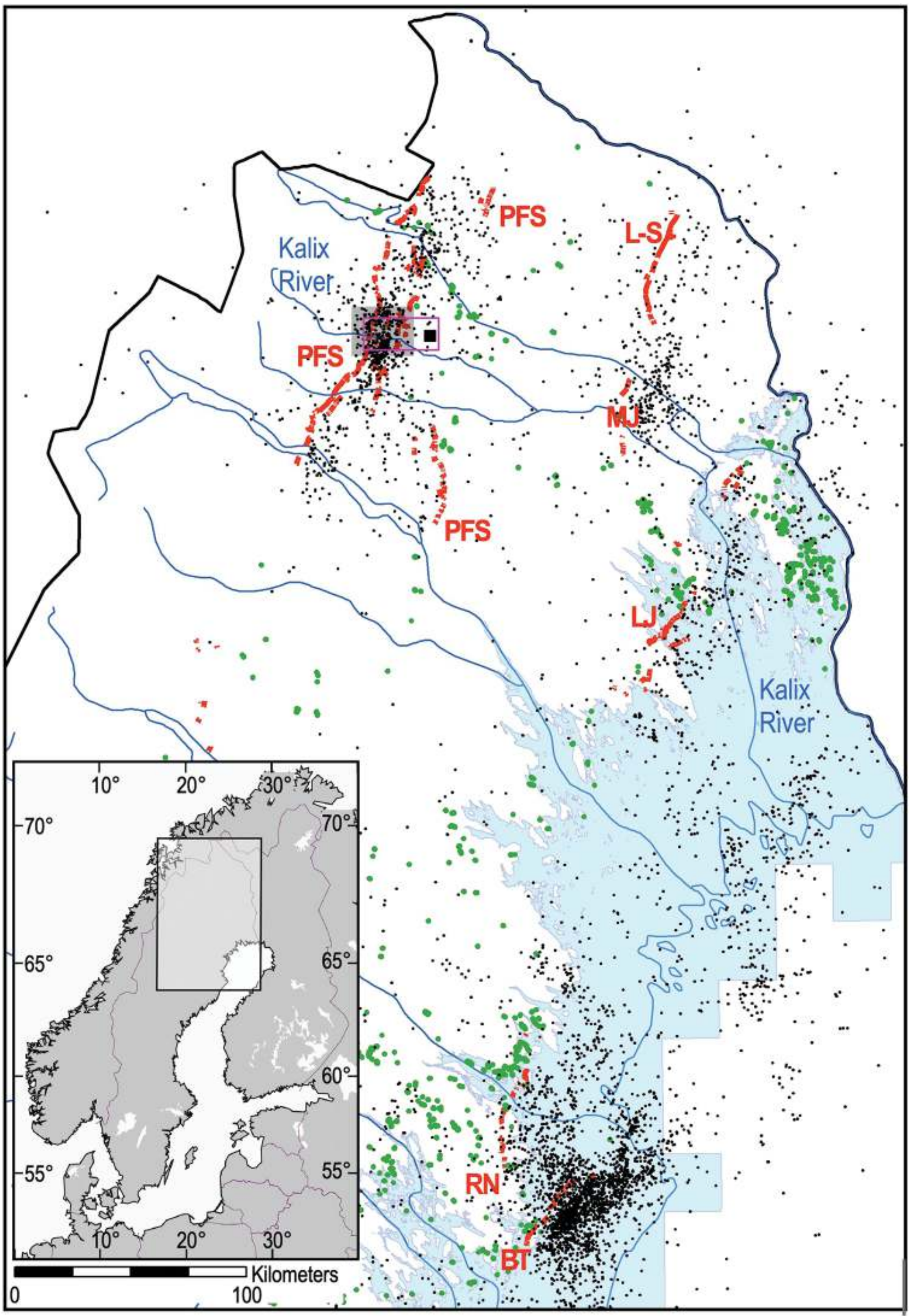

Figure 9.1

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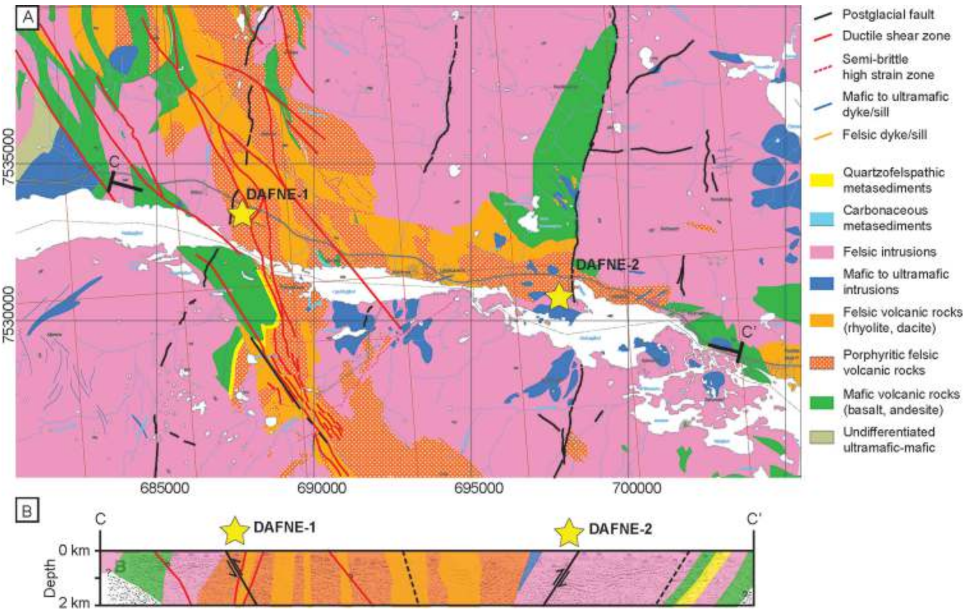

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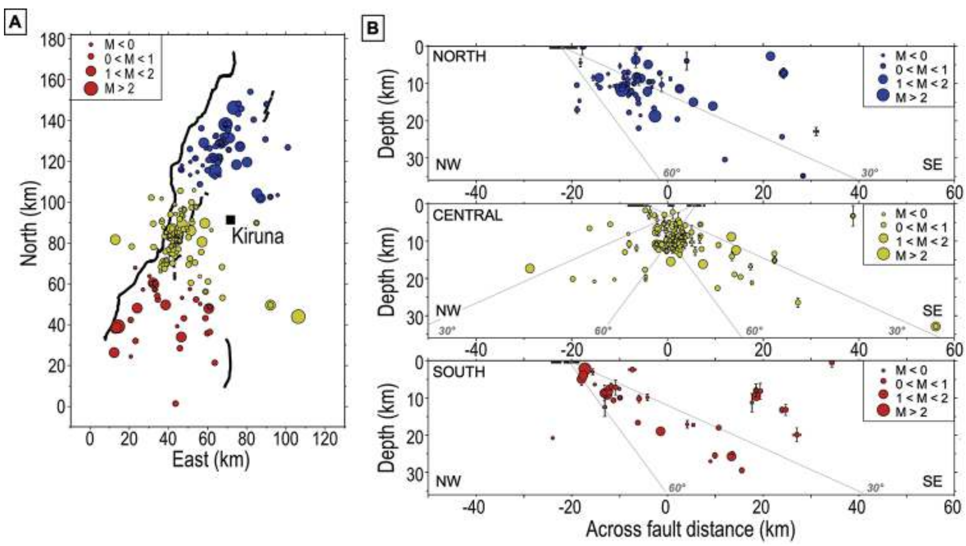

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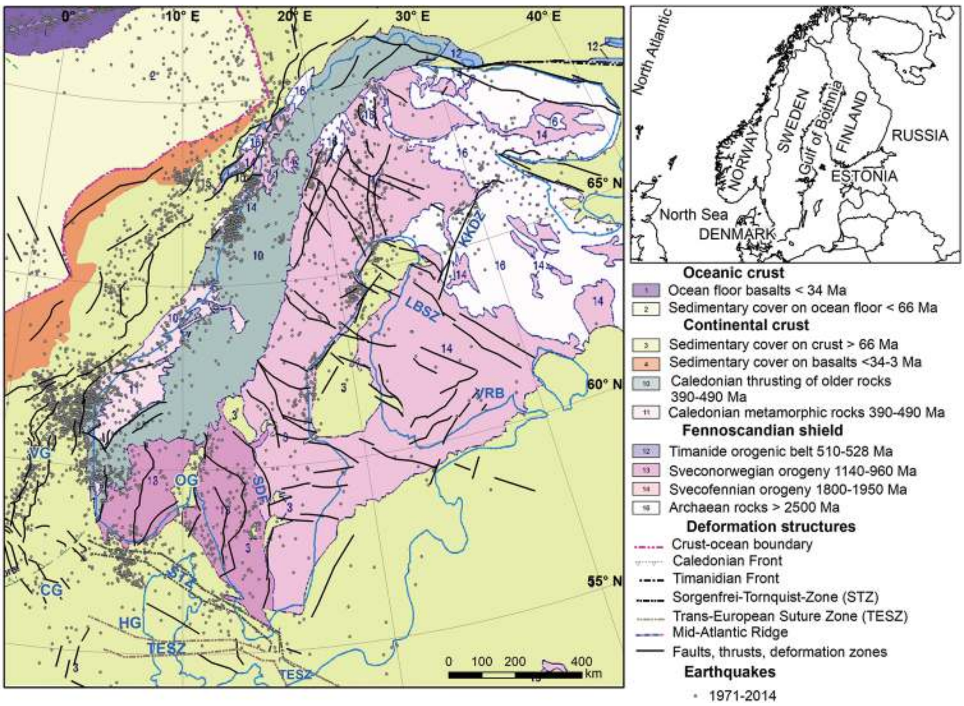

Figure 10.1

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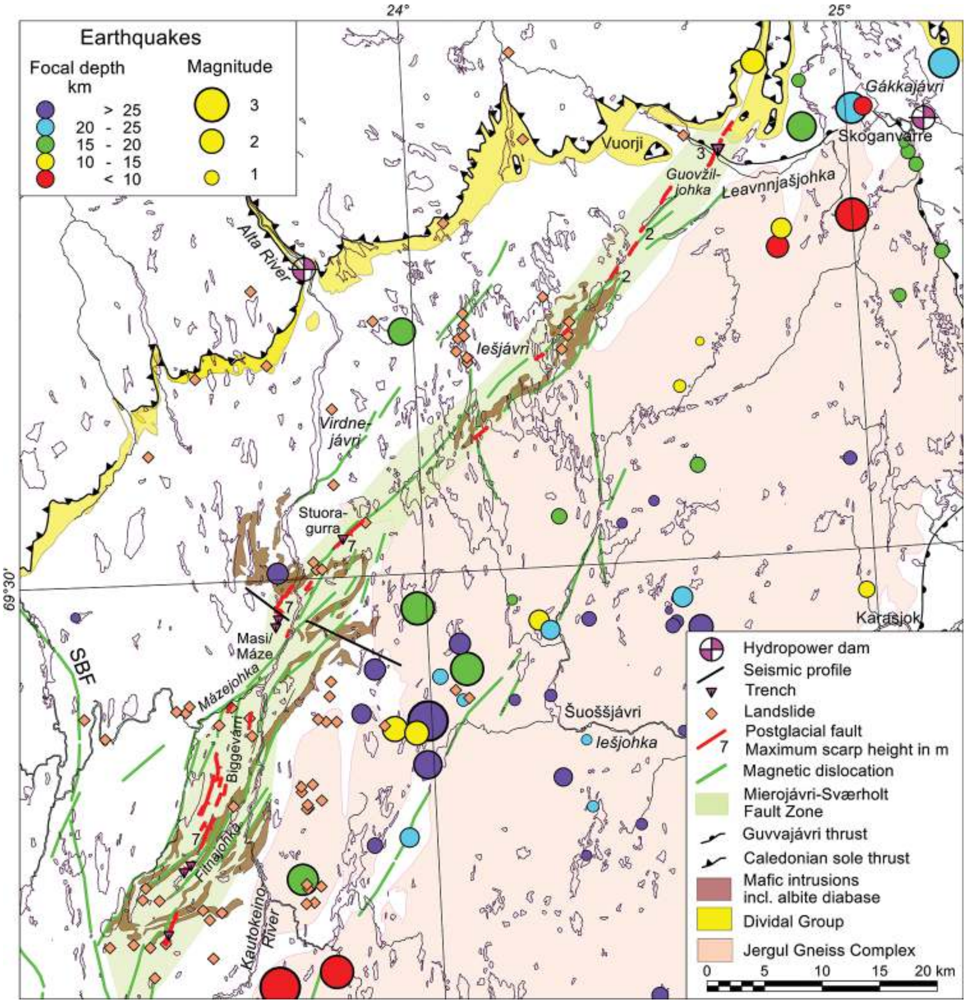

Figure 11.2

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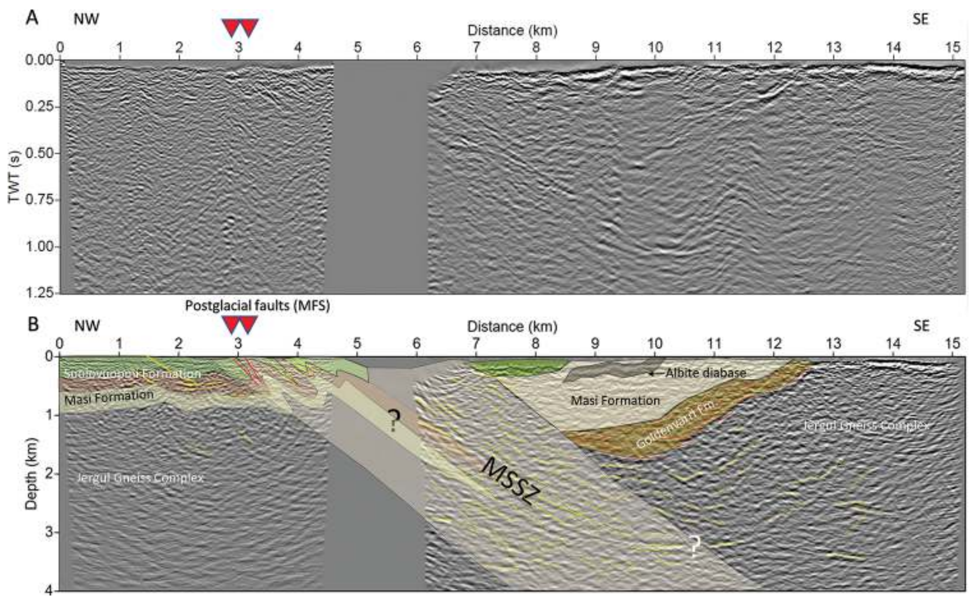

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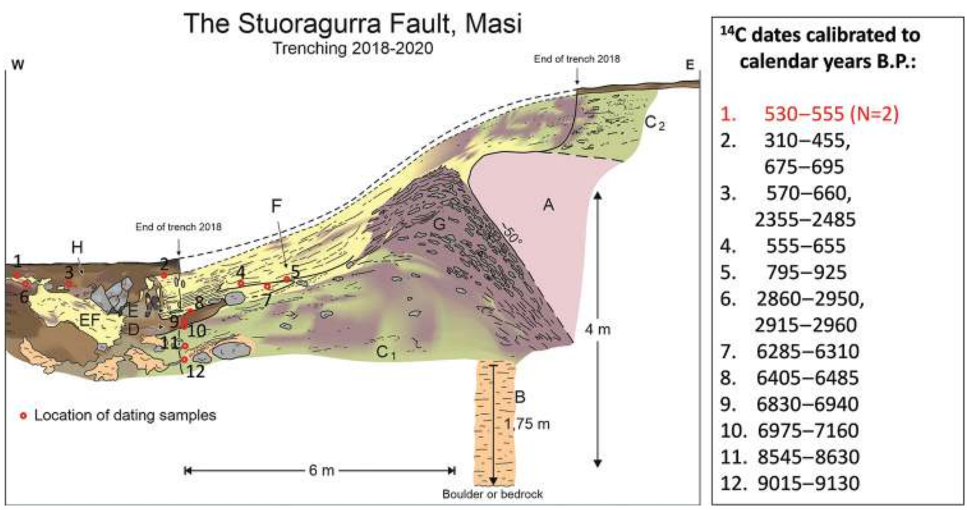

Figure 11.4

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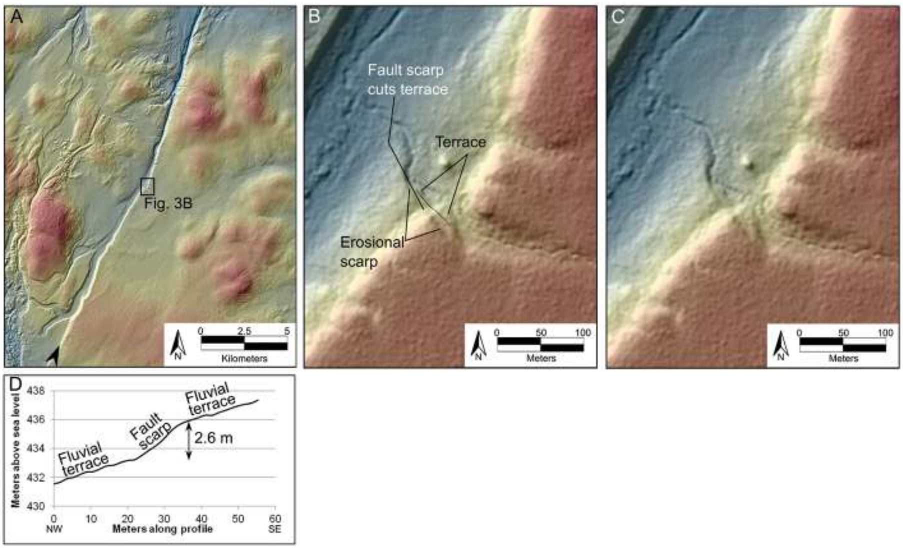

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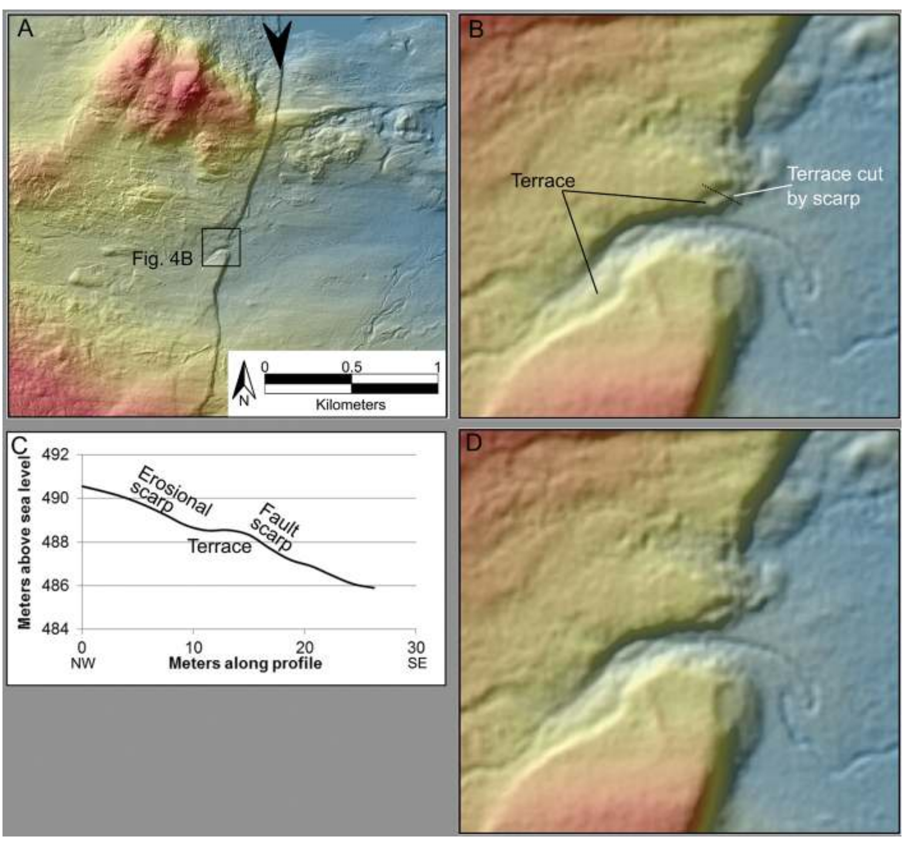

Figure 12.4

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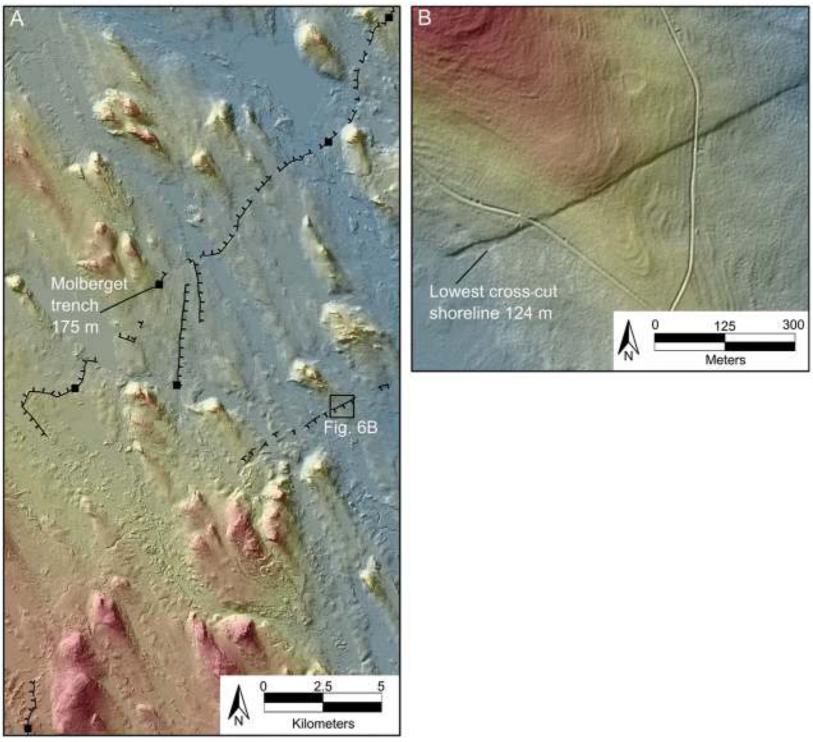

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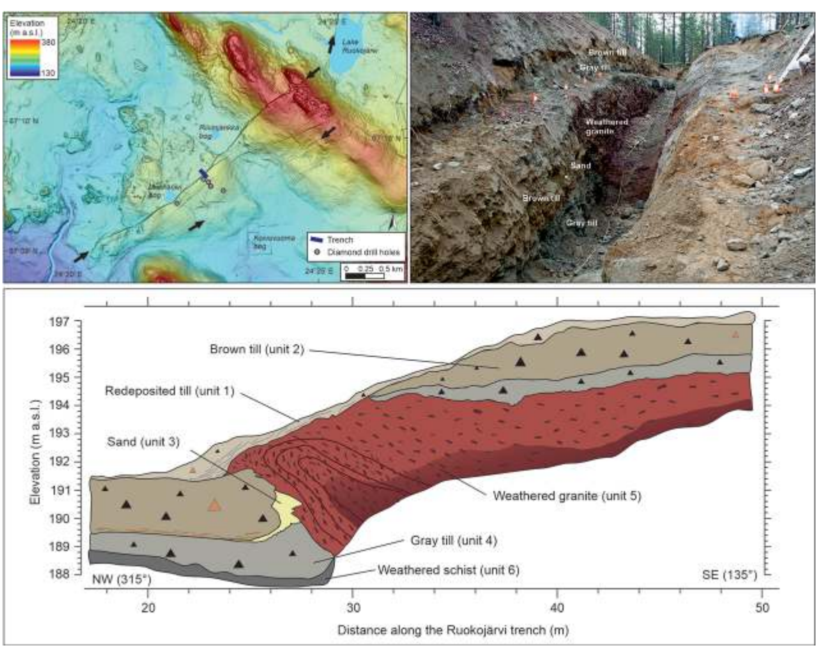

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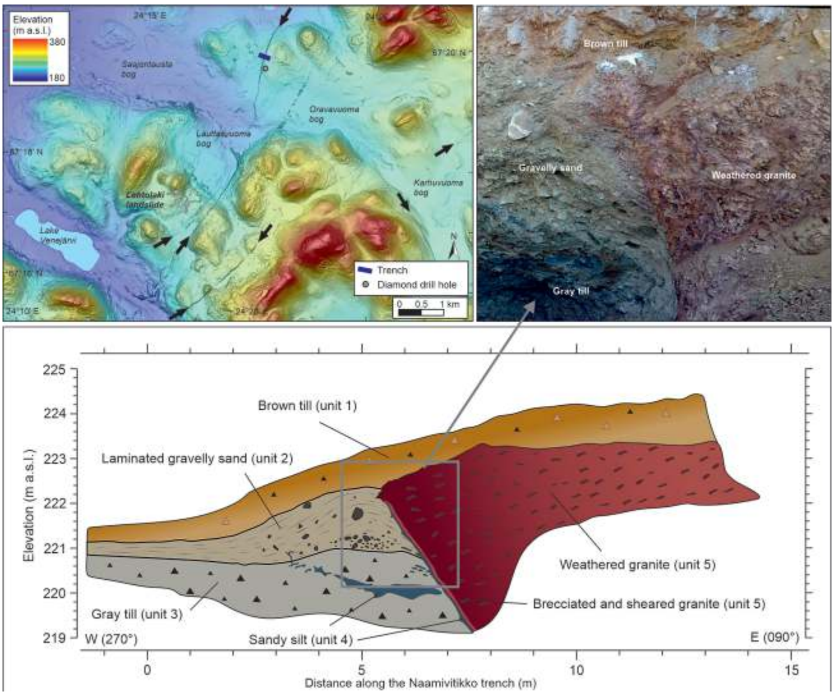

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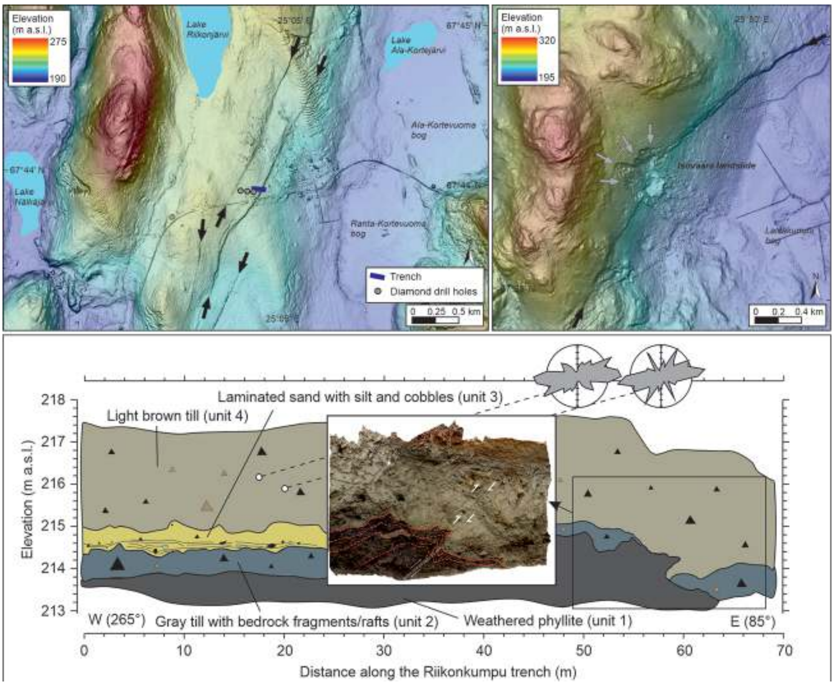
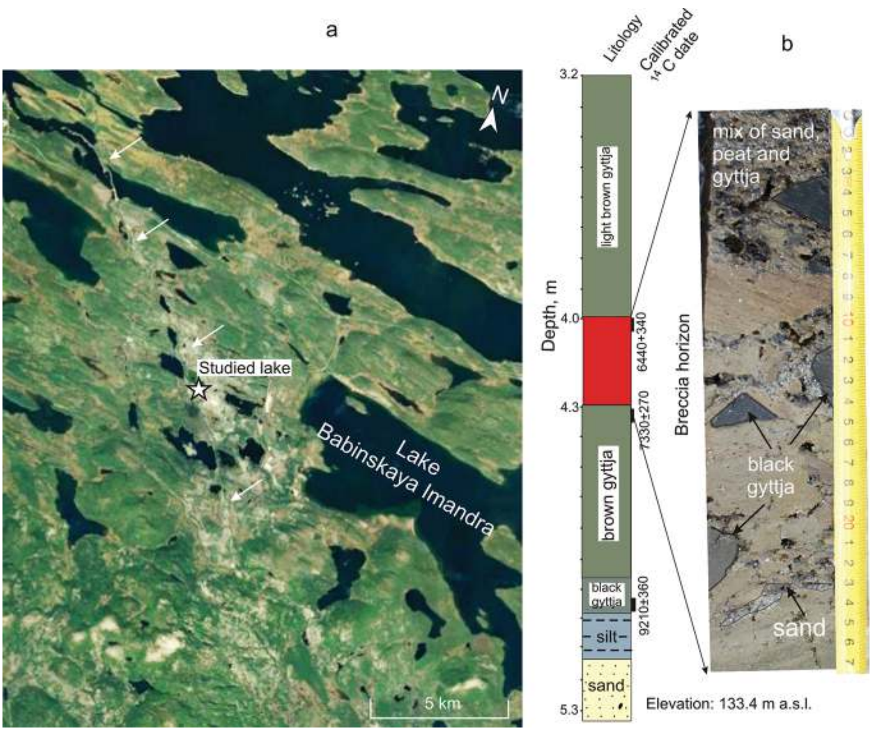


Figure 13.4

Figure 14.4



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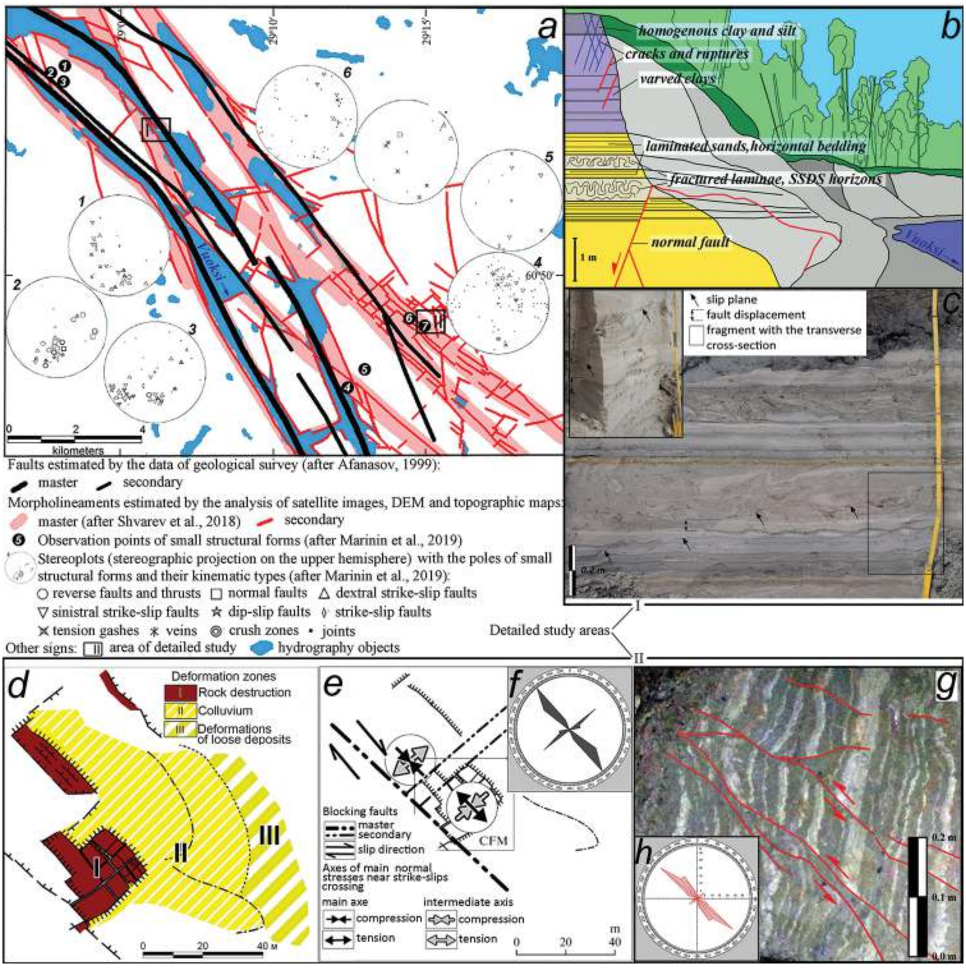
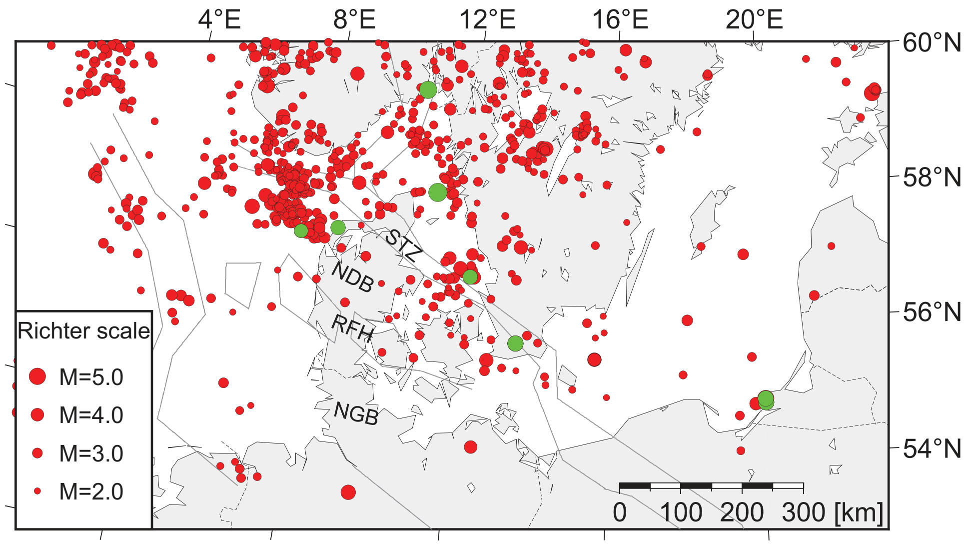

a
b
c
d
e
f
g
h
homogenous clay and silt
cracks and ruptures
varved clays
laminated sands, horizontal bedding
fractured laminae, SSDS horizons
normal fault
1 m
slip plane
fault displacement
fragment with the transverse cross-section
kilometers
Faults estimated by the data of geological survey (after Afanasov, 1999):
master
secondary
Morpholineaments estimated by the analysis of satellite images, DEM and topographic maps:
master (after Shvarev et al., 2018)
secondary
Observation points of small structural forms (after Marinin et al., 2019)
Stereoplots (stereographic projection on the upper hemisphere) with the poles of small structural forms and their kinematic types (after Marinin et al., 2019):
reverse faults and thrusts
normal faults
dextral strike-slip faults
sinistral strike-slip faults
dip-slip faults
strike-slip faults
tension gashes
veins
crush zones
joints
Other signs:
area of detailed study
hydrography objects
Detailed study areas
Deformation zones
Rock destruction
Colluvium
Deformations of loose deposits
Blocking faults
master
secondary
slip direction
Axes of main normal stresses near strike-slips crossing
main axe
compression
tension
intermediate axis
compression
tension
CFM

Figure 14.6



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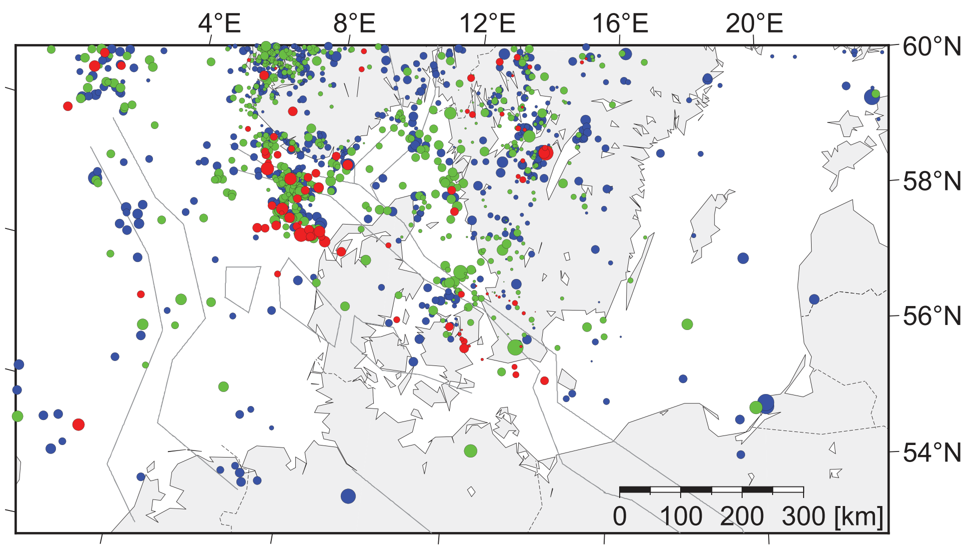

Figure 15.2

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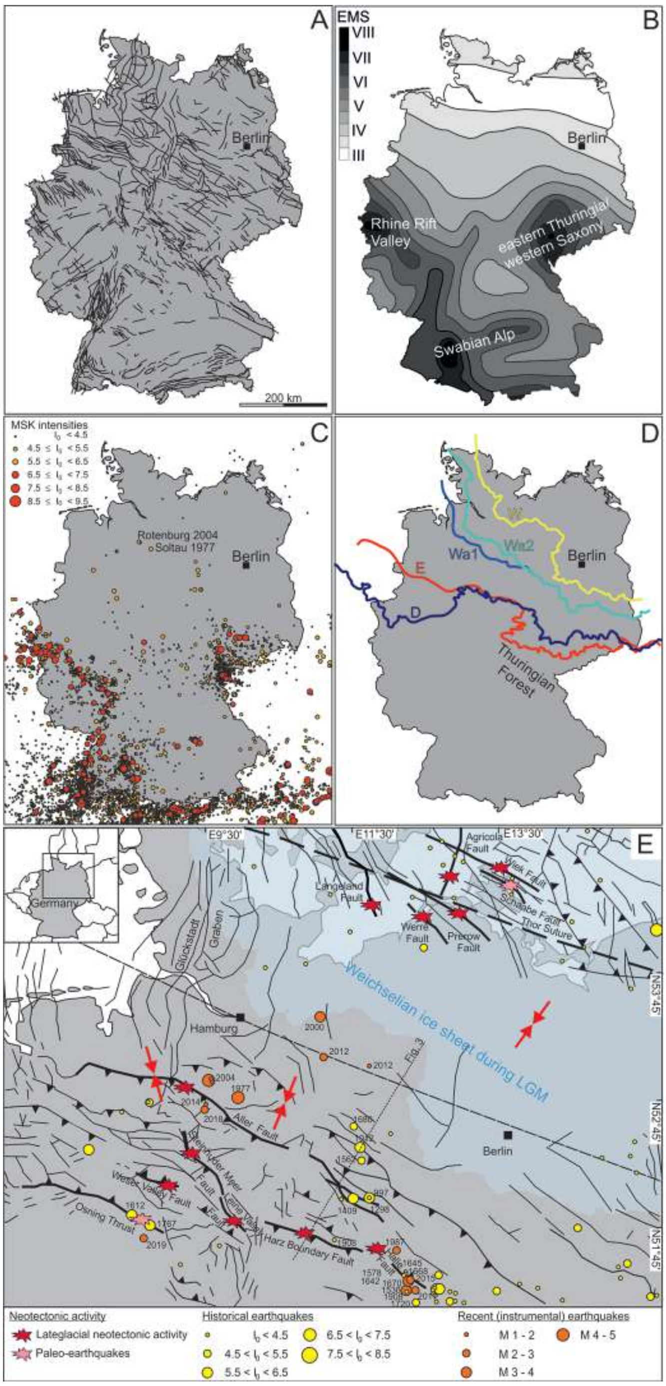

Figure 16.1

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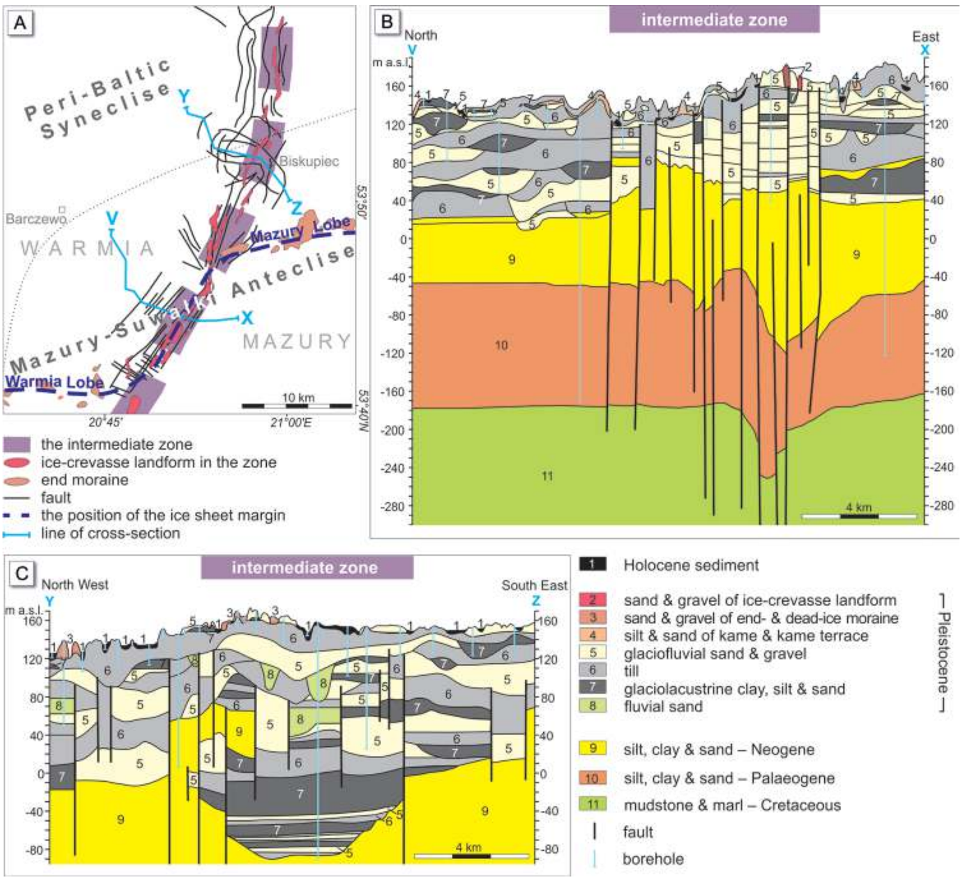

Figure 17.4

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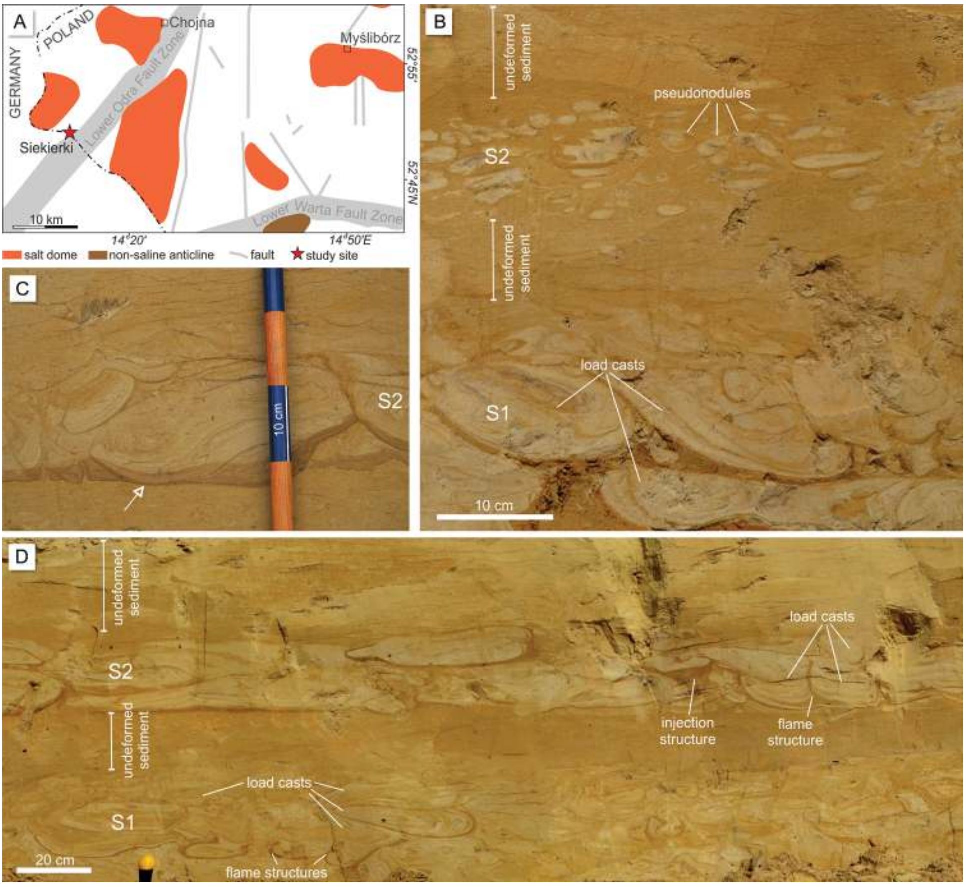

Figure 17.5

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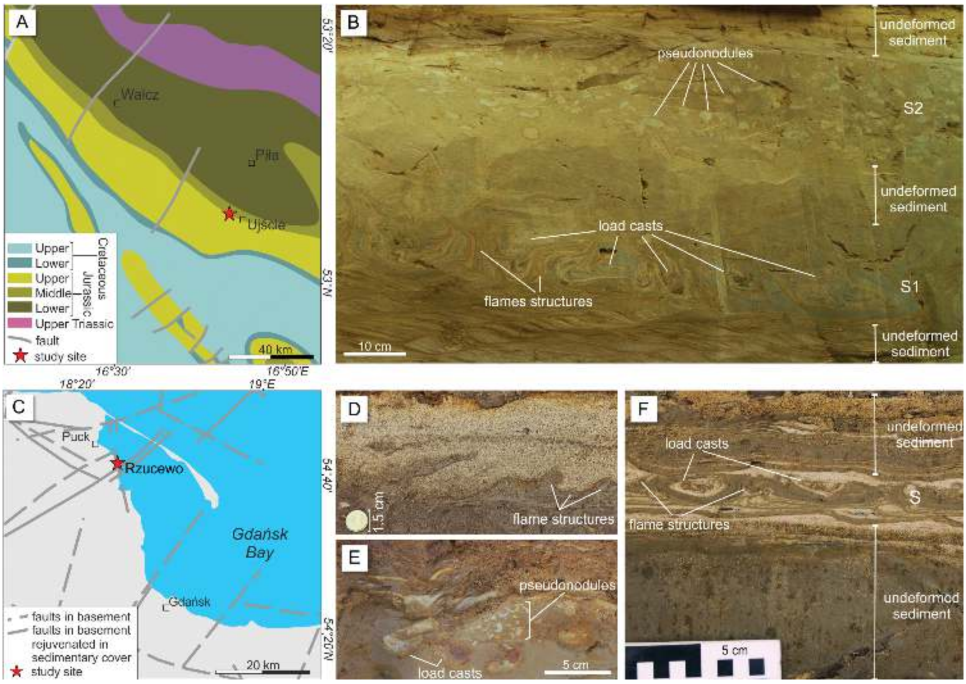

Figure 17.6

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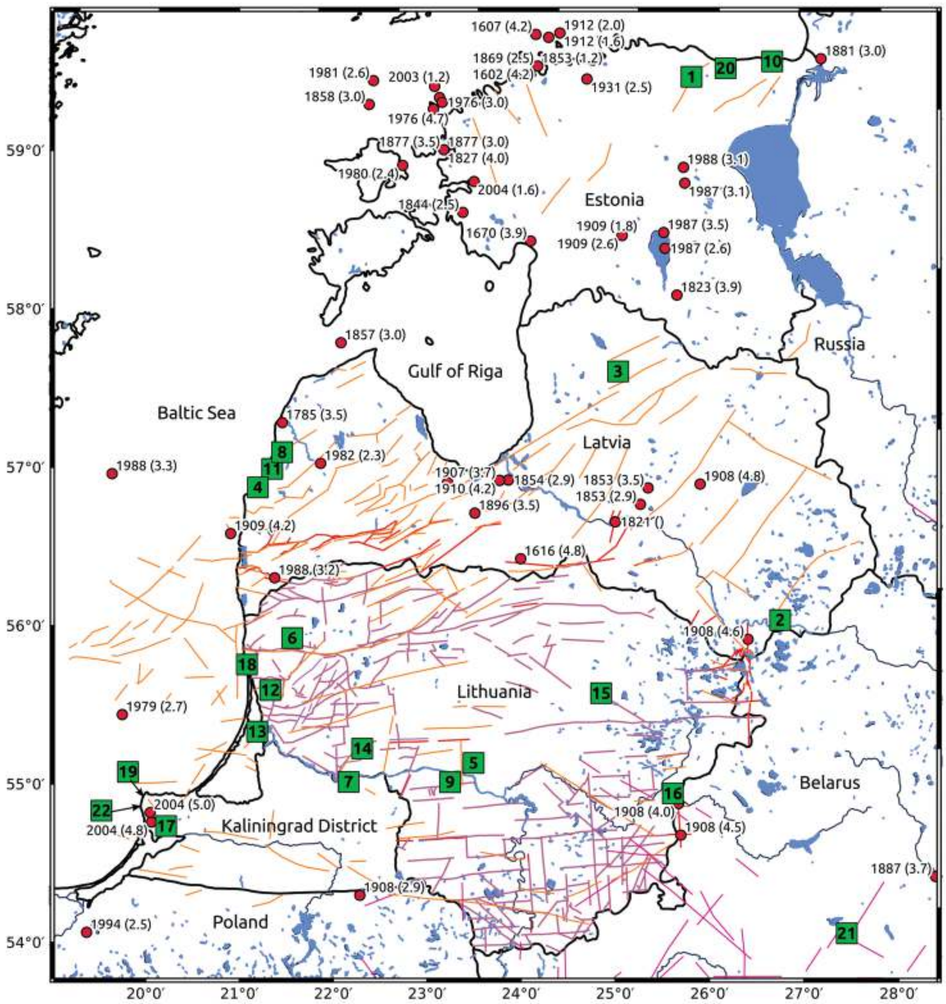

Figure 18.1

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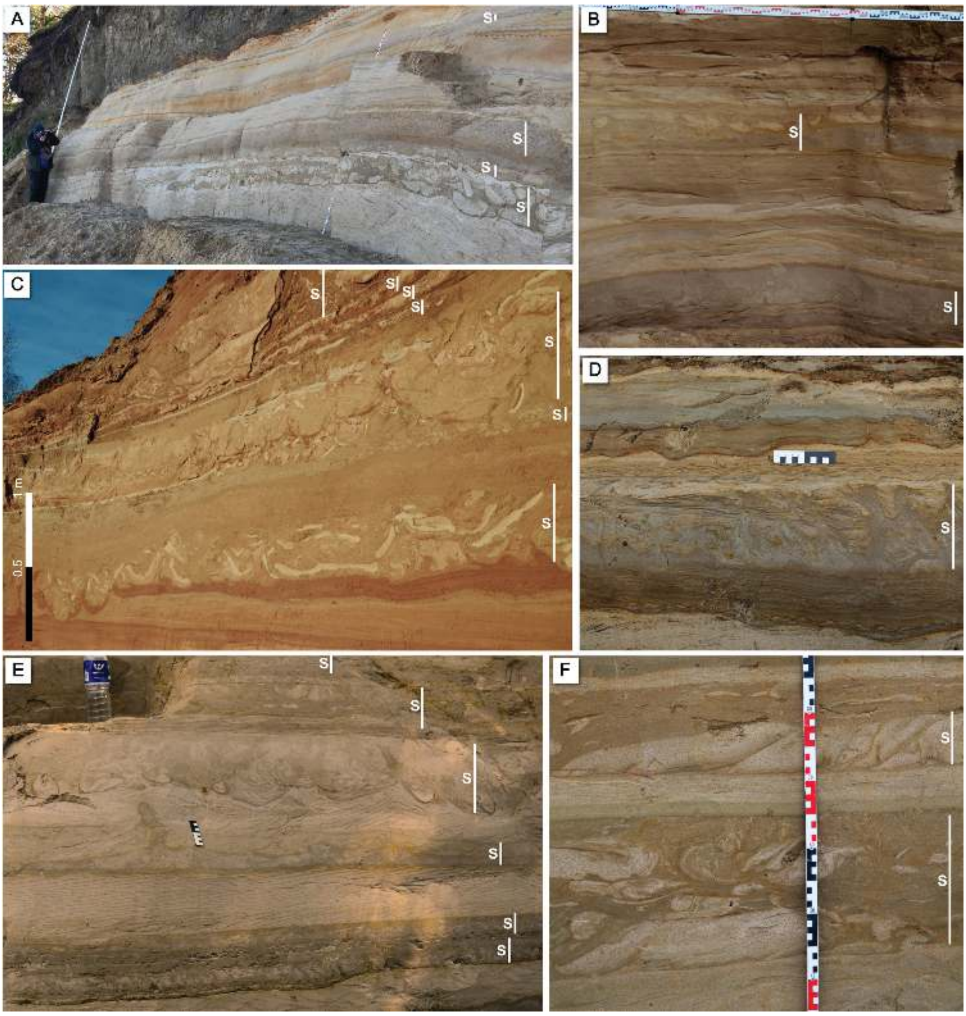

Figure 18.2

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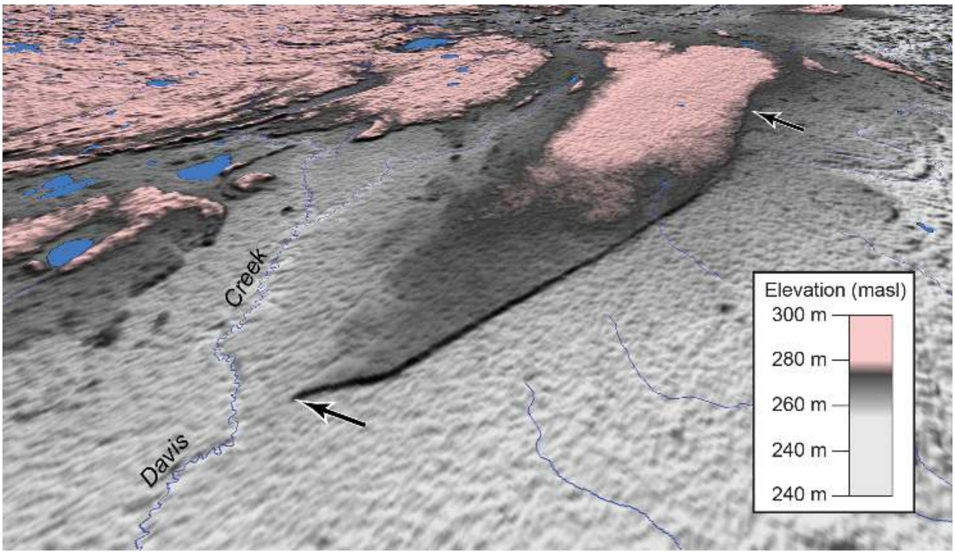

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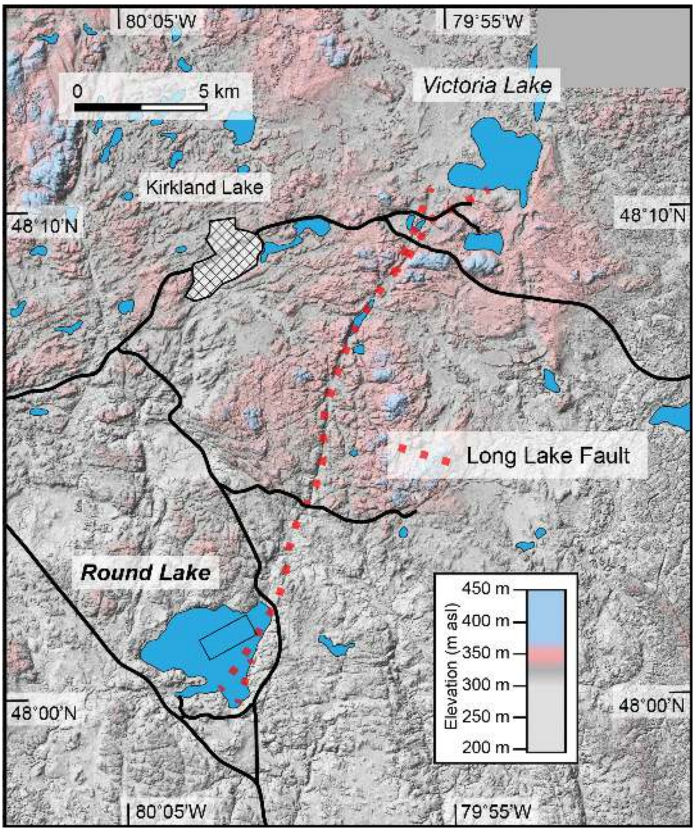

Figure 19.3

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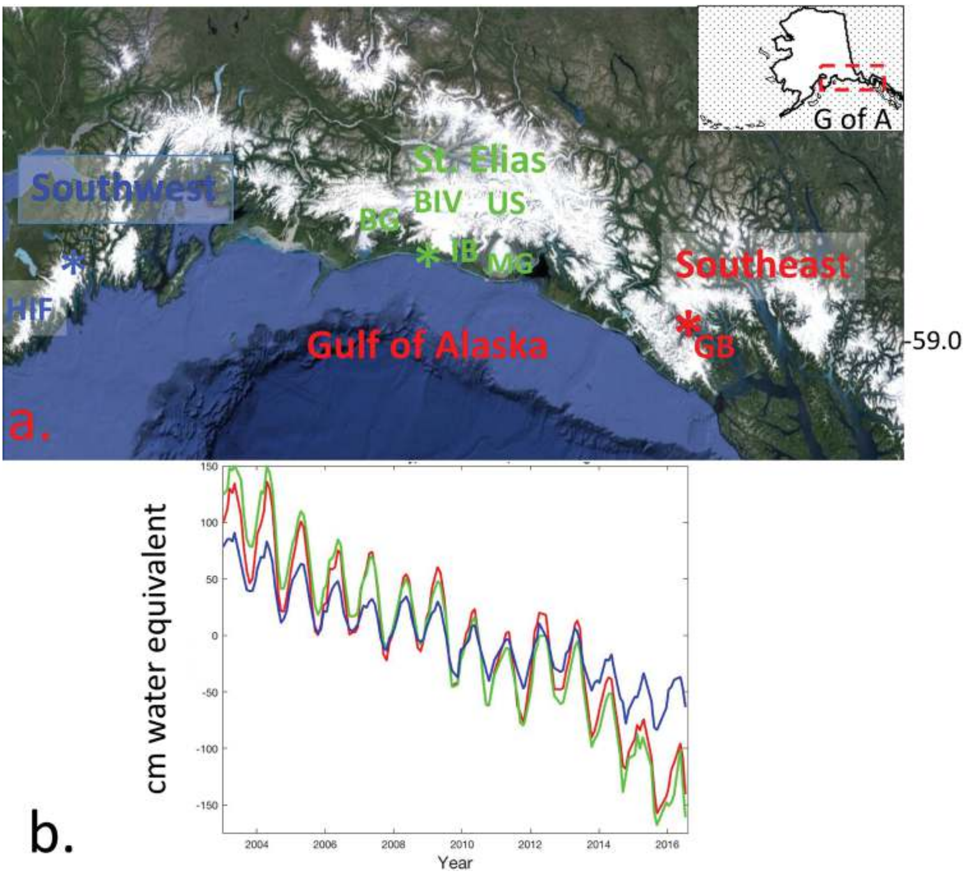

Figure 20.1

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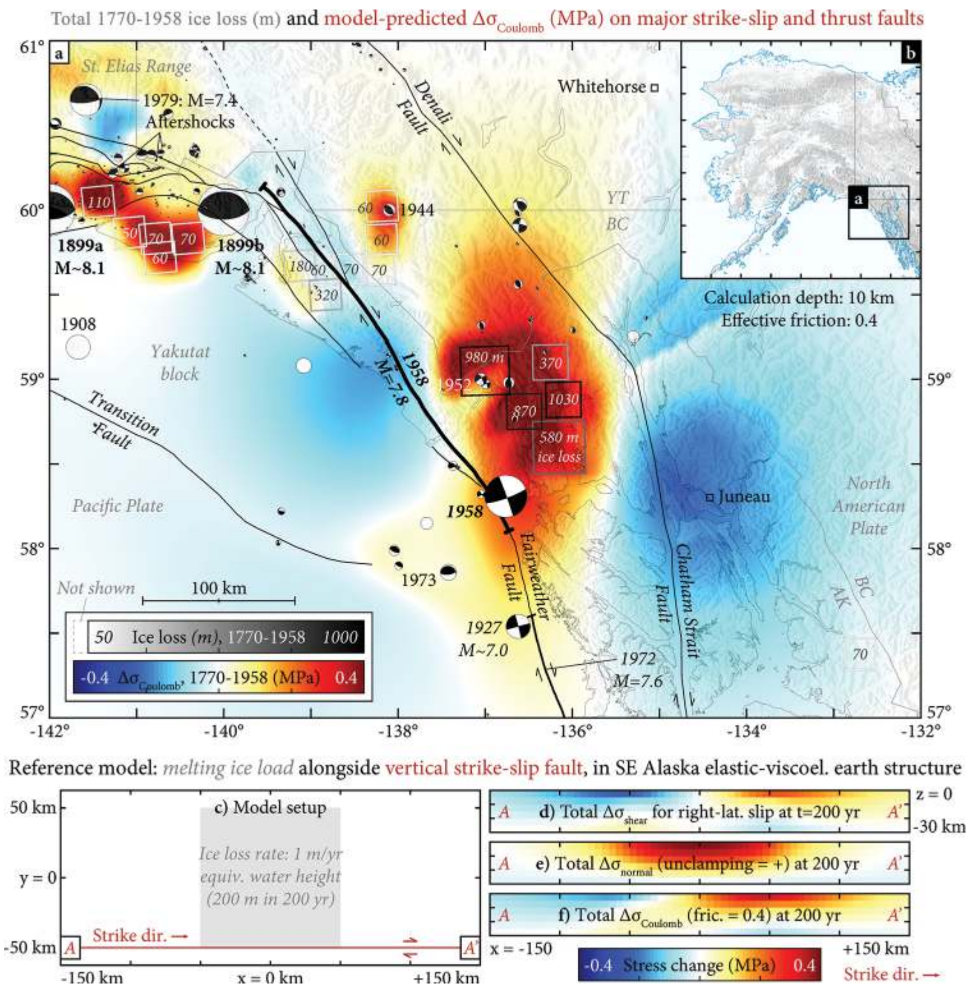

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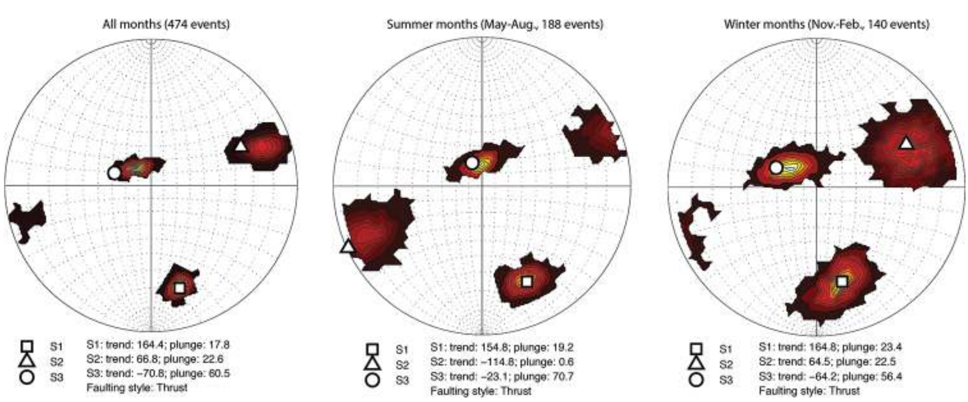

Figure 20.5

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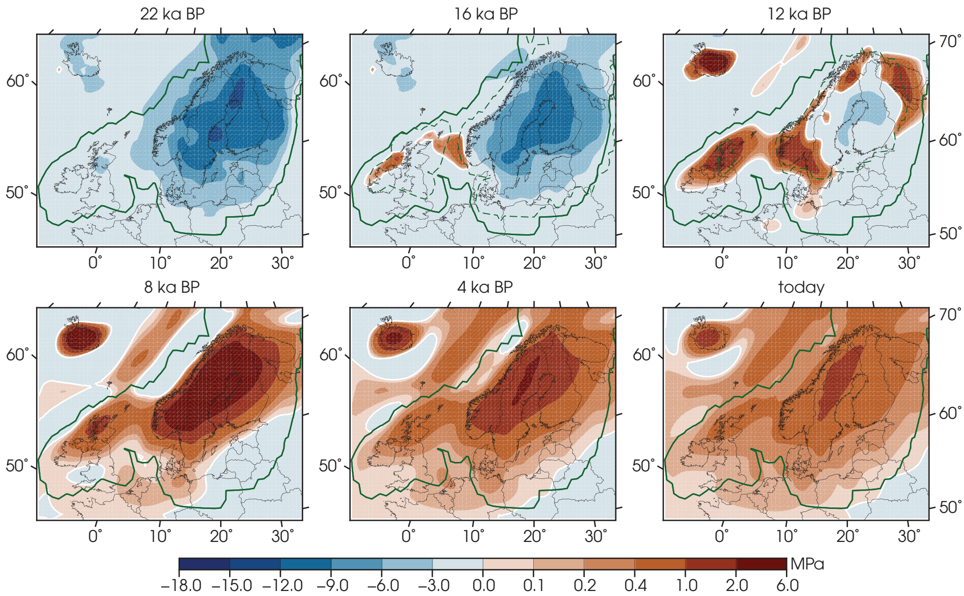

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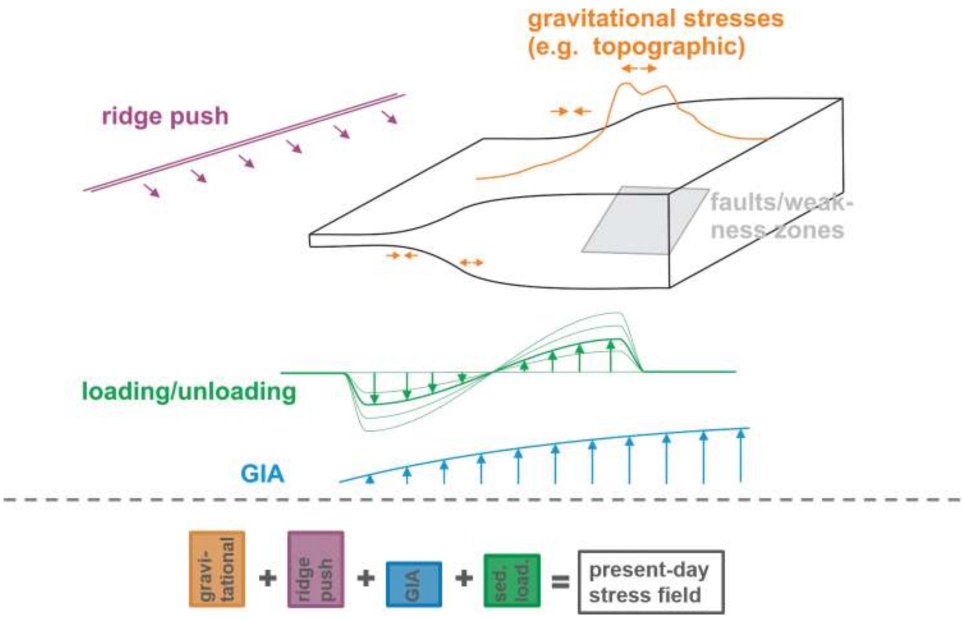

Figure 23.1

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