CRUSTAL MOVEMENTS AND RELATIVE SEA LEVEL CHANGES IN ICELAND

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INTRODUCTION

Relative sea level at a given location is influenced by a multitude of processes. They may be of oceanographic, meteorological or geological nature, and be of local origin or global extent. On a global scale sea level is dependent on the mass and volume of the sea water, and the volume of the basins that contain it. On a more local scale the relative sea level is influenced by ocean currents, wind pattern, and crustal movements. Crustal movements and the configuration and movements of lithospheric plates can also affect the volume of the oceanic container and thus cause global sea level changes. Major isostatic adjustments, like for example the ones following the last glacial period, may also lead to sea level changes on a global scale. In this paper a survey is made of crustal movements in Iceland that can affect relative sea level. An attempt is made to explain the causative processes, and relate them to the geological history of the area.

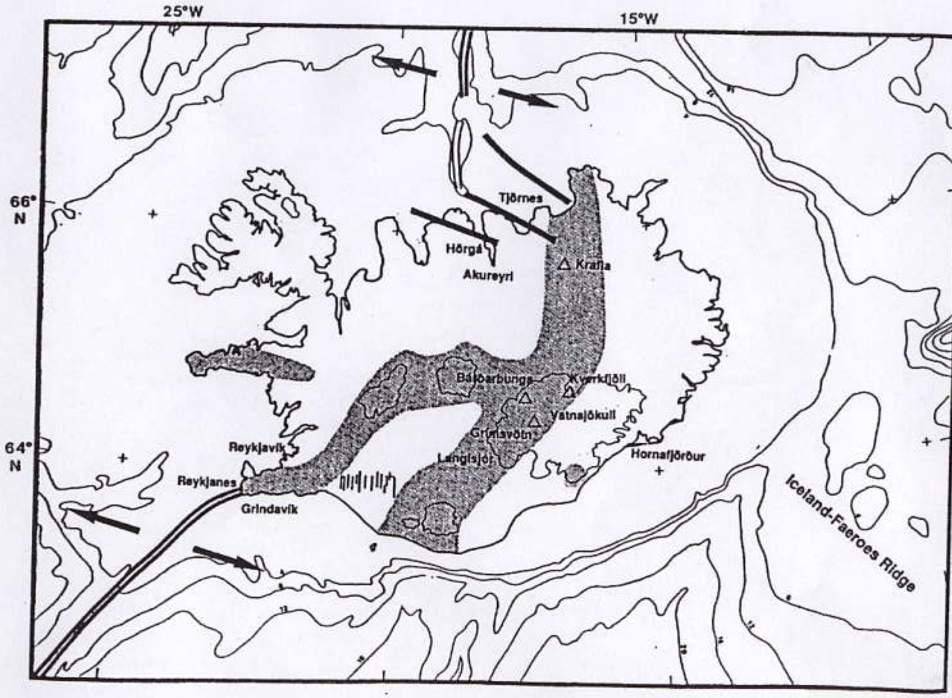

Figure 1. *Map of Iceland and bathymetry of the surrounding ocean. Plate boundaries are shown, with arrows indicating the direction of relative plate movements of the North American and Eurasian Plates. The shaded area marks the neovolcanic zones. Bathymetric contours are given every 400 m. The central area of the Iceland mantle plume is in the Bárðarbunga-Kverkfjöll-Grímsvötn region in east-central Iceland.*

Geologically, Iceland is located in a very dynamic environment. The country owes its existence to a major plume of upwelling mantle material. As the mantle material moves upwards its melting temperature is lowered. Partial melting occurs, and the melt is transported towards the surface to feed volcanism and form crust that is thicker than normal for the oceanic areas. The plume and the thick crust keep the country above sea level. In addition, Iceland is located on the boundary between two of the largest lithosphere plates, the North American Plate in the west and the Eurasian Plate to the east. The boundary crosses the island and is a source of volcanism and seismic activity (Fig. 1). New lithosphere is formed at the divergent plate boundary as the plates separate. These processes are accompanied by considerable vertical crustal movements. Variations in the intensity of the plume activity will lead to vertical movements, the cooling of the lithospheric plates with age causes subsidence, loading of the volcanic zones by new volcanic material added in the crust and at the surface ia accompanied by sinking. In addition, the crust responds to changing glacier load. Past, as well as current changes in glaciation lead to measurable crustal movements and sea level variations.

EARLY OPENING OF THE NE ATLANTIC AND THE BEGINNING OF ICELAND

The geological history of Iceland may be traced back about 60 million years, to the time when the Iceland mantle plume emerged and apparently intiated rifting of Greenland away from Norway. Since then the plume has seriously affected the history of plate movements in the North Atlantic. Greenland became attached to the Nort American Plate and has been moving away from the Eurasian Plate at a rate of about 2 cm per year. It is generally thought that a plume of rising material in the mantle forms a plume head. When the head arrives at the surface it leads to crustal

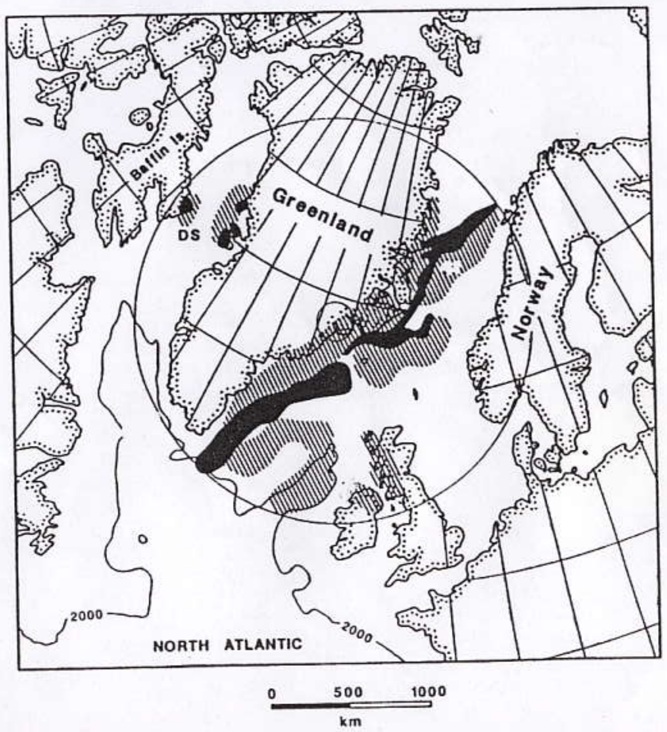

Figure 2. *Reconstruction of the North Atlantic region just after the onset of oceanic spreading between Greenland and Norway. Shading and hatching show the areas of volcanism associated with this event, and the small circle in E-Greenland shows the inferred position of the center of the plume at this time (from White and McKenzie, 1989).*

uplift and extensive and widespread volcanism (White and McKenzie, 1989). Remnants of this stage of the Iceland plume are found all around the continental margins in the NE-Atlantic, centered on the assumed plume center beneath E-Greenland (Fig. 2). After the initial intensive phase the volcanism continues above the plume, at a reduced rate, and leaves a trail of thickened crust on the plate above. The location and shape of the trail depends on the relative movement of the plate with respect to the plume. Iceland, the Iceland-Faeroes Ridge and the Iceland-Greenland Ridge are the expressions of this trail. The present location of the plume center is presumed to be beneath east-central Iceland, in the area of the subglacial volcanoes Bárðarbunga, Grímsvötn and Kverkfjöll (Fig. 1).

SUBSIDENCE AND LITHOSPHERIC COOLING

The lithosphere is formed at the divergent plate boundary by injection and extrusion of magma from the mantle. It then drifts away from the plate boundary and cools as it becomes older. The cooling leads to contraction and subsidence as the lithosphere adjusts isostatically with the asthenosphere below, and therefore a general relationship exists between water depth and the age of the lithosphere. A depth-age curve for the oceanic lithosphere is shown in Fig. 3. The crustal

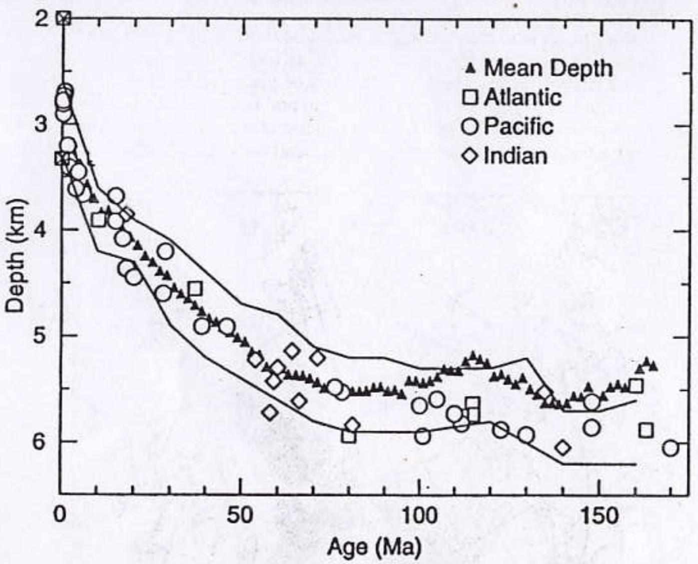

Figure 3. *Summary of global depth and age data from the ocean floors, demonstrating the subsidence of the oceanic crust in response to the cooling of the lithosphere with age. Modified from Carlson and Johnson (1994).*

subsidence during the first 10 million years is about 1000 m, and further 1700 m in the subsequent 50 million years. Close to the plate boundary the sea level can be expected to change by 0.1 mm per year due to this process. Although this rate hardly makes much difference in sea level on the human time scale of 1-1000 years, it certainly is of considerable importance on the geological time scale of 10^5 - 10^7 years. Thus the Iceland- Faroes Ridge (Fig. 1), now below present sea level, was formed by subaerial volcanism due to the Iceland plume. As this piece of crust drifted away from the plate boundary the lithosphere cooled and subsided. Parts of the ridge sank below sea level in the early Miocene (about 20 million years ago) and the highest parts no later than the Pliocene (about 8 million year ago) (Thiede and Eldholm, 1982).

UPLIFT DUE TO EROSION

Erosion of the land surface moves mass to the ocean, to be deposited at the ocean bottom. This mass removal leads to uplift of the crust by isostatic adjustment. Walker (1982) estimates that about 1000 m have been eroded off the 3-14.5 million years old basalt pile in Eastern Iceland. A large part of theis erosion is doubtless associated with the glacial period, i.e. the last 3 million years. Assuming crustal and mantle densities of 2600 kg/m^3 and 3000 kg/m^3, respectively, an isostatic adjustment leads to uplift of 870 m. Steinthórsson (1987) estimates the present erosion rates to be in the range 0.01-0.025 m^3 per year. Similar numbers are given by Gíslason and Arnórsson (1988) who include chemical denudation in their estimate. This corresponds to lowering of the land surface by 0.1-0.25 mm per year. Uplift due to isostatic adjustment would be in the range 0.09-0.22 mm per year. The effects of this process on present-time relative sea level in Iceland are thus of the same order of magnitude as that of lithosphere cooling, but with the opposite sign. They would tend to cancel each other.

PROCESSES AT THE PLATE BOUNDARY

The divergent Mid-Atlantic plate boundary crosses Iceland. The process of crustal accretion affects the plate boundary region, and is accompanied by substantial vertical movements of the crust. Relative sea level changes will result in the two areas where the plate boundary crosses the coast, i.e. in the SW corner of the country and in NE-Iceland (Fig. 1). These changes may be of considerable economic importance, since the city of Reykjavík and the towns of Hafnarfjörður, Vogar, Njarðvík, Keflavík, Sandgerði, Grindavík Þorlákshöfn and Kópasker are within the affected areas. At least three processes can be identified:

1. Volcanic loading of the neovolcanic zone. The plate divergence is accompanied by volcanism. New volcanic material is transported from depth and deposited at the surface or at shallow levels in the crust. This added surface load is accomodated by isostatic adjustment and subsidence. If a steady-state is assumed, the subsidence is equivalent to the average production rate of volcanic material. A model of this process developed by Pálmason (1973, 1980, 1986) assumes a production rate of $1.3 * 10^{-4}$ km^2/a per km length of the plate boundary, and derives a subsidence rate of a few mm per year, depending on the distribution of the load. This rate is consistent with drill hole data at Reykjanes. This subsidence rate is a long term average. On a shorter time scale the rate is not supposed to be constant. The volcanic production takes place in discrete events, but viscous layers at subcrustal depths respond slowly to the added crustal load. The relaxation time is of the order of hundreds of years. The subsidence is therefore expected to take place mostly in the first few hundred years after a volcanic event, and then gradually slow down. The Pálmason model assumes a direct relationship between loading and subsidence and does not take the finite strength of the plates. In reality the plates have a finite flexural rigidity and subsidence will occur also outside the areas of lava accumulation.

2. Elastic, continuous stretching of the plate boundary region. The divergent movements of the large lithospheric plates on either side of the boundary is more or less continuous. Therefore the crust in the plate boundary region between them is being continuously stretched. Its surface subsides due to elastic necking. The subsidence could amount to a few mm per year. This deformation is elastic and is recovered during rifting events when the crust fails along the plate boundary. On a short time scale this process may be difficult to separate from the loading process described above. Both lead to subsidence within the same areas and of the same order of magnitude. On a long time scale, however, they can be separated, since only the loading process contributes to the long time average subsidence. Continuing subsidence of the plate boundary region of several mm per year has been deduced from tilt measurements on the rift zone flanks by Tryggvason (e.g. 1974, 1982).

3. Rifting events. The crust in the plate boundary regions failes when the tensional stress exceeds the crustal strength, or when the strength is lowered due to e.g. inflow of magma in the crust. Failure occurs along one or more of the volcanic systems of the rift zone and is usually accompanied by magmatic activity. The volcano-tectonic events of the Krafla volcanic system in 1975-1989 (see e.g. Björnsson, 1985, Einarsson, 1991a, 1991b) provided a dramatic demonstration of such failure at the divergent boundary. Magma began flowing into a crustal magma chamber in the roots of the Krafla volcano in NE Iceland in 1974 or 1975. When the

pressure reached a critical level in the chamber, the surrounding crust failed and magma was injected into the fissure swarm that transects the volcano. A small volume of magma erupted to the surface, but most of it was injected laterally along the fissures. This magmatic event was accompanied by large crustal deformation near the coast where the fissure swarm enters the ocean. After this initial event the volcanic system went through a series of similar events. The Krafla volcano reinflated and injected new batches of magma into various section of the fissure swarm. Each time the fissure swarm was widened, its central part subsided and its flanks were uplifted (Fig. 4). The vertical displacements were of the order of tens of centimeters to a few meters.

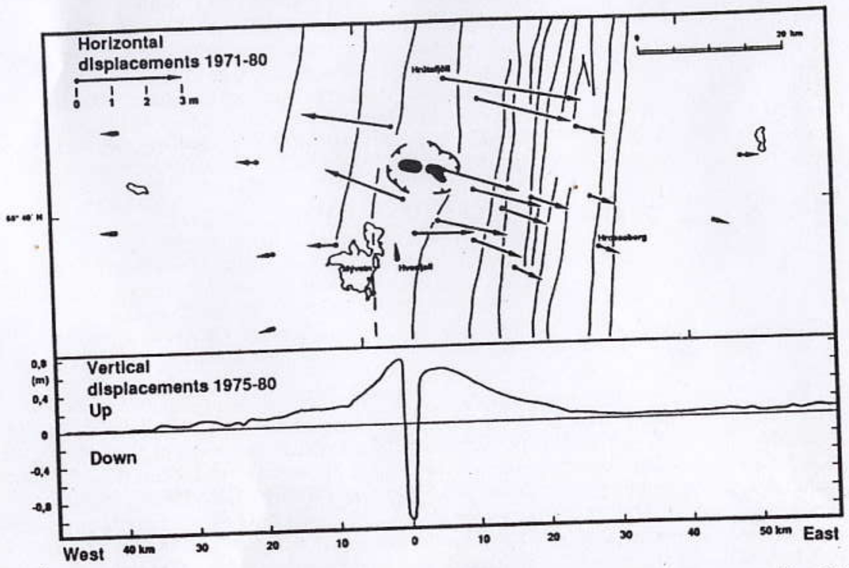

Figure 4. *Crustal movements during the Krafla volcano-tectonic episode as surveyed by Möller et al. (1982) and Kanngieser (1983). The Krafla fissure swarm widens and subsides, its flanks are contracted and uplifted. Modified from Einarsson (1991 b) and Björnsson (1985).*

About twenty such events were documented in this rifting episode of Krafla. Rifting episodes of this magnitude seem to occur in Iceland once or twice each century. Sudden sea level changes due to rifting and magma movements may occur where the volcanic systems intersect the coast, and may be of considerable economic importance. The harbor of Grindavík in SW-Iceland (Fig. 1) is particularly vulnerable, but other harbors and communities in SW- ans NE-Iceland may be affected. Holocene raised shorelines on post-glacial lavas on the Reykjanes Peninsula (see Sæmundsson and Einarsson, 1980) are possibly of this origin.

GLACIAL LOADING AND ISOSTACY

End of the last glaciation

The post-glacial geologic record of Iceland contains considerable evidence of rapid sea level changes at the end of the last glacial period. Raised beaches and terraces are found in numerous places in the lowlands, and generally become higher towards the interior of the country. In the coastal areas they are often at about 50 m elevation, but may reach over 100 m elevation further inland. These have generally been taken as evidence for isostatic adjustment of the crust due to the reduces load when the ice sheet disappeared. In contrast to Fennoscandia and Canada, where glacio-isostatic movements are still continuing, the Icelandic crust seems to have responded rather quickly to the load changes. Most of the movements were over in less than 1000 years. This fact

was used by Trausti Einarsson (1966) to conclude that the subcrustal viscosity in Iceland was an order of magnitude lower than beneath the continental shields. Sigmundsson (1990, 1991) reviewed the available geological evidence and constrained the subcrustal viscosity by model calculations. He determined a maximun viscosity of 10^{19} Pa s, and concluded that further refinement of this value required improved knowledge of the ice retreat history. This means that most of the crustal adjustments took place in about 400 years. This rapid adjustment is further supported by recent data on submerged coastal features and deltas in N-Iceland (Thors and Boulton, 1991) and SW-Iceland (Thors and Helgadóttir, 1991). These features are found at depths of 20-40 m. They are postglacial and must have been formed during the final stage of crustal adjustment to the reduced ice load, but at a low eustatic sea level. Combining their results from the Hörgá delta in N-Iceland with the eustatic curve of Fairbanks (1989), Thors and Boulton (1991) suggest the relative sea level curve shown in Fig. 5.

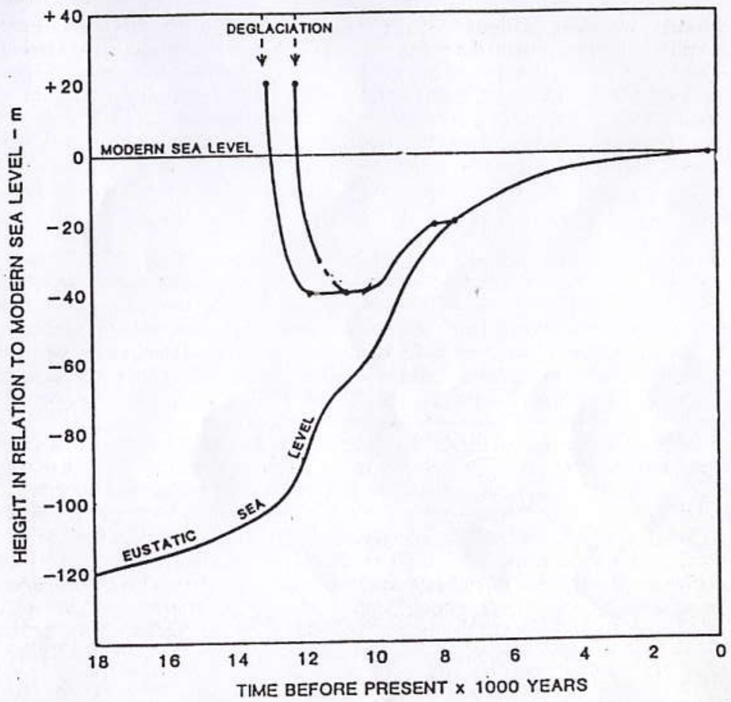

Figure 5. *Relative sea level curve for the Hörgá area in N-Iceland, as suggested by Thors and Boulton (1991). The sea level drops rapidly during and immediately after deglaciation, remains relatively stationary at -40 m around 10 000 years b.p., then begins to rise as the rate of eustatic sea level rise exceeds the rate of crustal rebound. After about 8000 years b.p. the curve follows the global eustatic curve of Fairbanks (1989).*

The disappearance of the major northern hemisphere ice sheets led to isostatic adjustments on a global scale. Peltier and Tushingham (1991) presented results of model calculations of the isostatic contribution to relative sea level changes. They assume a sperically symmetric earth model and calculate isostatic adjustments everywhere due to a given ice load history of the last deglaciation. These adjustments are still in progress and amount to a few mm per year over large parts of the globe, but Iceland is close to the zero-line. No changes are therefore expected in Iceland from this process.

Most of the gelogical evidence of sea level change during the Ice Age are from the end of the last glacial period. The Tjörnes sequence in N-Iceland (Fig. 1), however, contains evidence of repeated marine transgressions associated with earlier glacial periods (e.g. Th. Einarsson et al., 1967, Eiríksson, 1981). These changes are most likely associated with glacial loading, but it should also be considered that Tjörnes is located within an active transform zone of the plate boundary, the Tjörnes Fracture Zone. This is a complex zone of transform faulting and rifting (Saemundsson, 1974, Einarsson and Saemundsson, 1987, Einarsson, 1991 a), and the Tjörnes strata are located on a crustal block between two transcurrent features, The Húsavík Faults and the Grímsey seismic zone, and immediately west of the rift zone of North Iceland. The interaction of these zones may lead to complicated vertical crustal movements.

Present changes in ice load

The low subcrustal viscosity beneath Iceland, of the order of 10^{19} Pa s, suggests that changes in the ice load in this century might lead to detectable crustal movements. The volume of the Vatnajökull ice cap in SE-Iceland (Fig. 1) decreased by 180 km^3 in the warm period that began about 1920. Model calculations of Sigmundsson (1990, 1991) indicate that this decrease should lead to crustal uplift in the area around the glacier at a rate of 5-20 mm per year at the present time and tilt away from the glacier. An experiment to measure this uplift was conducted by Sigmundsson and Einarsson (1992) by measuring relative changes in lake level of the lake Langisjór near the SW edge of Vatnajökull. By comparing levelling results of 1991 to a previous survey in 1959, an uplift of about 12 cm was determined for a bench mark near the NE end of the lake with respect to a bench mark near the SW end of the lake, 15 km farther from the center of the glacier. This tilt rate constrains the subcrustal viscosity values to 10^{18} - $5x10^{19}$ Pa s, and confirms the results from the post-glacial uplift.

Further constraints will be provided by a new geodetic experiment initiated in 1991-1992 (Einarsson et al., in preparation). A geodetic network of 10 points was installed in the area SE of the glacier and measured by GPS geodesy. The network contains points near the glacier edge as well as a point at a distance of 50 km, i.e. outside the affected area. Uncertainties in relative positioning are estimated to be about 1 cm in the horizontal components and 2 cm in the vertical. Gravity was also measured at the geodetic points as well as several older gravity points in the area. Indications of gravity changes were found, consistent with uplift of several mm per year.

There is some evidence for rates even higher than the model calculations indicate. Jahn (1992) analysed the results of an extensive GPS-network in N - and E-Iceland in 1987. One of the points occupied was in Hornafjörður, at the SE edge of the glacier. The same point was occupied during a GPS-campaign in 1990. Comparison between the two determinations of the vector Akureyri-Hornafjörður showed changes that could be interpreted as the result of 8 cm uplift of the Hornafjörður point during the period 1987-1990. Imsland (1992, and this symposium) documents changes in relative sea level that could indicate similar rate of uplift. A small skerry, Hellnasker, within the Hornafjörður lagoon emerges about 75 cm above sea level at low tide at present. Before 1950 this skerry never showed, even at the lowest tidal levels. Crustal uplift is considered the most likely cause, even though this could conceiveably be caused by other changes in the lagoon. This observation indicates a rate of a little less than 2 cm per year.

There are also geological indications of earlier marine transgression in Hornafjörður (Jónsson, 1957, Sigmundsson and Einarsson, 1992, Imsland, 1992 and this symposium), presumably caused by increased ice load of Vatnajökull in the period 1400-1900 AD.

MEASUREMENTS OF SEA LEVEL

Reykjavík tide gauge record

The only tide gauge record in Iceland the exceeds a few years is from the Reykjavík harbour. The gauge was installed in 1956 and has been in operation with short interruptions since. The record was recently analysed by Guðmundsson and Einarsson (in preparation). Offsets and errors were corrected and about 60 harmonic coefficients determined. A record of yearly averages corrected for atmospheric pressure is shown in Fig. 6.

A trend of rising sea level is obvious. A straight line fit to the points gives the following equation:

$$H(t) = 2.183 + 0.0034\ (\text{year} - 1973)$$

The equation gives the height of the sea level in meters with respect to the Reykjavík harbor reference system. Relative sea level is rising in Reykjavík at a rate of 3.4 mm per year. Single yearly averages deviate as much as 11 cm from the best fitting line. The most conspicuous deviations occur during the period 1964-1968. By comparison with data from Grindavík in SW-Iceland and Thorshavn in the Faeroe Islands Guðmundsson and Einarsson (in prep.) argue that these deviations are of oceanographic origin. Regional conditions in the ocean appear to have been abnormal during this period. Sea ice, for example, drifted around large part of the Icelandic coast in 1965.

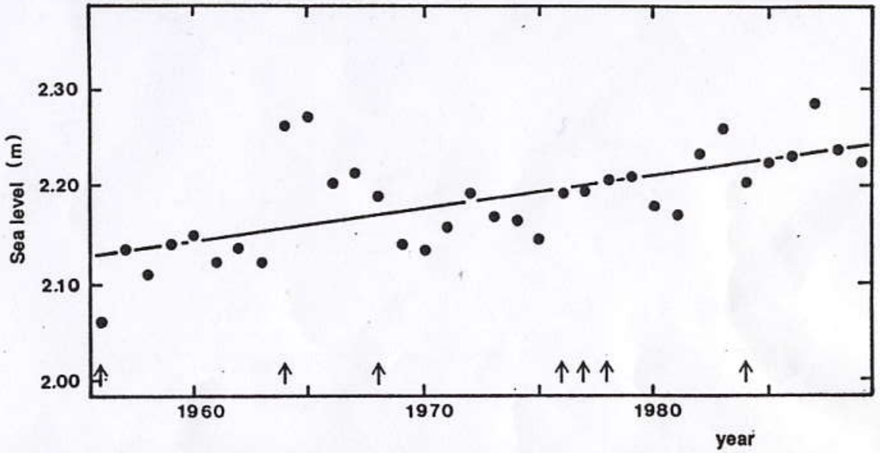

Figure 6. *Sea level in Reykjavík 1956-1989 from tide gauge data. Yearly averages are plotted as a function of time in years (modified from Guðmundsson and Einarsson, in preparation). The data have been corrected for atmospheric pressure. Arrows show times of leveling to a bench mark.*

The yearly averaging smooths out oscillations of shorter periods. The time series still contains longer tidal periods. Contributions from three long period tidal constituents of astronomical origin may still be assumed in the time series. These are the pole tide, generated by the Chandler wobble of 435 days period, the lunar perigee rotation with period of 8.8 years, and the lunar nodal regression with period of 18.6 years. Taking these constituents into account we are left with a secular sea level rise of 2.4 mm per year in Reykjavík.

The estimated secular sea level rise in Reykjavík is within the range of the global eustatic sea level rise of 2.4±0.9 mm per year given by Peltier and Tushingham (1991), and only slightly exceeds the 1.6-2 mm per year, as estimated e.g. by Trupin and Wahr (1990) and Douglas (1991). The data from Reykjavík therefore indicate that no significant crustal subsidence is occurring in Reykjavík at the present time. Slight subsidence is indicated only if the lowest estimates of eustatic sea level rise are taken. This conclusion is somewhat at odds with previously published views (e.g. Steinthórsson, 1987, Emery and Aubrey, 1991, Imsland and Th. Einarsson, 1991), who assume that crustal movements are the main contributor to the sea level rise in Reykjavík.

Short term sea level recordings

In addition to the Reykjavík tide gauge sea level has been measured at a number of locations over a short period. The Danish Navy recorded sea level at the beginning of this century at three locations in E-, W- and SW-Iceland. The data and reference points have not been found and are presumed to be lost. Measurements were done in Grindavík harbor in 1956-1964 by the Lamont Geological Observatory (now Lamont-Doherty Earth Observatory) in New York. The data are available, but the reference point has not been recovered. The Icelandic Harbor Authority made short-term sea-level determinations at 32 harbors around the country during the period 1975-1982 (Hafnamálastofnun ríkisins, 1980). The duration of the measurements was usually about 1-3 months at each location. Mean sea level and the principal harmonic coefficients were determined. Based on the deviations of monthly averages from the mean sea level curve in Reykjavík, the values of sea level obtained for these locations may be accurate to 10-20 cm.

Plan of future sea level monitoring

Iceland is rather well placed as a platform to observe changes in eustatic sea level. Monitoring of sea level rise in response to the anticipated global warming due to greenhouse gases in the atmosphere is becoming increasingly important. A plan to strengthen sea level monitoring in Iceland is being implemented. Gauges will be installed at 10-15 locations around the country in the next few years. An emphasis is put on long-term stability and regular calibration. A prototype instrument has been in operation in the Reykjavík harbor since early 1992. It consists of a pressure sensor and 3 sea surface sensors for automatic calibration. Digital data are collected into solid state memory, and the gauge contains no movable parts.

Absolute sea level determinations

Until recently sea level could only be determined with reference to land. The caused of relative sea level changes could therefore not be separated directly. A large global data set was needed to separate the contribution of global sea level change from the local effects of crustal movements. With the recent advent of space geodetic technology this is no longer necessary. With a combination of Very Long Baseline Interferometry (VLBI), Satellite Laser Ranging (SLR), GPS-geodesy (Global Positioning System), and other systems, it is possible to determine the position of geodetic points to within several millimeters in an absolute sense. Absolute sea level and absolute sea level changes can therefore be determined by a geodetic tie between sea level gauges and such absolute geodetic points. Absolute coordinates are available in two places in Iceland, near Reykjavík and Höfn (Magnússon et al., 1993). A DORIS determination and a GPS survey gave nearly identical results for a point near Reykjavík. A VLBI-point was established in Hornafjörður in 1992. Attempts are being made to connect these points to sea level in the vicinity.

CONCLUDING REMARKS

In this paper I have tried to give an overview of processes in the crust and upper mantle that can lead to changes in relative sea level. Examples have been given from the geological history of Iceland where these processes played a role. In spite of its short duration, only about 60 million years, this history is unusually rich and contains a broad range of these processes, e.g. uplift due to initiation of a mantle plume, rifting of a continent, subsidence due to lithosphere cooling, loading of the crust by volcanic production, stress changes at a plate boundary associated with plate movements and rifting, and loading and deloading of the crust by glaciers, denudation and sedimentation. These processes may play together in a complicated way and it may not always be obvious which one is the most important. Measurements of present sea level changes in Iceland will add important constraints to ideas on eustatic sea level, in particular when they can be made in conjunction with independent measurements of crustal movements, such as by GPS surveying, VLBI or other space-based techniques.

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THE DEVELOPMENT OF DOWNDRIFT EROSION

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INTRODUCTION

Downdrift or "leeside" erosion is an ancient feature on shores occurring where a headland, an inlet, a river, a bay, a canyon, a reef or other type of shoal blocked, wholly or in part, the natural longshore drift of materials, that is transport of sand and gravel, by waves and currents. This results in accumulations on the updrift, receiving, side and in the ocean, and a corresponding depletion of material, thereby causing erosion, on the downdrift side. The terminology "Littoral Drift Barrier" accordingly developed.

This paper which should be considered "an interim report", reviews practical cases of leeside erosion and makes an attempt to explain the development as function of time.

NATURAL LITTORAL DRIFT BARRIERS

Natural littoral drift barriers are found everywhere in the world and they have a determining influence on the development of shoreline configurations. Barriers may be more or less effective. As they store material on their updrift side and as the storage capacity usually is limited. some material will finally bypass the barrier and may then resume migration downdrift. Part of it may then be lost to the offshore bottom or be deposited in offshore shoals. Complete littoral drift barriers on a long uninterrupted shore are very rare, but they occur frequently as pocketed headland-shores, where the available materials are trapped between two headlands extending to greater depths making escape of materials from the pocket very difficult or impossible.

Man's influence on natural shores first started in rivers. A few thousand years ago he began developing ocean-going vessels. At the beginning they were based in river entrances, bays and other sheltered areas, but such favorable conditions were not always available. 3,000 to 4,000 years ago the Phoenicians built an open sea port at Tyre in the Eastern Mediterranean. Located on an open exposed littoral drift shore it was protected by breakwaters consisting of heavy concrete blocks. The Cretes and the Greeks followed the Phoenicians but their ports were situated on rocky shores with little littoral drift, if any at all. 2,000 years ago the Romans built a