Potential for landslide-related tsunami in the Dover Strait area (English Channel) based on numerical modeling.

J. Roger
Laboratoire de Recherche en Géosciences et Energies, Département de Physique, Université des Antilles et de la Guyane, Guadeloupe, France
École normale supérieure, Laboratoire de Géologie, CNRS UMR 6538, Paris, France

Y. Gunnell
Université de Lyon, CNRS UMR 5600, Lyon, France

Y. Krien
Bureau de Recherches Géologiques et Minières, Orléans, France

P. Ray
Institut Jean Le Rond dôAlembert, Paris, France

15 WCEE
LISBOA 2012

SUMMARY
Although earthquakes and tsunamis are less frequent in the Dover Strait than over active subduction zones, a plausible potential exists for intraplate earthquakes of magnitude Mw=6.9 generating a tsunami with damaging consequences. In April 1580, an earthquake shook the region violently and destructions were reported as far as London in the north and Rouen in the south. Despite fair weather conditions, a series of abnormal sea waves was reported in several harbours (Calais, Boulogne and Dover) on the same day. A first step was to produce a range of numerical coseismic tsunami simulations and to compare them with historical witness accounts. Results raise the question of whether such earthquakes could also trigger chalk flow-generated tsunamis along cliff lines on both sides of the Strait. Gravity-driven collapses affect the chalk cliffs periodically, but local tsunami waves caused by very large mass movements could reach heights of several meters and, for example, strike Dover.

Keywords: Dover Strait, chalk cliff, landslide, earthquake, tsunami

1. INTRODUCTION

On April 6, 1580, an earthquake occurred within the Dover Strait showing a maximum palaeointensity of VIII according to the isoseismal map constructed by Neilson et al. (1984) and the revision of their work by Melville et al. (1996). The estimated local magnitude of this event is 5.8 and makes it the most severe reported in this part of Europe. Nevertheless, such a magnitude appears to be inconsistent with the extent of the damaged area and with the generation of a coseismic local tsunami compatible with the witness accounts mentioned in available historical reports and simulated by Roger and Gunnell (2011).

Roger and Gunnell (2011) tested several coseismic rupture scenarios located within the Dover Strait with respect to the geology and to recent studies of intraplate shakes. Using numerical simulations, these authors showed that even a tsunami triggered by a Mw=6.9 earthquake, i.e. the maximum value potentially attainable in this region (Bakun and Scotti, 2006; Camelbeeck et al., 2007; Bungum et al., 2010), would not fit the historical accounts of tsunami wave heights and related destruction on April 1580. Historical documents indicate that this earthquake (on April 16 in the Gregorian time frame) was soon followed (but without precise detail on how soon) by a series of huge sea waves (Neilson et al. 1984, Melville et al. 1996). Given the fine weather, calm seas (Neilson et al. 1984, Lamb 1991) and neap tides at the time of the marine event, Roger and Gunnell (2011) indicate that the waves could conceivably be attributed to a tsunami. Reports of large waves reaching Kent and northern France at similar times are supported by up to six independent French, English and Flemish sources (Melville et al. 1996, Haslett and Bryant 2008). Flooding was more severe in Calais and Boulogne, but with 120 fatalities or more in Dover, additional deaths in France, and a minimum of 165 sunken ships reported.

The results of coseismic tsunami modeling, the fact that the outer wall of Dover castle collapsed with the cliff under it, reports from a boat passenger that his vessel touched the seafloor five times, and the mention by a mariner of a wave height of ~9 m (6 spear lengths) (Haslett and Bryant 2008 and references therein) converge towards a strong likelihood that a tsunami was associated with the 1580 earthquake. The wave height of ~7-10 m could, however, also be attributed to a tsunami generated by coastal mass movement interacting with the coastal waters. Such events are capable of causing large local wave amplitudes that attenuate rapidly with distance from the point source. Here we explore the potential of indirect tsunami generation by an earthquake-triggering landslide rather than by direct fault slip on the sea floor.

2. LANDSLIDE SCENARIOS

2.1. Generalities

The Dover Strait is a ~33 km-wide sea passage between France and England with a maximum water depth of ca. 60 m between Cap Gris-Nez and Dover, and is cut by a NW-SE-striking fault network (Fig. 1). As illustrated by the slope map (Fig. 2), it is also bordered by coastal chalk cliffs located within the maximum intensity zone proposed by Neilson et al. (1984) or Melville et al. (1996) for the 1580 event. The rupture processes appear to occur randomly on any of the numerous faults of the Weald-Artois shear zone complex, but particularly affects those that cut the coastal chalk cliffs. This situation is reminiscent of North America and Fennoscandia, where the local state of stress associated with recurrent intraplate seismicity has been attributed to postglacial rebound (Zoback and Grollimund, 2001; Camelbeeck et al., 2007; Mazzotti & Townend 2010). Such ruptures could directly generate a tsunami able to impact both sides of the Strait within a couple of minutes (tsunami travel times on Fig. 1, right), and/or affect the cliffs bordering the Strait, and trigger landslides. In terms of tsunami hazard prediction, the question is to determine the minimum volume of displaced material necessary to generate tsunami waves affecting one or both sides of the Dover Strait. This concern for volume thresholds carries implications for monitoring the potential for chalk mass movement along the English Channel coastlines, for example in the hinterland of the South Foreland where vertical chalk cliffs are currently undergoing the fastest rate of change in England and Wales (May, 2003).

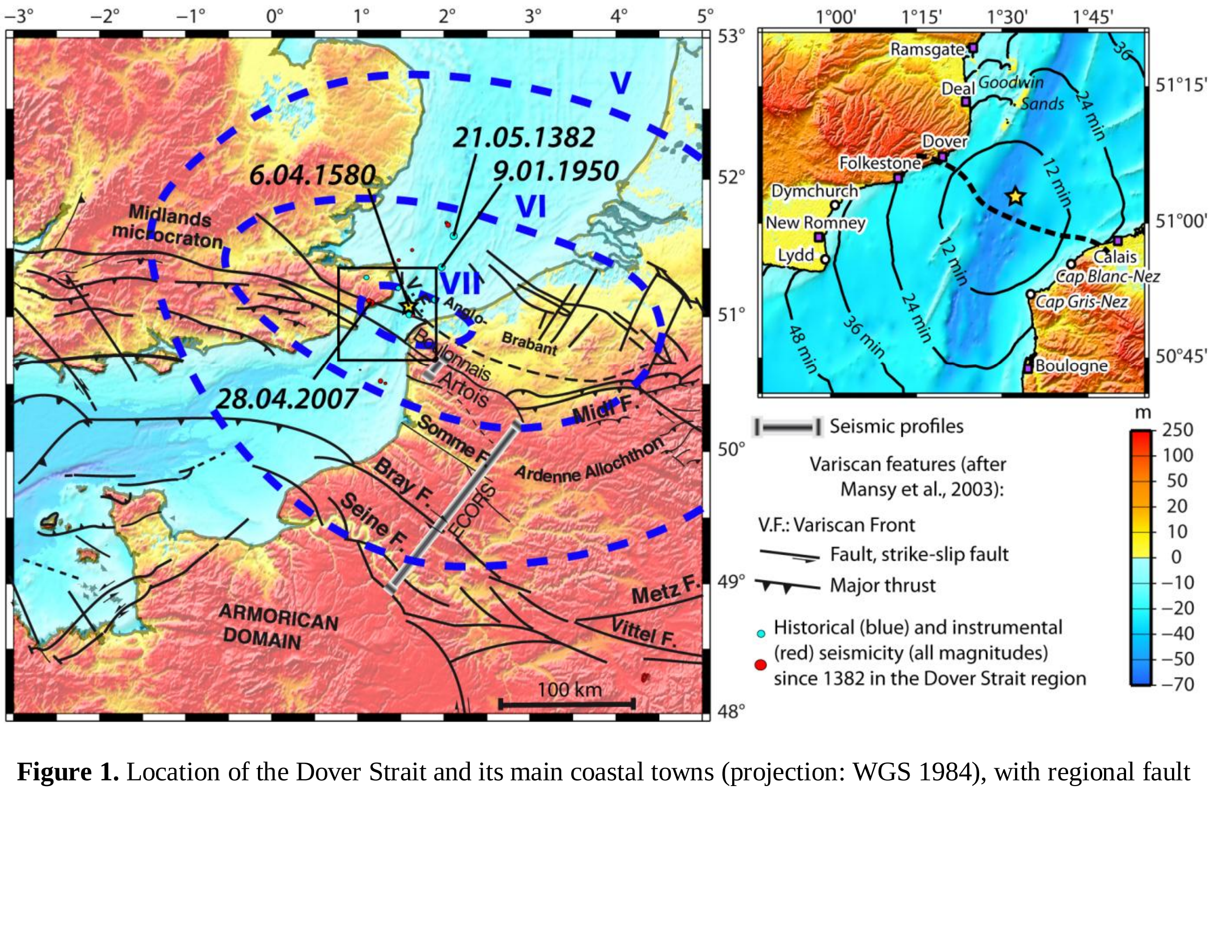

Figure 1. Location of the Dover Strait and its main coastal towns (projection: WGS 1984), with regional fault

pattern, historical seismicity and isoseismals for the 1580 earthquake. The relief grid combines Shuttle Radar Topography Mission data on land and a digitized, georeferenced and interpolated 1:115,000 scale bathymetric chart (Imray, 2007) of the Dover Strait. Blue dots are sites of seismicity recorded since 1973 (http://earthquake.usgs.gov/earthquakes/eqarchives/epic/). Tsunami travel times (TTT) of the first arrival are in minutes and are computed using the Mirone program (Luis, 2007). They are generated by a point source located at the center of the Dover Strait. Coasts where continental elevations exceed 30 m are dominated by sea-cliffs.

Being by far the thickest geological formation of the study area, reaching a maximum thickness of 270 m near Dover and standing higher than any other in the scenery, chalk is the dominant stratum of SE Kent. The upper dip slope of the North Downs chalk scarp corresponds to the older, Lower Chalk near Folkestone, and to the relatively resistant and whiter Middle Chalk closer to Dover. The latter still forms the cliff base at the South Foreland (Fig. 2), where it is capped by the feather edge of the softer Upper Chalk. These cliffs, particularly in Kent where the North Downs scarp terminates in the sea at Folkestone Warren, are prone to severe coastal erosion. Cliffs regularly recede by several meters, progressively or suddenly due to landslides or chalk flows of various sizes (Hutchinson, 1969, 2002; Dornbusch, 2006; May, 2003; Dornbusch et al., 2008). These chalk flows are due to the geological characteristics of the cliffs, to sea-level rise, and often occur in the aftermath of exceptionally high winter precipitation. Annual precipitation maxima in excess of 750 mm occur around the most elevated upland zone of the North Downs, in sharp contrast with the lowlands of NE Kent in the lee of SW winds, where annual totals fall to less than 625 mm (Coleman and Lukehurst, 1967). Finally the cliffs on both sides of the Dover Strait exhibit a range of conditions, e.g. slope angles, water saturation, foot-wall scouring, that are favourable to chalk flow generation (Middlemiss, 1983; May, 2003; Pierre and Lahousse, 2006). Chalk flows (Hutchinson, 2002) are flow slides, i.e. a high magnitude form of debris flow following from the structural collapse of rock or debris such as a rock avalanche. They develop under certain conditions in association with rock falls and hence are complex forms of mass movement. In NW Europe they affect >30 m-high soft chalk scarps with >40% porosity. The main characteristic is that the run-out over near-horizontal surfaces (for example the chalk platforms that are exposed at the base of some chalk cliffs at low tide) can be as much as 5 to 6 times the slope height. Their high energy and high velocity, involving momentary fluidization, affords them great destructive power.

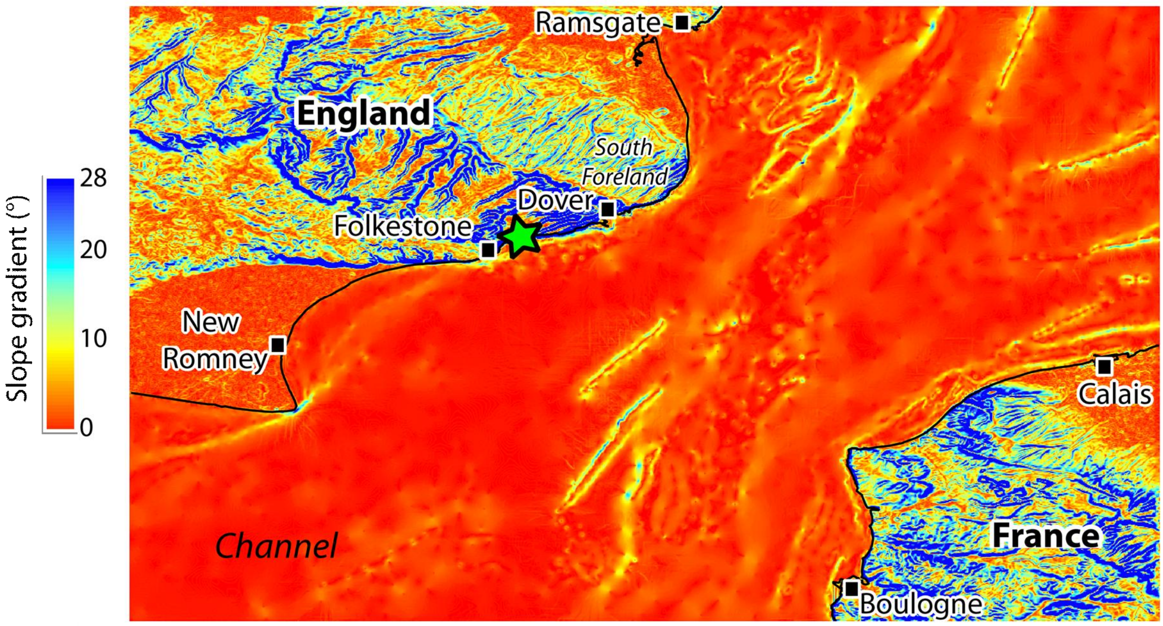

Figure 2. Slope gradient map of the Dover Strait region. Blue areas represent 28° slope angle and more. The green star indicates the position of the 1915 Folkestone Warren mass failure.

The processes driving chalk-cliff retreat in Sussex have been described in great detail by Dornbusch et al. (2008). Keefer (2000) underlines the fact that landslide occurrences correlate well with distance from the earthquake source and slope steepness, and in a more complex way, with rock type. We consider here that, within the framework of a hazard study, the behavior of chalk flows could be modeled in the same way as a classic rock landslide, as a first-order approximation.

2.2. Landslide tests

2.2.1. Scenarios

Depending on the geographical position of the cliffs relative to that of the rupture zone, we tested several hypothetical landslide scenarios. Due to the low depth below sea-level of the Strait floor (<< 60 m) and to sea-floor slope angles generally less than 10° except for small areas located along the shallow banks in the central Channel (Fig. 2), any eventuality of submarine landslides can be safely ruled out. According to Hutchinson (2002), subaerial landslide hotspots on both sides of the Channel occur west of Calais in the Cap Gris-Nez-Cap Blanc-Nez area and around Dover (Dornbusch, 2003; Dornbusch et al., 2008), i.e. where >100 m-high cliffs are located. In each tsunami model scenario, the slide parameters were constrained with data available from the literature (Table 1). We propose to test volumes of 0.5 million m^3 to 10 million m^3 always falling in 10 m water depth as preliminary approximation. This water depth is compatible with tidal ranges on both coastlines, where the highest tides in calm waters at Boulogne and Dover reach ~8 m and ~7.4 m, respectively, hence drowning the cliff-foot chalk platforms even at low tide (May, 2003). These volumes are calibrated on those proposed by Hutchinson (2002), who indicated a volume of ca. 1 million m^3 of debris involved in the Great Fall at Folkestone Warren (1915), which produced the largest run out (370 m seaward) ever reported in this region. The 10 million m^3 tested corresponds to the volume involved by the 1979 Nice submarine landslide (Assier-Rzadkiewicz et al., 2000), the biggest recorded along the French shores. These parameters are summarized in Table 1.

Table 1. Landslide scenario parameters.

Cliff segment	Landslide entry point		Angle of failure (counter-clockwise from north)	Water depth near landslide toe	Landslide volume	Landslide runout length in water	Landslide runout time in water	Landslide width at shoreline
	(decim. latitude)	(decim. longitude)	(degrees)	(m)	(m^3)	(m)	(s)	(m)
Shakespeare Cliff	51.1072	1.2893	200	10	$0.5 \cdot 10^6$	100	10	500
Langdon Stairs to St Margaret's	51.1330	1.3530	200	10	$1.0 \cdot 10^6$	200	20	1000
St Margaret's to Kingsdown-1	51.1615	1.4	240	10	$1.0 \cdot 10^6$	200	20	1000
St Margaret's to Kingsdown-2	51.1615	1.4	240	10	$1.0 \cdot 10^7$	600	60	2000
Cap Blanc-Nez	51.1072	1.2893	40	10	$1.0 \cdot 10^6$	200	20	1000
Cap Gris-Nez	50.8622	1.5720	90	10	$1.0 \cdot 10^6$	200	20	1000

2.2.2 Numerical modeling

We used the GEOWAVE model (Watts and Waythomas, 2003) to simulate the generation and propagation of landslide-generated tsunamis in the Dover Strait. The initial surface water deformation and velocity field are first computed using the TOPICS module (Walder et al., 2003). Wave propagation is then computed with the FUNWAVE module, based on fully non-linear Boussinesq equations accounting for frequency dispersion (Wei et al., 1995). Note that TOPICS does not deal with the splash zone, which is an area of complicated wave dynamics (e.g. Fritz et al., 2004). The characteristics of the tsunami wave in the near-shore zone might therefore be poorly reproduced. However, a full 3-D, multimaterial instationary model of a landslide-generated tsunami is beyond the scope of this paper. Given that very little is known about the geometry and dynamics of potential landslide-generated tsunami sources in the Dover Strait area, it would be unreasonable at this exploratory stage to expect more than order-of-magnitude precision for landslide-induced sea-surface elevation estimates. All the modelings have been done over a 150 m resolution grid of the Dover Strait which combines Shuttle Radar Topography Mission data on land and a digitized, georeferenced and interpolated 1:115,000 scale bathymetric chart (Imray, 2007) of the Dover Strait.

3. RESULTS AND DISCUSSION

Tsunami modeling results using landslide sources are shown in Fig. 3, which presents the maximum wave heights reached for each of the 6 scenarios described in Table 1.

The first assessment is that, considering a sliding direction perpendicular to the cliff (angle of failure in Table 1), slide location and slide volume play a major role in the energy distribution and thus in tsunami propagation paths. A volume of less than $1.0x10^6$ m^3 (Shakespeare Cliff in Fig. 3), i.e. the most common size of chalk flows, is unlikely to produce a widespread tsunami in the Dover Strait, due to the dispersion phenomenon associated with the propagation of these small wavelength / high frequency signals as explained by Ward (2001) or Harbitz et al. (2006). Typically, periods of 1-10 minutes are found for a landslide-related tsunami, compared with periods of ca. 1 hour for a coseismic tsunami. In contrast, volumes in excess of $1.0x10^6$ m^3, such as the Great Fall at Folkestone Warren (located to the west of the Langdon Stairs to St Margaret's area in Fig. 3), will locally produce wave heights exceeding 2 m and still preserve wave heights up to 0.5 m on the opposite side of the Strait. The worst case scenario, which remains hypothetical as it does not rely on any known historical event, corresponds here to the St Margaret's-Kingsdown test 2 (located at the highest point of the Dover cliffs), with a failing mass of $1.0x10^7$ m^3. Such an event generates waves more than 2 m high in near-field locations and is equally capable of reaching the French side with 2 m high waves at specific focal points. Given current cliff dynamics in the Dover area, where > 100 m of available cliff relief exists, and given the disconnection of some of these cliffs from the sea due to 20th century engineering developments at Dover docks and Shakespeare Cliff, the release of a chalk mass in excess of 0.01 km^3 in one single event is unlikely and model results do not support the possibility of such an event occurring today: if such a major event had occurred in 1580, it would have been mentioned in historical reports. Nevertheless it is still feasible that such an event happened in 1580. It is, for example, plausible that a seismic rupture set off several coastal chalk flows, each triggering local tsunamis on both sides of the Strait; this could have happened in 1580 and could explain the important waves and the reported destruction of boats in Dover, Calais and Boulogne at the same time.

4. CONCLUSION

Based on preliminary modeling results presented in this study, the hypothesis of a tsunami triggered by a single coastal landslide, itself generated by the 1580 "London" earthquake, seems to be impossible considering the minimum volume of debris identified as necessary to generate a Strait-crossing wave with sufficient amplitude on both sides of the Channel. Nevertheless, the sensitivity tests permitted by the modeling raise the prospect of tsunami generation by multiple chalk flows compounding the direct effects of coseismic rupture.

AKCNOWLEDGEMENT
This work has been funded by the French ANR MAREMOTI under contract ANR-08-RISKNAT-05-01c and by the TSUNAHOULE INTERREG IV project.

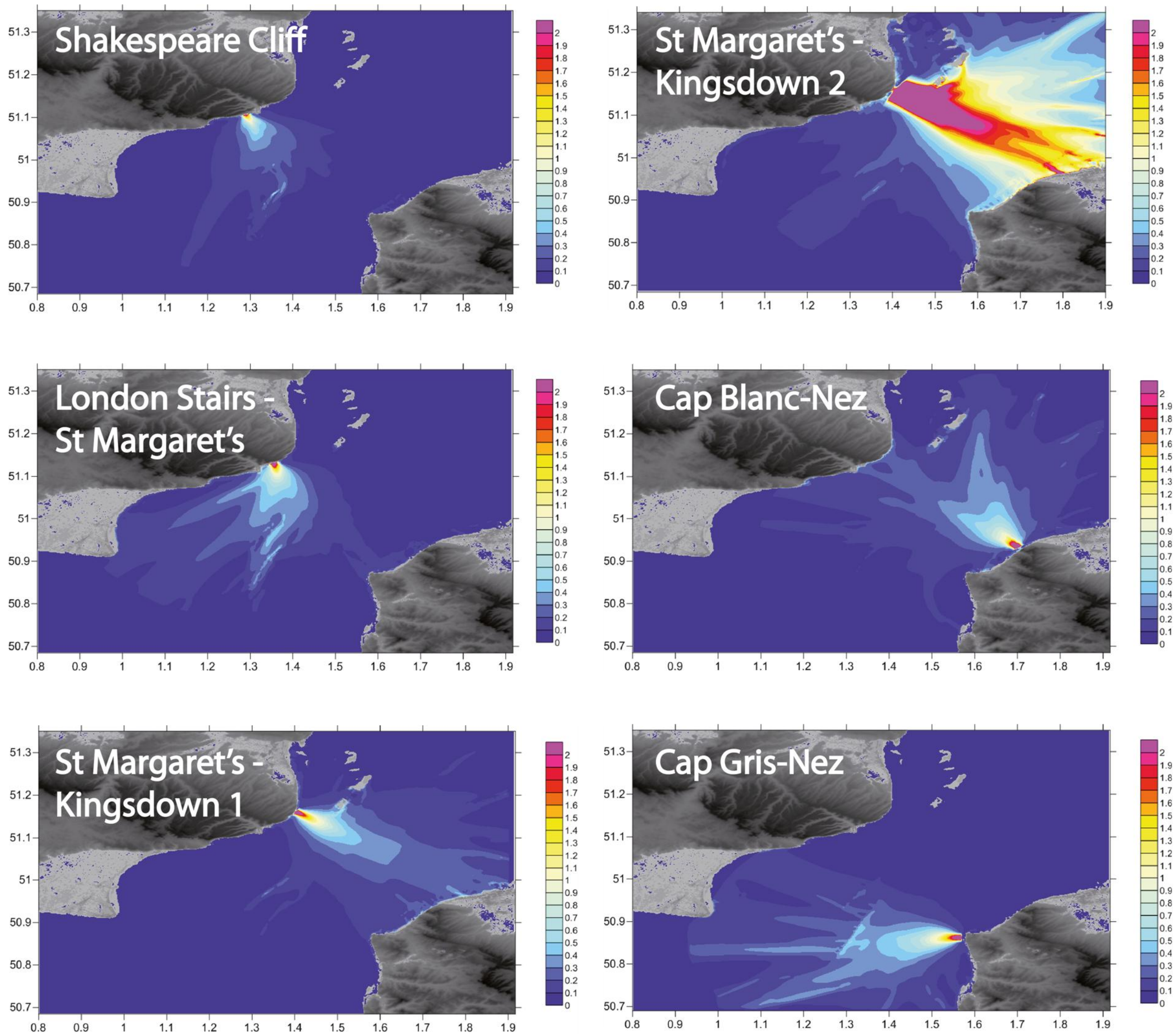

Figure 3. Maximum wave height maps of the Dover Strait obtained for each scenario of landslide-triggering tsunamis.

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