DynaQlim – Upper Mantle Dynamics and Quaternary Climate in Cratonic Areas

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Abstract The isostatic adjustment of the solid Earth to the glacial loading (GIA, Glacial Isostatic Adjustment) with its temporal signature offers a great opportunity to retrieve information of Earth's upper mantle to the changing mass of glaciers and ice sheets, which in turn is driven by variations in Quaternary climate. DynaQlim (Upper Mantle Dynamics and Quaternary Climate in Cratonic Areas) has its focus to study the relations between upper mantle dynamics, its composition and physical properties, temperature, rheology, and Quaternary climate. Its regional focus lies on the cratonic areas of northern Canada and Scandinavia.

Geodetic methods like repeated precise levelling, tide gauges, high-resolution observations of recent movements, gravity change and monitoring of postglacial faults have given information on the GIA process for more than 100 years. They are accompanied by more recent techniques like GPS observations and the GRACE and GOCE satellite missions which provide additional global and regional constraints on the gravity field. Combining geodetic observations with seismological investigations, studies of the postglacial faults and continuum mechanical modelling of GIA, DynaQlim offers new insights into properties of the lithosphere. Another step toward a better understanding of GIA has been the joint inversion of different types of observational data – preferentially connected with geological relative sea-level evidence of the Earth's rebound during the last 10,000 years.

Due to the changes in the lithospheric stress state large faults ruptured violently at the end of the last glaciation in large earthquakes, up to the magnitudes $M_W = 7–8$. Whether the rebound stress is still able to trigger a significant fraction of intraplate seismic events in these regions is not completely understood due to the complexity and spatial heterogeneity of the regional stress field. Understanding of this mechanism is of societal importance.

Glacial ice sheet dynamics are constrained by the coupled process of the deformation of the viscoelastic solid Earth, the ocean and climate variability. Exactly how the climate and oceans reorganize to sustain growth of ice sheets that ground to continents and shallow continental shelves is poorly understood. Incorporation of nonlinear feedback in modelling both ocean heat transport systems and atmospheric CO_2 is a major challenge. Climate-related loading cycles and episodes are expected to be important, hence also more short-term features of palaeoclimate should be explicitly treated.

Keywords GIA · Crustal deformation · Mantle dynamics · Quaternary climate

Introduction

The process of GIA with its characteristic temporal signatures is one of the great opportunities in geosciences to retrieve information about the Earth. It contains information about recent climate forcing, being dependent on the geologically recent on- and off-loading of ice sheets. It gives a unique chance to study the dynamics and rheology of the lithosphere and asthenosphere, and it is of fundamental importance in geodesy, since Earth rotation, polar motion and crustal

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S. Cloetingh, J. Negendank (eds.), *New Frontiers in Integrated Solid Earth Sciences*, International Year of Planet Earth, DOI 10.1007/978-90-481-2737-5_10,

deformation, and therefore the global reference frames are influenced by it.

Despite the existence of long and accurate time series and extensive data sets on GIA, there still exist many open questions related to upper mantle dynamics and composition, rebound mechanisms and uplift models, including the role of tectonic forces as well as ice thickness during the late Quaternary. DynaQlim aims to integrate existing data and models on GIA processes, including both geological and geodetic observations. The themes of DynaQlim include Quaternary climate and glaciation history, postglacial uplift and contemporary movements, ice-sheets dynamics and glaciology, postglacial faulting, rock rheology, mantle xenoliths, past and present thermal regime of the lithosphere, seismic structure of the lithosphere, and gravity field modelling.

DynaQlim will probably lead to a more comprehensive understanding of the Earth's response to glaciations, improved modelling of crustal and upper mantle dynamics as rheology structure. An important aspect is to construct and improve coupled models of glaciation and land-uplift history and their connection to the climate evolution on the time scale of glacial cycles.

Observational Basis

During the Pleistocene, quasi-periodic variations between glacial and interglacial intervals prevailed, with dominant periods closely related to those present in the Earth-Sun orbit and 25.8 kyr rotational precession of the Earth (Berger, 1984). These Milankovitch variations have played a key role in shaping the landscape and driving the geodynamic evolution of cratonic regions such as northern Eurasia and North America during the Quaternary.

Extensive and diverse sets of observations can be applied to study and understand the key processes involved, including geodetic land uplift measurements, geological observations of past sea-level changes, lateglacial faults, terminal moraines and other glacial deposits as well as various palaeoclimatological proxies. These observations have played a vital role in a number of recent studies that have improved our understanding of the structure and dynamics of cratonic regions and the influence of ice sheet variations.

Abundant data have been collected in various cratonic regions, including Antarctica, Laurentia and Fennoscandia. Laurentia and Fennoscandia have a similar glaciation history during the Quaternary, though their tectonic evolutions are different. In Antarctica the glaciation history is distinctly different. DynaQlim will collect and compile observational evidence predominantly from geodetic and geophysical methods.

Geodetic Observations

Geodetic methods provide accurate measurements of contemporary deformation and gravity change. There are systematic postglacial uplift observations for the last 100 years based on repeated precise levelling, tide gauges, geodetic high-resolution observations of recent movements, gravity change and monitoring of postglacial faults. Until recently, horizontal motions could not be observed accurately. However, current GNSS (Global Navigation Satellite Systems, including GPS) observations are accurate enough to observe even minor horizontal motions over distances of several hundreds of kilometres.

Maps of vertical motion have traditionally been based on long time series of tide gauges and repeated precise levellings over several decades. Tide gauge time series reflect both vertical motions of the land and variations of the surface of the sea. Maps of relative sea level change for Fennoscandia were published by e.g. Ekman (1996), Kakkuri (1997), Mäkinen and Saaranen (1998) and Saaranen and Mäkinen (2002). The latest uplift models, based on repeated precise levelling, tide gauge time series and geophysical modelling have been published by Vestøl (2006), and Ågren and Svensson (2007), Fig. 1.

In North America repeated levellings of the rebound area are confined to regions near Hudson Bay (Sella et al., 2007) or other coastal areas. Overall, levelling data are much more scattered than in Fennoscandia.

Space geodetic techniques, such as GNSS, allow the construction of 3-D motions from relatively short (less than 10 years) time series. The project BIFROST (Baseline Inferences for Fennoscandian Rebound Observations, Sea Level, and Tectonics) was initiated in 1993 taking advantage of tens of permanent GPS stations separated by a few hundreds of km both in Finland and Sweden. Results are discussed e.g. in Milne et al. (2001), Johansson et al. (2002), and

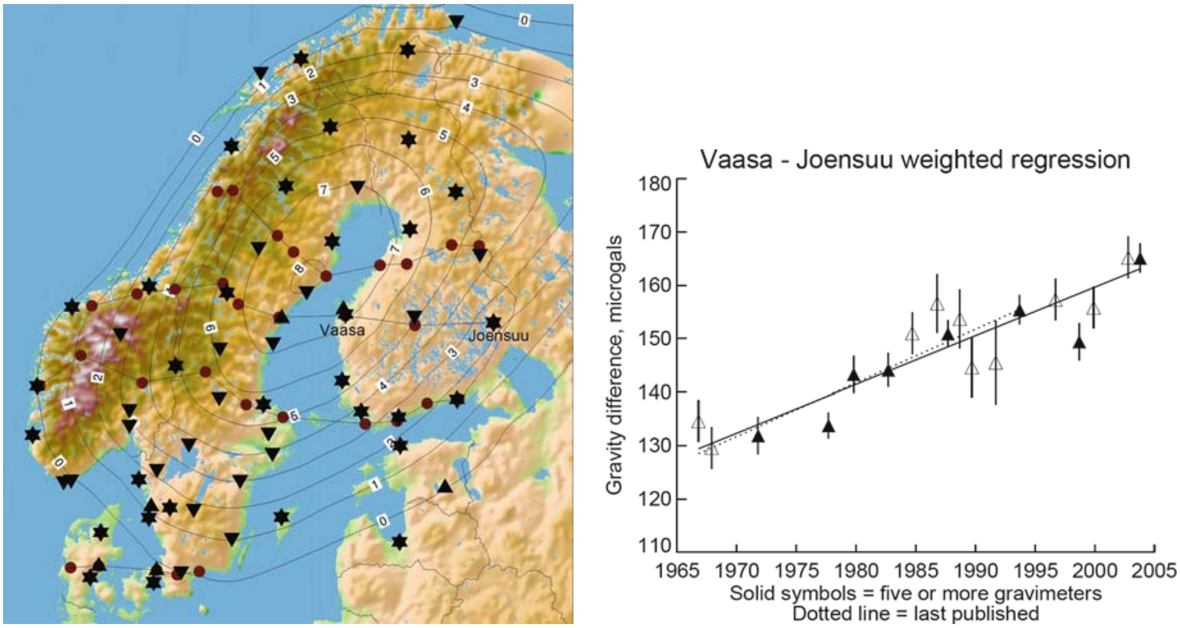

Fig. 1 GIA in Fennoscandia. *Left*: The upside-down triangles on the map are permanent GNSS stations, triangles stations where regular absolute gravity is measured as a part of the NGOS project, and dots with joining lines are the land uplift gravity lines, measured since the mid-1960s. Contour lines show the apparent land uplift relative to the Baltic mean sea level 1892–1991, based on Nordic uplift model NKG2005LU (Vestøl, 2006; Ågren and Svensson, 2007) *Right*: Diagram of the observed relative gravity change between Vaasa and Joensuu in Finland during 40 years of measurement on the land uplift gravity lines (Mäkinen et al., 2005)

Scherneck et al. (2002) (Fig. 3). Maps based on GPS time series were published e.g. by Mäkinen et al. (2003), Milne et al. (2001), Lidberg (2007), and Lidberg et al. (2007).

In North America several hundreds of continuous GPS stations have been used to compute contemporary velocities (e.g., Calais et al., 2006; Wolf et al., 2006; Sella et al., 2007). In Greenland a campaign with repeated GPS has been carried out over a period of close to 10 years (Dietrich et al., 2005) with uplift values of the order mm/year close to the ice cap.

The gravitational uplift signal can be detected by absolute and relative gravimetry (e.g., Ekman and Mäkinen, 1996; Mäkinen et al., 2005) or by the GRACE satellite mission (e.g. Wahr and Velicogna, 2003; Peltier, 2004; Tamisiea et al., 2007). The gravity satellites GRACE and GOCE are providing, or will provide, additional global and regional constraints on the gravity field (Pagiatakis and Salib, 2003; Müller et al., 2006). Recent studies have demonstrated that the GRACE data clearly show temporal gravity variations both in Fennoscandia and North America (Tamisiea et al., 2007; Ivins and Wolf, 2008; Steffen et al., 2008). The temporal trends and the uplift pattern retrieved from these data are in good agreement with previous studies and independent terrestrial data (Fig. 2).

The gravity change due to the postglacial rebound is about −2 μgal/cm of uplift relative to the Earth's centre of mass, or about −2 μgal/yr at the centre of the uplift area in Fennoscandia (Ekman and Mäkinen, 1996). Based on this, the peak geoid change rate is estimated to be 0.6 mm/yr. The results are based on land-uplift gravity lines in Fennoscandia (Fig. 1), observed regularly since the mid-1960s (Mäkinen et al., 2005). Currently, an increasing number of continuous GNSS sites are also monitored using repeated absolute gravity measurements.

Crustal deformation and sea level variation studies are based on stable reference frames. If effects at the 1 mm/yr level are to be studied, a stability of about 0.1 mm/yr in the reference frames is needed over several decades. Such stability is not yet achieved. Geodesy's response to this requirement is the Global Geodetic Observing System (GGOS), a new integral part of the International Association of Geodesy, (GGOS, 2008). There are several ongoing plans

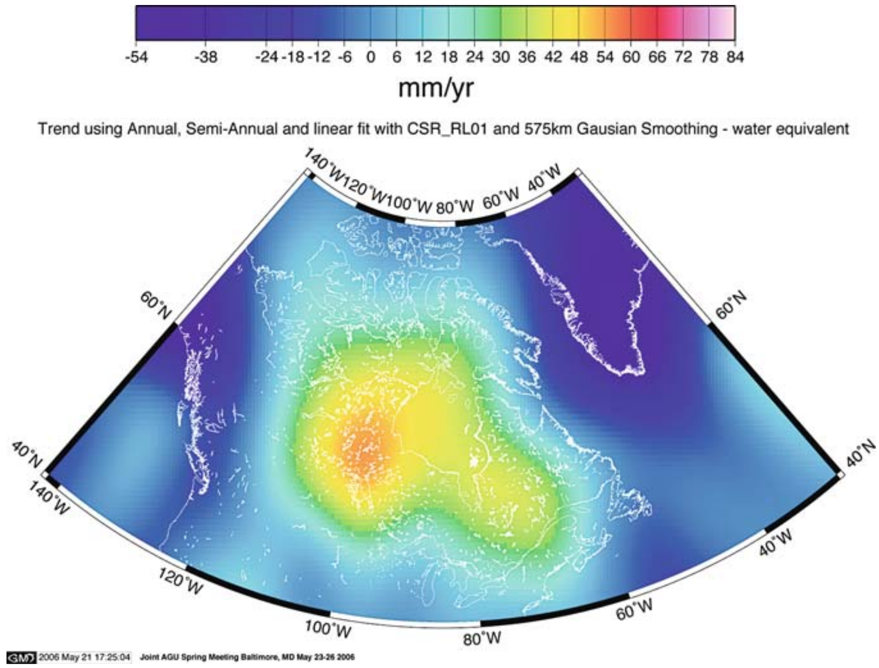

Fig. 2 GIA in North America shown as a GRACE-derived water-equivalent mass change. The GRACE signal is unfiltered by hydrological modeling. The GRACE Level 2 product employed is from Release 01 of the Center for Space Research (CSR) from the University of Texas at Austin, which uses the months January 2003 to December 2006, excluding July 2003. The harmonics are truncated at degree and order 60 and a Gauss filter of 575-km radius is applied. (Ivins and Wolf, 2008)

for regional implementation of GGOS, as an example the Nordic Geodetic Observing System, (NGOS, Poutanen et al., 2007). The NGOS plan includes also annual absolute gravity measurements at the permanent GNSS sites (Fig. 1).

Evidence from Geophysical Observations of Lithosphere Structure

Our present knowledge of the rheology and structure of the lithosphere is based on a combination of rock deformation experiments, petrophysical inference from seismology and heat flow (Blundell et al., 1992; Bürgmann and Dresen, 2008). Continuous GNSS observations of plate-wide strain, accompanied by seismological investigations, and followed by continuum mechanical modelling of GIA, studies of seismic source and wave propagation, and studies of the postglacial faults offer new insights into properties of the lithosphere. Observations and models of glacial and postglacial faulting can help to illuminate crustal stress fields and therefore crustal rheology issues.

Existing data on experimentally studied lower crustal and mantle composition and 3-D structure derived from xenolith data, lithospheric thermal models (Kukkonen et al., 2003; Hieronymus et al., 2007) and seismic studies (Bruneton et al., 2004; Sandoval et al., 2004; Yliniemi et al, 2004; Hjelt et al., 2006; Pedersen et al., 2006; Plomerova et al., 2006; Gregersen et al., 2006; Janik et al., 2007; Olsson et al., 2007) should be utilized for forward rheological modelling of the lithosphere and for testing of dynamic uplift models. The presence and volume of fluids in the upper mantle and the influence of fluids on the mantle rheology is an open question. As dissociated water may provide an effective mechanism for electrical conductivity in the upper mantle, important implications on mantle fluids and the lithosphere-asthenosphere system can be obtained from recent deep electromagnetic measurements (Korja et al., 2002; Hjelt et al., 2006; Korja 2007).

Inversion of deep temperature data in boreholes provides direct access to ground temperature histories during glaciation times (Kukkonen and Jõeleht, 2003). Kimberlite facies in crustal rocks contain mantle xenoliths and these provide a basis for extrapolating temperature and composition to larger depths using seismology (Stein et al., 1989; Kukkonen et al., 2003; Bruneton et al., 2004; Hjelt et al., 2006; Olsson et al., 2006, 2007; Pedersen et al., 2006). These results can be used to develop more realistic models of

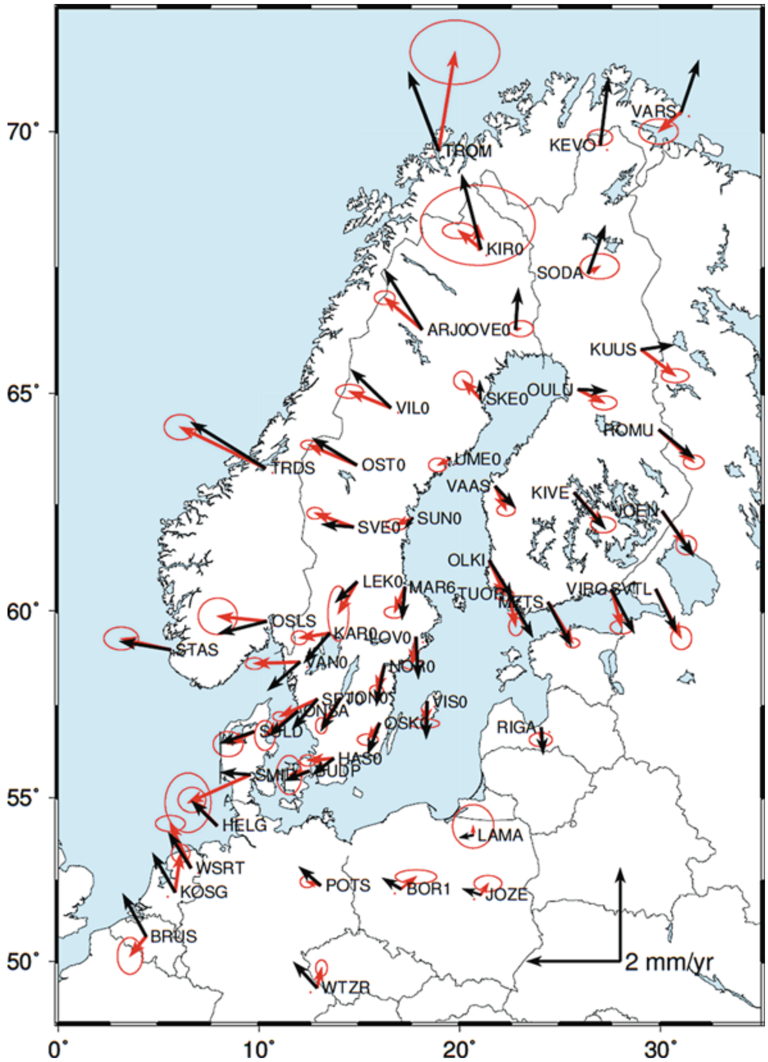

Fig. 3 Observed (*red*) and modelled (*black*) rates of horizontal displacement in Fennoscandia based on a model of Milne et al. (2001) and the GPS-derived velocity field of Lidberg et al., (2006), based on Nordic permanent GPS stations. (Lidberg et al., 2006)

mantle temperature and viscosity. These properties are key factors controlling the Earth's response to ice mass change.

Some of the largest fault scarps in northern Fennoscandia were formed at the end of the last glaciation (Kujansuu, 1964; Lagerbäck, 1979; Olesen, 1988), Fig. 4. These faults have lengths ranging from a few kilometers to 160 km and generally strike NNE, with maximum vertical offsets of 10–15 m. The faults generally dip to the east with downthrow to the west and they are almost exclusively reverse faults.

Quaternary deposits such as landslides and seismicity, trenching through the faults, dating using offset till sequences and radiocarbon dating of organic material, and geophysical investigations (e.g. Lagerbäck, 1979, 1990; Olesen, 1988, 1992; Bäckblom and Stanfors, 1989) have shown that the faults ruptured violently as large earthquakes. The magnitudes of these earthquakes is estimated to have reached MW 7–8, based on the distribution of triggered landslides, the distribution of current day seismicity and scaling relations for fault lengths (Lagerbäck, 1979; Arvidsson, 1996, Stewart et al., 2000).

As the faults are inferred to have ruptured just as the ice retreated from the respective area, these Glacially Induced Faults (GIFs) are frequently referred to as endglacial or postglacial, where the former is a more accurate description. The GIFs mostly ruptured through old zones of weakness (shear zones),

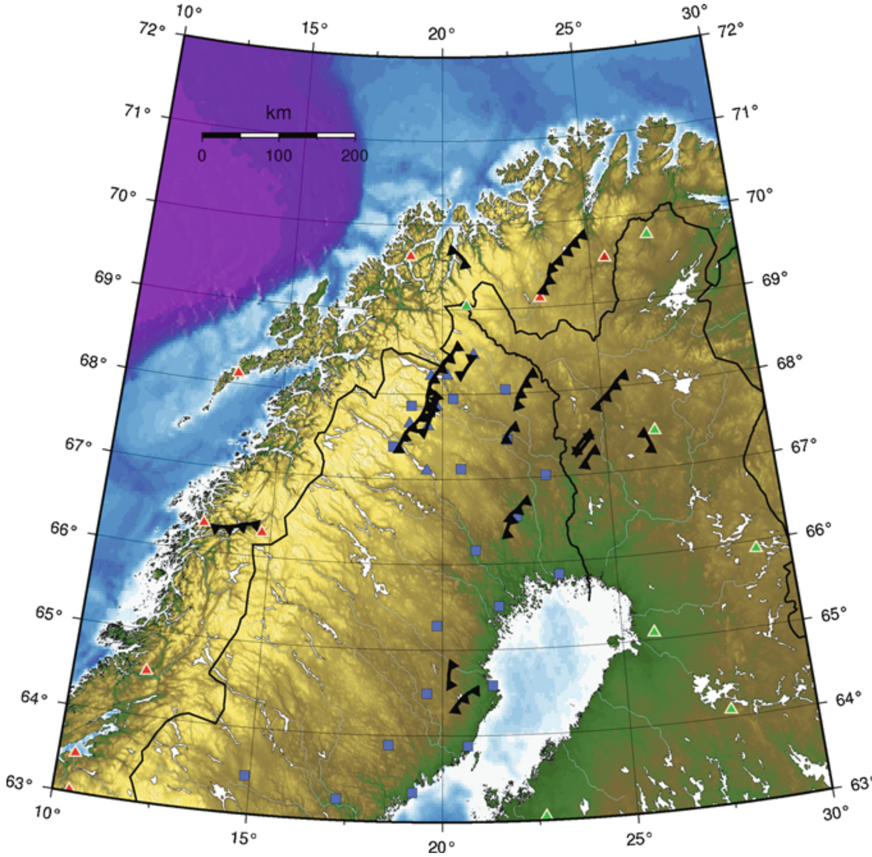

Fig. 4a Endglacial faults in Fennoscandia. *Blue squares* and *triangles* are Swedish permanent and temporary seismic stations, green triangles are Finnish seismic stations, and red triangles Norwegian seismic stations

Fig. 4b Example of an endglacial fault in Fennoscandia: the Stuorragura reverse fault of northern Norway. View is due to the E and scarp height is c. 7 m (Olesen et al., 1992)

not necessarily following one zone but instead jumping to another to comply with the restraints set by the causative stress field.

As seen in the map in Fig. 4, there is no clear relationship between the orientation of the faults and the centre of rebound (Figs. 3 and 9), the GIFs consistently strike NNE-NE and mostly dip to the east irrespective of their location west of, north of or at the center of rebound. In addition to the Fennoscandian GIFs, end- or postglacial faults have been identified in most previously glaciated areas. A comprehensive review is presented in Munier and Fenton (2004), with examples from North America, the British Isles and Russia.

Although much effort has been spent on investigating the faults, key questions concerning the formation and current status of the GIFs are still unresolved. These include fault geometry at depth, fault strength, the interplay between the glacially induced stress field and the tectonic stress at the time of rupture, the influence of pore pressure and current deformation rates. Especially intriguing is the fact that, excluding the Berill Fault located in southern Norway (Anda et al., 2002), large GIFs have been identified exclusively in northernmost Fennoscandia (Munier and Fenton, 2004; Lagerbäck and Sundh, 2008), an observation which poses a difficult challenge to models of GIF formation.

The reconstruction of the former sea level is important for both the quantification of GIA and the reconstruction of palaeo-environments. Geologically, the uplift is documented in ancient shorelines (e.g. Lambeck et al., 1998, Tikkanen and Oksanen, 2002), but the accuracy of the timing of the shorelines is a limiting factor. The time of formation of shorelines and other indications of former sea level spans over the last glacial cycle. The evidence is based on so called sea level indicators (SLI), such as fossil samples of shells or morphological features like ancient shorelines and isolation basins.

Over the last decades many attempts were made to reconstruct palaeo-shorelines worldwide. This is because of the implications of sea level changes for the densely populated areas near the coasts. This in turn gave rise to a number of international campaigns for the compilation and interpretation of these SLIs, e.g. IGCP 61, 200 (van de Plassche, 1986), 247 (van de Plassche et al., 1995) and 367 (Shennan et al., 1998). For GIA, SLIs are a unique data source, because they allow the isolation of sea level change produced by crustal deformations and unaffected by presently relevant processes, such as sea level rise due to global warming. However, due to their usually indirect indication of former sea level, they have to be used with care in constraining GIA models (Klemann and Wolf, 2007). In the periglacial regions the palaeo-sea level is dominated by regional lithospheric flexure due to glacial loadings. Therefore, a link between GIA and reconstructions of palaeoclimate especially in the coastal regions is of interest.

Figure 5 shows the reconstruction of the topography in northern Europe with respect to sea level near the end of the last glaciation. A residual ice sheet is still present. The extent of the sea shows parts of the English Channel being above sea level and the Baltic Sea flooding large parts of Finland and northern Sweden. Also visible is the isolation of the Baltic from the ocean at this stage which defines the region to be at the Baltic Ice-Lake stage. This fact complicates the use of a standard GIA model for the reconstruction of this lake stage. The lake level exceeded the mean sea level by up to 26 m before its drainage (Påsse, 1996).

A physical model for the reconstruction of the lake needs palaeoclimatic information on precipitation and evaporation. Coupling with a dynamic ice model allows quantification of the inflow of water from the melting ice sheet and therefore an assessment of the relative salinity of the environment and the height of this lake above eustatic sea level. This additional height acts as an additional load which deforms the earth. On the other hand, this additional amount of water does not contribute to the eustatic sea level during the lake stages of the Baltic Sea.

Realistic regional modelling will need considerable improvement in the reconstruction of the palaeotopography. In addition, a spherically heterogeneous earth model may become necessary. Such models are designed to simulate the present time sea-level and geoid variations on a global scale as recorded by the recent space missions CHAMP and GRACE.

For GIA two further aspects are important: First, the surface heat flow will constrain the dynamics of the ice sheet (Näslund et al., 2005) and thus the deglaciation history. Second, the viscoelastic response of the solid earth is influenced by the rheological behaviour of the lithosphere. Again, lateral variations play a crucial role in the inference of the regional palaeotopography.

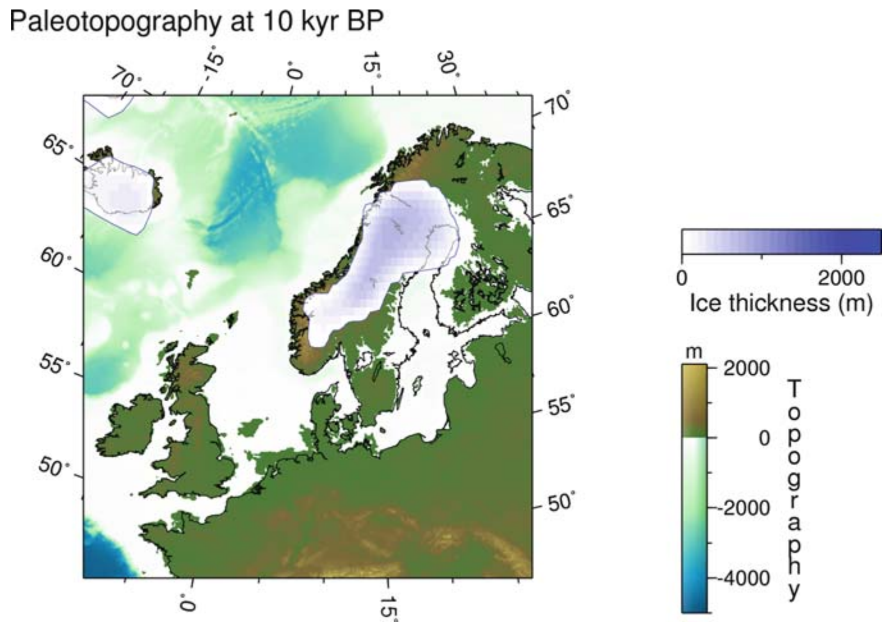

Fig. 5 Prediction of palaeo sea level near the end of the last glaciation phase in northern Europe based on a specific ice- and earth model

Seismicity and Stress-Field

The present-day seismicity in Fennoscandia as a whole is in general low to moderate in magnitude. Tectonic stress rates are small because this is an intraplate and cratonic region. This situation differs drastically from that at late-glacial times (i.e. 11–9 ky B.P.), when powerful earthquakes created impressive ground surface ruptures (e.g. Lagerbäck, 1979; Olesen, 1988). Presumably, the ice cap inhibited seismicity and strain-release during the Pleistocene glaciations. This caused earthquakes with magnitudes up to 8 when sudden global warming and ice-retreat occurred at the Pleistocene-Holocene transition (Johnston, 1989; Wu et al., 1999). A second mechanism, based on glacially induced stresses, was suggested by Wu and Hasegawa (1996)

The present-day seismicity in Fennoscandia is of intraplate type. The epicentres of these intraplate seismic events tend to be concentrated along ancient tectonic deformation zones. Due to the low seismicity level and relatively small number of permanent seismic stations in the past, studies of the sources of seismic events in Fennoscandia are relatively rare. The existing studies suggest that the sources are in areas of weakness in the crust which are favourably orientated with respect to the regional stress field and therefore can be reactivated (e.g. Slunga, 1991; Arvidsson and Kulhanek, 1994; Arvidsson, 1996; Uski et al., 2003). Thus studying both aspects (e.g. regional and local variations of stress field and distribution of zones weakness in the lithosphere capable to accumulate stresses) is important for the understanding of local seismicity and seismic hazard in Fennoscandia.

The stress state responsible for the observed seismicity appears to be the result of various stress-generating mechanisms (e.g. Fejerskov and Lindholm, 2000; Uski et al., 2003). In-situ stress measurements argue for relatively high magnitudes at shallow depths below the ground surface (Stephansson et al., 1986) The recent discovery of impressive stress-relief structures in different regions of Norway (Roberts, 2000; Roberts and Myrvang, 2004; Pascal et al., 2005) adds support to this conclusion. Observations to 6.5 km depth in the Siljan boreholes, central Sweden, suggest a strike-slip stress state at 5 km depth with maximum horizontal stress in the general direction NW-SE (Lund and Zoback, 1999). Although stress deviations are locally observed in Fennoscandia, maximum principal stress axes are in general horizontal and strike NW-SE (Slunga, 1991; Heidbach et al., 2008), suggesting ridge-push as the dominant mechanism (Fig. 6). A similar conclusion is reached for North America where the large majority of stress indicators show NE-SW compression in agreement with ridge-push (Adams and Basham, 1989).

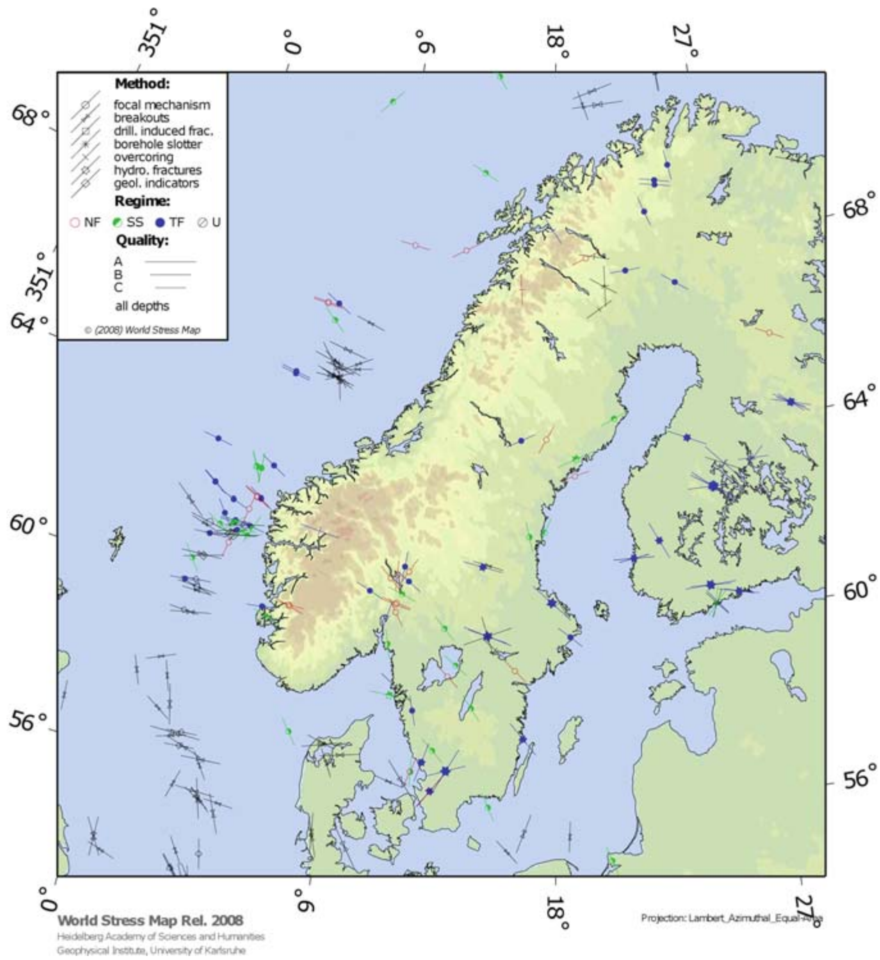

Fig. 6 Stress orientations in Fennoscandia and adjacent regions (World Stress Map website; Heidbach et al., 2008). Note the dominance of NW-SE maximum stress axes in agreement with ridge push force orientations in northern Europe

Postglacial rebound has often been advanced as a secondary source of stress modifying the tectonic stress. However, no clear radial pattern can be observed in the present-day stress compilations (Gregersen, 1992; Gregersen and Voss, 2009). This suggests that, in contrast with the situation that prevailed just after deglaciation (Wu, 1998), rebound stresses are nowadays in Fennoscandia, surpassed by plate motion forces or local stress sources.

A number of regions along the margins of the rebound domes in both Fennoscandia and Laurentia reveal complex and heterogeneous stress-fields. A similar situation exists in Antarctica and was possibly responsible for the $M_W \approx 8.1$ March 25, 1998 Balleny Is. earthquake in the Antarctic plate (Tsuboi et al., 2000; Ivins et al., 2003). This complexity is very unusual in the context of the homogeneity typically found in plate-tectonic stress fields. An additional stress source can be found in the anomalous elevation of southern and probably also northern Norway. It has been shown recently that the southern Scandes are prone to generate significant gravitational stresses acting on the adjacent regions (Pascal and Cloetingh 2009).

Due to the complexity and spatial heterogeneity of the regional stress field in Fennoscandia, it is not completely understood if the rebound stress (Fig. 7) is still able to trigger seismicity today. In addition, predictions of the onset time and mode of failure are very sensitive to the proper selection of models for ice sheet and mantle rheology (Wu et al., 1999; Klemann and Wolf, 1999; Lund and Näslund, 2008). However, the rebound processes seem to play a certain role in triggering seismicity in intraplate areas of northern America (see James and Bent, 1994; Wu and Johnston, 2000; Grollimund and Zoback, 2001). The same is claimed for Greenland by Chung and Gao (1997) and by Chung (2002).

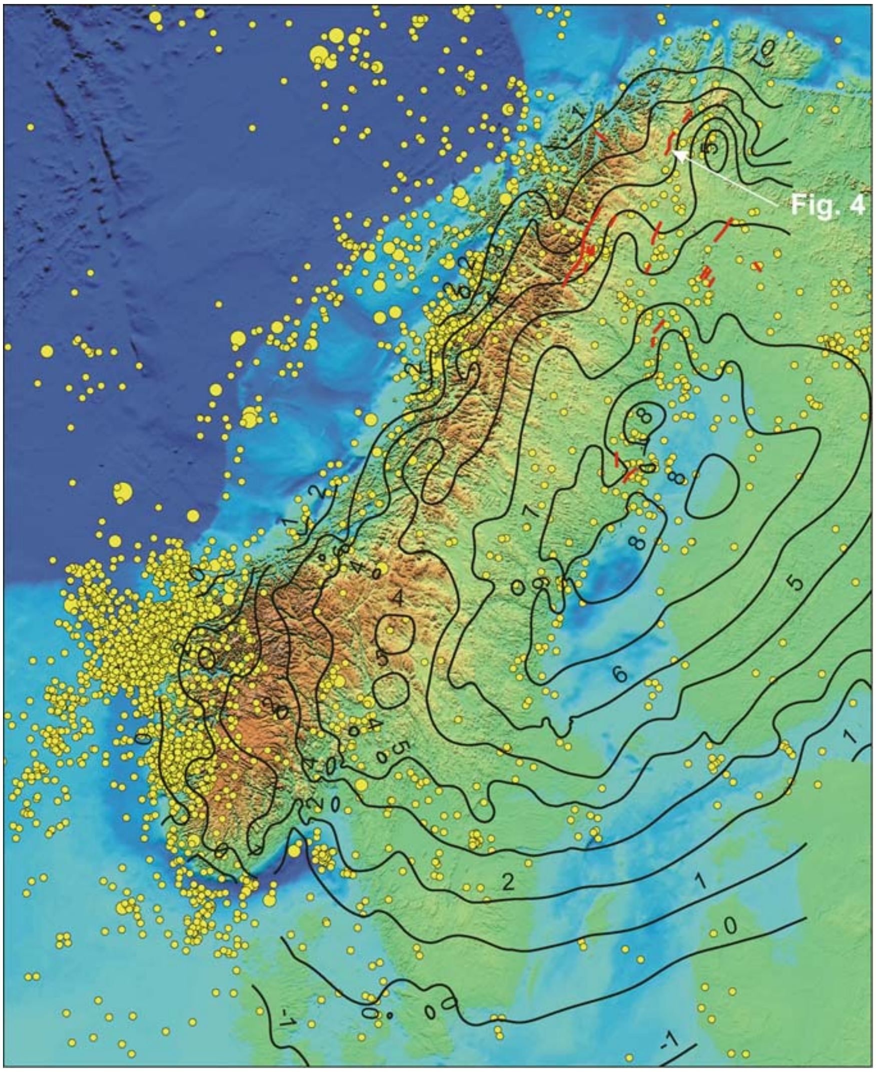

Fig. 7 Distribution of earthquakes in Fennoscandia superimposed on isolines of uplift in mm/yr (redrawn from Dehls et al., 2000)

The triggered earthquakes are concentrated either along zones of weakness or in regions of local stress concentrations. Recently van Lanen and Mooney (2007) have suggested that the structure of the lower crust, in particular the existence of deep faults rather than lateral variations in temperature, rheology or high pore pressure is a major factor controlling the spatial distribution of the intraplate seismicity in eastern North America. The faults are associated with ancient intracontinental rifts, palaeorifted margins or major terrain boundaries in the crust. These fossil structures can be easily reactivated by the rebound stresses. The importance of lateral variations in lithosphere structure for postglacial seismotectonics was also demonstrated by Wu and Mazzotti (2007), who showed that a narrow ductile zone that cuts vertically through the lower crust and lithospheric mantle generally has a larger effect on crustal motion and strain rates due to GIA than a horizontally uniform ductile layer.

Cryosphere and Palaeoclimate

Past and present changes in the mass balance of the Earth's ice sheets, ice caps and glaciers induce present-day deformation of the solid earth on spatial scales ranging from local to global. The Earth's deformational response to cryospheric change is complex due to a number of factors including complexities in the viscoelastic structure of the earth, the spatial and temporal variability of the ice mass changes and the interaction between the cryosphere and the ocean, which lead to a redistribution of cryospheric mass in a highly dynamic system. Both short and long term physical changes give important boundary conditions for the ice cap variations and their ablation. An example is provided by Greenland, where the ice sheet retreated after the last glacial maximum without disappearing. The associated sea level changes (e.g. Weidick 1995) supply important geophysical constraints together with those provided by GPS and gravity studies (e.g. Dietrich et al., 2005).

Recent advances in studies of the glacial history of northern Europe and Eurasia have significantly improved our understanding of the glaciation and deglaciation histories during the Weichselian and Holocene epochs over the past 100 kyr (Svendsen et al. 2004). In addition, the development of numerical modelling of glaciations has shed new light on the processes involved as well as their mutual couplings (Forsström, 2005; Zweck and Huybrechts, 2005; Näslund et al., 2005; SKB, 2005). As a result, the latest generation of ice sheet models is significantly better constrained and more realistic than previously.

Current Models and Problems to be Solved

The observations mentioned above can be used to constrain geophysical models of GIA using the observational evidence or to provide, e.g., GIA-induced uplift velocities for prespecified rheology (e.g. Lambeck et al., 1998). Milne et al. (2001) solved for Earth rheology using 3-D motion observations from the BIFROST continuous GPS network (Johansson et al., 2002) and the ice history of Lambeck. Current GIA models, however, suffer from regional deviations up to 0.2–0.5 mm/yr compared to the observed values. This will degrade the resolution of the past behaviour of the uplift, interpretation of the rheology, and therefore also the ice history.

The determination of the palaeoclimate and glaciation history of the Pleistocene are complex global problems and should therefore be solved in a joint effort. In view of the past observational and modelling activities, the two major formerly glaciated regions, Fennoscandia and Laurentia, are expected to be most promising. One main focus will be the global modelling of the climate forcing and the modifications due to the mechanical responses of the lithosphere and the asthenosphere (Bintanja et al., 2005; van den Berg et al., 2006).

Current GIA models are mostly based on radially stratified (3D) Earth models with linear rheology (e.g., Sabadini and Vermeersen, 2004). During the last few years progress has been made in the development of global, 3D-stratified earth modelling (Martinec, 2000; Wu and van der Wal, 2003; Whitehouse et al., 2006). However, due to computational restrictions, the latter models still have relatively low resolution.

High-resolution gravity data with spatial scales of 100 km or better and with nearly uniform quality will hopefully become available from the GOCE satellite mission scheduled for launch in 2009. The data are expected to reveal detailed 3D information on earth rheology and Late-Pleistocene ice sheet evolution (Vermeersen and Schotman, 2008). Collaborative projects have therefore been initiated on coupling interactively a regional (flat) 3D high-resolution finite-element model to a 3D thermomechanical ice-sheet model that includes dynamically consistent ice shelves. These models will especially be applied to northern Europe, where also information is available on mantle composition from xenoliths in the Caledonian lithosphere along the western margin of Northwest Europe. Simulation studies (Schotman et al., 2009) predict the appearance of distinct geoid anomaly patches expected to be visible in GOCE mission data for northwestern Europe, which in turn should provide valuable information on regional rheology and shallow Earth structure (Fig. 8).

Another task is to couple existing GIA uplift data, uplift models and the most recent geological and palaeoclimatological data on glaciation history. Northern Europe and Russia provide a study area with several recent contributions. Using land uplift models, the sensitivity of uplift data on variations in ice thickness and duration should be quantified, at least for the period of the last deglaciation, i.e. from the Last Glacial Maximum at about 22 kyr B.P. to the present time. Inverse modelling of the glaciation history may be a promising new approach previously not applied. The consideration of palaeotide models which are coupled to the land uplift history (e.g. Thomas and

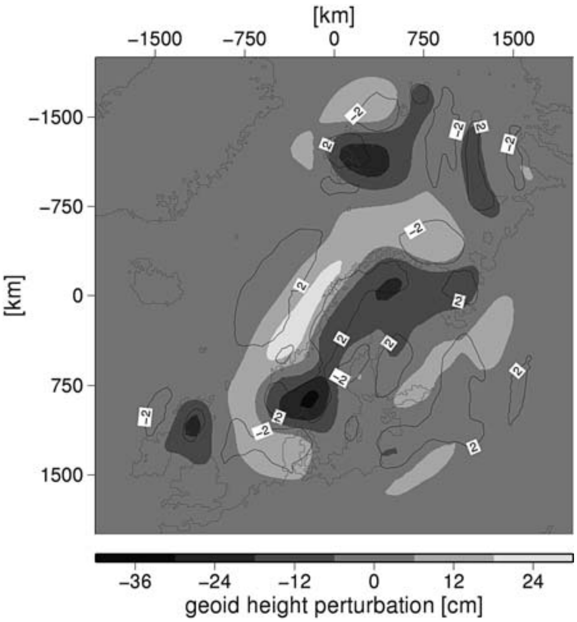

Fig. 8 Simulated geoid height perturbations induced by a crustal low-viscosity zone. For more details on models and parameters used, see Schotman et al. (2008). Such geoid anomalies should become detectable by the GOCE satellite mission in view of its expected resolution of the geoid down to characteristic spatial scales of 100 km and magnitudes of 1 cm

Sündermann, 1999) is expected to improve the reconstruction of palaeoenvironments and the interpretation of sea-level indicators (e.g. Klemann and Wolf, 2007). Also the influence of the Tornquist-Teysseire Zone (see Tesauro's map, this issue) on the uplift pattern during the last glaciation is of some evidence when focusing on sea-level variations in Denmark or southern Sweden.

An important step toward a better understanding of GIA in terms of the viscoelastic structure of the earth's lithosphere and mantle has been the joint inversion of different types of geodetic and gravimetric data related to GIA – preferentially connected with geological relative sea-level evidence of the earth' rebound during the last 10,000 years. An example of this is the study by Wolf et al. (2006), who reviewed and analysed the geological, tide-gauge, GPS and gravimetric evidence of GIA in the Churchill region of Hudson Bay, Canada. As a result, they were able to show that the different types of observable are consistent and lead to values of about 3.2×10^{20} Pa s and 1.6×10^{22} Pa s for the upper- and lower-mantle viscosities, respectively.

A similar situation exists in the region of Ny-Alesund, Svalbard, where, in addition to the data types mentioned above, also VLBI observations have been carried out (Hagedoorn and Wolf, 2003; Sato et al., 2006). However, in contrast to Hudson Bay, the situation on Svalbard is more complicated. The main reason is that, besides the contribution due to the Earth's GIA following the last Pleistocene deglaciation, the influence of the recent melting of the Arctic ice caps and glaciers cannot be neglected. Although some progress has been achieved in determining the recent glacial changes on Svalbard (e.g. Hagen et al., 2003; Pälli et al., 2003), uncertainties remain, which may explain the discrepancies existing between the different types of observation.

Whereas the spatial scale of Svalbard is too small for the detection of the recent glacial changes on the archipelago by the gravity satellite mission GRACE, the GIA signal over Canada recorded by GRACE during more than 5 years of operation is very prominent (Rangelova and Sideris, 2008), see Fig. 2. On the other hand, the density of the North American continuous GPS network is less than half of that of the Fennoscandian network, and the individual observation periods are usually less than half of those of the Fennoscandian stations (e.g. Sella et al., 2007), see also Fig. 9. Nevertheless, the joint inversion of Canada-wide geodetic and gravimetric data – supplemented by the vast geological evidence of postglacial relative sea-level change available – in terms of the Earth's viscoelastic structure is at the verge of becoming possible. At the same time, the need to consider lateral changes of this structure, such as variations of lithosphere thickness (e.g. Audet and Mareschal, 2004), or tectonic features becomes of ever increasing importance (e.g. Whitehouse et al., 2006; Klemann et al., 2008).

Models of the process of glacially induced faulting usually proceed along one of two lines. One is the strain accumulation mechanism suggested by Johnston (1989), whereby the glacial load inhibits earthquake strain release in the elastic part of the lithosphere and the accumulated strain is subsequently released instantaneously at deglaciation. A second mechanism, based on glacially induced stresses in the down-warped elastic lithosphere, was suggested by e.g. Wu and Hasegawa (1996).

Recent models of glacially induced faulting are based on the second mechanism, using increasingly

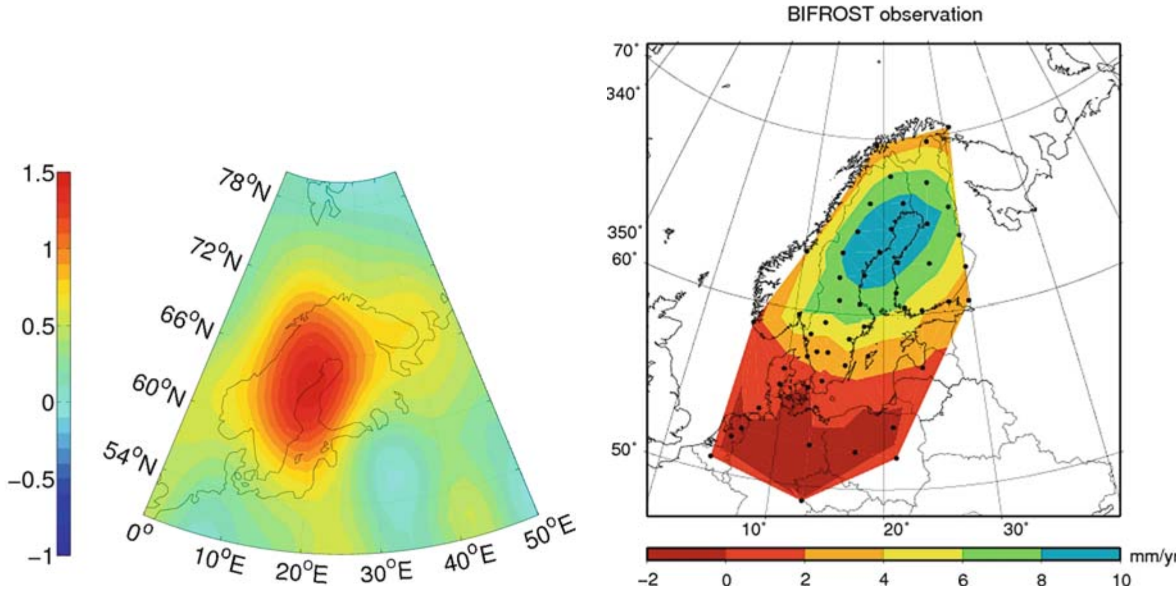

Fig. 9 Gravity trend from the GFZ GRACE monthly solution (01/2003–01/2008) and the GPS uplift velocity from Lidberg et al. (2007)

complex numerical models of the glacial history, earth rheology, rock mechanics and pore pressure (e.g. Wu et al., 1999; Hetzel and Hampel, 2005; Lund, 2005; Lund and Näslund, 2008). As alluded to above, these models are still in their infancy as they fail to predict many of the features of postglacial faulting observed.

One of the objectives of the DynaQlim initiative is to produce better models both of the glacial history and of the earth response in order to enhance our understanding of the phenomenon of glacially induced faulting and to better predict where and when it occurs. Not a purely academic exercise, as glacially induced faults are of utmost importance in the safety assessment for future nuclear waste repositories at northerly latitudes (e.g., Vidstrand et al., 2008).

Another important societal issue of the DynaQlim initiative is to investigate how much postglacial rebound stresses contribute to the triggering of seismicity that would eventually threaten human infrastructures. The largest historical earthquakes in Fennoscandia and the interior of North America occurred in the Rana region of northern Norway in 1819 with a magnitude of $M = 5.8$ (Bungum and Olesen, 2005), at mid-Mississippi River valley (New Madrid sequence) in 1811–1812 with M_W up to 8, and at Charleston in south-eastern USA in 1886 with $M_W \approx 7.7$ (Talwani, 1989). Because of a much denser infrastructure than in the 19th century, similar present-day earthquakes would be far more destructive.

As stated previously, the stress field responsible for the seismicity of formerly glaciated regions is dominated by ridge-push forces. It is however not known to what extent tectonic stresses interfere with postglacial rebound and other local stress sources to generate episodic destructive seismicity. Current models (e.g. Muir-Wood 2000) fail to account for the observations in a convincing way, probably because they do not incorporate reliable rheologies and do not involve the influence of other factors like topography (Pascal and Cloetingh, 2009) or glacial deep erosion and sediment loading (Stein et al., 1989). In brief, better illumination of this specific problem requires both more accurate rebound models and a more complete understanding of the pre-faulting stress regimes. Further aspects in the interference between GIA and surface processes are variations in river drainage during the glacial cycle which results in variations of sediment transport, trapping of sediments due to glacial variations and the resulting effects on palaeo-environments.

Climate

Ten Million Year Time Scale

The $\delta^{18}O$ isotopes deposited in ocean sediments record a long-term cooling trend sustained since the early Pliocene ($\sim$5.3 Ma to the present) (e.g., Raymo, 1994). There is great consensus that the ice sheet covering Antarctica evolved in concert with the opening of the Drake Passage 20–25 Ma that allowed the Antarctica ice sheet to become climatically buffered by a globally zonal current, the Antarctic Circumpolar Current, or ACC, the largest among all of the world ocean currents (Barker and Thomas, 2006). By direct inference, this is clearly an intersection of mantle convective and climate change time scales, and causality: plate tectonic reorganization ushering in a new stable state for the Earth's climate. What role the plate reorganization of about 4.5 Ma (Menard and Atwater, 1968) played in this context is poorly understood, although the physical isolation of the Pacific and Atlantic ocean water at the Panama isthmus likely occurred at this time (Haug and Tiedemann, 1998; Zachos, et al., 2001; Hay et al., 2002). Plate motions have been nearly constant since about 3–5 Ma (e.g., DeMets and Wilson, 2008).

Late-Pleistocene Ice Ages

Three orbital/rotational periodicities; eccentricity (400 and 100-kyr), obliquity (41-kyr) and rotational precession (23-kyr) that control summer insolation variability in the northern hemisphere, determine the time over which sustained growth and collapse of North American and Eurasian ice sheets can occur. Late summer insolation dominates the period at which glacial snow can melt during late summer. The shorter and less intense this period, the greater the chance that major ice sheets, ice caps and other large ice complexes may sustain secular growth. The relative importance of the three astronomical variables that form the "Milankovitch cycle", however, is poorly understood, in spite of the fact that the northern insolation variability may be calculated theoretically very accurately (e.g., Berger and Pestiaux, 1984). This is because the climate system has delicate and quite nonlinear feedback mechanisms at work, as well as late summer insolation variability. After about 4.5 Ma insolation minima cause prolonged ice ages.

While the waxing and waning of Northern hemispheric ice sheets are dominated by a 100-kyr cycle, benthic cyclicity in the $\delta^{18}O$ isotopic record is dominated by the 41-kyr-obliquity variability. Melt pulses of $\delta^{18}O$ from Antarctica, which are exactly out-of-phase with the Northern hemisphere at the 23-kyr precessional period, are likely the cause of this 41-kyr dominance in the benthic record, since the latter (obliquity variability) is always in-phase, and the former act to cancel one another. This fact is important because it means that both Northern and Southern hemispheric continental summer insolation is responsible for ice sheet growth and decay (Tziperman et al., 2006; Raymo et al., 2006) and that mean sea-level fluctuation from the last interglacial can be sourced to a continental origin with ever greater confidence. A substantial ramification may be the inference of 4–6 m higher last interglacial sea-level stand at Marine Isotope Stage 5e ($\sim$122 kyr BP). Rohling et al. (2008) suggests that the high-stand may be sourced to a substantial and rapid reduction of the volume of the Greenland ice sheet, well below its present day volume. At this time global mean surface temperatures were 2°C higher that today.

Jouzel et al. (2007) extracted high-resolution deuterium data (δD) from deep Antarctic ice cores (Dome C, EDC, and Dronning Maud Land hole, EDML) that are highly sensitive to the temperature of the ice that accumulated at the surface over the past 0.8 Ma. They show that the timing of warm and cold periods are tightly correlated with the global benthic $\delta^{18}O$ record determined by Lisiecke and Raymo (2005).

The ice core data also correlate with the more rapidly varying Dansgaar-Oschger (DO) events, short periods during which catastrophic marginal ice shelf breakup occurred, followed by massive expulsion of glacially eroded sediments into the ocean basins. The interval between these events is roughly $\sim$1500 yr (Hemming, 2004), and the coincident warming at discrete time intervals (<100 years) is also recorded in Greenland ice core $\delta^{18}O$ stratigraphy.

It is to be noted that the physical mechanism for the linkage of DO events with warm-cold oscillations is thought to be complex, involving internal ice sheet dynamics (e.g., Alley et al., 2001). The EDC and

EDML δD records also show a dominance of the in-phase obliquity cycle and support the bipolar seesaw hypothesis, in which thermohaline circulation cause warming and cooling in Greenland and Antarctica to be out-of-phase by roughly 0.5 kyr (Morgan et al., 2002). The Antarctic ice sheet is capable of providing the melt source for some of the millennial-scale benthic DO record (Blunier and Brook, 2001), a warming and melt pulsing that has probably been occurring since at least 2.2 Ma (Cowan et al., 2008).

A global ocean sediment record of the last 5.3 Ma studied by Lisiecki and Raymo (2005) reveals clear evidence of rhythmical deposition of ice rafted debris after 2.7 Ma. However, there is a time-dependence to the spectral trend in these global data, revealing stronger precessional and eccentricity oscillation later in the record, whereas prior to 1.6 Ma, the spectral power is dominated by the obliquity (41-kyr) variations. Raymo et al. (2006) used a coupled ice sheet-climate model that featured a dynamical Antarctic simulation that help explain the transition to relatively muted power in obliquity band since 1.6 Ma. A key model feature is that waxing and waning of ice grounded to the East Antarctic Ice Sheet (EIAS) margins provide huge mass contributions to ocean change during the northern hemispheric cooling associated with precessional (23-kyr) insolation changes, acting to cancel northern contributions.

The model EAIS volume changes would involve 30 m of eustatic sea-level change, about a factor of three larger than known for the last glaciation (Ivins and James, 2005). The ocean-modeled geographical distribution of marine sediment δ^{18}O observations supports these volumetrically equivalent contributions from north and south during 75–20 kyr (Rohling et al., 2004). The scenario proposed by Raymo et al. (2006) would explain the lack of 23-kyr benthic signature due to near complete cancellation between competing hemispheric responses to insolation forcing at the latter period. After 1.0 Ma, a longer-term cooling trend that is predicted by astronomical forcing (Pälike et al., 2001) drops solar insolation maxima at the coastal margins of the EAIS below a critical threshold, and the proximal ice sheet there experiences little terrestrial collapse during the warmest phases of post-1.0 Ma Milankovitch climate forcing. If the Raymo et al. (2006) scenario is correct, it has important implications for EAIS stability in a warming climate.

Last Ice Age, Postglacial Transition

Documentation of the Last Würm-Wisconsonian ice age has progressed rapidly during past 25 years. The greatest advance has come from Arctic Russia and Siberia (e.g., Serebryanny, 1985; Svendsen et al., 2004). One of the main upshots of the new glaciomoraine and other dated geomorphological indicators is the absence of large ice sheet complexes east of Severnaya Zemlya Is. and the Central Siberian Plateau at about 110°E longitude since about 55 kyr (e.g., Stauch et al., 2007). Additional constraints have become available from bathymetric and marine sediment core data and incorporated into comprehensive models that reconstruct the ice sheet that covered the Barents and Kara Seas at LGM (e.g., Elverhøi et al., 1993; Dowdeswell and Seigert, 1999). The Eurasian LGM features are now largely incorporated into the global ice models used for computation of GIA models for predicting solid earth deformation and sea-level variability (Lambeck et al., 2000; Peltier, 2004).

Exactly how the climate and oceans reorganize to sustain growth of ice sheets that ground to continents and shallow continental shelves continues to be poorly understood (Marshall et al., 2002; Zweck and Huybrechts, 2005) and incorporation of nonlinear feedback in modeling both ocean heat transport systems (e.g., Webb et al., 1997; Stastna and Peltier, 2007) and atmospheric CO_2 is a major challenge (e.g., Weaver et al., 1998). The entire stratigraphic column of δ^{18}O, δD and bubble-captured atmospheric greenhouse gases, CO_2, NO_2 and methane in the deep ice cores of Antarctica show that anomalies in these greenhouse gases are present during all major atmospheric temperature excursions, although their lags, leads, and, indeed, their causality is still in debate (Brook et al., 2000; Blunier and Brook, 2001; Wolff, 2005). Lacking the "data-training" of modern global climate simulation for contemporary weather analyses, numerical simulation of LGM climate is frustrated by the need to properly introduce the relatively sparse regional proxy climate data (pollen records, etc.) into such models, having grid resolution of roughly 250 km by 250 km (e.g., Bush, 2004).

During the last global glacial transition-time (LGT), a period corresponding to the collapse phase of continental ice sheets from LGM, roughly between 23 and 11 kyr, palaeosealevel, marine and ice core records, consistently show warming and rapid sea level rise during the initial phases of northern hemispheric summer insolation maxima. This initial phase, termed "Termination I" for the last glaciation, occurs over the entire northern hemisphere and is believed to be well underway by 21 kyr, accelerating in pace until about 19 kyr (e.g. Clark et al., 2002; MaCabe et al., 2007) when a large rise (Meltwater Pulse $1A_o$, or mwp$1A_o$) characterizes global sea level records (e.g., Lambeck et al., 2000; Peltier, 2004).

Trapped gas composition ratios (O_2/N_2) from the Dome Fuji and Vostok ice cores of Antarctica have been used to calibrate levels of local summer insolation by Kawamura et al. (2007). The data unequivocally show that the northern hemispheric insolation maxima lead Antarctic climate change, and that, therefore, northern hemispheric termination phases over the 3 global glacial cycles promote Antarctic ice sheet loss. Figure10 shows the power spectra that are recovered from the Dome Fuji local ΔT_{site} and $\delta^{18}O$ stratigraphy, as it clearly records all of the Milankovitch periodicity that correspond to northern hemispheric insolation curve. The cresendo of Antarctica's reaction to the Northern Hemisphere may have contributed in some measure to mwp$1A_o$ (Kawamura et al., 2007).

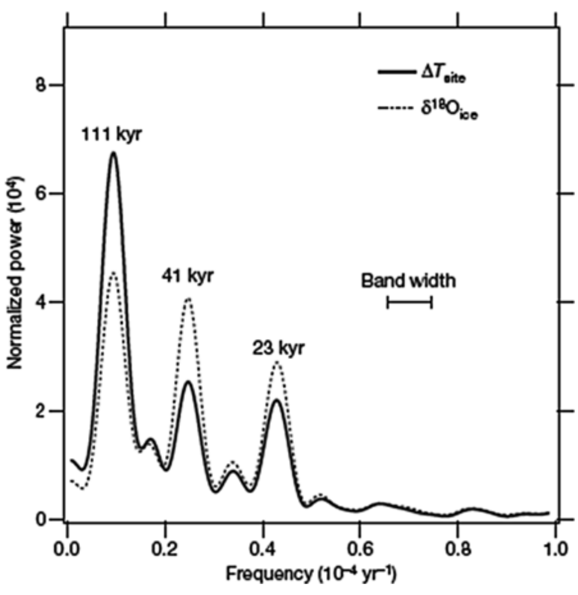

Fig. 10 Power spectra for temperature and oxygen isotopes anomalies at Dome Fuju ice core. Note that the amplitudes of the local 23-kyr and 41-kyr ΔT_{site} are quite similar. The $\delta^{18}O$ phase-lags are 1.0 ± 0.5 kyr and 2.1 ± 0.7 kyr, respectively, relative to precessional (23-kyr) and obliquity (41-kyr) Northern Hemispheric astronomical summer insolation bands. (From Kawamura et al., 2007)

After mwp$1A_o$, transition climate and sea level rise are characterized by a bi-polar seesaw, with southern warming/cooling episodes leading those of the northern hemisphere. One of the most interesting facts about transition-time and early Holocene climate that has been revealed in the Greenland ice core record is how rapid the mean atmospheric rise may occur (e.g., Steffensen et al., 2008). The largest freshwater pulses during glacial transition times affected global thermohaline circulation (THC) in the ocean, such that deepwater formation may be interrupted (e.g., Alley et al., 2005).

Overall, deepwater formation moderates climate swings and keeps the surface relatively warm. During times when deepwater formation is interrupted global cooling is a possible outcome (Knorr and Lohmann, 2007). This is likely what drove the Antarctic Cold Reversal (ACR) at 14.6–12.5 kyr as fresh water entered the oceans from northern hemispheric deglaciation (mwp1a), although a southern fresh source is not entirely ruled out (e.g., Alley et al., 2005). The precise relationships during transition times are not clear, although recent ice core evidence from Dronning Maud Land strongly points to the bi-polar climate seesaw operating during glacial times in a mode in which Antarctic warming leads each of the longer-interval (~1500 yr) DO events of Northern origin (EPICA Community, 2006).

An apparent overlap in the time period of "southern freeze" (ACR) with the period when northern latitudes were experiencing relative warmth: the Bolling–Allerod period (14.5–12.9 kyr), is also poorly understood and is likely to involve time-lags in ocean circulation phenomena of 500 years or more. A recent breakthrough by Muscheler et al. (2008) in improving the calibration of the radiocarbon ^{14}C time-scale using correlation of tree-ring carbon, and ^{10}Be in Greenland ice cores, reveals that a short-lived, but dramatic reduction in atmospheric carbon may have occurred near the beginning of the Younger-Dryas (YD) event. The latter is a dramatic millennial time-scale return to glacial (stadial) conditions between 10.8 and 10.0 kyr. A reduction in atmospheric carbon may have played a role in promoting the YD event, possibly acting in phase with a large fresh water pulse that curtailed the THC. Clearly, other greenhouse gases, such as CH_4 and NO_2 play an important role in feedbacks as is evidenced in deep ice cores of the past 650,000 yr (Mayewski et al., 2009).

Holocene and Neoglacial Change

Modern climate begins, essentially, at the end of the YD, and with the return of warm climate by nearly 6–7°C and substantial increases in eustatic sea-level. Prior to the YD event there is the large pulse of fresh water supply from the continents: mwp1b at about 11.3 kyr (Fleming et al., 1998; Peltier, 2004).

The chronology of the decay of the Laurentide ice sheet in North America is relatively well-understood during glacial transition time, but less so during the early Holocene, due to the relative abundance of ^{14}C dates on glacial moraine materials and within lake sediments (Dyke, 2004). The remaining Holocene Laurentide ice sheet consisted of the Keewatin, Baffin Is./Fox, Eastern Nunavut and Labrador ice domes (Dyke, 2004; Carlson et al., 2008). Cosmogenic exposure dating of the rocks proximal to the Labrador ice dome across a comprehensive transect normal to the eastern and western retreating fronts reveal rapid disintegration of the dome between 7.4 and 6.8 kyr (Carlson et al., 2007). One-σ errors on these ^{10}Be-calibrated dates are generally about $\pm$0.6 kyr, while the spatial resolution is 50 km along-transect.

Using a relatively primitive ice model of the dome and its shrinkage, Carlson et al. (2007) estimated the Labrador ice loss at this time accounted for about 3 m of eustatic sea level rise in a period in just 860 $\pm$ 170 years, and, thus, provides a plausible explanation for the source of the final meltwater pulse (mwp1c) that has recently been dated to 7.6 $\pm$ 0.4 kyr in southeastern Sweden by Yu et al. (2007). Loss of continental ice at this time is not too surprising due, in part, to the rise of northern summer insolation to their maximum during the Holocene “Optimum”. Should the model of Carslon et al. (2007; 2008) be correct, there is an important cautionary footnote, for the average rate of collapse of the Labrador and remaining Baffin Is. would have been large enough to cause eustatic sea-level rise at a rate or 3.5–5.25 mm/yr, corresponding to ice loss rates of 1250–1880 Gt/yr, or about 7–10 times the current rate of loss of ice from Greenland (Rignot and Kanagaratnam, 2006), which now experiences rapidly warming climate at a similar latitude in the climatic environment of the western North Atlantic (e.g., Sidall and Kaplan, 2008).

Connections to Upper Mantle Dynamics

What does this increasingly fine-tuned knowledge base of Quaternary climate have to do with mantle dynamics? Mantle time scale responses to loading have a wide viscoelastic relaxation spectra, and for phenomenon with low spherical degree order, and for polar-wander responses, loading cycles are quite important, hence these more refined features of palaeoclimate should be explicitly treated in future mantle response models. An optimum response time scale for the upper mantle is bracketed by 500–5,000 years for craton-like Earth structure, hence the differences in a load model that assume the ice sheet configuration differences between the left and right panels of Fig. 11 should be quite substantial.

There are also many examples of plate boundary responses to Little Ice Age climate changes in which low viscosity parameters are demonstrated to be sensitive to the last 400 years of loading-unloading sequences (e.g., Larsen et al., 2005). Apparently the coupling between atmospheric CO_2 and glacier/ice cap growth and diminution also occurs during the waxing and waning of the global Little Ice Age (Cox and Jones, 2008), with temperature leading CO_2 by 50 years. The ramifications for the mantle are that ocean and atmospheric temperature changes that now accelerate global ice loss will be accelerating for 50–100 years, regardless of how robust are the mitigating factors brought to bear against anthropogenic forcing.

Challenges with DynaQlim

DynaQlim (Upper Mantle Dynamics and Quaternary Climate in Cratonic Areas) is a regional co-ordination committee of the International Lithosphere Program (ILP) scheduled to last from late 2007 to 2012. The kick-off meeting was arranged in Copenhagen in February, 2008, followed by a session during the European Geosciences Union General Assembly in Vienna, April 2008 (DynaQlim, 2008). A key aim

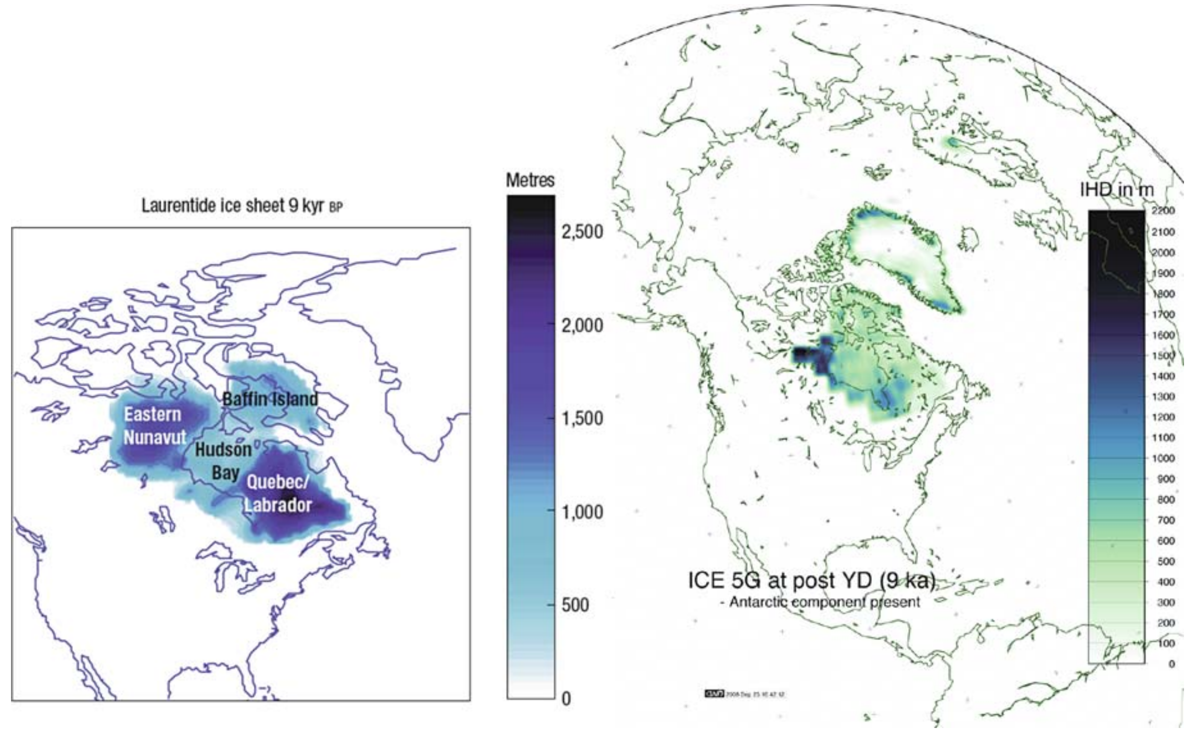

Fig. 11 Ice height differentials (IHD) at 9 kyr BP in a recent Laurentide ^{10}Be – ^{14}C based-reconstruction (left panel) and global ICE5G (Peltier, 2004) r.s.l. – ice dynamic based-reconstruction (right panel). Note that while both models portray the maximum Eastern Nunavut ice dome height amplitudes at the 1,500–2,000 m level, the Quebec/Labrador sectors differ by as much a 2 km in IHD. Left panel is from Carlson et al. (2008) right panel from a 256 degree-order spherical harmonic representation as discussed in Ivins and Wolf (2008)

is to facilitate the development of various models in order to generate more accurate predictions of Earth dynamics and ice sheet evolution during the Quaternary, and thus to understand the past and contemporaneous evolution of topography in previously glaciated terrains.

DynaQlim also provides a unique chance to study the dynamics and rheology of the lithosphere and asthenosphere in more detail. The results obtained during the running time of the project will have a number of applications, including present-day global change as well as future changes in response to a warming climate.

The tasks in DynaQlim have been divided into the following broad categories comprising the fields of

1. Geodesy, geodynamics, ocean dynamics;
2. Postglacial uplift, contemporary movements and gravity;
3. Dynamic ice sheets, glaciology;
4. Quaternary palaeoenvironments and climate;
5. Neotectonics and seismotectonics;
6. Dynamics, structure, properties and composition of the lithosphere;
7. IT, data management and outreach.

Access to relevant data is critical for successful scientific research. Concepts and technologies for an intelligent exchange of raw data (OGC, 2005), their desciription (OGC, 2007) as well as their representation in form of dynamic maps (OGC, 2006) exist.

For multidisciplinary projects conceptual challenges need to be tackled in order to make the data seamlessly accessible and portable for all users. Different disciplines (domains) describe the same datasets in their specific vocabulary, use their own data formats and organize them according to their requirements. In order to make a certain dataset originating from one domain usable to people from another discipline, the data need to be transformed and mapped between different profiles.

In the scope of the DynaQlim project, the exchange of data and information products among project partners needs to be realized based on sophisticated technologies. The installation of a web platform for the search and exchange of datasets and their descriptions is therefore intended.

Acknowledgments The research of Markku Poutanen is partly funded by the Academy of Finland, grant 120212. The research of Erik Ivins is funded by NASA's Earth Science Program, Solid Earth and Surface Processes Focus Area at the Jet Propulsion Laboratory, California Institution of Technology. The research of Jürgen Müller and Holger Steffen is funded by the Deutsche Forschungsgemeinschaft (DFG, German Research Foundation) through research grant MU1141/8-1 (SPP 1257) and that of Volker Klemann through the DFG research grant MA3432/2-2 (SPP1257).

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