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Post-glacial rebound of Iceland during the Holocene

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Abstract: The geodynamic context of Iceland provides the rare opportunity to analyse the combined effects of a ridge and a hotspot on deformation processes. From a digital elevation model, field analysis and a compilation of previous work, we propose a synthesis of vertical motions of Iceland during the Holocene with a focus on the post-glacial rebound. We determined two ancient marine limits, one above and one below the present sea level, estimated at 10 ka ± 300 years and 8150 ± 350 years BP, respectively. We calculated an uplift phase of 40–170 m with a rate of 2.1–9.2 cm a^{-1} between 10 ka ± 300 years and 8150 ± 350 years BP, corresponding to the post-glacial rebound of Iceland following the Weichselian glaciation. Spatial variations of the rebound are mainly related to the local glacial dynamics (ice load and deglaciation history) rather than the geodynamic context. However, the relaxation time deduced from uplift data is 4167 years in west Iceland and 2000 years in SSW Iceland. We estimated viscosity from relaxation time, ranging from 2.1×10^{19} Pa s to 3.2×10^{19} Pa s. The significant difference in the relaxation time is due to local variation of the lithospheric thickness as a result of rifting rather than because of a variation of the viscosity.

Because of its location both on a hotspot and on the Mid-Atlantic Ridge, Iceland has a specific rheological structure, thin lithosphere (Kaban *et al*. 2002) and low asthenospheric viscosity (Sigmundsson 1991), which modifies deformation processes on a lithospheric scale. Furthermore, its position in the middle of the North Atlantic Ocean makes it highly sensitive to climate fluctuations caused by oceanic and atmospheric circulation changes (Andrews 2005). Iceland is therefore subject to several deformation processes (magmatic, tectonic and glacial), each with a characteristic wavelength and time response. Thus it is interesting to study the coupling between this particular geodynamic context and the deformation patterns. Holocene vertical displacements are estimated around Iceland from palaeo-shoreline study. Digital elevation models, field data and previous studies allow us to map palaeo-shorelines all around Iceland. On this basis, we propose a quantitative synthesis of these displacements for the whole island, with a focus on the post-glacial rebound following the Weichselian glaciation, and discuss its spatial variations in terms of ice unloading and rheology.

Geological setting

Geological framework

Iceland is a young island, created by the interactions between the Mid-Atlantic Ridge and a mantle plume about 24 Ma ago (Einarsson 1994; Thordarson & Hoskuldsson 2002; Sigmundsson 2006). This specific context explains the anomalously thick crust and the intense tectonism and magmatism of the island. The spreading rate of the Mid-Atlantic Ridge is about 2 cm a^{-1} (DeMetz *et al*. 1994; Thordarson & Hoskuldsson 2002). Consequently, Iceland is crossed by a series of faults and volcanoes forming the Icelandic rift system at the junction between the Reykjanes Ridge in the south and the Kolbeinsey Ridge in the north. Three active zones composed of central volcanoes, rifts and fissure swarms accommodate spreading and magmatism (Fig. 1). The West Volcanic Zone in the SW is linked to the East Volcanic Zone by transform fault systems, the South Iceland Seismic Zone and the Mid Iceland Belt. The North Volcanic Zone extends towards the north to the Tjörnes Fracture Zone.

The Icelandic mantle plume enhances melt production under Iceland, generating an anomalously thick igneous crust. The crust is thickest above the centre of the plume (40–41 km), in the northwestern part of the Vatnajökull ice-cap. It thins away from the plume centre and is thinnest (<20 km) below the active rift zone, for instance in the northern part of the North Volcanic Zone and the SW of the West Volcanic Zone (Darbyshire *et al*. 2000). Therefore the Icelandic lithosphere differs drastically in thickness and in rheology from 'classical' oceanic lithosphere.

Vertical motions in Iceland

Because of its geodynamic and geographical context, Iceland is intensively subject to tectonic, magmatic and glacial processes generating vertical deformations of the lithosphere (Fig. 2). Duration, rate and wavelength of vertical motions differ depending on the active process (Dauteuil *et al*. 2005), as follows.

(1) Spreading of the mid-oceanic ridge generates centimetre-scale vertical displacements on a length scale of kilometres, such as tilted blocks parallel to the rift zone. Rates of vertical motion of 1–2 cm a^{-1} have been measured on the North Volcanic Zone borders (Hofton & Foulger 1996).

(2) Mantle upwelling under Iceland as a result of the hotspot causes a vertical and lateral redistribution of the overlying material, called 'dynamic topography'. It results in the formation of a large-scale bulge of the surface (several thousands of kilometres). The uplift rate of this process has been estimated about 0.2 mm a^{-1} in Hawaii (Zhong & Watts 2002).

(3) Local vertical deformations can be enhanced by volcanic processes such as variation of magma content inside a magma chamber. For example, withdrawal of magma from the magma chamber induces rapid and short-lived vertical displacement of the surface of up to 2 cm a^{-1} (e.g. at Krafla volcano, NE Iceland; Henriot *et al*. 2001).

(4) Post-glacial readjustment induces large-scale and fast vertical motions during short periods. Sinking of the lithosphere is enhanced by ice loading and its rebound by ice retreat. Several

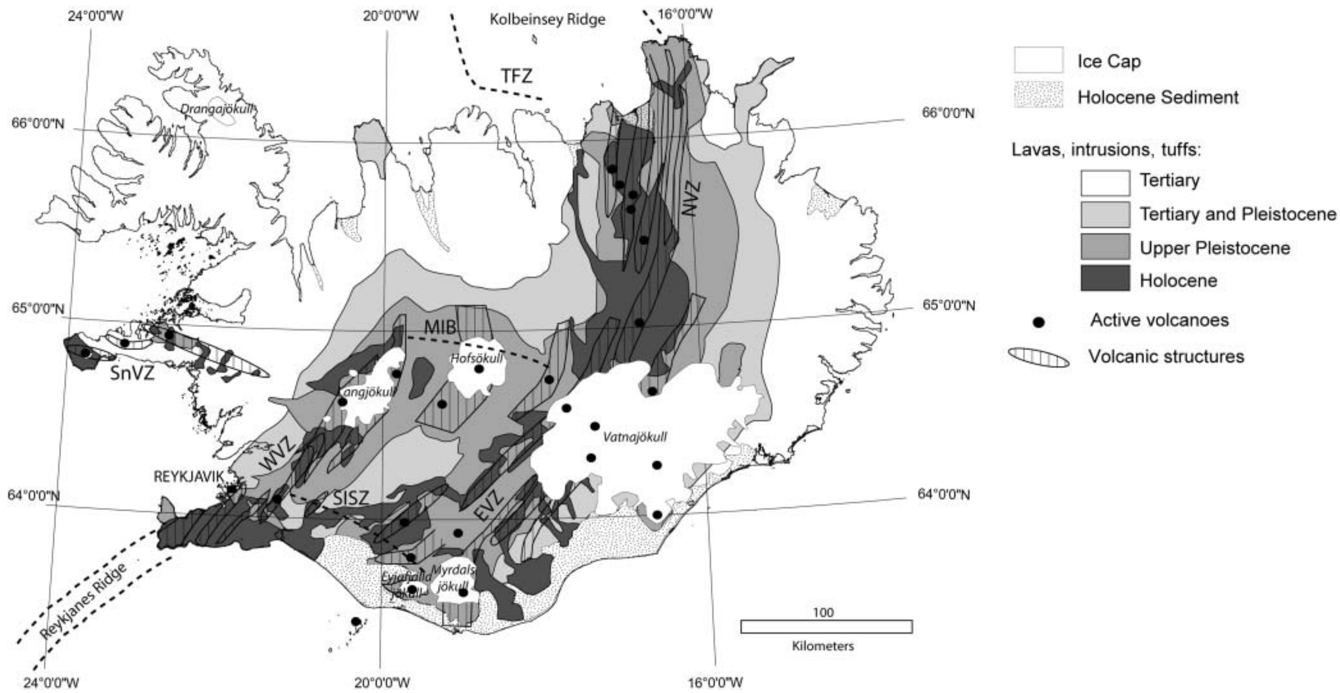

Fig. 1. Geological setting of Iceland based on the geological map of Iceland (Johannesson & Saemundsson 1988). Iceland is located both on the Mid-Atlantic Ridge and on a mantle plume. Three main rift zones link the Reykjanes Ridge in the SW and the Kolbeinsey Ridge in the north: the East Volcanic Zone (EVZ), the West Volcanic Zone (WVZ) and the North Volcanic Zone (NVZ). These rift zones are connected to each other by transform fault systems such as the South Iceland Seismic Zone (SISZ), the Mid Iceland Belt (MIB) and the Tjörnes Fracture Zone (TFZ). SnVZ, Snaefellsnes Volcanic Zone.

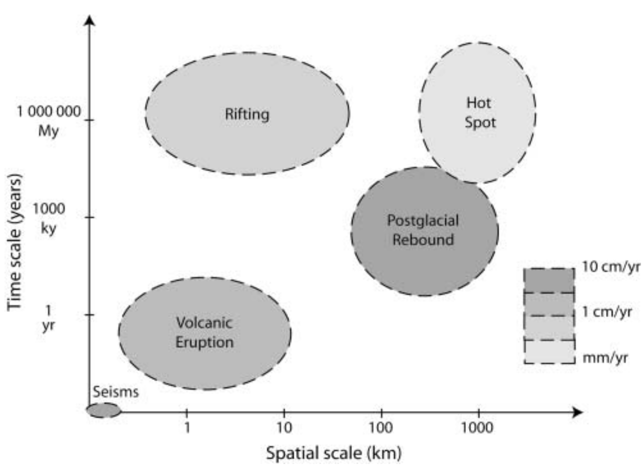

Fig. 2. Time v. space diagram of the processes generating vertical motion of Iceland (magmatic, tectonic and glacial processes). Change in the magmatic content of a magmatic chamber induces quick vertical displacement of the surface over short distances. Oceanic rifting of the mid-oceanic ridge generates vertical displacements, such as tilted blocks, at kilometre scale. The mantle upwelling underneath Iceland causes large-scale bulge of the surface (several thousands of kilometres) at a very slow rate (*c*. mm a^{-1}). Glacio-isostasy leads to fast rebound of the surface in response to ice unloading, which affects all of Iceland.

regions underwent a glacial isostatic rebound following the Weichselian glaciation, such as in North America (Mitrovica *et al*. 2000), northern Scotland (Firth & Stewart 2000) and Iceland, but the most remarkable and certainly the most studied example is the Fennoscandian uplift, which still continues (Gudmundsson 1999; Fjeldskaar 2000; Fjeldskaar *et al*. 2000). In Iceland, uplift rates have been calculated for some areas: *c*. 6.9 cm a^{-1} in Reykjavik between 10.3 ka and 9900 years BP (Ingolfsson *et al*. 1995), *c*. 7 cm a^{-1} in Berufjördur, eastern Iceland, between 10.3 ka and 9400 years BP (Norddahl & Einarsson 2001) and from 4.5 to 10.5 cm a^{-1} in the SSW between 10 ka and 8500 years BP (Biessy *et al*. 2008). The whole island underwent an isostatic rebound during this period but no quantitative synthesis for the whole island has been available until now. Recently, glacial-isostatic deformations around the Vatnajökull ice-cap induced by recent climate warming have been described (Sjörberg *et al*. 2004; Pagli *et al*. 2007). Global positioning system (GPS) measurements from 1994 to 2004 indicate vertical velocities around the ice-cap ranging from 9 to 25 mm a^{-1} (Pagli *et al*. 2007).

To quantify vertical motions of Iceland we used morphological markers located around the island as a reference of vertical position of the surface. Ancient shorelines were used as markers because they have been described in several places.

The Late Weichselian–Early Holocene period

The location of Iceland in the middle of the North Atlantic Ocean makes the ice extents highly sensitive to climate changes and consequently to glacial stages. According to Einarsson & Albertsson (1988), 15–23 glaciations affected Iceland during the past 3 Ma. The most recent glaciation, the Weichselian, took place after the Eemian period, approximately between 120 and 10 ka BP (Thordarson & Hoskuldsson 2002). The Last Glacial Maximum in Iceland is estimated between 20 and 17 ka BP (Van

Vliet Lanoë *et al*. 2006) with an ice-cap thickness up to 2000 m in the centre of the island (Norddahl & Pétursson 2005). The glacial extent during the Last Glacial Maximum is still controversial. Some workers have suggested an ice-cap extent to the shelf break (Olafsdottir 1975; Andrews 2005; Norddahl & Pétursson 2005; Hubbard *et al*. 2006) whereas others have suggested that the ice-cap was much less extensive (Van Vliet-Lanoë *et al*. 2006). Very abundant precipitation in the southern part of the island must have been responsible for the location in south Iceland of the thickest part of the ice sheet, as at present (Bourgeois *et al*. 2000).

Data on the deglaciation chronology and relative sea-level changes following the Weichselian ice-cap retreat can be obtained from palaeo-shoreline studies. The oldest dated marine shells found on raised beaches in western Iceland are dated about 12.8 ka $\pm$ 200 ^{14}C years BP (Ashwell 1975). This is the earliest record of a marine transgression on the coastal plains. Thus the warming started about 13 ka BP. Two to three glacial readvances have been described before the final deglaciation about 10 ka $\pm$ 300 years BP, at the beginning of the Holocene period (Ingolfsson 1988; Norddahl 1991; Norddahl & Pétursson 2005; Geirsdottir *et al*. 2009). The deglaciation started from the NW and continued gradually towards the SE, which is still covered by the Vatnajökull ice-cap. Thus the NW peninsula seems to have experienced a different glacial history from the rest of the island, with an independent ice-cap early in the deglaciation sequence (Hansom & Briggs 1991).

A review of the deglaciation history, the concurrent relative sea-level changes and the isostatic readjustment in various areas of Iceland is presented in Table 1 (locations of areas described in Table 1 are shown in Fig. 3). Two main conclusions can be drawn: (1) the deglaciation history is complex, with essentially two glacial readvances during the Older and Younger Dryas and a final deglaciation at about 10 ka BP (Preboreal) characterized by a relative high marine level; (2) a rapid isostatic readjustment of Iceland occurred after the ice unloading accompanied by a low relative sea level postdating the deglaciation (Boreal = Early Holocene).

We mapped and characterized through an analysis of digital numerical models the relative high and low marine levels on the coasts all around the island. We correlated our results with the previous review and field-work observations and measurements. Then we calculated the vertical motion of Iceland between the formation of these two stages to quantify the post-glacial isostatic readjustment.

Method and dataset

Quantification of post-glacial rebound can be obtained from raised marine terraces or beaches that form an elevation reference at a given time. If the relative sea level remains in steady state during a significant period, morphological markers (beaches or notches) record this stage. If the relative sea level changes by eustatic variation or land surface motion, these markers are raised or lowered. Consequently, they become progressively isolated and fossil marine features are created. Their ages can be obtained from radiocarbon dated marine shells and driftwood (Turcotte & Schubert 2002). If such fossil marine levels are located and dated, it is then possible to calculate the amplitude and rate of vertical displacement of the surface between these levels once their elevation have been corrected for eustatic changes. We used this method to quantify Holocene vertical motions around Iceland by following three steps: (1) location of palaeo-marine levels; (2) estimation of elevation and age of these levels; (3) calculation of the vertical displacement between these levels after correction for eustatic variations (Ingolfsson *et al*. 1995; Biessy *et al*. 2008).

Location of palaeo-marine levels

A marine surface can be defined as a sedimentary plain or as a marine notch. The genesis of a marine terrace leads to a slope break that is more or less significant according to the nature of the eroded substratum. When the latter is basaltic, the slope break can be significant, leading to the formation of marine cliffs, as it is commonly observed in Iceland. These slope breaks can be located on GPS profiles (Biessy *et al*. 2008) and on digital elevation models (DEMs). A DEM, even with a relatively low resolution, allows the mapping of slope breaks both onshore and offshore. In the case of a sedimentary plain, identification on a DEM is unlikely because of the lack of slope break, but most of the Icelandic coast is undergoing erosion rather than sedimentation, except for some areas that will be discussed below.

First, all the available data on Holocene palaeo-marine levels and glacial extents in Iceland were digitized and plotted on the DEM as a dataset. Then we mapped manually the first significant change that we observed in the slope of the DEMs onshore and offshore from the present coastline (Fig. 4). We produced several DEM profiles across the coast all around Iceland for a better interpretation of the slopes we mapped (Fig. 5).

Two DEMs were used, for the topography of Iceland and the bathymetry of the North Atlantic. The topographic DEM of Iceland is a compilation of digitized topographic maps (1:100 000) with a resolution of 3 arc second (40 m $\times$ 88 m) and reprojected to UTM (zone 27) WGS 1984 (kindly provided by R. Gloaguen and M. Heilbig). The vertical accuracy of the DEM is within 10 m. The bathymetric DEM was extracted from an ETOPO2 grid provided by the International Bathymetric Chart of the Arctic Ocean (IBCAO). It has a grid cell spacing of 2.5 km $\times$ 2.5 km, a North Pole Stereographic projection and WGS 1984 datum. This DEM was compiled from an accumulated database that contained all available bathymetric data at the time including soundings collected during historical and present-day expeditions as well as digitized isobaths and depth soundings from published maps (Jakobsson & Macnab 2006). Considering the echo-sounding data source we estimate a vertical accuracy of 5–10 m. Geographic information system (GIS) software calculates the DEM slopes automatically. Over a moving window of 3×3 cells, the maximum elevation change compared with a horizontal plane is calculated for each central cell relative to its eight neighbours. The maximum change identifies the steepest downhill descent from that cell. The lower the slope value, the flatter the terrain; the higher the slope value, the steeper the terrain.

Age of the palaeo-marine levels

Absolute ages of markers were found in the literature, such as radiocarbon ages of marine shells, tephrochronology and radiocarbon ages of lava flows (Fig. 3). A relative chronology was used to estimate the age of marine limits that do not contain radiocarbon dated marine shells: (1) if marine sedimentary features are preserved, it is not likely that they would have survived the advances and retreats of late Weichselian glaciers and consequently they are thought to postdate glaciation; (2) if subaerial lava flows cover the present-day coastal plain, this indicates that the sea level was lower at the time of the lava flow, and the absolute age of the lava flow gives a time boundary of

Table 1. *Synthesis of previous work describing the deglaciation history, concurrent relative sea-level changes and post-glacial isostatic adjustment of Iceland during the Late Weichselian–Early Holocene period*

Area	Deglaciation event	Palaeo-shoreline evolution		Post-glacial rebound	Reference
		Altitude	Age		
South-southwest					
Reykjavik area		30–35 m below sea level	Early Holocene	Rapid post-glacial isostatic rebound owing to low viscosity of Icelandic crust and upper mantle	Thors & Helgadottir 1991
	Extensive glaciation during Younger Dryas (11–10.3 ka BP). Post-glacial marine limit around Iceland reached between 10.3 ka and 9700 years BP, postdating Younger Dryas maximum glaciation	Relative sea-level change of *c.* 45 m, from +43 m a.s.l. to −2 m a.s.l.	Between 10.300 ka and 9400 years BP	Fast isostatic rebound of 6.9 cm a^{-1} controlled by rapid Preboreal deglaciation, together with low asthenospherie viscosities below Iceland and release of hydroisostatic stresses	Ingolfsson *et al*. 1995
From Reykjanes peninsula to Vik town		Relative high sea level from 50 to 100 m a.s.l.	10 ka BP	Rapid uplift rates (8–10 cm a^{-1}) above present active rift, surrounded by slower uplift rates: (1) 7 cm a^{-1} at young Reykjanes peninsula; (2) 4–6 cm a^{-1} in old basement area of Vik	Biessy *et al*. 2008
		40 m below sea level	8500 years BP		
West					
Borgarnes area	Glaciomarine environments with floating ice-shelf	Well above 80 m a.s.l.	Beginning of deglaciation 13–11 ka BP (Bölling–Alleröd time)		Ashwell 1975
SW of the Borgarfjödur	Two glacial readvances: (1) during Older Dryas (12–11.7 ka BP); (2) during Younger Dryas (11–10.3 ka BP)	Above 80 m a.s.l.	Bölling time (13–12 ka BP)		Ingolfsson 1987, 1988
		60–70 m a.s.l.	Final deglaciation 10.3 ka– 9700 years BP		
Northwest					
Vestfirdir peninsula	Separation of ice-cap from mainland ice mass during early deglaciation			Independent isostatic recovery history	Hansom & Briggs 1991
SE of Vestfirdir peninsula (Smahamrar)		Relative sea-level fall from 70 m to −2 m a.s.l.	Between 12 ka and 9000 years BP		
Northern coast of Vesfirdir peninsula (Hornstrandir)	High plateau (400–500 m) displays no markers of glacial erosion or deposition, suggesting a maximum ice-cap thickness of 500 m in this area				Hjort *et al*. 1995
North					
Eyjafjördur area	Two main stages: (1) ice-lake stage, with alternation of glacial readvances and ice retreats; (2) total retreat of ice at *c.* 10 ka BP				Norddahl 1991
		40 m below sea level	Postdating Late Weichselian glaciation		Thors & Boulton 1990

(*continued*)

Table 1. (*continued*)

Area	Deglaciation event	Palaeo-shoreline evolution		Post-glacial rebound	Reference
		Altitude	Age		
Northeast					
Melrakkasletta	Successive stages of glacial readvances during Older Dryas (*c.* 12 ka BP) and Younger Dryas (*c.* 11 ka BP) following by final deglaciation	Raised marine limit	10 ka BP		Pétursson 1991
East					
Berufjördur	Two glacial advances: (1) late Younger Dryas (*c.* 11 ka BP); (2) early Preboreal (*c.* 10 ka BP) with concurrent sea-level rise, culminating at *c.* 9900 years BP	Minimum position of relative sea level extrapolated at *c.* −30 m a.s.l.	*c.* 8600 years BP	Total glacio-isostatic rebound of 120 m, yielding 7.3 cm a^{-1} as a minimum rate of uplift between 10.3 ka and 9900 years BP	Norddahl & Einarsson 2001

Locations of study areas are shown in Figure 3.

this lower marine level. If some of the lava flows are now drowned by the sea, this reveals that the sea level has risen since their formation.

Calculation of vertical displacement

To calculate vertical motions of Iceland from palaeo-marine levels, it is necessary to take into account eustatic variations. The SPECMAP curve is a synthetic curve of the global sea-level changes for the last 400 ka, established by Imbrie *et al*. (1984) from $\delta^{18}O$ isotopic measurements in the North Atlantic. For this study, we focused on the eustatic variations over the last 20 ka. The SPECMAP curve is the most precise eustatic compilation for this time interval. Since the Last Glacial Maximum, global sea-level change has occurred in four stages: (1) from 20 to 16 ka BP the global sea level was stable at about −120 m (relative to the present-day sea level); (2) from 16 ka to 5000 years BP it rose at a rate of *c*. 1 cm a^{-1}; (3) from 5000 to 3000 years BP it fell slightly; (4) from 3000 years BP it has slowly risen (*c*. 0.4 cm a^{-1}) to its present-day level.

The vertical motion of Iceland ($\delta z_{\mathrm{Iceland}}$) between two palaeo-marine levels, t_1 and t_2 (with t_2 older than t_1), is determined by adding the present-day elevation variation (δz_{2-1}) between these two levels to the eustatic change (δe_{2-1}) between t_2 and t_1:

$$\delta z_{\mathrm{Iceland}} = \delta z_{2-1} + \delta e_{2-1} = (z_2 - z_1) + (e_2 - e_1). \qquad (1)$$

The SPECMAP curve shows that sea level has risen since the Last Glacial Maximum so $\delta e_{2-1} > 0$. Thus according to the elevation variation between palaeo-marine levels (δz_{2-1}), four cases are possible; (1) if $\delta z_{2-1} > 0$, then $\delta z_{\mathrm{Iceland}} > 0$, there is uplift of Iceland; (2) if $\delta z_{2-1} < 0$ and $|\delta z| < \delta e_{2-1}$, then $\delta z_{\mathrm{Iceland}} > 0$, there is uplift of Iceland but slower than the sea-level rise; (3) if $\delta z_{2-1} < 0$ and $|\delta z| > \delta e_{2-1}$, then $\delta z_{\mathrm{Iceland}} < 0$, there is subsidence of Iceland; (4) if $\delta z_{2-1} < 0$ and $|\delta z| = \delta e_{2-1}$, then $\delta z_{\mathrm{Iceland}} = 0$, there is no vertical motion.

Fieldwork

Fieldwork has been carried out to test the numerical method of palaeo-shoreline extraction. Figure 3 shows locations of the study areas. We chose areas where geological maps of 1:250 000 scale are available (Kjartansson 1969; Saemundsson 1977; Saemundsson & Einarsson 1980; Torfason 1985; Johannesson *et al*. 1990; Johannesson 1994) and where dated Holocene lava flows and marine shells have been described. It is therefore possible to constrain our observations and relative chronologies. First we observed morphological markers corresponding to sea levels and ice locations. Then we measured their elevation by GPS and compared our observations and measurements with previous work and our numerical data.

Palaeo-marine cliffs or notches characterizing palaeo-shorelines are visible on GPS profiles, as described in SW Iceland, between the Reykjanes Peninsula and the town of Vik, by Biessy *et al*. (2008). We made two profiles in the Borgarnes area (location shown in Fig. 3b).

Limits of the method

The validity of this numerical method can be debated: a slope break can be due to structures other than a palaeo-marine level, such as a moraine, delta or lava flow; also, palaeo-marine markers can be erased by glacial advance, debacle, flooding or volcanism after their formation. However, all these structures

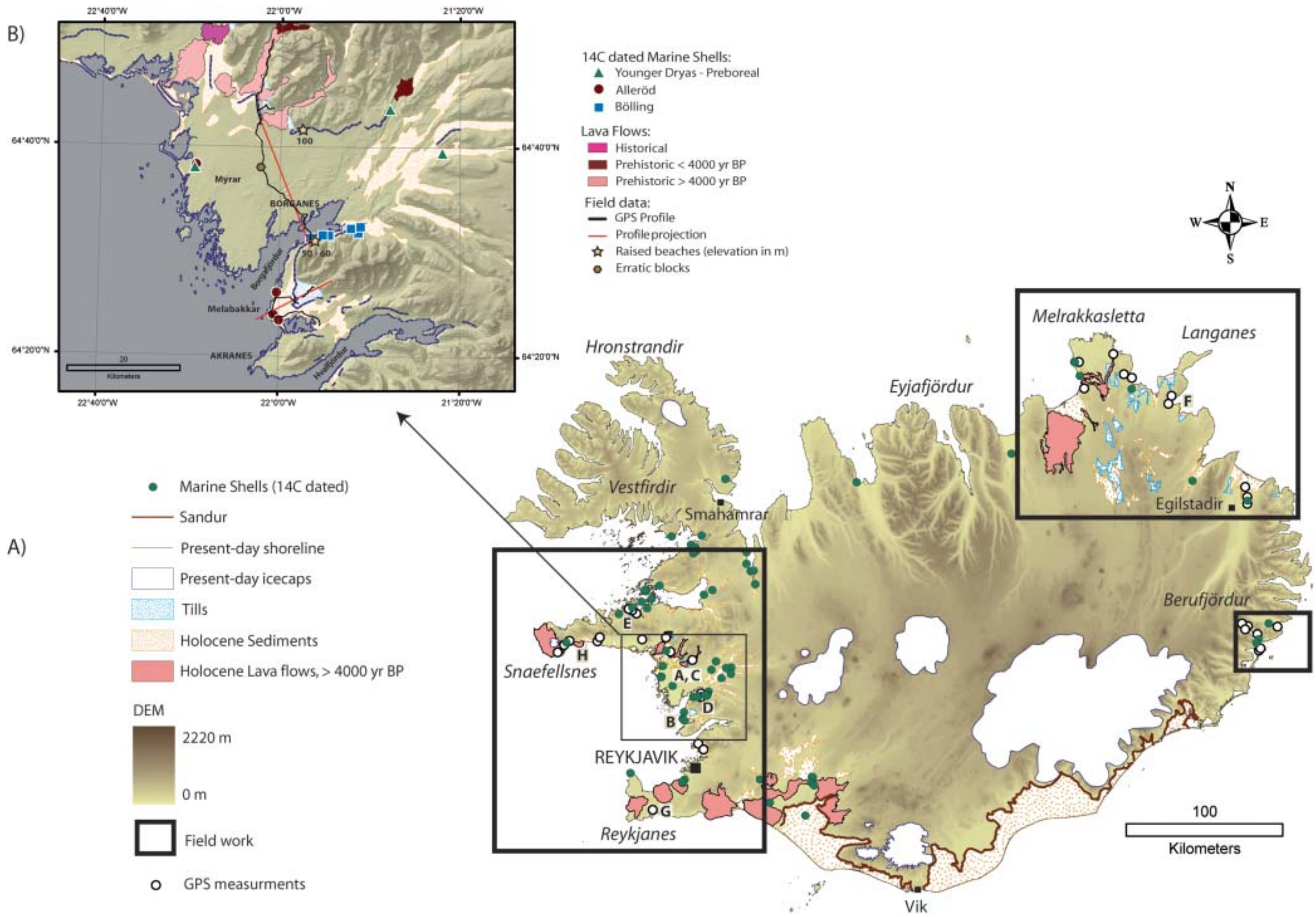

Fig. 3. (**a**) Location of geological data available from 1:250 000 geological maps (Holocene lava flows and marine shells), and location of fieldwork and GPS measurements. Letters A–H correspond to the location of photographs shown in Figure 6. (**b**) Geomorphological observations and GPS profiles in the Borgarnes area, western Iceland.

have been well described in Iceland in several earlier studies. Moraines formed during older glacial advances have been described in several places but their elevation is always higher than that of the final deglaciation marine level. Glacial advances younger than 10 ka, during the Little Ice Age, have been described (Geirsdottir *et al*. 2009). They were mainly inland and thus may not influence our results. The main problem is recent jökulhaup, flooding and glacial debacle south of Vatnajökull and in the great Jökulsa canyon NE of Iceland. Indeed, the southern coast of Iceland is covered by sandur, which may have erased and buried palaeo-marine markers in this area. In the NE, the lava flow stops abruptly, as a result of several flood events in the Jökulsa canyon, and thus a major slope break has been created close to the present-day shoreline. A significant delta has also been described offshore of this zone (Oxarfjördur) and disturbs the location of palaeo-marine levels. The results obtained in these two areas therefore have to be taken with caution.

Our method presents two possible sources of error. The first concerns the elevation (±10 m) and geographical location (±90 m) of the limits, which are both estimated from the accuracy of the DEMs. The second error source concerns the dating of these limits. Because of the lack of data in some areas, the age is the most difficult parameter to constrain and is the factor with the highest errors (±300 years BP).

Results

Palaeo-marine limits determined from DEM

As described above, the first slope break visible on the DEM was mapped both onshore and offshore. Thus we located two limits around Iceland: a relative high one onshore and a relative low one offshore compared with present-day sea level (Fig. 5). For accurate interpretation, we produced several profiles perpendicular to the coast. Different morphological coast types are present around Iceland: deeply eroded fjords, eroded plains (strandflat, e.g. Borgarnes area) and deposit plains (southern Iceland and Oxarfjördur area of northern Iceland). Transverse profiles in each of these coast types display these different morphologies (Fig. 5). Arrows on these profiles show where we mapped the line of the first visible slope breaks, which we assume to be the bottom of palaeo marine cliffs or palaeo marine notches.

In the NW (Vestfirdir) the coastal relief consists of fjords incised by the sea. It is difficult in some places to extract the high limit from the present-day sea level because of the steep slope. However, SW of this area, a slope break corresponding to the high limit was determined at +18 m above sea level (a.s.l.) (Fig. 5, profile A). This area has a particular morphology with numerous islands protected between two peninsulas, Snaefellsnes in the south and Vestfirdir in the north. The profile shows a

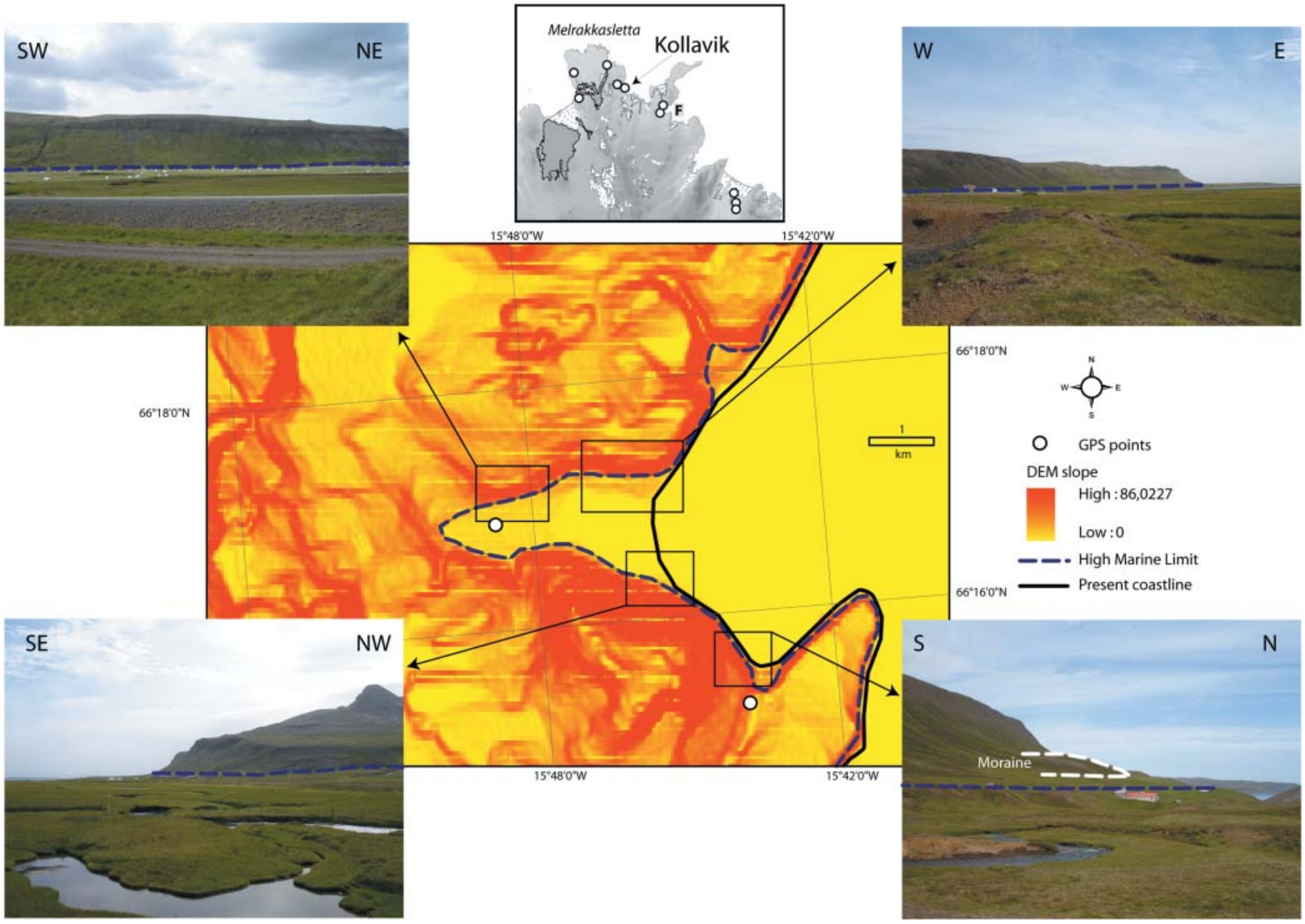

Fig. 4. Example of the marine limit extraction method from slope analysis on DEM in the Kollavik area, NE Iceland. The DEM slope is represented by a coloured gradient from yellow (low values) to red (high values). The first high difference observable in the slope (red) from the present-day shoreline was mapped (dashed blue line) and considered as a palaeo-shoreline with an elevation error of ±10 m. The line mapped numerically agrees very well with marine cliffs observed in the field (photographs of Fig. 6) as well as its elevation (DEM: *c*. 11 m; GPS: *c*. 17 m).

shallow coastal plain, which extends 30–40 km offshore to a slope break at about −22 m a.s.l.

South of the Vatnajökull ice-cap the coastal plain is covered by sandur, accumulation of fluvioglacial debris and sediments (Fig. 5, profile C). Seismic echo-soundings and seismic profiles indicate a sediment thickness of 80–100 m near the mouth of the Skeidararjökull glacier, increasing to about 250 m close to the coast (Gudmundsson *et al*. 2002). Two sandur units have been described, the first consisting of unconsolidated glaciofluvial Holocene sediments and the other of consolidated sedimentary rock of Pleistocene age. Consequently, the two limits located at +65 m and −46 m a.s.l. do not correspond to marine limits reached by the sea after the last glaciation. If we subtract the sediment thickness, the high level would be located at −35 m and the low one at −150 m. This provides a depth interval for the two limits, although it is impossible to determine the exact location of the palaeo-marine levels in this interval. This information was taken into account in the vertical motion calculation.

The determination of the slope break was also difficult in some eastern fjords. Nevertheless, in Berufjördur, the high limit is well defined and is located at +65 m a.s.l. (Fig. 5, profile B). In contrast to the Vestfirdir area in the NW, there is no extenive shallow coastal plain offshore. A slope break is observed offshore very close to the present shoreline and is very deep in some places (e.g. −104 m a.s.l. in the Berufjördur area).

Age and elevation of the palaeo-marine levels

The elevation of these two limits was determined on the DEM (±10 m). The high limit has a mean elevation of +37 m a.s.l., ranging from 0 to +225 m. The low limit has a mean elevation of −49 m a.s.l., ranging from 0 to −170 m a.s.l. (Fig. 5).

Absolute dates of lava flows and of marine shells are necessary to constrain the age of the two palaeo-marine levels. As explained above, several marine stages and glacial readvances occurred during the deglaciation. The marine level was lower during the final stage of the deglaciation (Young Preboreal) than at the beginning of the deglaciation (Bölling time). With our method we mapped the least elevated level (first slope break visible from the coastline), thus it should represent the final marine stage. Furthermore, the high marine limit we determined is sub-continuous all around Iceland, which suggests that it was formed during a single marine stage. Finally, the location of marine shells dated to Younger Dryas–Preboreal (10 000 ± 300 years BP) coincides with the location of our limit in several

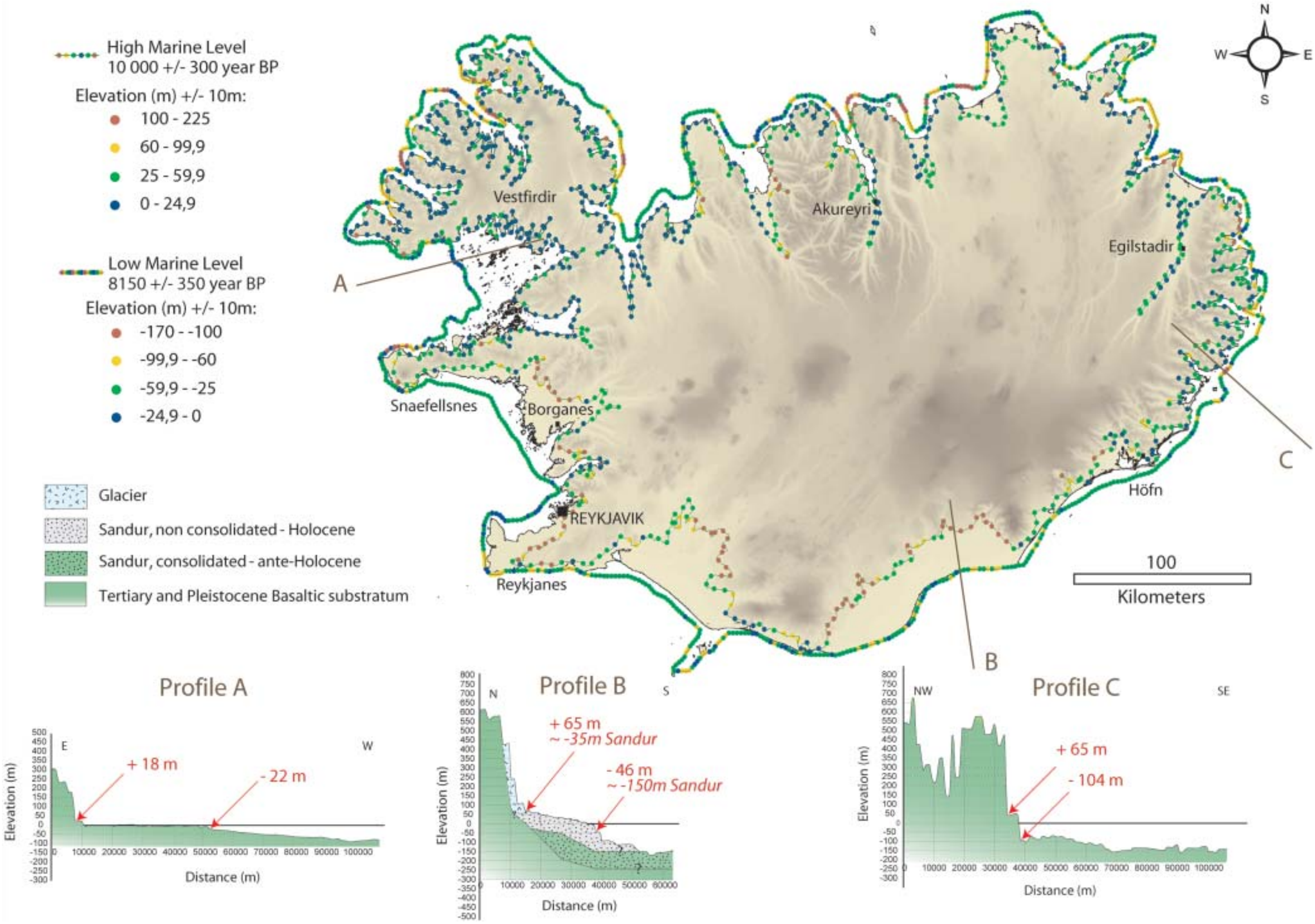

Fig. 5. Location of the two marine limits extracted from slope break analysis on DEM. Elevation and depth of these limits are indicated by coloured points. Transverse profiles show various Icelandic morphological coast types and the location of marine limits: profile A, fjords and shallow coastal plain extending offshore in the Vestfirdir area, northwestern Iceland; profile B, deposit plain (sandur) in the Vatnajökull area, southern Iceland; profile C, deeply eroded fjords in the Berufjördur area, eastern Iceland.

places around Iceland (WNW: Norddahl & Pétursson 2005; SSW: Hjartarson & Ingolfsson 1988; north: Norddahl & Pétursson 2005; NE: Pétursson 1991; east: Norddahl & Einarsson 2001; Norddahl & Pétursson 2005). We therefore assume the high marine level to be synchronous all around Iceland and dated to 10 ka ± 300 years BP, corresponding to the final deglaciation (Ingolfsson *et al*. 1995). The age of the low marine limit is difficult to constrain because of its location below the present sea level. It was formed when the subaerial Holocene lava flows located on the coastal plains were deposited. In south–central Iceland the Thjorsarhraun lava flow invaded the sea on the south coast at 7800 years BP (Hjartarson & Ingolfsson 1988), when sea level was located at about −15 m (Einarsson 1994). This lava flow covers a wide area (930 km^2) with a length of 120–140 km (Bergerat *et al*. 1998), indicating that the Icelandic inland ice sheet was vastly reduced at that time and that post-glacial glacio-isostatic uplift was completed at about 7800 ^{14}C years BP or slightly earlier (Norddahl & Einarsson 2001). Ingolfsson *et al*. (1995), Norddahl & Einarsson (2001) and Biessy *et al*. (2008) have estimated the age of relative low sea level reached after the isostatic adjustment at 8500 years BP. Therefore the low marine level that we have determined numerically was formed between 8500 and 7800 years BP when the glacio-isostatic uplift was completed, with a mean age of 8150 ± 350 years BP.

Field data analyses and correction of the previous numerical results

The coastal plain of Borgarnes, western Iceland, is formed of Tertiary basaltic flows. The morphology of this plain differs from that of the south coast, which is covered by sandur (thick accumulation of fluvioglacial debris and sediments). The coastal surface shows significant glacial erosion leading to the formation of roches moutonnées with well-developed glacial striae (Fig. 6a). Interglacial and glacial deposits locally covered this surface. The timing of these deposits suggests that this plain is old. No Weichselian moraines were observed on the plain, indicating that it was not greatly affected by the last glacial stage. However, a moraine outcrop is present on the south side of the Borgarnes fjord mouth (Fig. 6b). In some places, the plain is covered by fluvial deposits and erratic blocks resulting from jökulhaup or iceberg release (Fig. 6c). High marine cliffs delineate the plain and the fjord. A raised palaeo-marine level was observed at about 100 m a.s.l. in the north of the plain. Other raised marine terraces are located at the base of marine cliffs south of Borganes in Grjoteyri: here is a high level at about 90 m and a lower one around 60 m. This marine phase partially erased the deposits of the last glacial stage and is therefore considered as postdating the last glaciation. These observations are consistent with the descriptions of Ashwell (1975) and Ingolfsson (1987, 1988).

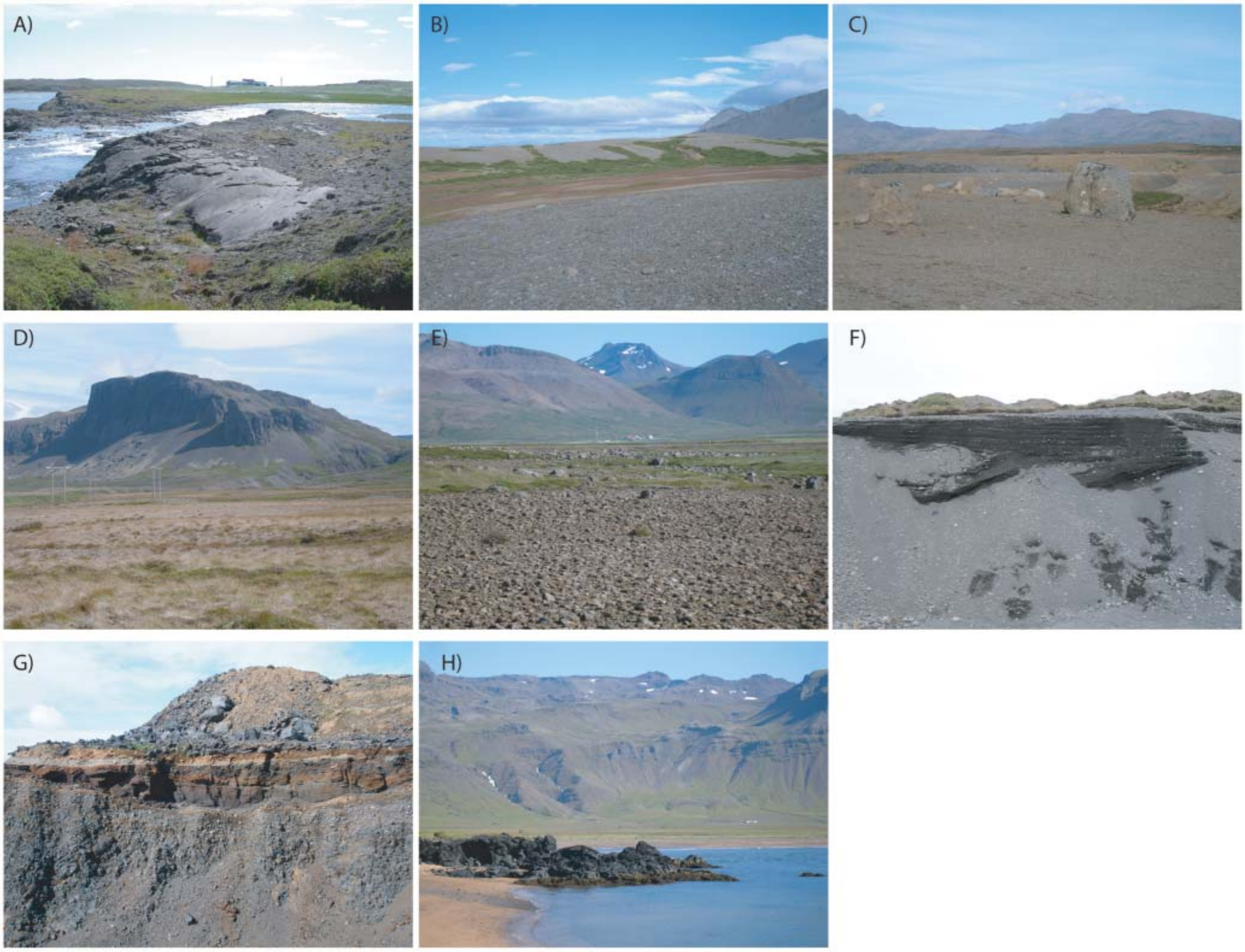

Fig. 6. Photographs from fieldwork observations (location shown in Fig. 3). (**a**) Glacial striae in the Myrar plain; (**b**) Skorholtsmelar moraine; (**c**) erratic blocks in the Myrar plain; (**d**) marine cliffs in Grjoteyri; (**e**) raised beach in Helgafellssveit area; (**f**) Langanesströnd Delta; (**g**) succession of pillow lavas, fluvial sediments and subaerial Holocene lava flow close to Grindavik; (**h**) partially drowned Budir lava flow.

Raised marine terraces, beaches and cliffs were observed on the Icelandic coast in all our other fieldwork areas (Fig. 6d–f). These marine markers seem to have not been affected by the last glacial stage. These features are evidence of a relative palaeo sea-level rise postdating the last glaciation, as in the Borganes area.

A succession of pillow lavas, fluvial sediments and subaerial Holocene lava flow was observed on the Reykjanes peninsula, southern Iceland, close to Grindavik (Fig. 6g). This sequence indicates a relative sea-level fall in this area. Furthermore, Holocene lava flows cover the coastal plain of Myrar, western Iceland, and some of them, such as the Eldborg and Budir lava flows, are partially drowned by the sea (Fig. 6h). Their location on the coastal plain reveals that the sea level was lower than the present-day sea level when they were deposited and their partial drowning indicates that the sea level has been rising since then. Icelandic coastal morphology is mainly characterized by marine cliffs formed after the last glaciation when the relative sea level rose. Palaeo-marine levels are represented by raised marine terraces at the base of the marine cliffs; these areas are generally used for farming.

Morphological observations and GPS profiles in the Borgarnes area, western Iceland, are consistent with the descriptions of Ashwell (1975) and Ingolfsson (1987, 1988). Considering these data, it is possible to produce a chronology of the marine limits formed during the last deglaciation in this area (Fig. 7). Ashwell (1975) described an ice-float environment with a high marine level (80–100 m) dated to 13–11.7 ka BP (Bölling–Older Dryas) (Fig. 7a). This environment is compatible with erratic blocks and the high marine level located at 100 m observed on the Myrar plain. Ingolfsson (1987, 1988) described the successive glacial advances in the area around Borgarfjördur and their concurrent sea-level changes. He described a glacial advance during the Older Dryas (12–11.7 ka BP) with a relative marine level at 80–90 m a.s.l. corresponding to the previous description of Ashwell (1975) and to the high marine level observed in Grjoteyri (*c.* 90 m). The moraine in Skorholtsmelar south of Borgarfjördur characterizes the latest glacial advance in the Younger Dryas. Ingolfsson (1987, 1988) also described a marine limit at 60–70 m a.s.l. corresponding to the final glacial retreat at about 10 ka ± 300 years BP. It is consistent

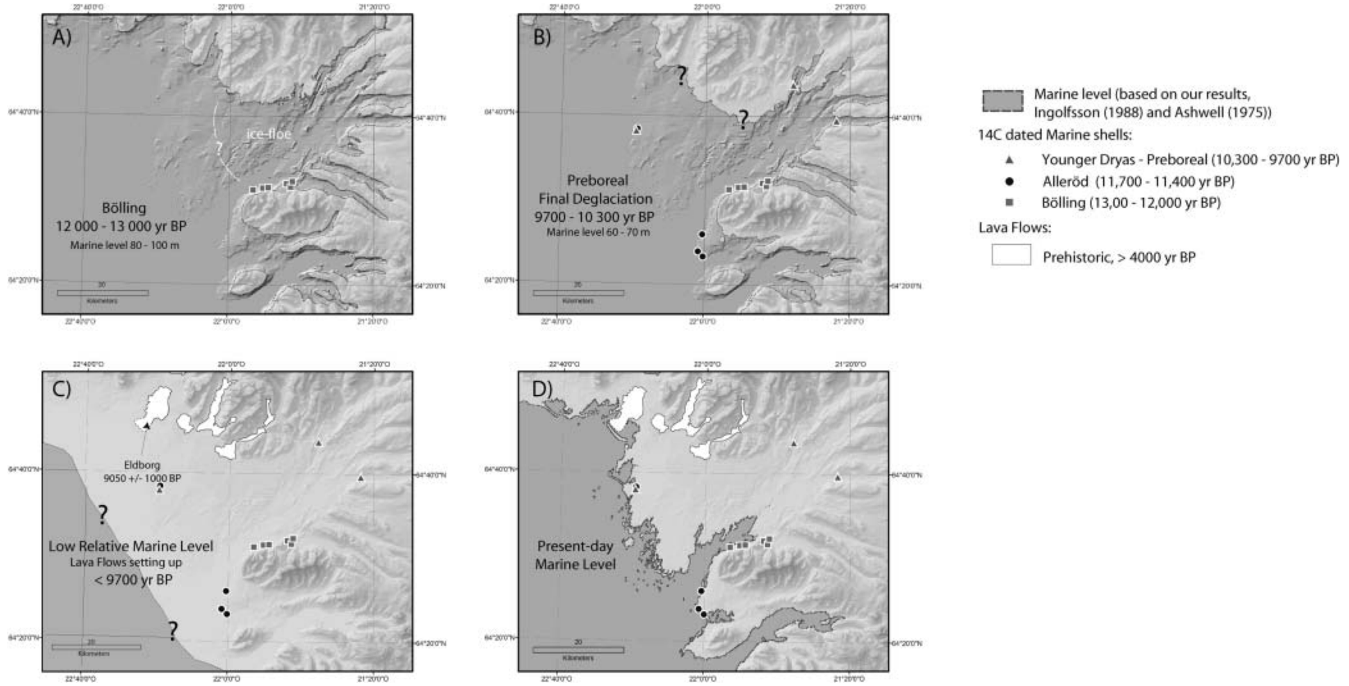

Fig. 7. Sea-level evolution during deglaciation in the Borgarnes–Myrar area, western Iceland: (**a**) in Bölling time (12–13 ka BP); (**b**) in Preboreal time (9700 years to 10.3 ka BP); (**c**) after the isostatic recovery, low relative sea level (>9700 years BP); (**d**) present-day shoreline. (Note that during the low sea level the area of the coastal plain had greatly increased.)

with our field observations and the low marine level observed in Grjoteyri (*c.* 60 m) (Fig. 7b). In the Myrar plain, only the high level (*c.* 100 m) is observed. This observation suggests that the post-glacial vertical motion has not been the same in the north of the Myrar plain as in the Borgafjördur area. The coastal plain extends offshore and islands are visible at 6–7 km from the present coastline, suggesting that the sea level was once lower than at present. Furthermore, the partially drowned subaerial lava flow at Eldborg, NW of the plain, indicates that the sea level was lower at the time of its formation (Fig. 7c). This lava flow is dated to 9050 ± 1000 years BP (Global Volcanism Program, www.volcano.si.edu). Since then the sea level has risen to its present level (Fig. 7d).

The high marine limit determined on the DEM corresponds to the 60–70 m high limit in Borgarfjördur and the 100 m high limit north of the Myrar coastal plain described above. This validates our numerical method for this area. To estimate the accuracy of the method in other areas, we measured by GPS the elevation of raised marine features observed in different places around Iceland, and compared these elevations with those obtained on the DEM. The majority of the GPS points are located on the high limit with an elevation difference ranging from 1 to 16 m, which supports the method we used. Some areas were difficult to interpret, such as the south of the Vatnajökull ice-cap and Oxarfjördur, north Iceland, because of sandur and recent jökulhaup, as mentioned above. Also, it was difficult to map slope breaks in flat areas such as the north of the Melrakkasletta peninsula, NE Iceland. Thus a correction of the high marine level elevation and location has been made before the calculation of the vertical displacement. The corrected line is shown in Figure 8. Other marine features were observed on the south side of the Snaefellsnes peninsula at a lower elevation but no data were available to date them.

It was not possible to observe the low marine limit because of its offshore location, but we consider that if the method is appropriate for the high marine level, it can be applied also for the low limit. Furthermore, the location and the depth of the numerical low limit are consistent with the observations on seismic profiles from Faxafloi Bay, SW Iceland, by Thors & Helgadottir (1991) (30–35 m on seismic profile, 30 m on DEM).

Vertical motions of Iceland were calculated (see equation (1)) for two periods: first between 10 ka ± 300 years BP and 8150 ± 350 years BP and second since 8150 years BP. Referring to the SPECMAP curve, the eustatic variations during these two time intervals are +18 m between 10 ka and 8150 years BP, and +36 m since 8150 years BP.

Uplift calculated between 10 ka and 8150 years BP

Vertical motions between 10 ka ± 300 years BP and 8150 ± 350 years BP indicate an uplift stage of Iceland (Fig. 8). The uplift amount varies from 40 to 170 m with a mean uplift amount of 102.5 ± 10 m. The rate of uplift is between 2.1 and 9.2 cm a^{-1} with an average of 5.5 ± 2.2 cm a^{-1}.

The histograms in Figure 8 show the distribution of uplift values. Three main zones of uplift amount can be distinguished: (1) low and slow uplift, from 40 to 80 m corresponding to 2.1–4.3 cm a^{-1}, mainly located in the NNW; (2) intermediate amount and rate of uplift, from 80 to 120 m corresponding to 4.3–6.4 cm a^{-1}, essentially located in the ESE; (3) high and fast uplift, from 120 m to 170 m corresponding to 6.4–9.2 cm a^{-1}, mainly located in the South Iceland Seismic Zone, SSW Iceland, between the West Volcanic Zone and East Volcanic Zone and on the SE of the Snaefellsnes peninsula, western Iceland.

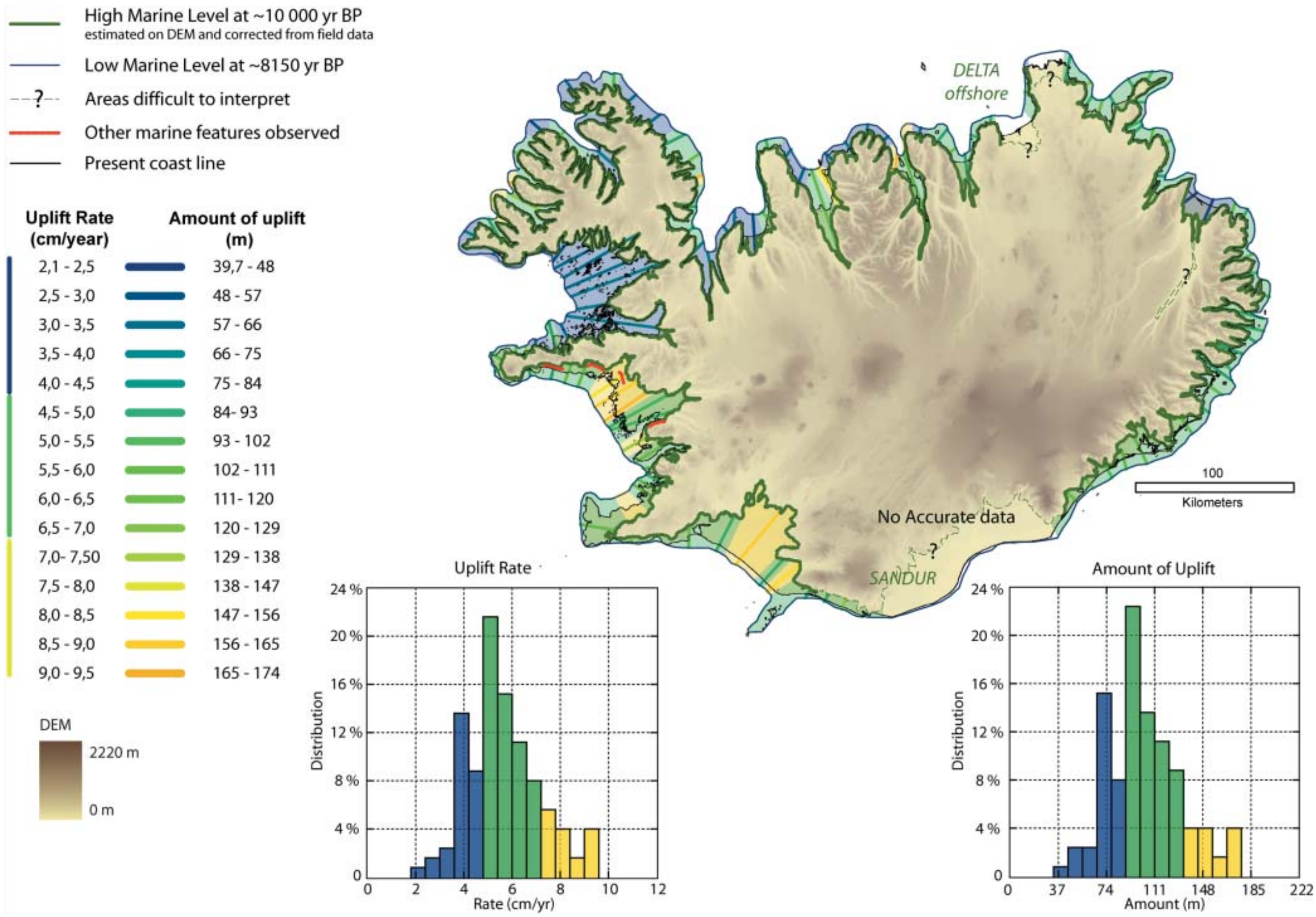

Fig. 8. Amount and rate of uplift calculated between 10 ka ± 300 years BP and 8150 ± 350 years BP. Eustatic variations between 10 ka and 8150 years BP were taken into account for the calculation and equal 18 m (SPECMAP curve). The high marine limit determined numerically was corrected from field observations (green bold line).

Vertical motion since 8150 years BP

The amount of vertical displacements of Iceland calculated between the low limit and the present coastline ranges from −136 m to +36 m in a period of 8150 years, providing rates of 0–1.5 cm a^{-1}. Approximately 46% of the values are between −136 m and −10 m, 40% between −10 m and +10 m, and 14% between +10 m and +36 m. Those between −10 m and +10 m are considered as null given that the error of the method is ±10 m. This means that most of the Icelandic coast (46%) has subsided (negative values), 40% has been stable and 14% has been uplifted at a slower rate than the eustatic rise (Fig. 9). However, these values are scattered around Iceland and no clear zones of subsidence or uplift can be distinguished.

Discussion

Validity of the results and the method

The results of uplift obtained between 10 ka ± 300 years BP and 8150 ± 350 years BP correspond to the post-glacial isostatic readjustment of Iceland following the Weichselian glaciation. Other processes such as rifting, hotspot bulge, etc. might play a part in the finite vertical motion, but the uplift rates calculated (>2 cm a^{-1}) are higher than the characteristic rates of these processes and are more consistent with isostatic rebound (Fig. 2), taking into account that they lasted some thousand years.

In the SSW, our results are very close to those calculated by Biessy *et al*. (2008): (1) 118 m against 108 m according to Biessy *et al*. (2008) in the Reykjanes peninsula segment; (2) 140 m for both studies in the Hveragerdi–Hvolsvollur segment; (3) 107 m against 100 m according to Biessy *et al*. (2008) in the Vik segment. On the other hand, the location of the palaeo-coastline differs, but the limit was difficult to map in this area because of the lack of large slope break as markers. Therefore the calculated uplift values are very satisfactory but the palaeo-coastline location is debatable. The uplift rates are slightly different because we did not use the same age for the Thjorsarhraun lava flow, as mentioned above (7800 years BP in this study against 8500 years BP according to Biessy *et al*. 2008).

In the Berufjördur area, Norddahl & Einarsson (2001) estimated by extrapolation an uplift rate of 7 cm a^{-1} between 10.3 ka and 9900 years BP. Ingolfsson *et al*. (1995) calculated an uplift rate of 6.9 cm a^{-1} in the Reykjavik area between 10.3 ka and 9400 years BP. These two uplift rates are slightly higher than those we determined in the same areas (5.7 and 6.1 cm a^{-1}, respectively). This difference can be explained by the fact that the ages used in the calculation were not the same. Our results represent an average of the post-glacial uplift between 10 ka and 8150 years BP. An isostatic readjustment is more important at the beginning of the unloading and decreases exponentially with

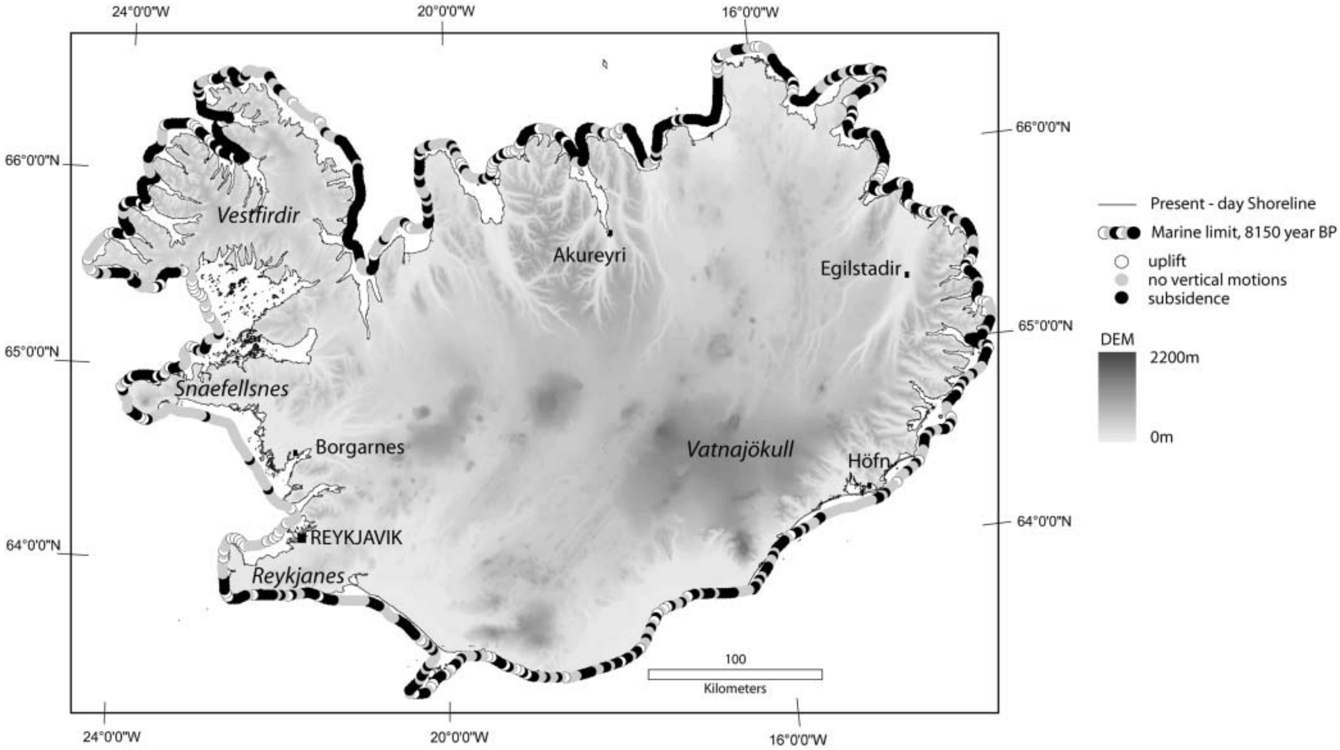

Fig. 9. Calculated vertical motion since 8150 years BP. Approximately 46% of the values are between −136 m and −10 m (blue points), 40% between −10 m and +10 m (green points), and 14% between +10 m and +36 m (red points). Those between −10 m and +10 m are considered as null (error is ±10 m). Positive values indicate uplift and negative values indicate subsidence.

time (Turcotte & Schubert 2002). Thus we suggest that the higher rates calculated by Norddahl & Einarsson (2001) and Ingolfsson *et al*. (1995) correspond to the fast rebound immediately following the deglaciation and that our values represent an estimation of the global post-glacial rebound for a longer period, and can be consider as a 'minimum' rate of the isostatic readjustment.

These results were obtained with a method based on the slope analysis of a DEM. This method is thus difficult to apply in flat areas or sedimentary plains; for example, south of Vatnajökull. However, the high limit determined by this method is well constrained by field observations and previous work, especially in the WSW and the eastern fjords area. Thus, it might be valid also in other areas such as NNW Iceland where the morphology is similar. This study is devoted to a global synthesis of the vertical motions that Iceland has undergone during Holocene time. Consequently, it may result in some differences at a very local scale.

Stages of vertical motion of Iceland

Two stages of vertical motion are determined. The first stage corresponds to the post-glacial rebound of Iceland between 10 ka ± 300 years BP and 8150 ± 350 years BP as described above. On average, Iceland underwent an uplift of 102.5 ± 10 m between 10 ka ± 300 years BP and 8150 ± 350 years BP, implying rates of 5.5 ± 2.2 cm a^{-1}. These values are higher than those described for other glacial areas; for example, 3.8–4.8 cm a^{-1} in the Hitra and Bjugn area of central Norway (Kjemperud 1986) and 3.3 cm a^{-1} on the east coast of Greenland (Björck *et al*. 1994). The position of Iceland in the middle of the North Atlantic Ocean makes the extent of glaciers in Iceland highly sensitive to atmospheric and oceanic temperatures. These high uplift rates are compatible with a fast glacial retreat as a result of the abrupt end of the cold Younger Dryas environmental conditions revealed by studies of Greenland ice cores (Alley *et al*. 1993; Dansgaard *et al*. 1993) and high-resolution North Atlantic deep-sea cores (Lehman & Keigwin 1992).

The second stage corresponds to the vertical motion of Iceland since 8150 years BP, once the isostatic rebound was completed. This vertical displacement can be related to (1) vertical displacements caused by the advance of the ice sheet during the little Ice Age (Geirsdottir *et al*. 2009), (2) rifting processes, or (3) thermal contraction of the crust. Norddahl & Einarsson (2001) suggested that the continuous cooling and thermal contraction of the bedrock in areas outside the volcanic zones has increased the density of the rocks, which may have led to isostatic readjustment and subsidence of the crust. Our results show mainly negative values, indicating subsidence, which may be in favour of this hypothesis. However, as has been mentioned above, no significant zone of subsidence is visible. Thus it is not possible to propose any correlation with geodynamic structures (rift, thermal anomalies or eruptive centres). Furthermore, these results depend only on the bathymetry DEM, which has a very low spatial resolution (2.5 km), and consequently have to be used with caution.

Ice thickness estimation

From the classical Airy isostatic model (compensation level in depth), it is possible to estimate the past ice thickness on the Icelandic coasts from the calculated uplift amount. With a mantle density of 3.1 kg dm^{-3} (ρ_m), an ice density of 1 kg dm^{-3} and an uplift of 40–170 m (h_{uplift}), we obtain values of ice thickness varying from 124 to 527 m ($h_{ice} = ((\rho_m \times h_{uplift})/\rho_{ice})$). These values are lower than but nevertheless consistent with previous estimations; for example, from 1000–1500 m maximum above the Vatnajökull area to 300–500 m along the coasts (Einarsson & Albertsson 1988; Norddahl 1990, 1991). Furthermore, our results might correspond to a minimum estimation of the rebound and therefore a minimum estimation of the ice thickness. Our results show that the Icelandic coasts were uplifted and ice covered from 100 to 500 m, which favours models of wide glacial extension over the whole island during the Last Glacial Maximum.

Origin of the spatial variations of the post-glacial rebound

Three main zones with different uplift rates can be distinguished: the NNW with a low uplift rate, the ESE with an intermediate rate, and the WSW with a high uplift rate.

The NNW area is characterized by low uplift rates (40–80 m; 2.1–4.3 cm a^{-1}). These low values can be explained by a thinner ice-cap on the Vestfirdir peninsula than on the mainland, and by a faster deglaciation (Hansom & Briggs 1991).

The ESE coast displays homogeneous uplift values (105 m and 5.6 cm a^{-1} on average). The ice thickness during the Weichselian glaciation was maximal in this area but we measured an intermediate amount of uplift here. However, this area was continuously and is still ice covered; if the deglaciation had been total, the amount of post-glacial rebound would have been much higher than the one we have measured.

High values of uplift rate are obtained in SSW Iceland between the two volcanic rift zones (the East Volcanic Zone and West Volcanic Zone), above the South Iceland Seismic Zone and in the Myrar plain. The rest of the high values are distributed around Iceland (scattered). The high rates in the north of the Myrar plain do not fit with any active rift zone but it is possible that the higher marine level mapped in this area could be an older marine level (perhaps of Bölling time) as observed in the surroundings (Grjoteyri, Borganes area). It is possible that the marine level at the final deglaciation time has not been marked because of the flat morphology of the plain. In this case the uplift would be less important in this area and more consistent with the values of the nearby areas. Hence a relation between high uplift rates and an active rift zone (South Iceland Seismic Zone) could be suggested. No high rates are found in the North Volcanic Zone but the calculation was difficult in the Oxarfjördur area because of jökulhaup in the great Jökulsa canyon.

It appears clearly that the history and timing of the deglaciation is the principal control on the post-glacial rebound amplitude, by applying a load during a more or less long period. This explains the main spatial variations of the rebound. The geodynamic context (rift) has only a local influence; for instance, in the southern area.

Relaxation time

From uplift data obtained on radiocarbon dated marine shells and dated lava flow, an uplift–time curve was drawn. Only two areas present enough data: the Borgarnes–Myrar area, western Iceland and the Hveragerdi–Hvolsvollur area, SSW Iceland located on the rift zone (Fig. 10). Uplift decreases exponentially with time according to the equation $\omega = \omega_m e^{-t/\tau}$, where ω is uplift, ω_m is uplift at $t = 0$, t is time, and τ is relaxation time (Watts 2001). The relaxation time can be calculated from this exponential function: $y = ae^{-bx}$ with $t = x$ and $\tau = 1/b$. Thus we obtain $\tau = 4167$ years for the western area (with a minimum of $\tau = 3733$ years and maximum of $\tau = 4725$ years) and $\tau = 2000$ years for the SSW area.

These values are of the same magnitude as but generally lower than those obtained in other glacial areas: 2000–3000 years for James Bay, North America (Mitrovica *et al*. 2000; Mitrovica & Forte 2004), 4000–7000 years for Richmond Gulf (Mitrovica *et al*. 2000; Mitrovica & Forte 2004), 4000–6000 years for Angerman River, Sweden (Turcotte & Schubert 2002; Mitrovica & Forte 2004), 1700 years in North America and 8000 years in Fennoscandia (Watts 2001). The relaxation time depends on several parameters such as the asthenospheric viscosity (μ), rock density (ρ), gravity (g) and wavelength of deformation (λ), as described by the equation

$$\tau = \frac{4\pi\mu}{\rho g \lambda} \tag{2}$$

of Turcotte & Schubert (2002). The relaxation time calculated in Iceland is generally lower than that in North America or Fennoscandia but the wavelength of deformation is also much lower.

Locally, the relaxation time is twice as fast in the southern area than in the west (2000 years v. 4725 years). The main parameters that control the relaxation time are the viscosity and the wavelength of deformation (see equation (2)). The viscosity varies with the temperature and the chemical composition. The wavelength of deformation depends directly on the lithospheric thickness, which is defined by the depth of the 1200 °C isotherm. In our calculation we consider the viscosity at the lithosphere–asthenosphere interface, so the temperature is the same for both areas (1200 °C). The Moho depth is similar below the two areas (20–25 km) (Darbyshire *et al*. 2000), thus the chemical composition and therefore the viscosity must also be similar. The Icelandic lithosphere is generally shallow along the axis of the oceanic ridges and deepens with distance from this axis (Kaban *et al*. 2002). Locally it is thicker below the SSW area (40–50 km) than below the western area (<40 km). A greater lithospheric thickness increases the deformation wavelength and thus reduces the relaxation time. Therefore the difference we obtained between the two relaxation times is due mainly to a variation of the lithospheric thickness rather than a variation of viscosity. The difference of lithospheric thickness is clearly due to the rift (Kaban *et al*. 2002), but curiously the thinnest lithosphere is found in the western zone, west of the West Volcanic Zone below the extinct spreading axis (Snaefellsnes Volcanic Zone), and not in the SSW area located between the West Volcanic Zone and East Volcanic Zone on the South Iceland Seismic Zone.

We estimated viscosities from the relaxation time with equation (2), using values of 3100 kg m^{-3} for mantle density (ρ) (Kaban *et al*. 2002), 9.81 m s^{-2} for gravity and different wavelengths for each area (90 km for the west and 140 km for the SSW, both corresponding to approximately three times the thickness of the lithosphere). Viscosity estimates are 2.1×10^{19} Pa.s in the SSW and 2.8×10^{19} Pa.s (2.5×10^{19} Pa s minimum to 3.2×0^{19} Pa s maximum) in the west. These estimates are close to those found by Sigmundsson (1991) (1×10^{19} Pa s), Sigmundsson & Einarsson (1992) (1×10^{18} to 5×10^{19} Pa s) and LaFemina *et al*. (2005) (4×10^{19} Pa s). They are slightly higher than those determined by, for example, Pagli *et al*. (2007) from present-day glacio-isostatic deformation in the SE of Iceland around the Vatnajökull ((8–10) $\times 10^{18}$ Pa s). However, in this area, the Moho is deeper

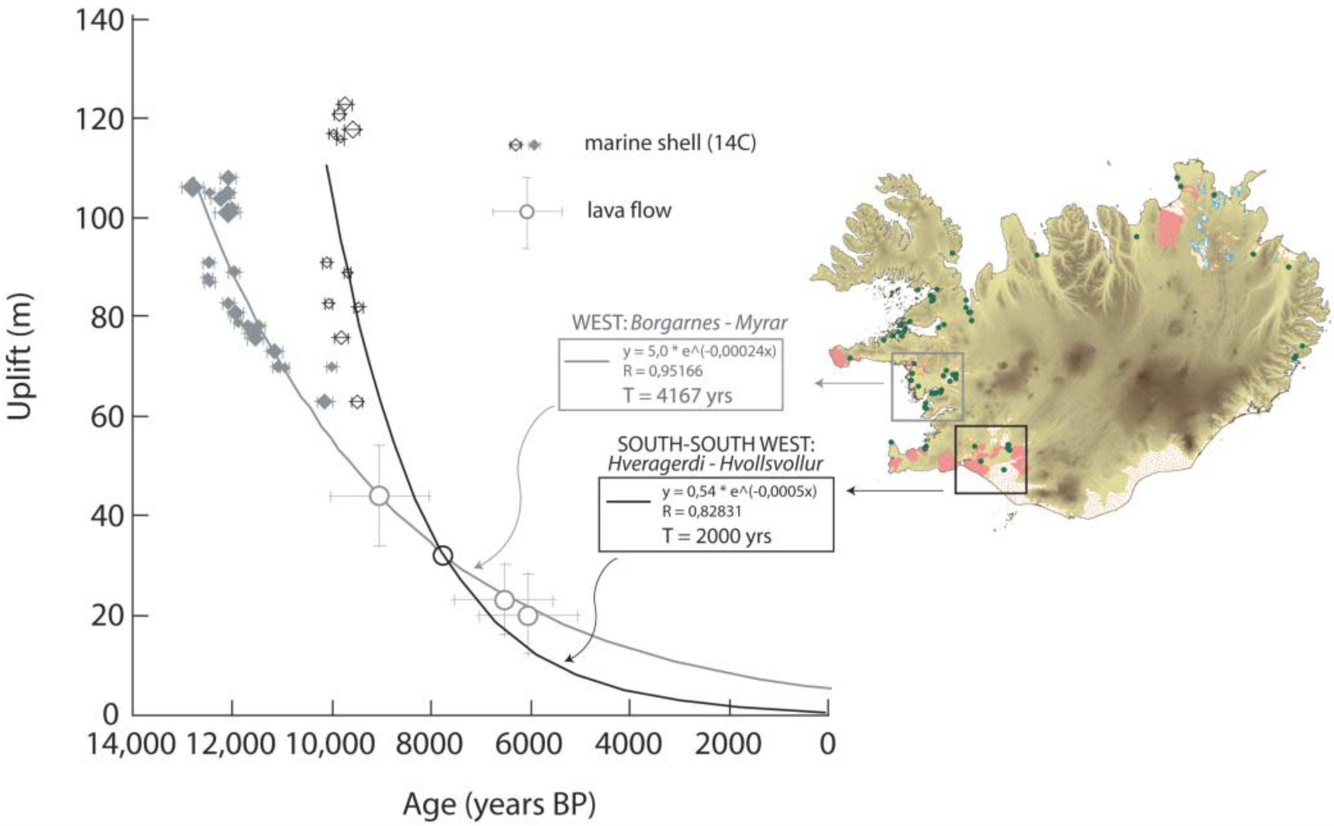

Fig. 10. Evolution of the uplift v. time from uplift data in the Borgarnes–Myrar area, western Iceland (external to the rift zone) and in the Hveragerdi–Hvolsvollur area, SSW Iceland (located on the South Iceland Seismic Zone), based on ^{14}C dated marine shells and drowned Holocene lava flows (SSW: Thjorsarhraun 7800 years BP (Hjartarson & Ingolfsson 1988; Norddahl & Einarsson 2001); west: Eldborg 9050 ± 1000 years BP, Budaklettur 6550 ± 1000 years BP and Ondverdarnesholar 6050 ± 1000 years BP (Global Volcanism Program, www.volcano.si.edu).

(25–35 km) than in the WSW, because of its location above the mantle plume centre.

Consequently, with this post-glacial study it is possible to illustrate the impact of the geodynamic context on the viscoelastic response of the lithosphere to ice unloading.

Conclusions

The aim of this work was to constrain the Holocene deformation affecting the Icelandic coast. From our study, we draw the following conclusions.

(1) From the palaeo-shoreline numerical extraction method, corrected from field observations, it is possible to establish a synthesis of vertical motions of Iceland during Holocene time.

(2) Two stages of vertical motion are determined: (a) a major uplift stage between 10 ka ± 300 years BP and 8150 ± 350 years BP corresponding to the post-glacial isostatic rebound of Iceland following the Weichselian glaciation; (b) a second stage of vertical displacement since 8150 years BP, mainly of subsidence, fluctuating between −136 m and +36 m, implying rates from 0 to 1.5 cm a^{-1}.

(3) The uplift amount ranges between 40 and 170 m (mean 102 m ± 10 m) leading to an uplift rate varying from 2.1 to 9.2 cm a^{-1} (mean 5.5 ± 2.2 cm a^{-1}). These values are higher than for other glacial areas. Our calculation is an average estimation of the total rebound. The rebound must have been even quicker at the beginning of the deglaciation and decreased exponentially. These data provide an estimation of the ice thickness of between 124 and 527 m along the coast.

(4) The post-glacial rebound shows spatial variations. Generally, there is an increase of uplift rate from NNW to SSE. The low uplift rate of the NNW is explained by a thinner ice-cap and a faster deglaciation than in the rest of the island. The eastern coast shows intermediate uplift but the deglaciation was not total in this area, which was continuously covered by an ice-cap. High uplift rates are visible above the southern rift zone, explained by the influence of rifting on the lithosphere.

(5) Relaxation times, calculated from uplift data for two areas, are $\tau = 4167$ years in the west and $\tau = 2000$ years in the SSW. This local difference indicates that rebound relaxation following unloading is a complex process, which is not homogeneous and can vary over short distances. Viscosity estimates from these relaxation times range from 2.1×10^{19} Pa s to 3.2×10^{19} Pa s. The difference in relaxation time is therefore due to a difference in lithospheric thickness below these two areas and not to a variation of viscosity, and shows the influence of the rift on the viscoelastic response of the lithosphere to ice unloading.

We demonstrate with this study that it is possible to combine a traditional Quaternary stratigraphy approach and a digital approach to obtain an overview of the post-glacial rebound of Iceland during Holocene time. This study shows the importance of glacial isostasy as a process generating vertical deformation in comparison with tectonic or magmatic processes.

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