Tsunamis: Seismological and Disaster Prevention Studies

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Japanese tsunami studies in the 1980's are reviewed. Because of the multi-disciplinary nature of the phenomenon, tsunamis are studied by researchers from various fields, including seismologists, geologists, oceanographers and coastal engineers. The tsunami from the 1983 Japan Sea earthquake caused 100 fatalities, and stimulated tsunami research, cooperation of workers from various fields for comprehensive work, and development of a quick tsunami warning system. Recent research developments include descriptive studies of the size and location of past tsunamis, tsunami observational systems, the generation and propagation of tsunamis, coastal effects, and a tsunami warning system.

1. Introduction

Most large shallow earthquakes around Japan occur beneath the ocean and generate sea waves, or tsunamis, as well as seismic waves. Tsunami, meaning harbor-wave, is one of a small number of Japanese words that have become international. Because of the multi-disciplinary nature of the phenomenon, tsunamis have been studied by geologists, seismologists, oceanographers and coastal engineers. The operation of tsunami warning systems requires knowledge about all of the above fields and rapid communication involving international cooperation.

The study of tsunamis has a long history in Japan. In the past, theoretical studies of tsunami generation and propagation were emphasized (e.g., Takahashi, 1942; Kajiura, 1963). In recent years, more practical and comprehensive studies from generation to coastal run-up have been developed. One of the catalysts for such comprehensive and cooperative work was the 1983 Japan Sea earthquake of May 26, 1983 (*M* 7.7), which generated a large tsunami that caused 100 fatalities. This was the largest tsunami disaster in Japan since 1960, when the tsunami from the Chilean earthquake propagated across the Pacific Ocean and killed about 140 people.

In this paper, we review recent developments in the broad area of tsunami research

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in Japan, and illustrate how cooperation has arisen among researchers from various fields. We also examine how the 1983 Japan Sea earthquake tsunami stimulated further developments in tsunami research. After this paper was submitted in early 1992, several damaging tsunamis have occurred in the world, including that from the 1993 Southwest Hokkaido earthquake ($M 7.8$). Various research on these tsunamis is still being conducted and the coverage on these studies are incomplete.

2. Descriptive Studies of Tsunamis

2.1 Location and size of tsunamis

The tsunami source area can be estimated in a manner similar to that used to locate earthquakes using S - P time, except that the time difference between tsunami arrival time and the origin time is used. Because of the very slow propagation velocity of tsunamis compared to that of seismic waves, a finite source area is usually estimated. The wavefront, or ray path, can be drawn from each station backwards to the source area. Such diagrams are called inverse refraction diagrams and are extensively used in Japan (Miyabe, 1934; Hatori, 1969; Abe, 1973). Figure 1 shows an example for the 1968 Tokachi-oki earthquake. Wavefronts corresponding to travel times for each station are shown as arcs in the center panel. The location of each station is shown in the left panel with numbers. The tsunami source area estimated in this way roughly corresponds to the aftershock area as shown in the left panel.

The size of tsunamis has been described by magnitude scales. The one longest in use is the Imamura-Iida scale, m (Hatori, 1979, 1986 a). It starts at -1 and is roughly equal to $\log_2 H$, where H is the maximum tsunami height. The definition of the height is ambiguous; it can be run-up height or amplitude on tide-gauge records. This scale is especially convenient for old tsunamis for which no instrumental records exist.

Fig. 1. The tsunami source area (center panel) and crustal deformation pattern (right panel) for the 1968 Tokachi-oki earthquake. In the center panel, the numbers beside each travel-time arc indicate the tide gauge station shown in the left panel. Solid and dashed arcs indicate positive and negative first arrivals, respectively. Modified from Abe (1973).

Recently another magnitude scale M_t , called tsunami magnitude, has been defined and assigned for many earthquakes by Abe (Abe, 1979, 1981, 1989 b). The definition for a trans-Pacific tsunami is:

$$M_t = \log H + C + 9.1$$

and for a regional ($100 \text{ km} < \Delta < 3,500 \text{ km}$) tsunami in the northwestern Pacific ocean is:

$$M_t = \log H + \log \Delta + 5.8,$$

where H is a maximum amplitude on tide gauges in m, C is a distance factor depending on a combination of the source and the observation points, and Δ is the actual distance in km. The above formulas were calibrated with the moment magnitude scale, M_w , of earthquakes. The 1968 Tokachi-oki earthquake tsunami (Fig. 1) was $m=2$ and $M_t=8.2$. The 1983 Japan Sea earthquake tsunami was $m=3$ and $M_t=8.1$, and the 1993 SW Hokkaido earthquake tsunami was $m=3$ and $M_t=8.1$.

2.2 Field surveys to document run-up heights

After damaging tsunamis occurred, field surveys are usually carried out to measure run-up heights and to collect eyewitness accounts. Preliminary survey reports for the recent tsunamis were made by Satake *et al.* (1993) for the 1992 Nicaragua tsunamis, by Yeh *et al.* (1993) for the 1992 Indonesia tsunamis, and by the Hokkaido Tsunami Survey Group (1993) for the 1993 SW Hokkaido tsunamis.

For historical tsunamis, the run-up heights can be estimated from historical literature such as official diaries. Such literature studies have provided much important information about long-term seismicity and recurrence intervals of earthquakes (Usami, 1988) and tsunamis (Hatori, 1977; Tsuji, 1983). Tsuji (1987) estimated the damage distribution for historical tsunamis on the basis of necrologies archived at temples. Historical documents usually give tsunami height from local reference points; complementary field surveys are sometimes necessary to measure the run-up heights from sea level (Hatori *et al.*, 1983; Shuto *et al.*, 1987).

Geological methods also have become popular in studying old earthquakes. Minoura and Nakaya (1991) drilled boreholes to detect tsunami deposits. Sedimentological and geochemical analyses have made it possible to estimate dates and run-up heights of prehistoric tsunamis and also the recurrence intervals and sizes of large earthquakes. For example, the recurrence interval of tsunamis that arrive at Sendai, northeastern Japan, is estimated at about 800 years. Investigation of archaeological relics, whether buried by running water or not, also helps to document tsunami run-up heights for past tsunamis (Abe *et al.*, 1990). Kato and Kimura (1983) estimated the age of a boulder in Ishigaki Island, presumably transported by a tsunami in 1771, and claimed it to be the highest tsunami record in Japan.

2.3 Tsunami catalog

The location, size, run-up heights and damages of past tsunamis in and around Japan were compiled as a catalog by Watanabe (1985). This catalog contains 177 tsunamis that occurred between 684 and 1984 near Japan, and 40 trans-Pacific tsunamis that were observed in Japan with descriptions of tsunami travel times, run-up heights,

damage, and earthquake parameters including intensity distributions.

2.4 Tsunami probability

Using the tsunami run-up heights from past tsunamis in the last 400 years, Aida (1988) estimated cumulative tsunami energy for each 20 km interval of the Japanese coast. Rikitake and Aida (1988) further estimated tsunami hazard probabilities for the period between 2000 and 2010. They showed that the probability of a tsunami exceeding 5 m is as high as 41% in some parts of central Japan.

3. Tsunami Observation System

3.1 Tide gauge systems

Tsunamis are recorded on tide gauges. About 500 tide gauge stations are currently in operation in Japan for the purpose of monitoring ocean tides (GSI, 1989). Tsunami waveforms as old as those from the 1896 Sanriku earthquake have been recorded on these tide gauges.

Figure 2 (upper left) shows the typical structure of a Japanese tide gauge. The gauge is set in a well which is connected to the outer sea by an intake pipe. Because of the severe winter weather in northern Japan, particularly on the Japan Sea coast, some stations have mechanisms to filter out shorter-period components such as surf or ocean beat. Because the period of tsunami waves is also shorter than that of ocean tides, tsunamis are also affected by the tide gauge structure. For example, the visually observed tsunami height from the 1983 Japan Sea earthquake was reported to be more than twice that measured on tide-gauge records at some nearby stations.

After the 1983 tsunami, systematic studies including theoretical computations and

Fig. 2. Schematic pictures of tsunami recording systems. The upper left shows a traditional tide gauge. The newly developed tsunami gauges include the ocean bottom tsunami gauge (upper right) and acoustic gauges (lower two).

in situ measurements of tide gauge responses were carried out (Murakami, 1983; Okada, 1985; Satake *et al.*, 1988). The results showed that at some stations the response was very slow because of the intake pipe. If the tsunami period is shorter than the response time of the tide-gauge system, the tide-gauge system cannot follow the actual sea water movement of the tsunami, causing the recorded waves to be distorted. If the response of the system is known, tide-gauge records can be corrected to estimate the real tsunami waveforms.

3.2 Tsunami gauges

In recent years, efforts have been made to develop and deploy instruments specially designed to record tsunamis (e.g., Aida *et al.*, 1981; Tada *et al.*, 1991). These are called tsunami gauges or wave gauges. New types of tsunami gauge include pressure gauge and acoustic gauge (Fig. 2). The pressure gauges are set on the ocean bottom. The acoustic gauges are set either above the water level, measuring the travel time of acoustic wave in the air, or on the sea bottom and measures the travel time in the water. These are more suitable to measure the shorter period waves than conventional tide gauges. Wave gauges, however, do not continuously measure water heights. For tsunami observation, it is necessary and important to make continuous measurements. Ultrasonic wave gauges were installed in Kesennuma bay by the city office for the purpose of tsunami warning systems (Sato *et al.*, 1992). These gauges recorded the 1992 Nicaragua tsunamis across the Pacific ocean, tide gauge in Japan could not detect such a small tsunami signal. Most of these gauges are set in shallow (up to 100 m) water.

The Japan Meteorological Agency (JMA) has been operating two ocean bottom cables that telemeter seismic and tsunami data in real time from permanent ocean bottom seismographs and pressure gauge sensors (MRI, 1980). Four pressure gauges are installed over 100 km offshore Honshu, where the ocean depth is more than 2,000 m (Fig. 3). This system provides valuable data for tsunami research (Okada, 1991; Satake

Fig. 3. The location of permanent ocean bottom seismograph system. Seismographs are laid at the open and closed circles and tsunami gauges are installed at the closes ones. From Okada (1991).

et al., 1992; Imamura and Shuto, 1993), and is also used as a part of a tsunami warning system, as we discuss below.

4. Tsunami Generation

4.1 Earthquake fault model

Tsunamis are generated from disturbance of the ocean surface or bottom. Among various geological phenomena that cause tsunamis, earthquake tsunamis are the most frequent and well studied. Abe (1973) showed that the pattern of water surface disturbance, either uplift or subsidence, in the tsunami source area for the 1968 Tokachi-oki earthquake was consistent with the crustal deformation pattern estimated from seismological analyses (Fig. 1).

Because of the long wavelength of crustal deformation compared to the water depth, the water surface disturbance can be assumed to be the same as the crustal deformation on the ocean bottom (Kajiura, 1963). Kajiura (1970) examined the efficiency of tsunami generation in terms of energy exchange and showed that source motion can be considered abrupt if the duration is less than several minutes. Kajiura (1981) also examined the relationship between fault parameters and tsunami energy.

Tsunami generation and propagation from fault motions have been calculated analytically (Yamashita and Sato, 1974), using normal mode theory (Ward, 1980, 1981, 1982; Comer, 1984; Okal, 1988) or numerically (Aida, 1974). Aida (e.g., Aida, 1978) made numerical computations for many earthquakes. For each event, he computed tsunami waveforms for several fault models and compared them with the observed tsunami waveforms. Aida (1977b) further used an empirical relationship between the computed tsunami wave height at 200 m depth and the average inundation heights (the latter is about two to three times larger than the former) to estimate the fault parameters of old earthquakes for which no tsunami waveforms are available.

4.2 Inversion for heterogeneous fault motion

Since tsunami waveforms vary with the fault parameters, earthquake fault parameters can be estimated from the observed tsunami waveforms. Aida (1972) tried to estimate the initial water disturbance from the observed tsunamis waveforms. In the 1980's, seismologists found that fault motion of large earthquakes is not uniform but has heterogeneities described as asperities or barriers. Satake (1989) inverted tsunami waveforms in the same way as seismic waves to estimate the heterogeneous slip distribution on the fault for the 1968 Tokachi-oki and 1983 Japan Sea earthquakes (Fig. 4). Figure 5 shows a comparison of observed tsunami waveforms with the computed ones for a heterogeneous slip distribution for the 1983 Japan Sea earthquake. Satake (1993) further examined the slip distribution for the 1944 Tonankai and 1946 Nankaido earthquakes. For such old earthquakes, tsunami data provide better information about the source motion than seismic waves.

4.3 Tsunami earthquakes

Some earthquakes, e.g., the 1896 Sanriku or 1946 Aleutian earthquakes, caused abnormally larger tsunamis than expected from seismic waves. Kanamori (1972) named

Fig. 4. Schematic illustration showing the inversion of tsunami waveforms to estimate the slip distribution on an earthquake fault.

Fig. 5. Observed tide-gauge records from the 1983 Japan Sea earthquake tsunami and the computed one from the source model obtained by the inversion. From Satake (1989).

such earthquakes "tsunami earthquakes." Abe (1989 a, b) quantitatively defined a tsunami earthquake an event whose tsunami magnitude (M_t) exceeds the surface wave magnitude (M_s) by more than 0.5. Kanamori (1972) proposed that an unusually slow source process would excite relatively large tsunamis compared to seismic waves. Other models have been also proposed; Fukao (1979) proposed that a high-angle fault in the shallow sediment layer caused larger tsunami than seismic waves for the 1975 Kuril earthquake. Satake and Kanamori (1991) showed that the unusually large tsunami from the 1984 Torishima earthquake was due to abnormal tsunami propagation along the ridge system as well as an unusual source mechanism of the earthquake, probably the intrusion of magma into sediments. The 1992 Nicaragua earthquake was also identified as a tsunami earthquake (Ide *et al.*, 1993; Imamura *et al.*, 1993; Kanamori and Kikuchi, 1993; Satake *et al.*, 1993).

4.4 Tsunamis from other causes

Volcanic eruptions also generate tsunamis. Although they are less frequent than earthquake tsunamis, volcanic tsunamis have caused tremendous damage. A few analyses or model calculations have been done for tsunamis from volcanic origin. The eruption of Unzen in 1792 (which also erupted in 1990), caused a collapse of Mayuyama, which caused a huge tsunami and killed 15,000 people (Aida, 1975). The 1883 eruption of Krakatau, probably the largest eruption in the world in recent centuries, also caused a tsunami and claimed about 36,000 lives. The generation mechanism of this tsunami has been debated for the last decade (Yokoyama, 1981, 1987; Latter, 1981; Self and Rampino, 1982; Camus and Vincent, 1983; Francis, 1985). Recently, more numerical computations (Nishimura and Satake, 1993; Nomanbhoy and Satake, 1995; Kawamata *et al.*, 1993) and hydraulic experiment (Togashi and Hirayama, 1993) have been made to model these volcanic tsunamis.

The 1741 tsunami generated near Oshima-Ohshima, between the 1983 Japan Sea and 1993 SW Hokkaido earthquakes was enigmatic; the tsunami damage was too extensive to be of volcanic origin (Aida, 1984) but there is no record of an earthquake in historical literature, in contrast to many records describing the volcanic eruption.

5. Tsunami Propagation

5.1 Governing equation

The wavelength of tsunamis is usually much longer than the water depth. Such a wave can be treated as a long-wave (also known as shallow water wave) in hydrodynamics. The most general equation for a long-wave is the Boussinesq equation. For one-dimensional case, it is given as

$$M_t + g h \eta_x + g \eta \eta_x + (M^2 / (h + \eta))_x + g n^2 M^2 / (h + \eta)^{7/3} = h^2 M_{xxt} / 3,$$

where M is the water discharge (horizontal velocity integrated from bottom to the surface), η is the water height above still water level, h is the still water depth, g is the gravitational acceleration, n is Manning's roughness coefficient, and subscript shows temporal (t) or spatial (x) derivatives.

The term on the right hand side shows dispersion. Dispersion becomes important

only when a tsunami propagates a long distance on a mild slope or in a river. The last term in the left hand side accounts for bottom friction, which is often expressed by Manning's roughness coefficient. The third and fourth terms are for non-linear motion of water. The non-linear and friction terms become important only for shallow water depth. For tsunami propagation in deep water, only the linear terms (the first two terms in the left hand side) need to be considered. In such a case, the equation becomes the linear long-wave equation whose velocity is given as $\sqrt{gh}$. Since the velocity depends on ocean depth, which is usually well known, tsunami propagation can be accurately computed. Satake (1988) applied ray-tracing to tsunamis and showed that the effect of bathymetry is very important for tsunami propagation.

5.2 Numerical computation of tsunamis

For tsunami propagation, the finite-difference method with a staggered leap-frog scheme has been most commonly used (Aida, 1969; Hwang *et al.*, 1972). The crustal deformation pattern due to faulting can be used as the initial condition. At the boundary to the open sea, a radiation condition, in which tsunami waves go out of the computation region without changing their shape, is usually assumed. Different land boundary conditions are used depending on the purpose of the tsunami computation. When tsunami behavior in the deep ocean is examined, total reflections on a fixed coast can be assumed. When the evaluation of tsunami run-up is the main purpose of the computation, the ocean-land boundary cannot be fixed. Several methods have been proposed for a moving boundary (Aida, 1977a; Iwasaki and Mano, 1979) (Fig. 6).

The grid size for numerical computation is important for both stability and accuracy of computations. The grid size and computational time step must satisfy the stability condition,

$$K = C\Delta t/\Delta x \leq 1.0,$$

where K is the Courant number, C is the wave velocity and Δx and Δt are the spatial grid size and time step.

Fig. 6. Moving boundary conditions on the ocean-land boundary. That above is adopted by Aida (1977 a) with a weir-type formula and that below is by Iwasaki and Mano (1979) considering a relationship between velocity and water depth at the wave front.

Sayama *et al.* (1988) examined the computational error on tsunami refraction due to the digitization of bathymetry. Imamura and Goto (1988) examined the truncation error associated with the finite-difference method and showed that a coarse grid size can cause numerical dispersion. For accurate computation of tsunamis, 20 to 30 grid points per wavelength are preferable (Shuto *et al.*, 1986). For the tsunami run-up problem, Goto and Shuto (1983) examined the effects of grid size, wave period, slope and the moving boundary conditions on the computation.

For practical tsunami computation, the linear long-wave equation is used in the deeper ocean, with a larger grid size. In the near shore region, the non-linear equation must be used with a smaller grid size. Hence computations are usually made with variable grid size and governing equations.

5.3 Numerical computation of trans-pacific tsunamis

For tsunamis propagating a very long distance such as across the Pacific ocean, the effects of Earth's sphericity and rotation (Coriolis force) must be also included. Further, physical dispersion cannot be neglected (Kajiura, 1970). Imamura *et al.* (1990) showed that an appropriate choice of grid size can make the effects of numerical and physical dispersion equal. In other words, the finite difference computation for a linear long-wave can simulate the linear Boussinesq equation which includes a dispersion term. The choice of grid size is determined by the Imamura number, Im , defined as

$$Im = \Delta x \sqrt{1 - K^2} / 2h,$$

where h is the average water depth. The value of Im must be about 1. Figure 7 shows the effect of Im for the propagation of the 1964 Alaskan earthquake tsunami.

5.4 Edge wave and free oscillation of Japan Sea

When a tsunami source is located on the continental shelf or a ridge system, the energy of a propagating tsunami is trapped on the shelf or ridge (Abe and Ishii, 1989; Satake *et al.*, 1992). This is known as an edge wave in oceanography.

The tsunami from the 1983 Japan Sea earthquake was observed for as long as two days after the earthquake on the Japan Sea coast. This is due to the multiple reflection of tsunamis in the almost-closed Japan Sea (Hatori, 1986 b; Abe and Ishii, 1987; Abe, 1991). Satake and Shimazaki (1988 a, b) treated such tsunamis as a normal mode of

Fig. 7. Comparison of tsunami profiles computed for the 1964 Alaskan tsunami using the linear long-wave equation (solid lines) and the linear Boussinesq equation (dotted line). The profile from the linear long-wave equation with $Im=1.04$ agrees well with that from the linear Boussinesq equation.

Fig. 8. Eigenvalues, or eigenperiods, and the corresponding eigenvectors, or water height patterns, for the three oscillation modes of the Japan Sea computed by Satake and Shimazaki (1988b).

the ocean basin. Figure 8 shows the periods and oscillation patterns for the three lowest fundamental modes of the Japan Sea. Satake and Shimazaki (1988 a, b) showed that a tsunami source in the deep ocean, such as the 1983 Japan Sea earthquake tsunami, excited free oscillations of the whole Japan Sea; whereas those occurring on the continental shelf, such as the 1964 Niigata earthquake tsunami, excited only regional oscillations.

6. Coastal Effects

6.1 *Flooding and damages to houses*

Takeda and Tsuji (1987) used LANDSAT data to estimate tsunami run-up heights and flooded areas for the 1983 Japan Sea earthquake tsunami by means of vegetation change. In order to make accurate numerical computations of tsunami flooding, the wave front must be determined and the friction law on the dry bed must be known. Fujima and Shuto (1990) conducted hydraulic experiments to estimate the friction law. Their result shows that Manning's law with a proper coefficient is sufficient for determining the maximum run-up. However, this law may not be applicable to the surge front of a tsunami.

When a tsunami runs over the land, the houses and other structures along the shoreline are damaged or washed away by the force of the tsunami. To evaluate such a wave impact force, the wave profile, pressure and particle velocity at the front must be accurately known. Matsutomi (1990) attempted to develop a procedure for predicting the maximum impulsive force due to floating timbers, and experimentally presented the apparent inertia coefficient and the impact duration time. Goto (1983) developed a numerical method to estimate the spread of timbers. Hatori (1984) found that damage to wooden houses is generally proportional to a drag force, defined as the product of the tsunami height and the square of water velocity.

6.2 *Current due to tsunami*

The current due to tsunamis is not well known or recorded except for eyewitness accounts. A series of aerial photographs was taken at the time of the 1960 Chilean tsunami on Sanriku coast. From an analysis of these photographs, the current velocity on the water surface was estimated (Takahashi *et al.*, 1993). The numerical computation of current requires finer grids than the standard tsunami computation for water height, since the direction and magnitude of the current are sensitive to the local topography. By comparison of experiments and numerical simulations, Tomizawa *et al.* (1990) showed that horizontal eddy viscosity must be included for accurate computation of tsunami current velocity.

Pearl culture was heavily damaged due to the current of the 1960 Chilean tsunami in Japan. Data on damage were collected in Matoya bay, Kii Peninsula. Sato (1960) investigated the degree of damage done to pearl culture rafts. A comparison of numerical computation to damage data showed that pearl culture rafts are safe as long as the maximum velocity is less than 1 m/s (Nagano *et al.*, 1991).

6.3 *Sand transport*

As we have seen before, geologists study tsunami deposits to get important information about past tsunamis and earthquakes (Abe *et al.*, 1990; Dawson *et al.*, 1991; Minoura and Nakaya, 1991). Possible disasters from future tsunamis, on the other hand, include the erosion near the foundation of a structure or deposition of sediments that may block the inflow mouth to an electric power plant. At this stage, it is difficult to compute sand transport due to tsunamis, because the evaluation of velocity, shear stress and transport formula have not been established. Future developments are expected because of the scientific and engineering applications.

Deformation of the sea bottom associated with the 1960 Chilean tsunami was observed in some harbors on the Sanriku coast. The change in the bottom profile due to the tsunami shows a mode similar to the annual change. The erosion in a river-harbor and a long and narrow harbor are remarkable whereas the erosion in a wide harbor is indistinct (Kawamura and Mogi, 1961). These data give valuable information which can be used to establish the sand transport simulation.

6.4 *Tsunami defense work*

Structures such as seawalls, tsunami breakwaters and tsunami gates provide a direct way to prevent tsunami hazards. The height of a sea wall is usually designed based on the highest water level from past tsunamis. Aida *et al.* (1988) examined by numerical experiments the behavior of tsunamis exceeding the design height. The result shows that when a tsunami has the same height as the sea wall, the water level near the wall becomes higher than the wall, because of the reflection. In this case, the water level at some places in the town becomes so high as to give rise to human and material damage in spite of the construction of the sea wall.

High seawalls are sometimes not welcomed because of inconveniences placed on the daily life of local residents. Alternatively, a tsunami-resistant building may be considered as a prevention structure. The tsunami force on this building and the reduction

of tsunami effects on the houses behind it should be evaluated for the effective use of a tsunami-resisting building zone. A tsunami-control forest can also be expected to have a similar effect. Shuto (1987a) estimated the effectiveness and limit of a tsunami-control forest in terms of the cumulative tree diameters and the density of undergrowth.

Recent growth in industrial activity has increased tsunami disaster potential. For instance, if an oil spill is accompanied by fire, the effect is devastating. Examples are found in Seward, Valdez and Whittier in Alaska (from the 1964 Alaskan earthquake), Niigata (from the 1964 Niigata earthquake) and Okushiri Island (from the 1993 SW Hokkaido earthquake). Numerical simulation can roughly estimate oil spread and burn area, and is applied to Miyako bay, the northern Sanriku coast (Goto, 1985; Shuto, 1987b).

7. Tsunami Warning Systems

Tsunami warning systems utilize the velocity difference between seismic and tsunami waves. The arrival time difference of these waves is obviously longer for larger distances. The Pacific-wide tsunami warning system is in operation through international cooperation. JMA is exchanging seismic and tide gauge data with foreign countries including the United States (Alaska and Hawaii), Micronesia (Palau Is.), Russia (Khabarovsk), South Korea and Hong Kong through satellite or telephone communications.

A tsunami warning system for local tsunamis is more challenging (Katsumata, 1987). The first arrival of the tsunami from the 1983 Japan Sea earthquake reached the coast 7 min after the earthquake and killed school children. The tsunami warning was issued 14 min after the origin time. Although this was still much shorter than the 20 min, required by regulation, further developments have been made to issue tsunami warnings in shorter times. Currently, JMA has been equipped with the Earthquake Phenomena Observation System (EPOS; Fig. 9), which consists of super-mini computers, telecommunication links, and telemetry facilities connecting more than 60 seismic stations including two permanent ocean bottom seismographs (Yokota and Yamamoto, 1989). The EPOS automatically detects an event, picks P and S wave arrivals, estimates the hypocenter and magnitude, and issues a tsunami warning based on the criteria of location, magnitude and depth. Based on the judgment of the duty seismologist, the warning messages are distributed to the government, TV and radio stations. Because of this new system, the tsunami warning was issued in 5 min after the 1993 Southwest Hokkaido earthquake.

Zama and Yoshihara (1987) developed a simpler system, in which the location and magnitude are estimated from a three-component seismograph, and tsunami arrival times are then calculated from submarine earthquakes.

Seismological studies for more accurate and detailed tsunami warnings are still in progress. Assuming that tsunami and seismic magnitudes are equal and using the definition of tsunami magnitude, Abe (1989a) showed that tsunami heights can be predicted from an earthquake magnitude. Fukushima *et al.* (1989) showed the possibility of real-time moment tensor inversion using local data. Izutani and Hirasawa (1987) developed a simple method to estimate fault parameters from strong motion data. If

Fig. 9. Schematic illustration of current tsunami warning system of the Japan Meteorological Agency.

all of the fault parameters of earthquakes could be estimated in real time, the numerical computation of tsunamis could be made before the actual tsunami arrives at coastal regions, and more detailed warnings could be issued (Shuto *et al.*, 1990). Iwasaki (1991) showed that measurements of the T-phase, not only on seismographs, but also reported from ships as sea-shocks, can be used to estimate the tsunami size.

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