RE-APPRAISAL OF TSUNAMI HAZARD IN THE NORTH ATLANTIC OCEAN: A CASE STUDY FOR ICELAND

J. Eliasson[1] and R Sigbjörnsson[2]

[1] *Professor, Dept. of Civil Engineering, Engineering Research Institute, University of Iceland.*
[2] *Professor, EERC, School of Engineering and Natural Sciences, University of Iceland*
Email, jonase@hi.is, ragnar.sigbjornsson@hi.is

ABSTRACT :

Iceland lies in the middle of the North Atlantic Ocean as a superstructural part of the Mid-Atlantic Ridge, the boundary between the North American and Eurasian tectonic plates. Towards the south the coast is exposed to significant tsunami hazard, according to some recent studies. The objective of the present paper is to review and reassess this potential hazard and put it into context with other earthquake-induced threats on the island. Most tsunamis are generated by shallow earthquakes in subduction zones. The only subduction zones around the northern part of the Atlantic Ocean are the Puerto Rico Trench and the Antilles Subduction Zone around the eastern Caribbean. Earthquake generated tsunamis have hit Puerto Rico and the Virgin Islands several times in recorded history, most recently in 1918. However, by far the most noteworthy Atlantic tsunami was generated by the great Lisbon earthquake of 1 November 1755, with a magnitude in the order of 9 it left Lisbon in ruins and damage was reported as far north as Ireland. According to some estimates the waves reached the south coast of Iceland. In addition, landslide and glacial flood generated tsunamis are judged to be a threat to Iceland. The tsunami hazard on the south coast of Iceland is quantified in terms of hazard curves utilising numerical modelling and computational fluid dynamics (CFD) simulations. The run-up heights are dealt with emphasising the effects of the coastal geometry and the shape of the continental shelf. Potential tsunami induced damage is addressed and the villages and coastal infrastructures under threat identified. The main finding is that the tsunami hazard in Iceland is quantified as moderate, which appears to be in line with historic recordings. Furthermore, we find that some of the recent studies may have overestimated the hazard.

KEYWORDS: Waves, tsunami, landslide

1. INTRODUCTION

In recent years great effort has been invested in analysing the risk of serious tsunami hazards in all of the major oceans in the world. This paper deals exclusively with the North Atlantic Ocean south of Iceland. It utilises scientific methods, risk assessments and known source locations that have been reported up to 2006.

Tsunamis usually originate from earthquakes that cause a seafloor displacement and are more often than not strengthened by landslides triggered by the earthquake. Often these triggering earthquakes do not contribute any significant amount of energy to the tsunami wave and this seems to be the general case in the North Atlantic. There have also been speculations in the scientific community about the danger of tsunamis from gigantic glacial flood waves, jökulhlaups, of volcanic origin emerging on the south coast of Iceland. Here there is a well defined probability of occurrence (Eliasson et al., 2006) and clear geological evidence of the volcanic events (Eliasson, 2008) but practically no historical evidence of a tsunami.

The main difficulty in tsunami hazard assessment is the generation of wave energy at the source. In many cases this is simply solved by estimating the ocean surface amplitude at the source (Day, 2003) but the estimate of the amplitudes involves many problems and is very uncertain. For the landslide tsunamis, block slide models are popular (Tinti et al., 1997). In the block slide models the blocks must reach very high velocities to create a serious tsunami. Here we introduce a new translatory wave model for this purpose (Eliasson, 2008). It is based on the assumption that sliding blocks break up and become debris flows when the velocity is high enough.

Once the initial wave is estimated, the energy transmission from the source to the point of impact can be quite accurately modelled in CFD models (Kerridge, 2005; Ward and Asphaug, 2003; Tinti et al., 1999) as long as the shallow water wave equations are in a stable domain. Instability and wave breaking, or reflection, sets in when the wave crest approaches a beach. On a flat beach there is practically no reflection but heavy wave breaking and energy dissipation. Few reports are available on how this problem is handled in practical hazard assessment, but most often this is simply done by making all coasts completely reflecting as this is on the safe side in run-up estimation. But it is an open question how the correct boundary conditions are, or should be, in terms of reflection in the various models reported. The final phase of the re-appraisal is an estimate for the tsunami risk in the North Atlantic from known sources in terms of return periods or annual probability of exceedance.

2. REPORTED OCCURRENCES OF TSUNAMIS IN THE NORTH ATLANTIC

In the following we shall focus on the eastern region of the North Atlantic Ocean. Tsunamis are considered rare in these waters. In the National Oceanic and Atmospheric Administration (NOAA)’s National Geophysical Data Centre (NGDC) Natural Hazards Database are found 122 run up events from 67 sources in the period after the year 1100 AD. The majority of these are reported with zero run-up height and of the rest, many are suspected to be storm floods without any tectonic events involved. If the zero run-up events are disregarded, 27 remain, of these 15 are reported with non-tectonic sources. Then we are left with 12 tsunami events, seven of unknown sources (Table 1) and five from known earthquake sources (Table 2).

It looks as though four of the events in Table 1 could be real tsunamis, indicating that the other ones would therefore be floods with other causes (e.g. storm floods). We are then left with nine tsunami events in the region in the 250 year period since 1755. In the 600 years before 1755, 17 tsunami events are in the NGDC database. The event frequency for these two periods is somewhat similar, or 33 years between events.

The NGDC data of tsunami action in the eastern North Atlantic Ocean indicate a mainly volcanic origin. They are listed in Table 3. There is one event common for Tables 2 and 3 on 21 November 1984. This event is estimated to come from a volcanic landslide in the Charley-Gibbs Fracture Zone.

Table 1 Tsunamis in the eastern North Atlantic from unknown causes (NGDC)

Date			Tsunami Source Location		Max Water Height (m)	No. of run-up's reported
Year	Mo	Dy	Country	Name		
1755	11	5	UK	ENGLAND	0.9	1
1756	2	27	UK	ENGLAND	1.8	1
1762	9	27	UK	ENGLAND	3	1
1763	9	18	UK	ENGLAND	3	1
1767	9	5	IRELAND	IRELAND	1.5	1
1855	2	17	PORTUGAL	AZORES	10	1
1950	8	15	UK	NORWAY & ENGLAND	0.5	26

Table 2 Tsunamis in the eastern North Atlantic caused by earthquakes (NGDC)

Date			Tsunami Source Location		Max Water Height (m)	No. of runups reported
Year	Mo	Dy	Country	Name		
1755	11	1	PORTUGAL	LISBON	30	52
1761	3	31	PORTUGAL	LISBON	2.4	11
1765	7	23	SWEDEN	SWEDEN	1.2	1
1969	2	28	PORTUGAL	PORTUGAL	1.14	3
1894	11	21	ATLANTIC OCEAN	ATLANTIC OCEAN	12.2	1

Table 3 Tsunamis in the eastern North Atlantic caused by volcanic activity in the Atlantic Ocean at Lat. 49.000 Long. -34.300 (Charlie Gibbs Fracture Zone) (NGDC)

Date			Tsunami Source Location		Max Water Height (m)	No. of runups reported
Year	Mo	Dy	Country	Name		
1884	12	29	ATLANTIC OCEAN	ATLANTIC OCEAN		
1894	11	21	ATLANTIC OCEAN	ATLANTIC OCEAN	12.20	

Especially the 1894 event is described rather dramatically in Berninghausen's (1968) *Tsunamis and seismic seiches reported from the western north and south Atlantic and the coastal waters of northwestern Europe*. The 1884 event is more uncertain.

3. WAVE HEIGHT SCALING

The result of the modelling and calculations is that tsunami wave formation (see appendix) and tracking is a quasi linear process where linear scales can be used, if it is done carefully. This means that wave heights from an observed event can be scaled:

$$H_{ijk} = \lambda_{ij} H_{jk} \tag{1}$$

This means that tsunami wave height in a reference location i from an event number k in a source location j is proportional to the near-field wave with a proportionality coefficient independent of the event number. There are two processes to consider here, the displacement wave in the near field and an eventual oscillatory wave in the far field. Eq. 1 should hold for both of them, except the breaker height of the oscillatory wave. This means it should hold where ever shallow water wave velocity is larger than the particle velocity of the wave itself.

Table 4. Reported tsunami sources in the North Atlantic Ocean

	UK coastal waters	NW Europe conti-nental slope	Plate boundary area west of Gibraltar	Canary Islands	Mid-Atlantic Ridge	Eastern North America contine-ntal slope	Iceland south coast	Carib-bean
Tsunami source	Earth-quake	Slide	Ms > 7 Earth-quake	Slide	Ms > 6.5 Earthquake	7.5>Ms>7 Earthquake	~ $3x10^5$ m^3/s glacial flood	Ms > 8 Earthquake
Historic large event	Dogger Bank 1931	Storegga	Lisbon 1755	Local small	C. G. zone 1894 ?	Grand Bank 1929	Katla volcano	Virgin Island 1867
Evidence	Seismo-logic	Geo-logical	Local wave 6 – 13 m	Geo-logical	NGCD 12,20 m	Seismo-logic	Historical & geological	Seismo-logic
Event observa-tions	British Geolog. Survey	None	British Geological Survey	None	ERI Univ. Iceland	??	(Eliasson et al., 2006)	Eye witness reports

4. FREQUENCY OF TSUNAMIS

The time between tectonic events is a random function (Eliasson et al., 2006):

$$t = t_a + s\, g_k(0,1) \tag{2}$$

where t_a is the mean time between events, s their standard deviation and $g_k(0,1)$ a random variable with zero mean and unit standard deviation and a proper distribution. Eq. 2 shows that the probability of an event occurring next year depends a little bit on the history. Nevertheless it is customary in risk analysis to use the average probability as derived from Eq. 2 as the probability of occurrence for an event. This means that a probability of occurrence of a tsunami wave from source j is:

$$P_j(H_j \le x) = F_j(H_j \le x)/T_{r,j} \tag{3}$$

Here $T_r = t_a$ in Eq. 2. The probability of a tsunami wave of height H at a site i when the possible sources are J will then be:

$$P(H_i < x) = 1 - \prod_{j=1}^{j=J}\left(1 - F_j(H_j/\lambda_{ji} < x)/T_{r,j}\right) \tag{4}$$

This is the probability of a tsunami event with wave a height less than or equal to the threshold or state variable x. This variable has to be calculated in a point j that is a near-field point for the beach in question and a far-field point for all the sources, except the jökulhlaup, next to last column in Table 4, where we can use the results of Eliasson (2008). To assess whether or not the tsunami is dangerous, the value of the amplification factor λ and the run up coefficient, i.e. the shoaling coefficient from j to the beach, must be known.

5. REFERENCE LOCATION AND THE SCALING FACTORS

We have chosen 62°30' N and 18°20′ W as a reference location for the tsunami threat on the south coast of Iceland. As source locations we have used the locations listed in Table 4.

The tidal range on this coast is 3 m. The harbour areas and the lowest inhabited parts of the country are 2 m above that, so a tsunami wave must have a run-up height 2 m or more to do any significant damage. This leaves out following source locations:

- ✓ Caribbean: To far away to evoke a tsunami > 2 m in the reference point
- ✓ UK coastal waters: Source tsunami too small
- ✓ Mid-Atlantic Ridge: Tsunami risk estimated as very small except volcanic slides in the Charley – Gibbs Fracture Zone (NCGD 1894 report) and earthquakes in the plate boundary west of Gibraltar (e.g. Lisbon 1755).

Table 5. Reference point wave height estimations in the North Atlantic Ocean

Parameter	NW Europe continental slope, big slide	NW Europe continental slope small slide	Plate boundary area west of Gi-braltar	Canary Islands	North Atlantic Ocean Charley -Gibbs Zone	Iceland south coast
Source distance, km	1,000	1,000	2,800	3,780	1,820	0
Av. track depth D,m	2,500	2,500	3,500	3,500	2,500	N/A
Opening β	3.14	3.14	6.28	3.14	3.14	N/A
Slide scar, m^2	27,000	9,000	N/A	10,000	10,000	N/A
Slide width B, km	133	45	N/A	40	40	N/A
Slide slope ΔH, m	500	500	N/A	2,000	2,000	N/A
Slide slope ΔL, km	100	100	N/A	20	20	N/A
Head wall y, final, m	200	100	N/A	200	200	N/A
Wave speed c f.f.	158	158	187	187	158	N/A
Travel time, hr	4.4	4.4	5.2	5.2	4.4	N/A
Scar volume, m^3	2,700,000	450,000	N/A	1,000,000	1,000,000	N/A
Near-field radius	656	450	15	205	189	N/A
Slide slope I ‰	5.	5	N/A	100.	100	N/A
Slide velocity V, m/s	3.2	2.2	N/A	14.1	14.1	N/A
Slide flow q, m^2/s	632	224	N/A	2828	2828	N/A
Near-field H	4.0	1.4	13	15.1	17.9	N/A
Scaling factor λ	0.58	0.34	0.07	0.16	0.20	
H farfield (en. flux)	2.3	0.5	1.0	2.5	4.2	2
Return period, yr	100,000	100,000	400	100,000	?	200

The scaling factors for slide tsunamis are estimated using the translatory wave theory (Eliasson et al., 2007) for wave heights in the near-field. The translatory wave for a submarine slide is a little different from the mega-flood waves described by Eliasson et al. (2007) but in principle the same. Two things cause the velocities in a submarine debris flow wave to be much lower than in a similar water wave on dry land. Firstly a much heavier bottom friction develops in debris flows than in fluid flows and secondly, the flow resistance term between the slide and the water. Consequently, the velocity coefficients are much lower. The initial tsunami wave generation is modelled as a displacement wave running away with the shallow water wave velocity at a particular depth. Then a wave energy transmission model (Eliasson, 2008) is used to find the scaling factors. The calculations are fired up using the data in the yellow fields. This data is estimated using contour maps of the ocean floor and data available from Kerridge (2005). Further description of the estimation process is not possible within the prescribed limits.

6. TSUNAMIS WAVE HEIGHT FREQUENCY

Equations 3 and 4 are used to find the approximate exceedance frequencies of the various wave heights listed in Table 5. There is large uncertainty in the estimation of the source wave heights, as previously stated, and some uncertainties in the scaling coefficients too. There is also an additional uncertainty associated with the fact that the distribution g_k in Eq. 2 is not known either.

The results are displayed in Fig. 1. The calculated points are shown in blue and a logarithmic trend line in red.

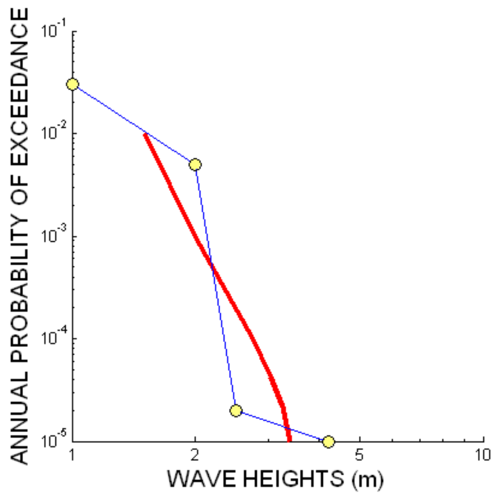

Fig. 1 Tsunami wave height in the reference point versus annual probability of exceedance (1/return period)

It can be seen that tsunamis greater than 1.5 m can occur every 100 years, while potentially dangerous tsunamis (> 2 m) occur only with a return period of 1000 years or so on average. Hence, we conclude that the risk is in the range of low to moderate. The wave heights around 3.5 m seem to occur with a return period a little lower than 100,000 years. These are very dangerous tsunamis that would devastate several places on the coastline with possible deaths, and they would do so at any sea level of the tide.

6. RUN-UP HEIGHTS AND POPULATION CENTERS UNDER TREAT

6.1 The situation in South Iceland

Approaching the coast from the reference point the wave runs into a new near-field process, or the shoaling, breaking and run-up of the tsunami wave. This is a non-linear process that is difficult to estimate, especially the run-up wave height. Several mathematical solutions exist, both analytical and numerical from CFD. The analytical methods utilise mostly the conservation of momentum, or transported energy but they are for two dimensional waves only.

We must distinguish between two cases: The first one is the displacement wave, which is a translatory in nature with properties different from the oscillatory wave. The second case is the oscillatory wave. When we have big tsunamis, it will be the displacement wave that hits the nearby coasts. Farther away from the source it will become an oscillatory wave, which is the situation for the earthquake generated tsunami in our case. We will investigate these two cases on a slowly varying bottom.

6.2 Displacement wave

The particle velocity will be increased as the square root of the depth ratio. The water particles cannot overtake the wave front, so a bore, H meters high above still water level, will be formed and this bore travels ashore as a breaking wave and inundates the land. According to linear theory the bore is formed when u = c, this leads to H = D for the breaking point of the wave, but from higher order theories and practical experience for long wave breaking on a beach, H = 0.7 D is closer to the true value. If the beach slope is so small that traditional long wave mild slope equations (Svendsen, 2006) are valid (e.g. river estuaries), then the inundation will be up to a level R = H above still water level, or less. If the slope is large so reflection (full reflection or partial) sets in we will have:

$$R = H + \frac{u^2}{2g} = H + \frac{H^2\left(\sqrt{g(H/0{,}7}\right)^2}{(H/0{,}7)^2\,2g} = 1.35H \qquad (5)$$

Fig. 2 Iceland and surrounding seas, depth scale by deeper blue for each 200 m. Population centres in red.

6.3 Oscillatory wave

The analytical methods use the various wave theories, Jonsson (1995) utilises the energy flux and set-down to predict shoaling of finite amplitude waves with and without a net volume flux. The predictions are carried into the point of breaking of the wave front, here the theory breaks down. Jonsson (1995) uses Stokes fifth order theory. For waves of very small steepness Stokes first order theory is quite as good, as can be seen from Fig. 5 in Jonsson's paper. Carrier et al. (2003) use a semi-analytic solution technique and find the inundation depth on the shore, their method is also close to linear though the starting point is the nonlinear wave equation. Guard (2005) presents a general solution for the problem with similar properties. All these results indicate that the shoaling process is near linear for tsunami waves with a very small steepness. Linear theory for shoaling means keeping the energy flux constant until the point of breaking. Then we find:

$$\frac{H_b(r)}{H} = a = \left(\frac{0{,}7\,D}{H}\right)^{1/5} \tag{6}$$

Here $H_b(r)$ is the breaker height of the radial wave and the amplification factor due to shoaling of the radial wave is denoted a. For waves around 1 m coming in from the deep regions of the North Atlantic Ocean it can become a = 2 – 5, for waves in the range of few centimetres it can become a = 5 – 10.

When $H_b(r)$ is found the run-up will be the same as in the case of the displacement wave, 1.0 – 1.35 times the breaking wave height.

6. CONCLUSION

There are five source areas that could evoke dangerous tsunamis for South Iceland. These are the submarine landslides as follows (popular event names are added for clarity):

NW Europe continental slope (Storegga)
Canary Islands (La Palma)
North Atlantic Ocean, slides in the C.G. Zone (volcanically or earthquake triggered)
Iceland south coast (jökulhlaup mega-flood)
Plate boundary area west of Gibraltar (Lisbon earthquake).

When the probabilities of tsunami heights from these sources are weighted together the return periods shown in Table 5 emerge.

Table 5. Reference point wave height estimations North Atlantic Ocean

H, m	1.5	2	2.3 – 3.2	2.5 -4.3
Return period, yrs	100	1,000	10,000	100,000

It must be noted that the uncertainty in H in Table 5 is considerable. The limits indicated are the authors' first estimates only. It is, however, interesting that an event in the Charley-Gibbs fracture is decisive for the maximum value. Practically no data exist upon events there, so it may be hoped that future research will provide useful data on volcanism and observable slide scars in this region.

Earthquakes will be the triggering mechanism for all these events, but the only one so big it that it can be used as a warning signal is the event at the plate boundary area west of Gibraltar. A further complication in a warning system design is the fact that the ocean south of Iceland, especially the areas in the southwest, has very high and frequent storm waves so the possibility in loosing the warning signal are very high.

The tsunami wave heights in Table 5 are moderate only so when a tsunami hits the coast the tide will have a great effect. Highest tidal levels on the south coast (2.5 – 3.5 above mean low water spring tide) occur in mid-winter (January – February), the same time as the greatest storms. The possibility for a severe event that has inundation up to 6 – 7 m above highest water spring tide with bad weather and high seas does exist. The probability of such an event would presumably be very low, $P = 10^{-7} – 10^{-8}$ per year or less.

Finally it must be noted that the eastern part of Iceland is different from the south coast. It faces the NW Europe continental big slide slopes and has many long fjords where amplification of oscillatory waves is possible.

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