Was Grand Banks event of 1929 a slump spreading in two directions?

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Abstract

This paper describes a kinematic model of tsunami generated by submarine slides and slumps spreading in two orthogonal directions. This model is a generalization of our previously studied models spreading in one direction. We show that focusing and amplification of tsunami amplitudes can occur in an arbitrary direction, determined by the velocities of spreading. This kinematic model is used to interpret the asymmetric distribution of observed tsunami amplitudes following the Grand Banks earthquake—slump of 1929. © 2002 Elsevier Science Ltd. All rights reserved.

Keywords: Tsunami; Slides; Slumps; Grand Banks earthquake; Grand Banks slide

1. Introduction

Generation of tsunami by vertical motion of the sea floor during earthquakes has been studied extensively [1–6]. While it has been suggested that submarine slumps and slides may also generate tsunami [7], only recently it was recognized that those generation mechanisms are common, and that their study is important for the estimation of hazards from inundation.

There are many differences in the effective tsunami source parameters between earthquakes and submarine slides and slumps [8]. One of the more important differences is the velocity with which the motion of the sea floor spreads out [9]. During earthquakes, this velocity is of the order of kilometers per second and the areas of the uplifted sea floor can exceed $10^7$ km$^2$. Submarine landslides and slumps can also cover very large areas ($10^7$ km$^2$), but are usually smaller ($10^5$ km$^2$) and spread with smaller velocities (tens to hundreds of meters per second [8]). As a consequence, the near-field motion in the fluid surrounding the source cannot be approximated by ‘transferring’ the displacements of the bottom to the water surface, to describe the initial conditions for initiation of numerical analyses of tsunami propagation.

Recently, we investigated the complexities of the water wave motion in the near-field by considering a simple two-dimensional kinematic representation of slides and slumps, with the vertical motion of the sea floor spreading with constant velocity in one direction only [8,9]. This paper shows how such calculations can be extended to two-dimensional spreading, and illustrates the consequences on the near-field waves.

There are many examples of earthquake source areas, long in the direction perpendicular to the principal direction of motion. For example, large subduction zones produce earthquakes with elongated fault planes, hundreds of kilometers long (parallel to the axis of the subduction zone) and only 50–300 km wide along the fault dip [10]. Faulting on such surfaces will spread along and perpendicular to the arch axis, with velocities in the range 1–3 km/s. For very long periods, such sources can be approximated by a moving point source [11], but for intermediate and short periods, the motion may appear as multiple source processes [12].

There are many submarine landslides and slumps, which are narrow and long [13,14]. To describe the tsunami source for such slides and slumps, it becomes essential to assume that the sliding spreads at least in two directions, one uphill or downhill and the other one in the direction parallel to the coastline. For modeling long period seismic waves, such sources have been approximated by a single couple source, parallel to the principal direction of sliding [15,16], but to model intermediate and short periods, and to study near-field motions, it becomes necessary to assume that the source is spreading in at least two directions.

A detailed and complete presentation of tsunami waveforms for sources spreading in two dimensions is complex...
and too voluminous to be shown. This is because the number of cases required to illustrate the variations with respect to source geometry and nature of spreading are prohibitively large. In this work, we consider an example of an elementary kinematic source model for which the motions spread out in two perpendicular directions (along the length and width of a rectangular source region). We illustrate near-field waveforms from this source model for different ratios of the velocities of spreading, and compare the peak tsunami amplitudes with those from sources spreading in one dimension. Finally, we discuss how this model could be used to explain the asymmetry of tsunami radiation following the Grand Banks earthquake of 1929.

2. Mathematical model

2.1. Model of tsunami generation and propagation in two dimensions

The model we consider (Fig. 1) is a fluid layer of constant depth, \( h \), bounded by the rigid ocean floor at \( z = -h \) and by the free surface at \( z = 0 \), and excited by a ‘small’ uplift, \( \zeta(x, y; t) \), at \( z = -h \). The uplift of the free surface is subject to the following three boundary conditions

\[
\nabla^2 \phi(x, y, z; t) = 0, \quad (1)
\]

subject to the following three boundary conditions

\[
\phi_z(x, y, z; t) - \eta_z(x, y, t) = 0 \quad \text{at } z = 0 \quad (2a)
\]

\[
\phi_y(x, y, z; t) - g \eta_z(x, y, t) = 0 \quad \text{at } z = 0, \quad (2b)
\]

and

\[
\phi_z(x, y, z; t) - \zeta_z(x, y, t) = 0 \quad \text{at } z = -h. \quad (2c)
\]

A linearized shallow water solution (for water depth much smaller than the tsunami wavelength) can be obtained by the Fourier–Laplace transform, defined as

\[
\tilde{\eta}(k; s) = \int_{-\infty}^{\infty} \tilde{\zeta}(x; s) \int_{-\infty}^{\infty} e^{-iky} \int_{-\infty}^{\infty} e^{-s\tau} f(x, y; \tau) d\tau dy. \quad (3)
\]

Transformation of the equation of motion and boundary conditions, and the assumptions of linearity and shallow water lead to the following solution for the transform of \( \tilde{\eta}(k; s) \), in terms of the transform of \( \zeta \), \( \tilde{\zeta}(k; s) \) [3,9]

\[
\tilde{\eta}(k; s) = \frac{s^2 \tilde{\zeta}(k; s)}{s^2 + \omega^2 \cosh kh},
\]

where \( \omega \) is circular frequency and

\[
\omega^2 = |gk \tanh kh|.
\]

Then, a solution for \( \eta(x, y; t) \) is obtained as follows: (1) \( \zeta(x, y; t) \) is transformed to obtain \( \tilde{\zeta}(k; s) \), (2) \( \tilde{\eta}(k; s) \) is computed from Eq. (4) after substitution of \( \tilde{\zeta}(k; s) \), and (3) \( \eta(x, y; t) \) is computed by performing inverse transform. For the models we consider in this paper, the forward and inverse Laplace transforms are evaluated analytically using tables, and the Fourier transforms are evaluated numerically using fast Fourier transform (FFT).

The above constitutes a linearized solution, known as the ‘shallow water solution’. It works well if the water depth is much smaller than the wavelength of the water waves [3]. Inclusion of the neglected nonlinear terms in the boundary conditions would have permitted a solution involving a solitary wave [17]. Other simplifications adopted in the above solution involve the assumption that the ocean floor is rigid and that \( \zeta(x, y; t) \) may be assumed a priori. This assumption thus neglects the coupling of submarine slides and surface waves [18].

2.2. Model of a tsunami source spreading in two dimensions

A detailed and complete presentation of tsunami waveforms for sources spreading in two dimensions is complex and too voluminous to be shown. This is because the number of cases required to illustrate the variations with respect to source geometry and nature of spreading are prohibitively large. In this paper, we consider one such source model as shown in Fig. 2. We chose this model as an example, and to describe the general characteristics of the tsunami following the Grand Banks event of 1929.

The model in Fig. 2 is a two-dimensional generalization of Model 1B in Ref. [8]. It represents mass movement initiated at \( x = 0 \) and \( y = -W/2 \) and spreading bilaterally in the \( x \)-direction and unilaterally in the \( y \)-direction. The accumulation zone spreads with velocity \( c_R \) in the positive \( x \)-direction (downhill), the depletion zone spreads with velocity \( c_L \) in the positive \( x \)-direction and with velocity \( c_{1L} \) in the negative \( x \)-direction (uphill), and both zones spread in the positive \( y \)-direction with velocity \( c_p \). In the \( x \)-direction, the spreading stops at time \( t''_L \), which is the time when the accumulation zone reaches \( x = L_R \) and the depletion zone reaches \( x = -L_{1L} \), and in the \( y \)-direction it stops at time \( t''_p \) when the slide reaches the line \( y = W/2 \). Fig. 2 shows the end configuration of the slide, for the case when \( t''_L < t''_p \), and for uniform vertical motion amplitudes of the slide, \( \zeta_0 \) for the accumulation zone and \( \zeta_1 \) for the depletion zone. The length of the depletion zone and the amplitudes \( \zeta_0 \) and \( \zeta_1 \) are determined by specifying the ratios \( c_R/c_p, c_R/c_{1L}, c_{1L}/c_R \), and \( c_L/c_R \), the depth of the fluid layer \( h \),
the width of the slide \( W \), and the characteristic length in \( x, L_R \).

In general the sliding can go on up to the time of maximum source duration \( t = L/c_R + W/c_y \) (then the plan view of the slide becomes a rectangle), but in most of the examples shown in this paper, we will 'stop' the source motion and analyze the tsunami amplitudes at time \( t' = W/c_y \) (Fig. 2).

The slide in Fig. 2 is equivalent to a superposition of three spreading uplifts, \( \zeta^{(R)} \), \( \zeta^{(L)} \) and \( \zeta^{(LL)} \), defined as

\[
\zeta^{(R)}(x, y; t) = \zeta_0 H(t - \frac{y + W/2}{c_y} - \frac{x}{c_R})
\]

(6)

\[
\zeta^{(L)}(x, y; t) = -(\zeta_0 + \zeta_1) H(t - \frac{y + W/2}{c_y} - \frac{x}{c_L})
\]

(7)

and

\[
\zeta^{(LL)}(x, y; t) = -\zeta_1 H(t - \frac{y + W/2}{c_y} + \frac{x}{c_{LL}}).
\]

(8)

The Laplace–Fourier transform of such \( \zeta(x, y; t) \) is then

\[
\bar{\zeta}(k_x; s) = \int_{-W/2}^{W/2} e^{ik_x y} \left\{ \int_0^{L_R} e^{i\pi k_x x} \left( \int_0^\infty e^{-\pi k_x^2} \zeta^{(R)}(x, y; t) dt \right) dx + \int_0^{c_L/c_y} e^{ik_x x} \left( \int_0^\infty e^{-\pi k_x^2} \zeta^{(L)}(x, y; t) dt \right) dx + \int_{-c_{LL}/c_y}^0 e^{ik_x x} \left( \int_0^\infty e^{-\pi k_x^2} \zeta^{(LL)}(x, y; t) dt \right) dx \right\} dy.
\]

(9)

Both the forward and inverse Laplace Transforms involved in the evaluation of \( \bar{\zeta}(k_x; s) \) and of \( \eta(x, y; t) \) can be evaluated analytically, but the expressions are lengthy and the reader is referred to Ref. [19] for further details.

3. Results

In the following, we vary the parameters of the tsunami source model in Fig. 2 and present the resulting waveforms and peak amplitudes in the near-field. The purpose is to illustrate the general characteristics of tsunami generated by this source model (spreading in two dimensions), before it is used to interpret the tsunami following the 1929 Grand Banks earthquake.

3.1. Tsunami waveforms for different velocities of spreading

The combinations of ratios \( c_R/c_y \) and \( c_R/c_T \) of the model in Fig. 2, which we consider in this study, are summarized in Table 1 (second row and second column). The first row and first column show indices \( J \) and \( I \) associated with the selected values of \( c_R/c_y \) and \( c_R/c_T \). Common parameters in all these examples are: the water depth \( h = 2 \) km (implying \( c_T = \sqrt{gh} = 0.14 \) km/s), the total length of sliding in the positive \( x \)-direction, \( L_R = 82 \) km (equal to part of the length of the region of depletion and the entire length of the region of accumulation), the extent of the region of depletion in the negative \( x \)-direction, \( L_{LL} = 50 \) km, and the width of the slide \( W = 265 \) km. Also, in all the examples, \( c_{LL} = 0.61c_R \) and \( c_L = 0.1c_R \).

The maximum positive and minimum negative tsunami amplitudes for all of these cases, computed at the end of sliding [19], are shown in Table 1 and Fig. 3(a) and (b). The dots in Fig. 3(a) and (b) correspond to the combinations of values of \( c_R/c_y \) and \( c_R/c_T \), and the numbers next to the dots show the peak wave amplitudes. The weak solid lines show contours of the peak wave amplitudes. Figs. 4–10 show the tsunami waveforms at the end of sliding (at time \( t' = W/c_y \)) only for selected characteristic cases: for a ‘fast’ spreading of the slide (Fig. 4), for ‘slow’ spreading of the slide (Figs. 8 and 10), and for \( c_D \) and \( c_A \) such that focusing amplification occurs along the spreading edge AB and/or BC (Fig. 2).

3.2. Peak tsunami amplitudes as function of \( c_A/c_T \) and \( c_D/c_T \)—amplification caused by focusing

Trifunac et al. [8] presented numerous examples and
Fig. 3. (a) Contours $(\eta/\zeta_0)_{\text{max}}$ versus $c_R/c_y$ listed in Table 1. (b) Contours $(\eta/\zeta_0)_{\text{min}}$ versus $c_R/c_y$ listed in Table 1.
described how the amplitude of the leading tsunami could be amplified when \( c_R/c_T \sim 1 \), for slide and slump sources which spread only in \( x \)-direction. Their results correspond to \( c_T \to \infty \) and for amplification along the slide edges CD and EA in Fig. 2. Here we show how the same phenomenon occurs along the spreading edge AB (for the depletion zone) moving out with velocity \( c_D \) and spreading edge BC moving out with velocity \( c_A \). From Fig. 2, it is seen that

\[
\beta = \tan^{-1} \frac{c_{LL}}{c_y},
\]

\( c_R/c_T = 1 \)
\( c_R/c_T = 5 \)
\( h=2 \) km
\( W=265 \) km
\( L_{LL}=82 \) km
\( L_{LL}=50 \) km
\( \eta/\zeta_0 = 1.17 \)
\( (\eta/\zeta_0)_{max} = 1.17 \)
\( \eta/\zeta_0 = 2.35 \)
\( (\eta/\zeta_0)_{max} = 2.35 \)

and

\[
\alpha = \tan^{-1} \frac{c_R}{c_y}.
\]

For focusing amplification of wave amplitudes to occur, it is necessary to have \( c_D/c_T \to 1 \) and or \( c_A/c_T \to 1 \) where

\[
c_D = c_{LL} \cos \beta \quad \text{or} \quad c_D = c_y \sin \beta,
\]

and

\[
c_A = c_R \cos \alpha \quad \text{or} \quad c_A = c_y \sin \alpha.
\]

---

**Table 1**

Summary of \( (\eta/\zeta_0)_{max} \) and \( (\eta/\zeta_0)_{min} \) versus \( c_R/c_y \) and \( c_R/c_T \) for 0.3 \( \leq c_R/c_T \leq 50 \) (\( J = 0, 1, 2, \ldots, 9 \)) and 0.5 \( \leq c_R \leq 20 \) (\( J = 0, 1, 2, \ldots, 8 \)), for the source model in Fig. 2, with \( W = 265 \) km, \( L_R = 82 \) km, \( h = 2 \) km, \( c_{LL} = 0.61c_R \) and \( c_I = 0.1c_R \) at time \( t^* = W/c_y \).

<table>
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<th>( J = )</th>
<th>( c_R/c_T )</th>
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<th>3</th>
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<td>8</td>
<td>20</td>
<td>1.42</td>
<td>9.75</td>
<td>0.11</td>
<td>–2.04 –9.94 –0.11</td>
<td></td>
<td></td>
<td></td>
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<td>1.48</td>
<td>8.46</td>
<td>0.21</td>
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<td>6</td>
<td>5</td>
<td>1.17</td>
<td>1.27</td>
<td>9.40</td>
<td>0.22</td>
<td>–2.35 –2.36 –9.90 –0.23</td>
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<td></td>
<td></td>
<td></td>
</tr>
<tr>
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<td>2</td>
<td>2.24</td>
<td>3.07</td>
<td>4.26</td>
<td>4.03</td>
<td>0.12</td>
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<tr>
<td>4</td>
<td>1.5</td>
<td>2.34</td>
<td>3.40</td>
<td>3.06</td>
<td>4.75</td>
<td>2.44</td>
<td>0.69</td>
<td>–4.34 –4.58 –4.18 –3.22 –4.37 –9.75</td>
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<td>3</td>
<td>1</td>
<td>4.64</td>
<td>4.80</td>
<td>2.66</td>
<td>1.26</td>
<td>0.36</td>
<td>–2.52 –4.30 –2.91 –2.25 –0.39</td>
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<tr>
<td>2</td>
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<td>1.13</td>
<td>0.85</td>
<td>0.87</td>
<td>0.32</td>
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<td>0.54</td>
<td>–0.22 –0.74</td>
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**Fig. 4.** Normalized surface displacement \( \eta/\zeta_0 \) at time \( t = t^* = W/c_y \), for \( c_R/c_T = 1.5 \) and \( c_R/c_T = 5 \).

**Fig. 5.** Normalized surface displacement \( \eta/\zeta_0 \) at time \( t = t^* = W/c_y \), for \( c_R/c_T = 1 \) and \( c_R/c_T = 2 \).
Defining
\[ \frac{c_R}{c_T} = \eta, \]
we obtain
\[ \frac{c_D}{c_T} = \frac{c_D}{c_R} \eta = \frac{c_{LL}}{c_R} \eta \cos \beta, \]
(12)
\[ \frac{c_A}{c_T} = \frac{c_A}{c_R} \frac{\eta}{\eta} \cos \alpha. \]
(13)

For a given \( \alpha \) and \( \beta \), we can compute \( c_R/c_T \), which will result in amplification by focusing along AB or BC (Fig. 2)

\[ c_R/c_T = 2 \quad \text{and} \quad c_R/c_T = 2. \]

Fig. 6. Normalized surface displacement \( \eta/\zeta_0 \) at time \( t' = W/c_T \), for \( c_R/c_T = 2 \) and \( c_R/c_T = 2 \).

Fig. 7. Normalized surface displacement \( \eta/\zeta_0 \) at time \( t' = W/c_T \), for \( c_R/c_T = 2 \) and \( c_R/c_T = 2 \).

Fig. 8. Normalized surface displacement \( \eta/\zeta_0 \) at time \( t' = W/c_T \), for \( c_R/c_T = 5 \) and \( c_R/c_T = 2 \).

Fig. 9. Normalized surface displacement \( \eta/\zeta_0 \) at time \( t' = W/c_T \), for \( c_R/c_T = 0.5 \) and \( c_R/c_T = 1 \).
respectively. Table 2 lists the selected values of ‘depletion’ zones, i.e. to edges BC and AB in Fig. 2, where subscripts ‘A’ and ‘D’ refer to ‘accumulation’ and... for the model in Fig. 2 for two examples: (1) \( c_R/c_T = 0.3 \) and \( c_R/c_T = 1 \) and (2) \( c_R/c_T = 20 \) and \( c_R/c_T = 20 \).

\( (\alpha \rightarrow 90^\circ) \), the equivalent \( \text{L} \rightarrow \text{W} = 265 \text{~km} \), when \( c_R/c_T \rightarrow 0 \) \( (\alpha \rightarrow 0^\circ) \), \( \text{L} \rightarrow \text{L}_R = 82 \text{~km} \) (Fig. 2), and \( h = 2 \text{~km} \).

**Fig. 11.** Peak tsunami amplitude \( \eta_{\max}/s_0 \) versus \( \text{L}/h \) for the model in Fig. 2 superimposed on the curves of Ref. [8] corresponding to \( \text{W}/\text{L} = 0.5 \) and to \( \text{W}/\text{L} \geq 0.5 \). The two points correspond to \( c_R/c_T = 20 \) (implying \( \alpha = 87.1^\circ \)) and equivalent \( \text{L} = \text{W} \sin(87.1^\circ) = (265)(0.999) = 278.5 \text{~km} \) and \( c_R/c_T = 20 \) and to \( c_R/c_T = 0.3 \) (implying \( \alpha = 16.7^\circ \)) and equivalent \( \text{L} = \text{L}_R \cos(16.7^\circ) = (82)(0.757) = 76.5 \text{~km} \) and \( c_R/c_T = 1 \) and \( h = 2 \text{~km} \).

It is seen that both points approach (from below) the results for the two-dimensional model spreading in \( x \)-direction only. The geometrical features of the model in Fig. 2, of course, change continuously with \( \alpha \) and \( \beta \) that is with \( c_R/c_T \), but it is clear that along \( n_A \) and \( n_D \), versus \( \xi = c_R/c_T \) (Fig. 3(a) and (b)), the upper bound on \( \eta_{\max}/s_0 \) (Eq. (15)) and lower bound on \( \eta_{\min}/s_0 \) (Eq. (16)) can be approximated by the results of \( \eta/s_0 \) versus \( \text{L}/h \) for the two-dimensional models. These results suggest that, locally, for any submarine slide, slump or uplift of ocean bottom by earthquake dislocation, any portion of the spreading uplift which is ‘straight’ and which spreads out for distance \( \text{L} \), the peak tsunami amplitudes will grow versus \( \text{L}/h \) as shown in Fig. 11.

**3.3. Peak tsunami amplitudes as functions of \( \text{L}/h \)***

Trifunac et al. [8] showed that, for their model of a slide spreading only in the \( x \)-direction, the peak tsunami amplitudes \( \eta_{\max}/s_0 \) and \( \eta_{\min}/s_0 \) (computed for \( c_R/c_T = 1 \) and for \( \text{W}/\text{L} > 0.25 \), where \( \text{L} \) is the length of the accumulation zone) grow monotonically versus \( \text{L}/h \). In the examples presented in this paper, when \( c_R/c_T \rightarrow \infty \)

\[
\eta_A = \frac{c_R}{c_T} \cos \alpha, \quad \eta_D = \frac{1}{c_{LL} \cos \beta},
\]

where subscripts ‘A’ and ‘D’ refer to ‘accumulation’ and ‘depletion’ zones, i.e. to edges BC and AB in Fig. 2, respectively. Table 2 lists the selected values of \( \eta_A \) and \( \eta_D \) for \( c_R/c_{LL} = 1/0.61 \). The thick lines in Fig. 3(a) and (b) show Eqs. (15) and (16), superimposed on the contours of the peak tsunami amplitudes (maxima and minima in Fig. 3(a) and (b), respectively) computed at the end of sliding for all the combinations of \( c_R/c_T \) and \( c_R/c_T \) considered in this paper.

**Table 2**

<table>
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<th>( \xi = c_R/c_T )</th>
<th>0.3</th>
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<th>0.7</th>
<th>1.0</th>
<th>1.5</th>
<th>2.0</th>
<th>5.0</th>
<th>10.0</th>
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<td>1.22</td>
<td>1.41</td>
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<td>5.10</td>
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<td>( \beta )</td>
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<td>23.1</td>
<td>31.4</td>
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<td>50.7</td>
<td>71.9</td>
<td>80.7</td>
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<td>( \eta_D )</td>
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<td>1.78</td>
<td>1.92</td>
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<td>2.58</td>
<td>5.25</td>
<td>10.12</td>
<td>20.06</td>
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4. Case study: tsunami following the 1929 Grand Banks earthquake

In the following, we first review the data and selected previous studies of the submarine slide of 1929 associated with the Grand Banks earthquake in Atlantic Canada. Then we suggest that the tsunami it generated could be interpreted to have resulted from a slide spreading in two directions.

4.1. Data and a review of selected previous studies

4.1.1. The earthquake

The Grand Banks earthquake occurred on 18 November 1929 at 20 h, 32 m, 00 s at 44.69°N and 56.00°W [20]. It had surface wave magnitude $M_S = 7.2$ and body wave magnitude $m_B = 7.2$ (Fig. 12). It has been suggested that the earthquake occurred along a landward extension of the New Foundland Fracture zone (a transform plate boundary, [21]), or along strike slip faults [22]. It was also suggested [23] that the source mechanism could be represented by just a slump. Hasegawa and Kanamori [15] investigated the strike–slip, thrust and single-force mechanisms, using seismograms at 14 stations, and concluded that a slump with a single-force model is most compatible with the recorded seismological data. In their model, about 550 km$^3$ of sediments is assumed to have slumped creating a single force of about $1.4 \times 10^{20}$ dyn. They describe a single couple source model, lasting about 50 s, and with net impulse of this force equal to zero.

4.1.2. The slump

The aerial extent of the slump (shown by dashed lines in Fig. 12) is based on seismic reflection work, core samples, and the area outlined by instantaneous breaks of the submarine telegraph cables [24–28]. This slump is about 260 km wide and 135 km long. Hasegawa and Kanamori [15] estimated its area as 37,500 km$^2$ and the average thickness of the moving material as 15 m.

The distribution of instantaneous cable breaks (shown by solid points in Fig. 13) defines approximately the area where the sediment failure was triggered by the earthquake. The upper part of the continental slope was not affected by the failure, but it was dissected by head wall scarps at 600 m water depth. On St. Pierre Bank there was no sediment failure on the upper slope. Headwall scarps begin at about 600 m water depth, and continue down slope into an area of extensive sediment failure, which has affected only the uppermost few meters of sediment [27]. Assuming the area of $20 \times 10^9$ m$^2$ and average thickness of 5 m, the total volume of sediments slumped from the continental slope is 100 km$^3$, consisting of gravel, sand, and coarse silt (30%), and of fine silt and clay (70%). The upper fan valleys, with total area of $4 \times 10^9$ m$^2$, had unknown thickness of erosion, but if it is assumed that this was 40 m, then this would have contributed 160 km$^3$ of sand and gravel [27].

4.1.3. The turbidity current

The slump following the 1929 Grand Banks earthquake triggered the first well-documented turbidity current on the
ocean floor [25]. This current moved as far as 1700 km, rupturing the trans-Atlantic telegraph cables at 28 places [29]. It traveled down the continental rise and then along the ocean floor, at first with velocity of 55 knots (28.3 m/s) near the foot of the continental slope, slowing down to 11 knots (5.6 m/s), some 300 nautical miles (555.6 km) further south [25]. It is estimated that it deposited 24 km$^3$ of sand and gravel over the area of \(8 \times 10^9\) m$^2$ of fan valleys (assuming 3 m thickness), and 160 km$^3$ of sand and coarse silt (95%) and fine silt and clay (5%), over the northern Sohm Abyssal Plain (shown by gray zone in Fig. 13 [27]).

To explain the excess sand on the Sohm Abyssal Plain, Piper and Aksu [27] suggest that “…removal of an average thickness of 30–40 m of coarse sediment from the upper valleys would be required. Failure of this coarse sediment in 1929 may have been by liquefaction and retrogressive sliding: process that if reasonably efficient would not leave any clear evidence in the failure area…”

4.1.4. The tsunami

The Grand Banks event of 1929 generated a tsunami with at least 12.2 m amplitude in Burin inlet [30] and killed 26 people. Gregory [31] estimated the maximum amplitude in Burin inlet to be 30.5 m. At Lamaline, the tsunami amplitude was 4.6 m [32]. No tsunami was observed at Sable Island [30]. At Halifax, the amplitude was only 0.5 m (Fig. 14), and at Atlantic City it was 0.3 m. Murty and Wigen [33] showed that resonance in the V-shaped Burin inlet could explain amplification of tsunami in the inlet. Travel time curves for this tsunami [34] show that “the tsunami energy traveled preferentially toward south coast of Newfoundland” (Fig. 15).

4.2. An interpretation by a source model spreading in two dimensions

To show how the model in Fig. 2 could be used to explain the observed tsunami amplitudes, we consider the model geometry shown in Fig. 12. For this model, the x-axis is oriented in the S30°E direction, and the positive y-axis in the E30°N direction. The combined length of the final depletion...
and accumulation zones is assumed to be 50 and 82 km, respectively. The assumed width of the slide is 265 km. The movement is initiated at \( x = 0 \) and \( y = -W/2 \) (at \( y = -128 \) km). We assume that the accumulation zone then spreads downhill with constant velocities, \( c_R \) at the tip of the slide and \( c_L = 0.1c_R \) at its tail. The tip of the depletion zone is assumed to spread uphill with velocity \( c_{RL} = 0 \); \( c_R \) (retrogressive sliding), and with average amplitudes \( \zeta_0 \) and \( \zeta_1 \) (Fig. 2), chosen such that the cross-section areas of the accumulation and depletion zones \( (A_1 \) and \( A_2) \) are the same.

The largest tsunami amplitudes from this source will be directed towards Burin when \( b \) is close to 42\(^\circ\) (Fig. 2). This occurs close to the case \( c_R/c_y = 1.5 \) (with \( \beta = 42.5\)\(^\circ\), see Table 2), and will be associated with large tsunami amplitudes when \( \eta_D = (0.61 \cos \beta)^{-1} \), near \( c_{R} / c_T = 2 \) (Fig. 3(a) and (b)). If \( h = 2000 \) m, then \( c_T = 0.141 \) km/s, \( c_R = 0.280 \) km/s and \( c_y = 0.187 \) km/s. With these velocities, it takes 292 s (4.88 min) for the slide to extend to \( L_R = 82 \) km (down slope), and 1417 s (23.6 min) to spread over its entire width of 265 km. These durations are not inconsistent with the reported times of cable breaks (Fig. 12), but are longer than the source durations, of about 50 s, discussed by Hasegawa and Kanamori [15].

Fig. 16 shows tsunami amplitudes plotted versus normalized time \( t/t^* \left(t^* = W/c_y \right) \), for \( c_R/c_y = 1.5 \), \( c_R/c_T = 2 \) and \( c_T = 0.141 \) km/s (i.e. \( h = 2000 \) m). The maximum and minimum amplitudes, assuming \( \zeta_0 = 3 \) m, are shown in Fig. 14 and are compared with the observed amplitudes. The value of \( \zeta_0 = 3 \) m was not fitted to the reported peak amplitudes, because the azimuthal coverage is limited and the number of observation points is also small. It was chosen approximately, to be consistent with the observed amplitudes at Halifax and Lamaline, and to underestimate the observed amplitude at Burin, which may have been amplified by the geometry of the Burin inlet [33].

The published estimates of the slide thickness \( (\zeta_0) \) and volume \( (V) \) vary over a broad range. Hasegawa and Kanamori [15] give \( \zeta_0 \sim 15 \) m and \( V = 550 \) km\(^3\). Piper and Aksu [27] give \( V = 100 \) km\(^3\) and \( \zeta_0 \sim 5 \) m, while Driscoll et al. [35] give \( V = 185 \) km\(^3\), which would imply \( \zeta_0 \sim 5 \) m.

5. Discussion and conclusions

In this paper, we presented a tsunami source model which is a rectangular slide (slump) spreading in both \( x \) and \( y \) directions. This model is a generalization of Model 1B in Ref. [8], which can spread only in the \( x \)-direction.
consequence of this generalization is that focusing now can occur in any direction, determined by the ratios of the velocities of sliding in the x and y directions. In the example we chose for this presentation, the spreading is bilateral in the x-direction (both the accumulation and depletion zones spread out with different velocities) and unilateral in the y-direction. Our model can readily be used to compute tsunami waveforms from a bi-directional slide spreading in the y direction, but we chose to restrict the examples to unilateral spreading to simplify the presentation, and because we were interested to explore the applicability of such models to the Grand Banks tsunami of 1929.

Following the Grand Banks tsunami of 1929, six submarine telegraph cables, lying on the continental slope between 150 fathoms (83 m) and 1800 fathoms (1000 m), and between 54.5°W and 57.5°W, were broken ‘instantaneously’ by ground shaking or by almost instantaneous and between 54.5°W and 57.5°W, were broken ‘instantaneously’ by ground shaking or by almost instantaneous

breaks occurred 59 min after the earthquake (Breaks No. 17 and 18, for Imperial Cable Company Halifax—Tayal at 43°37′N, 55°15′W and at 43°15′N and 56°07′W, shown in Figs. 12 and 13 [25]).

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