TSUNAMIS

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Contents

Summary

Tsunamis are often triggered by submarine earthquakes. In order to estimate an initial profile of a tsunami from an earthquake's fault parameters, the elastic dislocation model is usually used with the assumption that the slip occurs uniformly on the fault plane in a semi-infinite elastic body. Actual tsunami is sometimes quite different from this assumed model profile. Efforts have been made to fill this gap. In the sea deeper than 200 m, a tsunami can be modeled reasonably with the linear long wave theory. In the shallow sea, nonlinear effects become nonnegligible and the fully nonlinear shallow-water theory is used. Tsunamis increase their height in the shallow sea, due to the shoaling, focusing, resonance, and dispersion effects. Tsunami "magnitude" is a measure to express the total energy of a tsunami, while tsunami "intensity" is a measure to express the strength of a tsunami at a given location. Damages to human lives and properties are given as a function of tsunami intensity. Fires induced by a tsunami often give devastating damages, especially when they are associated with oil spill. Tsunami countermeasures consist of three parts: hardware such as sea walls, city planning such as relocation of residence and introduction of tsunami-resistant building zones, and software such as forecasting, evacuation drills, and continuation of disaster culture.

1. Introduction

Earthquakes are the primary cause of most tsunamis. But tsunamis are a more infrequent phenomenon than are earthquakes. A weak earthquake does not generate a tsunami. A great earthquake with the hypocenter at a deep location does not generate a tsunami. Neither does an earthquake caused by strike-slip fault movement generate a tsunami. The recurrence interval of huge tsunamis is usually much longer than a life span of human beings. Even a small tsunami occurs very infrequently. Difficulties in understanding tsunamis come from this infrequent occurrence.

It is difficult to understand the whole picture of a tsunami. Each tsunami is quite different from another. Each tsunami behaves in quite different movements at different places, corresponding to different topography. Tide records provide quantitative time-series data, but there are two problems. One is the distribution and number of tide stations. Compared to the spatial extent of a tsunami, tide stations are often too sparse. In addition, tsunami data taken by tide gages are biased and deteriorated by the hydraulic characteristics of the tide well. High-frequency components of the water-surface fluctuations are cut off or filtered. Other kinds of data, more rich in number, are tsunami traces measured in the post-tsunami survey. They usually show the highest water level made by the tsunami but can not suggest any hydrodynamics or time histories of the tsunami runup. They are “fossils of tsunami." Hence, difficulties in understanding tsunamis also come from the lack of reliable data.

Once it has occurred, a tsunami gives devastating effects on the coastal community. The last and best way to save human lives is an early evacuation guided by forecasting and warning. The second best is to make a coastal city tsunami resistant. It is quite difficult for coastal residents to continue to be alert to tsunami hazards based on former experiences. At such places as the west coast of the United States where there are no written records at all about huge tsunamis some 300 years ago, except for one proof (i.e., sediments transported by the tsunami), it is impossible for the residents to learn of the former event. If a newly developed coastal resort community, without prior tsunami records, finds one day that they are located in the tsunami risk area in front of the subduction zone, how can they prepare for possible tsunamis?

During these 30 years, the areas of tsunami science and engineering made big progress, assisted by advancement in seismology and computer science. Seismology made it possible to estimate the initial profile of tsunamis from fault parameters determined by seismic information. The nonlinear shallow-water theory is used to solve for tsunamis on complicated topography with the aid of a big high-speed computer. The technique thus developed is now applied to practical defense works in many fields, forecasting, hazard maps, design of defense structures, and so on.

2. Causes of Tsunamis

The sea surface always fluctuates. There are three causes for wave motions:

meteorological, astronomical, and the rest. Waves generated by winds become swells after they leave the wind field. Typical wave periods of wind waves and swells are 3 to 30 seconds. Moving atmospheric low pressure together with the winds accompanying it causes the sea-surface rise, called the storm surge. A typical storm surge period is several hours to half a day. Tides are more-regular oscillations caused by the pulling gravitational forces of the moon and the sun. The fundamental period of tides caused by the moon is 12 hours and 25 minutes. Water waves generated by other causes are tsunamis. Its period ranges from a few minutes to two hours.

The first person who recorded a tsunami and thought that the tsunami was generated by an earthquake was Thucydides, a Greek historian. In summer of 426 BC during the Peloponnesian war, the sea receded after an earthquake, and then a huge wave hit the city of Orobiae at the northwestern coast of Euboea Island. A part of the city subsided into the sea. The island of Atlanta at the other side of the strait was also hit by the tsunami. A part of Athenian fortifications was swept away. Thucydides considered that the full force of the earthquake drew the seawater from the shore and then the sea suddenly swept back again even more violently.

Most of the causes of tsunamis are submarine earthquakes. However, not the ground shaking but the vertical sea-bottom deformation generates a tsunami. The greater the earthquake is, the larger the vertical displacement of sea bottom is; hence the greater the tsunami that is generated. An earthquake that triggers a tsunami is often called the tsunamigenic earthquake.

There are exceptions for this rule. Much larger tsunamis than expected from the earthquakes' seismic waves can be generated. This type is called the tsunami earthquake. After several weak earthquakes on 15 June 1896, another weak earthquake was felt along the shore of the Sanriku District, Japan, at 1930 (local time). No one paid special attention to this earthquake and tried to evacuate. Half an hour later, a giant tsunami hit the coast and claimed more than 22 000 human lives. The highest runup was nearly 40 m. This was a typical tsunami earthquake. The mechanism of the tsunami earthquake in the Sanriku coast was thought to be a large but slow rupture of the earthquake fault.

A huge tsunami can cause damage to remote places after travelling across the ocean. This is called a distant tsunami, remote-source tsunami, or far-field tsunami. At 1911, on 22 May (GMT) 1960, an earthquake of $M_s = 8.5$ or $M_w = 9.5$ struck off the Chilean coast. A giant tsunami was generated and hit the Chilean Coast first as a local tsunami or a near-field tsunami. Its maximum runup height was estimated as high as 20–25 m. The tsunami spread over the Pacific Ocean, gave damage to Hawaii, then concentrated toward Japan at the antipode of the tsunami source after a 22.5-hour journey. The whole coast on the Pacific Ocean side of Japan, more than 3000 km long, were affected and damaged. Its tsunami height was 3–6 m.

Landslides can also generate tsunamis. Lituya Bay in Alaska repeatedly experienced huge local tsunamis in 1958, 1936, 1899, 1874, 1853–1854, and also probably in 1900. This bay is about 11 km long, 1 km wide, and 160 m deep. On 10 July 1958, an earthquake caused 30×10^6 m^3 of rocks weighing 90×10^6 tons to slide from the northern shore from an average height of 600 m with the dimensions of 700 m to 900 m and an average thickness of 90 m. The slide forced the water surge up to the height of 520 m on the opposite shore. Then, the water ran down into the bay to form a huge tsunami higher than 30 m in the bay.

Another mechanism is volcanic actions. In May 1883, the volcanic eruption of Krakatau in the Sunda Strait between Java and Sumatra Islands, Indonesia began. On 27 August, a giant tsunami was generated. Because of thickly falling ashes, no one could see the tsunami offshore. When coastal residents noticed the white cap of the tsunami, the tsunami was just in front of them. There was no time for evacuation. It claimed more than 36 000 human lives. Its maximum runup was higher than 30 m. The comparison of the topographies before and after the eruption indicates that two-thirds of the original Krakatau island was blown away, leaving an area 200 m deep and about 10 km wide. Among several generation mechanisms proposed, caldera formation is the most probable.

3. Hydrodynamics of Tsunami from Generation to Coastal Effects

3.1. Generation

A fault movement generates shaking (earthquake) and displacement of ground. If a submarine fault is shallower than 80 km, the displacement might penetrate to the surface of sea bottom. In order to imagine the generation of a giant tsunami, let us suppose that an area of sea bottom, a few hundred kilometers long and several tens to 100 kilometers wide, moves vertically by several meters within 100 seconds. The water a few kilometers thick above the area has no time to flow outward and consequently the water surface will show the same vertical displacement as the sea bottom. In this way a tsunami is born.

No person could and/or will be able to measure the tsunami initial profile, but one can estimate it by calculation or, very rarely, by direct measurement of the sea bottom displacement. The most popular way of estimating an initial tsunami profile is to assume that the sea bottom displacement is a result of the fault movement in a semi-infinite, elastic, homogeneous body. A fault movement is described by its location including its depth, geometrical characteristics (strike, dip, and slip angles of the fault plane), physical characteristics (length, width, and dislocation of the fault plane) and dynamic characteristics (rupture direction, rupture velocity, and rise time of the fault movement). With fault parameters (except for dynamic characteristics), the static displacement of sea bottom can be computed.

The assumption of a homogeneous movement in a fault plane leads us to a simple tsunami profile: one crest and one trough in the area of the generation. Efforts have been made to obtain more realistic initial profiles. If a rupture process is well recorded and analyzed, plural fault planes, if they exist, can be determined with fault parameters for each plane. Heterogeneous movement in a fault plane is estimated from inversion of seismic data or of tsunami data with geodetic data. In place of the semi-infinite body model, the multilayered model can be introduced and numerically solved. The introduction of heterogeneity changes the simple initial profile (one crest and one trough) to a complicated one.

A tsunami initial profile determined with seismic data alone usually does not explain the tsunami and total tsunami energy. It is necessary to modify this first solution with results of tsunami simulation. First, compute a tsunami for the initial profile determined from seismic data and output the tsunami heights along the 200 m water-depth contour near the shore. Compare these results with the measured runup heights averaged in the interval of about 15 km along the open coast. The latter is usually two to three times the former, according to accumulated experiences in numerical simulation. Adjust the initial height to satisfy this condition. This method is widely used to determine the initial condition.

There is no accurate measurement of the vertical displacement of sea bottom surface caused by a submarine earthquake except for the case of the 1964 Great Alaska earthquake. The deformation of ground was reconstructed by the displacement on islands and by comparison with a pre-earthquake topography. Along the direction normal to the long axis of the deformation area, the vertical displacement (or the two-dimensional tsunami profile) shows a gentle wavy shape 450 km long with one crest and one trough. Its trough-to-crest wave height was about 6 m. Near its crest, there was a sharp rise about 6 m high and about 30 km wide at its base. This rise was concluded as a result of a subfault developed in the accretionary prism. It is impossible to detect this kind of subfault from seismic information at present, although its contribution to tsunami heights is important.

There are several efforts to estimate tsunamis generated by landslides, submarine landslides, volcanic action, and others. Most of them are carried out based upon the measured tsunami data. Geodetic data alone are insufficient in accuracy to estimate the generation mechanism because the movements of landslides are the key factor to determine the tsunami generation efficiency.

3.2. Propagation

When generated, a tsunami has a wavelength several tens of kilometers long that is much longer than the water depth, at most a few kilometers. For example, the average water depth in the Pacific is about 4.2 km. This water wave is categorized as long waves, for which the hydrostatic water pressure is a good first-order approximation. The initial height of the tsunami several meters high is quite small compared to the water depth and the wavelength. This water waves belongs to small amplitude waves, for which the linear wave theory is applicable.

A near-field tsunami in the ocean deeper than approximately 200 m can be analyzed with the linear long-wave theory. The velocity of its energy propagation is the same as its phase velocity that is the square root of the product of the gravitational acceleration and the water depth for the first-order approximation. Every component of different frequencies is approximated to propagate with the same velocity. A linear long wave propagating on the water of constant depth, therefore, does not change its wave profile.

For a higher-order approximation, the phase velocity is influenced by the dispersion effect depending upon frequency. Wave components of different frequencies propagate with different velocities. This difference, although very small, results in a non-negligible deformation in wave profile, if the travel time becomes long as in case of a far-field tsunami. A parameter p_a is used to judge whether the dispersion effect should be included or not:

$$p_a = (6h/R)^{1/3}(a/h)$$

where h is the water depth, a the horizontal dimension of the tsunami source, and R the distance from the source. If $p_a < 4$, the dispersion effect should not be neglected. Under this condition, the linearized Boussinesq equation that includes the first-order effect of the phase dispersion should be used. The equation should also be modified by including the Coriolis effects and expressed with the spherical coordinates.

For a huge tsunami, such as the 1960 Chilean tsunami, the Pacific Ocean behaves like a small pond. Average water depth of the Pacific Ocean, 4.2 km, gives the tsunami a propagation velocity faster than 730 km hr^{-1}. The tsunami traveled 17 000 km from the source off Chilean coast to Japan within 23 hours. It started toward Japan with the crest at its front but arrived at Japan with a big ebb. The first crest became unrecognizably small and the following trough began to grow near the Hawaiian Islands. This change was the result of the dispersion due to the Coriolis effect.

The energy of the 1996 Irian Jaya tsunami was effectively transported to Japan although Japan is not located on the major direction of initial tsunami energy radiation. This is explained by the existence of the south Honshu ridge that acted as an effective wave guide. The shallower the water is, the slower is the wave propagation. Waves change their propagation direction toward the shallower ridge crest. This refraction causes concentrating and effectively transporting tsunami energy along an oceanic ridge.

Similar effects can be expected on the continental shelf, which is another wave guide. Tsunamis that enter the sea on the continental shelf are refracted toward the shore, reflected from the shore to the sea, and refracted again toward the shore. Typical oscillation characteristics of tsunamis thus propagating along a sea ridge or a continental shelf as edge waves are the beat in the time history that the tsunami wave height gradually increases and the highest wave arrives later. For the farther tsunami source, the later the highest wave appears.

3.3. Tsunamis in the Shallow Sea

The tsunami energy propagates in relation to the water depth. The rate of energy transmission (the product of wave energy per unit sea-surface area and the energy propagation velocity) is constant, if there is no energy loss due, for example, to sea-bottom friction. The shallower the water depth is, the slower the velocity is, and therefore, the higher the wave height becomes. This is the shoaling effect.

Consider a bay with a broad entrance and narrowing towards inland. Tsunamis are usually much longer than the bay’s length. Tsunami energy concentrates toward the head of the narrow bay, hence tsunami height increases. This is the focusing effect. The similar effect occurs between two wave rays. (The ray is the trajectory of wave propagation.) The linear long wave theory gives a simple relation of the Green formula, among wave height H, water depth h, and width of the bay or normal distance between two wave rays b,

$$Hh^{1/4}b^{1/2} = \text{constant}.$$

Another important amplification mechanism is the resonance. If an external force is applied to the water in a bay and then removed, the water begins to oscillate with its natural period and the oscillation gradually diminishes due to energy dissipation. The fundamental period of the natural oscillation, T, for a rectangular shaped bay is given by

$$T = 4l/(gh)^{1/2},$$

where l and h are the length and water depth of the bay, and g the gravitational acceleration. If a tsunami having the same period as the natural period enters a bay, energy is stored and the resonance occurs in the bay. Three waves in succession are enough to establish the resonance for tsunamis.

Even if a tsunami starts with a simple initial profile of one crest and one trough, the tsunami shows very complicated movements on the shore, because of refraction, reflection, and diffraction caused by the topography. In 1983, the Nihonkai Chubu (Middle Japan Sea) earthquake tsunami hit the north Akita coast, which has a smooth shoreline 55 km long. The relatively straight coast is bounded at the both ends by a

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