Chapter 2 Tsunami Generation

2.1 Tsunami–Earthquake Energy Relations, Source Areas

Tsunami Magnitude in Relation to Other Parameters

Iida (1970) followed the earlier work by Imamura (1949) to define tsunami magnitude, $m$, in reference to Japan, as:

$$ m = \log_2 \eta_{\text{max}} $$

where $\eta_{\text{max}}$ is the maximum height in meters measured at a coast 10–300 km from the tsunami origin. Iida (1956) gave the tsunami grade scale based on $\eta$ (Table 2.1). Iida (1970) cataloged the earthquakes of magnitude, $M$, greater than 5.8 accompanied by tsunamis, in and near Japan from 1900 to 1968. The geographic distribution of the epicenters of the tsunamigenic earthquakes (classified according to tsunami magnitude, $m$) shows that most epicenters lie on the Pacific Ocean side of Japan (not in the Sea of Japan).

Table 2.1. Tsunami grade scale of Iida and Imamura. Depending on the requirement, this scale has been extended occasionally from $-3$ to $5$. (Iida 1956)

<table>
<thead>
<tr>
<th>Grade scale</th>
<th>Significance</th>
</tr>
</thead>
<tbody>
<tr>
<td>-1</td>
<td>Minor tsunami with max. wave height ($h$) less than $\frac{1}{2}$ m</td>
</tr>
<tr>
<td>0</td>
<td>$h$ is of the order of 1 m; no damage</td>
</tr>
<tr>
<td>1</td>
<td>$h$ is of the order of 2 m; house damage along coast; ships washed ashore</td>
</tr>
<tr>
<td>2</td>
<td>$h$ is of the order of 4–6 m; some destruction of houses; considerable loss of life</td>
</tr>
<tr>
<td>3</td>
<td>$h$ is of the order of 10–20 m; damaged area along coast about 400 km</td>
</tr>
<tr>
<td>4</td>
<td>$h$ is greater than 50 m; damaged area along coast more than 500 km</td>
</tr>
</tbody>
</table>

Iida (1963b) studied about 100 tsunamigenic earthquakes that occurred from 1700 to 1960 and produced a diagram (not shown here) showing the relation between tsunami magnitude, $m$, and water depth, $D$, at the epicenter. From this diagram, the following upperbound on tsunami magnitude can be expressed:

$$ m = 1.66 \log D - 1.62 $$

where $D$ is expressed in meters.
Iida (1963b) also investigated the relation between $m$ and $\log S$ where $S$ is the ratio of the water depth at the epicenter to the distance between the epicenter and the point on the coast where $m$ is determined. This is expressed as:

$$\log S = 0.12 m + 2.12$$  \hspace{1cm} (2.3)

or inversely

$$m = 8.33 \log S - 17.0$$ \hspace{1cm} (2.4)

These relations show that tsunami magnitude is proportional to the slope, $S$, of the sea bottom at the epicenter and that the water depth, $D$, at the epicenter has more influence on $m$ than the distance of the epicenter to the coast.

The tsunami grade magnitude scale discussed above was originally devised for tsunamis with sources within 600 km of the Sanriku coast. This scale is not linear, even after the logarithmic transformation. Adams (1974) devised a new tsunami magnitude scale, based on the logarithm to the base of 2, of the run-up expected at 1000 km from the epicenter. He called this the “logarithmically linear scale” of tsunami magnitude. This scale is better than the tsunami grade magnitude scale, because data are reduced to a fixed epicentral distance, even if not observed there. The linearly logarithmic scale also includes a correction for geometrical spreading.

**The Concept of “Tsunami Intensity”**

Soloviev (1970, p. 152) pointed out the inappropriateness of the usage “tsunami magnitude.” He stated:

“If seismological terminology is applied to description of tsunamis, the grades of the Imamura-Iida scale must be designated as the intensity of the tsunami and not its magnitude. This is because the latter value must characterize dynamically the processes in the source of the phenomenon and the first one must characterize it at some observational point … the nearest point to the source included.”

Another important point made by Soloviev was the distinction between the mean height, $\bar{\eta}$, and the maximum height, $\eta_{\text{max}}$, of tsunami inundation. This distinction is necessary because, although it is the mean height that is relevant for the estimation of tsunami energy, in older descriptions of tsunami emphasis was given to the maximum height. The differences between mean height and maximum height could be mostly attributed to topography. Figure 2.1 shows the relation between $\bar{\eta}$ and $\eta_{\text{max}}$ for different tsunamis.

Soloviev defined tsunami intensity, $i$, as:

$$i = \log_2 (\sqrt{2} \bar{\eta})$$ \hspace{1cm} (2.5)

Compared to (2.1), (2.5) differs in three respects. First, instead of magnitude, $m$, intensity, $i$, appears. Second, the maximum height, $\eta_{\text{max}}$, is replaced by the average height, $\bar{\eta}$. Third, the factor, $\sqrt{2}$, was introduced to account for the average difference in the maximum and mean heights of tsunamis of different intensities.

**Tsunamigenic and Nontsunamigenic Earthquakes**

Iida (1970) studied the relation between earthquake magnitude, $M$, and focal
depth, $H$, of submarine earthquakes that occurred from 1926 to 1968 in and near Japan (Fig. 2.2). The solid line represents the boundary between tsunamigenic and nontsunamigenic earthquakes, i.e. tsunamigenic earthquakes occur to the right side of this line. This (lower) limiting magnitude for tsunamigenic earthquakes is expressed as:

$$M = 6.3 + 0.005 H$$

where the focal depth, $H$, is expressed in km. However, there are three tsunamigenic earthquakes to the left of the solid line (not shown here). If these are included then the smallest magnitude of a tsunamigenic earthquake is given by:

$$M = 5.6 + 0.01 H$$

Iida (1970) also gave the following relation for the lower limit for the magnitude of earthquakes which have generated disastrous tsunamis (i.e. tsunamis of magnitude $> 2$):

$$M = 7.7 + 0.05 H$$
This limit is shown by the dashed line in Fig. 2.2.

In the data represented in Fig. 2.2, there were 79 earthquakes with magnitudes greater than given by (2.6) and yet they were not followed by tsunamis. Of these, 32 were aftershocks and 15 were relatively deep-focus shocks. Both types are unlikely to produce tsunamis. Iida used the concept of faulting to explain the fact that the remaining 32 earthquakes were nontsunamigenic.

For these 32 earthquakes, Iida considered focal-plane solutions. Radiation patterns from both P and S waves were used, and faulting along the nodal plane was assumed, to calculate the dip and strike components of the unit vector of motion because of faulting. The faulting is classified as dip-slip type if the dip component exceeds the strike component. The reverse situation is called strike-slip. Iida found that for more than 60% of the tsunamigenic earthquakes studied, faulting was the dip-slip type.

Iida (1970) investigated the relation between earthquake magnitude and tsunami magnitude classified according to the type of faulting. He concluded that tsunami magnitude is proportional to earthquake magnitude and that most large tsunamis result from earthquakes associated with dip-slip faulting. For other detailed studies on Japanese tsunamis see Hatori (1970) and Watanabe (1970). For the source mechanism of tsunamigenic and nontsunamigenic earthquakes in the northwest part of the Pacific Ocean see Balakina (1970).

**Seismic Energy Versus Tsunami Energy**

Iida (1963b) gave the following relation between earthquake magnitude, M, and tsunami magnitude, m:

\[ m = (2.61 \pm 0.22) M - (18.44 \pm 0.52) \]  

(2.9)

Gutenberg and Richter (1956) gave the following relation between M and seismic energy, \( E_s \) (expressed in ergs):

\[ \log E_s = 11.8 + 1.5 M \]  

(2.10)
From these two equations, one can write:

\[
\log E_s = 22.4 + 0.6 \, m
\]

or

\[
E_s = E_0 \times 10^{0.6m}
\]

with

\[
E_0 = 2.5 \times 10^{22} \text{ ergs}
\]

Takahasi (1951) gave the following relation between tsunami magnitude, \( m \), and tsunami energy, \( E_t \):

\[
E_t = E'_0 \times 10^{0.6m}
\]

with

\[
E'_0 = 2.5 \times 10^{21} \text{ ergs}
\]

In Takahasi's method, the tsunami energy is estimated from the following relation (based on symmetry, for example Equation (1.143)):

\[
E_t = \pi \rho g r \sqrt{gD} \left( \overline{\eta} \right)^2 \tau
\]

where \( g \) is gravity, \( \rho \) is water density, \( r \) is the distance between the observation location and tsunami source, \( D \) is the water depth at the observation point, \( \overline{\eta} \) is the average observed height of tsunami, and \( \tau \) is the duration of tsunami.

From (2.9) and (2.13)

\[
\log E_t = 21.4 + 0.6 \, m = 10.3 + 1.5 \, M
\]

Hence, comparing (2.11) with (2.15)

\[
\frac{E_s}{E_t} = \frac{E_0}{E'_0} = 10
\]

Thus, about a tenth of the seismic wave energy appears as tsunami energy. Table 2.2 supports this statement quite well (actually \( E_s / E_t \) has been observed to vary over a substantial range with an average value of about 10).

<table>
<thead>
<tr>
<th>Earthquake</th>
<th>Magnitude (M)</th>
<th>Seismic wave energy ( (E_s) ) ( 10^{23} \text{ ergs} )</th>
<th>Tsunami energy ( (E_t) ) ( 10^{23} \text{ ergs} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Chile, May 22, 1960</td>
<td>8.5</td>
<td>35.5</td>
<td>3.0</td>
</tr>
<tr>
<td>Sanriku, Mar. 2, 1933</td>
<td>8.3</td>
<td>17.8</td>
<td>1.7</td>
</tr>
<tr>
<td>Nankaido, Dec. 20, 1946</td>
<td>8.1</td>
<td>8.9</td>
<td>0.8</td>
</tr>
<tr>
<td>Tokachi, Mar. 4, 1952</td>
<td>8.1</td>
<td>8.9</td>
<td>0.8</td>
</tr>
<tr>
<td>Tonankai, Dec. 7, 1944</td>
<td>8.0</td>
<td>6.3</td>
<td>0.79</td>
</tr>
<tr>
<td>Aomori, Feb. 10, 1945</td>
<td>7.3</td>
<td>0.56</td>
<td>0.004</td>
</tr>
</tbody>
</table>
Iida (1963a) gave the following empirical relation between earthquake magnitude, M, and the area, A, of aftershock activity (expressed in \( \text{km}^2 \)).

\[
\log A = 0.9 M - 3.0 \tag{2.17}
\]

He also deduced that the area over which crustal deformation occurs (which might lead to tsunami generation) is equal to the area of aftershock activity. However, care should be taken to treat this as a specific rather than as a universal result.

For Pacific Ocean tsunamis, Soloviev (1970) gave the following relation between \( \tau \) (duration of tsunami), its intensity, \( i \), and earthquake magnitude, M:

\[
\log \tau = -0.6 + 0.12 i + 0.24 M \tag{2.18}
\]

Soloviev also deduced the following relation between tsunami frequency, \( n \), and tsunami intensity, \( i \):

\[
\log n = k_1 - k_2 i \tag{2.19}
\]

where \( k_2 = 0.31 \). He prepared a table (Table 2.3) and listed the coefficients \( a \) and \( b \) (for different geographic regions on the globe) of the following formula which expresses the relation between earthquake magnitude, M, and earthquake frequency, N.

\[
\log N = a - bM \tag{2.20}
\]

<table>
<thead>
<tr>
<th>Zone</th>
<th>( a )</th>
<th>( b )</th>
<th>( T )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Aleutian Is. &amp; southwest Alaska</td>
<td>8.43</td>
<td>0.92</td>
<td>-0.15</td>
</tr>
<tr>
<td>Kamchatka &amp; Kuril Is.</td>
<td>8.72</td>
<td>0.96</td>
<td>-0.2</td>
</tr>
<tr>
<td>Hokkaido</td>
<td>8.90</td>
<td>0.82</td>
<td>-0.1</td>
</tr>
<tr>
<td>East Honshu</td>
<td>9.21</td>
<td>1.08</td>
<td>-0.1</td>
</tr>
<tr>
<td>South Japan</td>
<td>6.80</td>
<td>0.80</td>
<td>-0.1</td>
</tr>
<tr>
<td>Japan Sea</td>
<td>7.07</td>
<td>0.86</td>
<td>+0.1</td>
</tr>
<tr>
<td>Ryukyu Is.</td>
<td>7.68</td>
<td>0.92</td>
<td>-0.4</td>
</tr>
<tr>
<td>Taiwan</td>
<td>8.24</td>
<td>1.20</td>
<td>-0.6</td>
</tr>
<tr>
<td>Luzon</td>
<td>7.84</td>
<td>0.96</td>
<td>0.0</td>
</tr>
<tr>
<td>Southwest Philippines</td>
<td>5.27</td>
<td>0.60</td>
<td>+0.1</td>
</tr>
<tr>
<td>Philippine Deep</td>
<td>7.05</td>
<td>0.80</td>
<td>-0.1</td>
</tr>
<tr>
<td>Talaud &amp; Sangihe Is.</td>
<td>5.52</td>
<td>0.60</td>
<td>-0.1</td>
</tr>
<tr>
<td>Banda Sea</td>
<td>6.06</td>
<td>0.70</td>
<td>+0.2</td>
</tr>
<tr>
<td>Sulawesi &amp; Kalimantan</td>
<td>7.86</td>
<td>1.00</td>
<td>+0.4</td>
</tr>
<tr>
<td>Djawa &amp; Lesser Sunda Is.</td>
<td>8.63</td>
<td>1.10</td>
<td>-0.1</td>
</tr>
<tr>
<td>Sumatera</td>
<td>6.23</td>
<td>0.68</td>
<td>0.0</td>
</tr>
<tr>
<td>Irian</td>
<td>5.75</td>
<td>0.68</td>
<td>+0.1</td>
</tr>
<tr>
<td>New Britain &amp; North Solomon Is.</td>
<td>7.35</td>
<td>0.80</td>
<td>-0.3</td>
</tr>
<tr>
<td>South Solomon Is.</td>
<td>6.67</td>
<td>0.75</td>
<td>-0.1</td>
</tr>
<tr>
<td>Santa Cruz &amp; New Hebrides Is.</td>
<td>8.04</td>
<td>0.90</td>
<td>-0.2</td>
</tr>
<tr>
<td>Fiji Is.</td>
<td>7.00</td>
<td>0.84</td>
<td>0.0</td>
</tr>
<tr>
<td>Samoa, Tonga, Kermadec Is.</td>
<td>10.70</td>
<td>1.30</td>
<td>-0.45</td>
</tr>
<tr>
<td>New Zealand, North Is.</td>
<td>7.58</td>
<td>0.94</td>
<td>0.0</td>
</tr>
<tr>
<td>Chile &amp; Peru</td>
<td>7.67</td>
<td>0.85</td>
<td>+0.15</td>
</tr>
<tr>
<td>Central America &amp; Mexico</td>
<td>7.85</td>
<td>0.84</td>
<td>-0.35</td>
</tr>
<tr>
<td>California</td>
<td>7.20</td>
<td>0.90</td>
<td>0.0</td>
</tr>
<tr>
<td>Canada</td>
<td>7.94</td>
<td>1.00</td>
<td>0.0</td>
</tr>
<tr>
<td>Hawaiian Is.</td>
<td>7.00</td>
<td>0.92</td>
<td>+0.5</td>
</tr>
</tbody>
</table>
Soloviev also defined the following parameter:

\[
T = \frac{\log n(0)}{N(7\frac{1}{2})}
\]  

(2.21)

where \( n(0) \) is the frequency of tsunamis with zero intensity, and \( N(7\frac{1}{2}) \) expresses the frequency of earthquakes with magnitude \( 7\frac{1}{2} \). Soloviev interpreted the parameter, \( T \), as the ability of earthquakes in a given region to generate tsunamis. Table 2.3 lists values of \( T \) for various regions.

According to Soloviev, the difference in tsunami probability among different regions can probably be accounted for by the peculiarity of tectonics of zones. In zones of block tectonics with rigid superficial layers and superficial seismic activity, the probability of an earthquake exciting a tsunami is higher than in zones of so-called arc tectonics with relatively mild superficial layers and somewhat buried seismic activity.

**Methods of Estimating Tsunami Energy**

Generally speaking there are four methods to estimate tsunami energy: (1) based on propagation of tsunami waves from the generation area, (2) knowing the vertical displacement at the generation area, (3) from the empirical relation between earthquake and tsunami magnitude, and (4) knowing the maximum predominant period of a tsunami.

The first (Takahasi) method has been described earlier. In the second method, the principle that the earthquake causes the sea bottom to move vertically is used, and the work partly goes into increasing the potential energy of sea water while the rest goes into generation of tsunami. The work done against the hydrostatic bottom pressure, \( \rho g D \), is:

\[
E_1 = \rho g \sum S b D dS
\]  

(2.22)

where \( D \) is the water depth, \( b \) is the vertical displacement of a small portion, \( dS \), of the sea bottom. The work done that goes into increasing the potential energy of the water is:

\[
E_2 = \rho g \sum S b \left( D - \frac{b}{2} \right) dS
\]  

(2.23)

Hence, the energy, \( E \), transformed into tsunami is:

\[
E = E_1 - E_2 = \frac{1}{2} \rho g \sum S b^2 dS
\]  

(2.24)

observations showed that \( E \) determined from (2.24) is probably an overestimate.

In an average sense, one can write:

\[
\sum S b^2 dS = b_m^2 A
\]  

(2.25)

where \( b_m \) is the average vertical displacement and \( A \) is the area of dislocation at the tsunami origin. From (2.24) (2.25)

\[
E = \frac{1}{2} \rho g b_m^2 A
\]  

(2.26)
At this stage, Iida (1963a) assumed that the area, $A$, of bottom dislocation is approximately equal to the area, $A_s$, of aftershock activity, which of course is a highly questionable assumption. He gave the following relation between $A$ and earthquake magnitude, $M$:

$$\log A = 0.9M - 7 \quad (2.27)$$

This formula disagrees with the relation he gave in his 1963b paper (Equation (2.17)). From (2.26) (2.27):

$$\log E = 9.69 + 0.9M + 2\log b_m \quad (2.28)$$

Here $b_m$ is expressed in cm. Table 2.4 lists tsunami energy computed from this formula.

<table>
<thead>
<tr>
<th>Earthquake</th>
<th>Vertical displacement (m)</th>
<th>Tsunami energy ($10^{22}$ ergs)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mar. 3, 1933</td>
<td>6.5</td>
<td>18</td>
</tr>
<tr>
<td>Nov. 4, 1952</td>
<td>10.7</td>
<td>14</td>
</tr>
<tr>
<td>Dec. 21, 1946</td>
<td>9.5</td>
<td>8</td>
</tr>
<tr>
<td>Mar. 2, 1952</td>
<td>6.5</td>
<td>4</td>
</tr>
<tr>
<td>Dec. 7, 1944</td>
<td>10.0</td>
<td>8</td>
</tr>
<tr>
<td>Nov. 3, 1936</td>
<td>7.1</td>
<td>2.1</td>
</tr>
</tbody>
</table>

In the third method, the empirical relations between earthquake energy and tsunami energy were used to estimate the latter (Equation (2.13)).

In the fourth method, the empirical relation between the predominant period, $T_p$, of a tsunami and tsunami energy, $E_r$, is used:

$$\log E_r = 11.7 + 6.8 \log T_p \quad (2.29)$$

The predominant period, $T_p$, of a tsunami is defined by Iida (1963a) as that period with the maximum spectral intensity which can be determined from tide-gage records. Table 2.5 compares energy of different tsunamis determined by the four methods.

Grigorash and Korneva (1970) studied the effect of refraction to estimate tsunami energy based on tide-gage records. The tsunami they studied was in the Black Sea after the earthquake at Anapa July 12, 1966. The small tsunami caused by this earthquake was recorded at several tide gages on the Crimean and Caucasian Black Sea coasts. In their analysis, the tidal record from Gelendzhik was mainly used. They assumed that the average speed of the tsunami was $180$ km/h$^{-1}$ and the period of the tsunami was 38 min. When refraction near the epicenter was ignored, the energy, $E_r$, of the tsunami was estimated to be $6.5 \times 10^{20}$ ergs, whereas
when refraction is included, the energy was estimated as $2 \times 10^{19}$ ergs. Thus, ignoring refraction, one can overestimate the energy by 30 times.

<table>
<thead>
<tr>
<th>Earthquake</th>
<th>Tsunami energy ($10^{12}$ ergs) estimated from</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Location</td>
</tr>
<tr>
<td>Mar. 3, 1933</td>
<td>off Sanriku</td>
</tr>
<tr>
<td>Dec. 7, 1944</td>
<td>off Tonankai</td>
</tr>
<tr>
<td>Dec. 21, 1946</td>
<td>off Nankaido</td>
</tr>
<tr>
<td>Mar. 2, 1952</td>
<td>off Tokachi</td>
</tr>
<tr>
<td>Nov. 4, 1952</td>
<td>Kamchatka</td>
</tr>
</tbody>
</table>

**TSUNAMI SOURCE MOTION**

Adams and Furumoto (1970) recognized the binary relation between earthquakes and tsunamis. "...either a very large tsunami is generated or none at all...."

Japanese tsunamists have paid significant attention to the association in the area of tsunami generation and area of aftershock activity (e.g. Hatori 1963). One important point made by Hatori is that the tsunami-generating areas are usually elliptical. Hatori (1970) found that the features of tsunami sources for northeastern and southwestern Japan are different. In the northeast, sources of small tsunamis are mostly located in the sea, at depths shallower than 2 km, and sources of large tsunamis lie on the steep continental slope near the trench. In southwestern Japan, sources of large tsunamis lie near the coast. The major axes of source areas are parallel to the island arcs. The dimension of the source area of a tsunami generated by a deep earthquake is considerably smaller than that from a shallow earthquake. Watanabe (1970) studied the tsunamigenic earthquakes in and near Japan for 1900–68 and placed special emphasis on the tsunami source, area of aftershocks, magnitude of earthquakes, and type of faulting.

Van Dorn (1964) determined the source mechanism of the Alaskan earthquake tsunami of March 1964, based on crustal dislocation following the earthquake, the behavior of the water immediately following the earthquake, and the tsunami record at Wake Island. He found that enormous areas have undergone relatively large vertical changes in elevation relative to sea level. Virtually the entire Kenai Peninsula, from the Turnagain Arm of Cook Inlet, including the Kenai–Kodiak Ridge, and Kodiak Island itself, appears to have subsided by 0.6–1.8 m (Fig. 2.3). At the same time, most land areas along the sea coast from the Yakutat area to the center of Prince William Sound have been elevated by similar amounts. The axis about which this dipolar movement occurred is shown in Fig. 2.3.
Regarding the pattern of water movement, Van Dorn stated that there was relatively little exchange of energy between Prince William Sound and the Gulf of Alaska, because the Sound is virtually cut off from the Gulf by a group of islands. Van Dorn analyzed the tsunami record at Wake Island (6052 km from the epicenter) and concluded that a large fraction of the tsunami energy was contained in a solitary wave formed in the Gulf of Alaska.

Pararas-Carayannis (1967) studied the source mechanism of the 1964 Alaskan earthquake tsunami somewhat along the same lines as Van Dorn (1964). Although his results generally agree with those of Van Dorn, there are some differences. He estimated the total area of displacement by the earthquake to be 215,000 km². Van Dorn has not given any corresponding value, at least explicitly. The energies of the tsunami were estimated by the authors as $5.88 \times 10^{21}$ ergs (Pararas-Carayannis) and $2.3 \times 10^{21}$ ergs (Van Dorn).
Pararas-Carayannis also supported the contention of the Japanese tsunamists that the area of major aftershock activity and the generation area of tsunamis roughly agree. To support this, he stated that about 52 aftershocks occurred following the 1964 Alaskan earthquake, the areas agreeing with tsunami source area.

Van Dorn attributed the tsunami to the tectonic movement caused by the earthquake, but Pararas-Carayannis, although agreeing with Van Dorn about the tsunami in the Gulf of Alaska, argued that in Prince William Sound waves were generated by slumping and landslides, in addition to the tectonic mechanism. He reasoned that if the tsunami was only generated by tectonic movement, then the first wave would be the highest. However, observations showed that at Uzinki, Kodiak City, Women's Bay, and other places on Kodiak Island, the third and fourth waves were highest. In fact, Pararas-Carayannis identified two different major tsunamis in Prince William Sound, one at Montague Island and the other at Port Valdez.

Pararas-Carayannis (1967, p. 308), however, was careful not to attach too much importance to the second mechanism. He stated:

“Slumps or avalanches, similar to the ones that occurred in Prince William Sound, are usually localized; they can produce no large tsunami that would travel across wide portions of the ocean. According to Wiegel (1954), not more than 2% of the potential energy of a falling or sliding body is converted into wave energy. In Prince William Sound however, slumping and sliding when added to tectonic movements created tsunami waves of very large energy, but their effect was catastrophic only locally; very little of the energy escaped the Sound.”

Van Dorn (1963) determined the source mechanism of the tsunami caused by the Aleutian earthquake of Mar. 9, 1957. The principle used was to take the wave record at Wake Island and calculate its spectrum. By inversion of the Kanzer-Keller solution (Chapter 1) the spectrum at the source was obtained. He found that the tsunami source was a circular depression and had displaced a total volume of $2.3 \times 10^{10}$ m$^3$. About 4% of the seismic energy had gone into tsunami energy.

Sokolowski and Miller (1970) developed a simple technique for identification of tsunami source region from a single seismic record. Although this method might not work universally, it has the merit that (where it works) it can be used in real time prediction. Sokolowski and Miller (1970, p. 135) described the principle of their method:

“In selected cases, there appears to be a similarity between seismic records of earthquakes which occur within a small source region. This similarity exists only between the record of a given seismograph station, not between the records of earthquakes recorded at different stations. If this similarity is consistent, it may be possible to identify in real time the particular fault system in which an earthquake occurs, assuming a sufficient amount of reference data exists. As the net ground motion along a given fault is typically consistent, it should be possible to improve the estimation of tsunami generation.”

In Chapter 3, the T-waves produced by an earthquake will be considered. Johnson (1970) proposed a method to determine the rupture length following an earthquake, by T-phase data. Johnson believed that rupture length is better than
earthquake magnitude as an indicator of the tsunami-generating potential of an earthquake. Certainly this presents a different viewpoint and should be seriously considered.

Johnson and Norris (1968), during their study of a series of Aleutian earthquakes starting Feb. 4, 1965, showed that T-phase source solutions appeared in groups about some points along the Aleutian ridge. These cluster points were referred to as T-phase radiators and, according to the authors, a major earthquake will excite T-phase radiators in a sequence as the rupture progresses along the fault. Eaton et al. (1961) proposed that the duration of the T-phase can be used to estimate the rupture length.

Johnson and Norris (1968) studied the Fox Island earthquake swarm of Dec. 21, 1962, and showed, through study of the T-phases, that each earthquake in the swarm affected the same region of the ocean floor, but with differing strengths. They concluded that the region affected was too confined and could not generate a tsunami and their surmise proved to be correct.

It might not be feasible to collect the T-phase from different stations and coordinate it properly for practical prediction in the short time available. To overcome this difficulty, Johnson (1970) proposed that an array of hydrophones be used at a single station and the azimuth of the signal received could be calculated through cross-correlation analysis.

There might be some difficulty in determining precisely the duration of the T-phase (defined as the time between the arrivals of the T-phase from the first and last radiator excited by the main shock) when aftershocks are present. Johnson proposed to rectify this by a correlogram that would show a monotonic variation in the azimuth of the peaks.

Ionospheric recording of Rayleigh waves due to earthquakes (e.g. Yuen et al. 1969) can be used to determine the source mechanism. This method will be especially advantageous for the real time tsunami prediction problem, because other seismic methods of determining the source mechanism through P or S wave analysis might be impractical in the short time available.

Furumoto (1970) proposed that Rayleigh waves recorded by ionospheric Doppler effects could be studied through the surface wave analysis technique developed by Brune et al. (1960), to yield information on the source mechanism. The success of this method will depend on a prior knowledge of the dispersive properties of the medium between the epicenter and the seismic recording station. For details of the surface wave analysis see Brune et al. (1960).

2.2 Tsunami Generation by Earthquakes

In Section 1.2 the classical C-P problem and the theories of Kranzer and Keller, and Kajura were considered. In this section, the application of the C-P problem to tsunami generation following occurrence of an earthquake will be discussed. Special emphasis will be given to numerical models that simulate tsunami generation much more realistically than analytical models.
**Analytical Models**

*Modifications to the classical C-P problem* — Carrier (1971) developed an analytical theory for the generation and propagation of tsunamis and applied this to the 1964 Alaskan earthquake tsunami. Two important results emerged: (1) for tsunami source areas of considerable breadth dispersion is not important, whereas for narrow initial displacements it is extremely important; (2) if the statement that the second or third crest is higher than the first is valid, then the ground displacement must consist of elevation and subsidence and not just elevation.

Podyapolsky (1970) developed a theory for tsunami generation by treating the crust of the earth as an elastic solid half-space and treating the ocean as an elastic liquid in a uniform field of gravity. Displacement of the ocean bottom following the earthquake was considered small compared to the uniform depth of the ocean. He concluded that the differences between his theory and the relatively simpler classical theory are insignificant for periods of the order of $10^4$ s, but for tsunamis with periods up to $10^3$ s, the classical theory is adequate.

Cherkesov (1965, 1966a, 1968) and Fedosenko and Cherkesov (1968) studied the effect of inhomogeneity of the ocean on tsunami waves. They showed that the effect of stratification on tsunamis is not significant. One interesting conclusion of Cherkesov's 1965 paper is that secondary tsunamis could be caused by nonuniformity such as depth discontinuities, when a density discontinuity occurs simultaneously. Cherkesov (1970) investigated the effect of viscosity on tsunami generation and propagation and showed that viscosity reduces the maximum wave amplitude at some point far from the source area; this reduction factor depends on shallow-water and deepwater depths. Also, viscosity influences low water after the arrival of the main crest of the tsunami.

**Numerical Models for Tsunami Generation**

Compared to analytical theories available for tsunami generation, numerical models aimed at simulating tsunami generation are few. Aida (1969a, b) used numerical modeling to study the tsunamis caused by the 1964 Niigata and 1968 Tokachi-Oki earthquakes. The computed water levels at two stations agreed reasonably well with observed levels (Fig. 2.4). Aida (1969a) assumed instantaneous deformation of the bottom, whereas in his 1969b paper he considered progressive deformation. These numerical experiments threw some light on the important question of directivity coefficient, defined as the ratio of the heights of the waves radiated in opposite directions. The directivity coefficient depends on the speed of the bottom deformation in the horizontal plane. If the depth decreases in the direction of progress of the bottom deformation, then the directivity coefficient increases in that direction.

Hwang and Divoky (1970, 1971, 1972) numerically simulated the tsunami generation and propagation following the 1964 Alaskan earthquake. Hwang et al. (1970a) performed numerical experiments to simulate the tsunami generation caused by the underground nuclear explosions "Milrow" and "Cannikan" and the Rat Island earthquakes. Hwang et al. (1970b; 1972a, b) simulated the tsunamis due to the Alaskan earthquake of March 1964, the Chilean earthquake of May 1960,
and the Andreanof earthquake of 1957.

Hwang et al. (1970a) assumed that a portion of the sea bottom moves as:

\[ Z = -\xi(x,y,t) \]  \hspace{1cm} (2.30)

The continuity equation is:

\[
\left(\frac{\partial \eta}{\partial t} - \frac{\partial \xi}{\partial t}\right) + \frac{\partial}{\partial x} \left((D - \xi + \eta) \, U\right) + \frac{\partial}{\partial y} \left((D - \xi + \eta) \, V\right) = 0 \]  \hspace{1cm} (2.31)

where \( U \) and \( V \) are the vertically averaged velocity components in the \( x \) and \( y \) directions, \( D \) is the water depth, \( \eta \) is the water level relative to its equilibrium position, and \( \xi \) is the displacement of the ocean bottom. The horizontal equations of motion, with the inclusion of the convective acceleration terms but the neglect of friction and Coriolis terms (in later models they included Coriolis terms and spherical polar coordinates), along with the continuity equation, form a set of three equations for the three variables \( U \), \( V \), and \( \eta \).

The finite difference forms of these equations, suitable for numerical integration in time, were adapted from Leendertse (1967). Each time step was split in half. In the first half step, \( \eta \) and \( U \) were computed implicitly whereas \( V \) was computed explicitly and this process goes from time step \( n \Delta t \) to \( (n + \frac{1}{2}) \, \Delta t \). During the progression to \( (n + 1) \, \Delta t \), \( \eta \) and \( V \) are computed implicitly and \( U \) is computed explicitly. Inclusion of the Coriolis terms needs special attention because \( U \) term occurs in the \( V \) momentum equation and vice versa.
Hwang and Divoky (1970) numerically simulated the tsunami created by the 1964 Alaskan earthquake and prescribed the following bottom displacement: let \((a,b)\) and \((i,j)\) denote the coordinates (with reference to a Cartesian system) of the epicenter of the main shock and some point in the source region. Then the vertical motion at \((i,j)\) is assumed to start at time

\[
t_{i,j} = \frac{\Delta x}{V_p} \left\{ (a - i)^2 + (b - j)^2 \right\}^{\frac{1}{2}}
\]

where \(\Delta x\) is the grid spacing and \(V_p\) is the velocity with which the rupture propagates. Plafker and Savage (1970) estimated \(V_p\) to be 3–4 km/s.

The bottom displacement is prescribed as follows:

\[
\xi_{i,j}(t) = \begin{cases} 
0 & \text{for } t \leq t_{i,j} \\
\xi'_{i,j} \sin^2 \left( \frac{\pi(t-t_{i,j})}{2\tau} \right) & \text{for } t_{i,j} \leq t \leq t_{i,j} + \tau \\
\xi'_{i,j} & \text{for } t \geq t_{i,j} + \tau
\end{cases}
\]

Here \(\tau\) is a characteristic time scale of ground motion. Berg and Housner (1961) estimated \(\tau \sim 10\) s generally. The parameter \(\xi'_{i,j}\) is the permanent displacement at \(i,j\) following the earthquake.

Figure 2.5 shows the positions of the leading wave of the tsunami at successive times (in seconds). The dotted area represents the tsunami source region. For large travel distances, instead of a Cartesian system, spherical polar coordinates were used (Hwang and Divoky 1975).

\[\text{FIG. 2.5. Position of leading disturbance of the Alaskan tsunami at successive times (in seconds) (region of ground motion shaded). (Hwang et al. 1970b)}\]
Although use of a spherical polar coordinate system makes the model more sophisticated, nevertheless, there are some uncertainties in the boundary conditions. Usually, a condition of complete reflection is adopted at closed boundaries, mainly for simplicity. This is not a good assumption because limited evidence indicates tsunami energy is lost at a rate of roughly 60% per shoreline interaction (Munk et al. 1962). A better condition should incorporate such a dissipative factor into the model. The condition at open boundaries is more difficult to specify. The physical requirement is that the wave be totally transmitted through the boundary, but this cannot be achieved without computation beyond the boundary. An artifice sufficiently accurate is to assume that the wave profile travels without change of form across the final space grid point. Figure 2.6 shows the position of the leading wave of the tsunami at several times, following the Chilean earthquake of May 1960. The dotted area shows the tsunami source region.

![Fig. 2.6. Computed position of leading disturbance at various times. Chile 1960; region of ground displacement shaded. (Hwang and Divoky 1975)](image)

These numerical experiments also confirmed the phenomenon of energy radiation preferentially directed normal to the major axis of the source. This directivity of radiation is dependent on the elongation of the source region. It appears that sources for large tsunamis are typically quite elongated with major axes paralleling the local coast, trench, or island arc so that generally the dominant flow of energy is seaward.

Garcia (1972) studied the generation of tsunamis due to motion on the eastern end of the Mendocino Escarpment near California. He used the complete Navier-Stokes equations and restricted his study to the first 200 s after generation. An initial single crest split in two and moved in opposite directions. He verified his numerical results against his laboratory experiment which will be briefly discussed later.
2.3 Tsunami Generation by Nonseismic Causes

Tsunami generation caused by sources other than earthquakes (such as volcanic eruptions, nuclear explosions, and landslides) will be considered.

Application of C-P Problem to Explosions at or Near the Water Surface

Terazawa (1915) studied an extended initial disturbance below the surface to simulate an underwater explosion. He assumed that the water body had infinite horizontal dimensions and infinite depth. Two different forms of initial conditions were made: he allowed first an initial surface displacement without an initial velocity, and second, an initial surface impulse without an initial surface displacement. Terazawa (1917) also considered the case when an explosion occurred below the free surface at a finite depth. However, he made at least two unrealistic assumptions: first, that either the explosion took place at such great depth or was so weak that it would not break the water surface by ejecting a column of water upward; and second, that the pressure pulse generated by the explosion was assumed to travel with infinite speed. The second assumption enabled him to use the method of images, widely used in electrostatics.

Nomura (1938) extended Terazawa's work to include water of finite depth. Unoki and Nakano (1953a, b) applied the C-P problem to the tsunamis generated by the Myojin volcanic explosions. Keller (1963) showed that a displacement at the surface is more efficient in producing water waves than an identical displacement at the ocean bottom. Levy and Keller (1963) considered instantaneous vs. noninstantaneous sources, point sources vs. extended sources, and symmetric vs. asymmetric sources.

Tsunamis Due to Volcanic Eruptions

The volcanic Myojin Reef (32°N, 140°E) erupted several times in September 1952, and generated three different tsunamis (Nakano et al. 1954). The first tsunami occurred Sept. 16; the second, Sept. 24, and was probably responsible for the destruction of the survey ship *Kayo-Maru No. 5*, belonging to the Hydrographic Service of Japan. A third tsunami occurred Sept. 26. Lane (1965, p. 252) stated that the tsunamis attained maximum heights of 7 m.

When the volcano Sakurajima erupted in January 1914, it proved to be one of the largest volcanic eruptions in the history of Japan. The tsunami following the major eruption attained a height of about 3 m. There appears to be an association between volcanic eruptions and earthquake swarms. One example is the Ebino earthquake swarm in 1968 and its relation to the Kirisima volcanoes. These swarms generated tsunamis that caused some damage to the coasts of Miyazaki and Kooti prefectures.

The volcanic Mount St. Augustin, Alaska, erupted on Oct. 6, 1883, at about 0800 h (local time). Twenty five min after the eruption, a tsunami 7.6 ± 0.9 m high invaded Port Graham (Davidson 1884). Two waves 5.5 and 4.6 m high followed at 5 min intervals.

One of the most powerful volcanic eruptions appears to have occurred around 1450 B.C. (suggested through radio carbon dating). The volcanic island Santorin...
(now called Thera) in the Aegean Sea erupted. The tsunami generated was estimated to have been over 100 m high and inundated the shorelines of the eastern Mediterranean Sea (Marinatos and Imboden 1972). According to the authors, the Santorin eruption was more violent than that of Krakatoa (83 km² of erupted island vs. 21 km²).

A most devastating tsunami due to nonseismic causes occurred when the volcano Krakatoa (in the Sunda Strait between Java and Sumatra) erupted in August 1883. The major eruption occurred at 1055 h (Batavia time) on Aug. 27, and was the largest explosion in recorded history. The tsunamis following this underwater eruption were up to 30 m high and caused extensive damage on the coasts of Java and Sumatra (Lane 1965). About 36,000 people died, 295 villages and towns were inundated, and 5000 ships were destroyed. The tsunami appears to have traveled at speeds from 560 to 640 km/h. The tsunami was noticed at Port Elizabeth (Gill 1883) in South Africa (8045 km away) and there are indications that it might have arrived in the English Channel (17,700 km from Krakatoa). For details on the tsunami heights at various locations see Verbeek (1884). The period of the tsunami varied from 75 min at Negapatam (on the southeast coast of India) to 24 min at Aden. The wave amplitude varied from 0.56 m at Negapatam to 0.23 m at Aden. No tide-gage disturbances were observed at Singapore, only one-third the distance of Negapatam. There appears to be two reasons for this. First, the directivity of the tsunami was such that its energy was concentrated toward the Indian Ocean (and not toward Singapore) and second, many shallow areas and islands between Krakatoa and Singapore could have dissipated the tsunami.

The tsunami was also recorded (Boutelle 1884) at Kodiak, Alaska, and at Saucelito, near San Francisco. One curious result was that, although Kodiak is 2370 km closer to Krakatoa than San Francisco, it felt less impact of the tsunami. Also, the tsunami appeared to have arrived earlier at San Francisco than at Kodiak, which probably can be explained by the different paths traveled by the tsunami.

Earlier explanations of the tide-gage disturbances following the Krakatoa eruption were based on the travel of the tsunami as a long gravity wave, whereas later theories also included the important phenomenon of atmospheric-oceanic coupling. The Royal Society prepared an extensive report on the various aspects of the Krakatoa eruption (Symons 1888) and this report formed the basis for several investigations.

Ewing and Press (1953, 1955) explained the tide-gage disturbances through the concept of coupling between the atmosphere and the ocean. In their 1953 paper (p. 3) they described some important consequences of the eruption. A barometric disturbance traveled outward from the volcano and was recorded at many stations on at least three passages around the earth. Ewing and Press specifically examined the following conclusions arrived at by Captain W. J. L. Wharton (in the Royal Society Report) on the tsunami following the eruption: (1) to the north and east in the Java Sea the waves could be traced for 724 km, then were reduced to small amplitudes; (2) to the south and east the propagation was limited, not extending beyond the west coast of Australia; (3) to the west the waves traveled over great distances, reaching Cape Horn and possibly the English Channel; and (4) distur-
bances in the Pacific and Caribbean had no connection with Krakatoa, but were the results of some seismic activities.

Ewing and Press concluded that, whereas the tide-gage disturbances in the Indian Ocean could be explained as due to the tsunami, disturbances in the other oceans could be accounted for by the coupling of the barometric disturbance to the ocean. This latter assumption was based on the fact that the times of tide-gage disturbances correlated well with arrival times of either the first or second airwave.

Ewing and Press (1953) gave a table listing some stations outside the Indian Ocean where the tide-gage disturbances were observed, with the local times of arrival of the airwave, and the expected arrival time of the tsunami generated by atmospheric coupling. A correction (based on Wharton's work) was applied to the travel times to take into account the shallow region near the tide-gage station. The differences between the observed time of tide-gage disturbance and expected time of occurrence based on atmospheric coupling were generally less than 1 h. Only at St. Paul (Kodiak Island) was the difference as large as 4.5 h. Ewing and Press hypothesized that the Alaskan Peninsula and the mainland of North America, respectively, could have blocked the first and second waves.

Although Ewing and Press (1953, 1955) proposed the mechanism of air coupling they did not resolve the subsequent problems their hypothesis raised. They mentioned they were investigating whether a free wave in the ocean with a phase speed of 1100 km/h (the phase speed of the Krakatoa airwave) is possible and whether significant coupling of atmosphere to ocean could occur even though there is no resonance.

Press and Harkrider (1966) considered this problem and calculated the acoustic-gravity modes (described in Chapter 6) for an ocean of uniform depth of 5 km, underlying an atmosphere with many isothermal layers. This calculation (Press and Harkrider 1962; Harkrider 1964) revealed that free waves with phase velocities close to long gravity waves in the ocean do exist in the atmosphere and these could efficiently transfer energy to the oceanic modes.

To prove that the atmospheric mode did excite the ocean mode, Press and Harkrider (1966) plotted the ratio of the water level deviation to the atmospheric pressure at sea level as a function of the phase velocity, and found resonant peaks corresponding to all gravity modes.

Garrett (1970), although agreeing with Harkrider and Press (1966, 1967) on some counts, disagreed on others and offered, in my opinion, a correct explanation. Garrett (1970, p. 44) questioned the possibility of the airwave having the same phase speed as the ocean wave generated by the Krakatoa eruption, and elaborated on his skepticism as follows:

"There are certain difficulties associated with the mechanism proposed by Harkrider and Press, notably those connected with the difference between the phase and group velocities of the atmospheric gravity wave, and with horizontal variations in ocean depth, but it suffices to point out that their theory is inconsistent with many of the observations."

Garrett proposed that free ocean waves are produced by the atmospheric pressure pulse at any abrupt change in the depth of the ocean (e.g. the continental slope), and he satisfactorily accounted for the observed tide-gage disturbances.
According to the inverse barometer effect, 1 millibar (mb) of surface pressure corresponds to an amplitude of about 1 cm. The atmospheric pressure wave from Krakatoa was probably a few mb, but as the tide gages are situated usually in shallow water, the forced displacement of the sea surface due to the pressure wave would be much less than 1 cm, and undetectable.

Garrett then considered free waves generated at a depth discontinuity. Compared to the wavelength of the airwaves from the Krakatoa eruption, the continental slope can be treated as an abrupt depth discontinuity, rather than a gradual variation of depth. Thus, in the course of traveling over the ocean to the coast with an amplitude of a few mb, a pressure pulse can produce a transmitted free wave with an amplitude of several cm and can be detected. Shallowing and resonance, could amplify this even further.

Before comparison with observed data, the highlights of Garrett’s theory will be briefly summarized. The atmospheric pressure pulse from Krakatoa could generate at the continental slope free ocean waves of sufficient amplitude to be detectable by coastal tide gages. When the airwave arrives from the ocean to the land, the first sea wave could be thought of as a free wave generated at the continental slope, but when the airwave travels from the land to the ocean, the sea wave is hardly detectable (available data indicate this). Sea waves arriving at later stages may be explained as free waves generated either at the opposite side of the ocean or at some other major depth changes in the ocean so these depth changes may focus the long waves at certain places, enhancing their amplitude.

Let $t_a$, $t_s$, and $t_a$, respectively, be the times when the airwaves pass the tide gage, the travel time of the sea wave from the continental slope to the tide gage, and the travel time of the airwave from the continental slope to the tide gage. Then the expected time of the sea wave at the tide gage is $t_o + t_s - t_a$ which is strictly true only for normal incidence. But as $t_a << t_s$, the sea wave is expected approximately at time, $t_s$, after the arrival of the airwave. This agrees with the conclusions arrived at by Ewing and Press (1953) based on Wharton’s data, bearing in mind that Ewing and Press considered only the first arrivals. Also, some identifications of the disturbances by Wharton on the tide-gage records are open to doubt, according to Garrett.

Garrett (1970) was surprised that the extensive Royal Society Report by Symons (1888) contained no information on the pressure amplitude of the airwaves. However, Scott (1883) showed that in Europe the first two airwaves (one direct and the other through antipodes) had an average range of 3 mb. As Europe is roughly 90° from Krakatoa, the airwave was not the largest. It is quite possible that, with pressure amplitudes in excess of 3 mb, the airwaves could have caused the observed tide-gage disturbances.

Another conclusion of Ewing and Prés (1953) was that no tide-gage disturbances were observed when the airwave traveled from the continents to the oceans. This is in agreement with Garrett's theory, because the sea wave in this case should only have $\frac{1}{4}$ the amplitude of the reverse situation, at least for normal incidence. According to Garrett, this wave should disturb the tide gage at $t_s + t_a$ after the passage of the airwave over the gage.
Figure 2.7 shows the tidal record for Devonport where the generated sea wave (when the airwaves traveled from land to ocean) was detectable. In this diagram, $A^I$ and $A^{II}$ are the arrival times of the direct airwave and the airwave through the antipodes. $A^I$ is offshore and $A^{II}$ is onshore and both were approximately normal to the continental slope. The first arriving sea wave associated with $A^{II}$ is marked $S^{II}$ whereas $S^I$ is the first arriving sea wave associated with $A^I$. Garrett explained that the sea wave at San Francisco which arrived $5_2^h$ after the first (onshore airwave) probably was generated either near the Philippines, or as the airwave was passing over the Mariana Ridge south of Japan. Some contribution to the San Francisco record might have come from the refracted free waves generated by airwaves traveling over the Hawaiian Ridge and the Aleutians. The main sea wave at St. Paul (Kodiak) arrived some 13 h after the arrival of the first wave and 2 h before the arrival of the second. This sea wave might have been generated at the Philippines and then reflected off North America.

**TSUNAMIS DUE TO NUCLEAR EXPLOSIONS**

Garrett (1969, 1970) tried to detect the sea waves arising from the atmospheric pressure pulse caused by the large Soviet nuclear test on Oct. 30, 1961, over Novaya Zemlaya, when a $5.7 \times 10^{10}$ kg bomb was detonated. No significant sea waves were detected either associated with this test or any other atmospheric nuclear test (underground nuclear tests will be considered later). To understand the energies involved in nuclear explosions and volcanic eruptions, according to Press and Harkrider (1966), a surface explosion of $100-150 \times 10^{10}$ kgm would be required to duplicate the pressure pulses generated by the Krakatoa eruption.

Van Dorn (1961) made wave measurements at Eniwetok Atoll, Wake Island, and Johnston Island during the underground nuclear explosions "Cherokee," "Zuni,"
“Navaho,” and “Tewa” conducted at Bikini Atoll in 1956. Figure 2.8 shows locations of the tests and observing stations, and the number below the names of observing stations are the great circle distances between that station and Bikini Atoll in nautical miles.

From the analysis of his data, Van Dorn found that the linear theory was adequate to account for the observed dispersion of these gravity waves as well as their decay with increasing distance. Van Dorn also studied the dispersion of the tsunami of Mar. 9, 1957, and it was quite similar to that of the explosion-generated waves. Thus, Van Dorn’s study gave support to the contention that for a centered wave system phase dispersion does not depend on the nature of the source. Another interesting result from Van Dorn’s study is that even small islands can effectively scatter the waves and a correction has to be applied in calculating

**FIG. 2.8.** Distribution of wave recording stations and great circle distances in kilometers from Bikini Atoll. Contours give instantaneous positions of wave front originating at Bikini at 20-min intervals. (Van Dorn 1961)

**FIG. 2.9.** Waves generated by the underground nuclear test Oct. 1969, Amchitka Island. (Vitousek and Miller 1970)
the rate of amplitude decay. For details on the scattering correction see Van Dorn (1961, fig. 15).

Vitousek and Miller (1970) described an instrumentation system that could be used to measure tsunami waves in the open ocean. They claimed this system is capable of functioning to a depth of 6 km. Here, only the measurements made by this system on the tsunami due to the “Milrow” underground nuclear explosion on Amchitka Island, Oct. 2, 1969, will be briefly discussed. Figure 2.9 shows the envelope of waves generated. This test, which had a yield of about $10^9$ kg, produced waves only with amplitudes of a few cm. The “Cannikan” explosion on Amchitka Island in 1971, with a yield of about $.5 \times 10^{10}$ kg, did not create a significant tsunami.

Hwang et al. (1970a) performed numerical experiments to simulate the tsunamis generated by the underground nuclear explosions “Milrow” and “Cannikan.” Figure 2.10 shows the contours of the water level of the tsunami generated by the Milrow

![Fig. 2.10. Surface displacement contours (in cm) resulting from nuclear test on Amchitka Island. (Hwang et al. 1970a)](image-url)
test, as simulated by a numerical experiment. In this diagram, the epicenter refers to the site of the explosion.

**Turbidity Currents and Landslides**

Although turbidity currents and tsunamis are not directly related, nevertheless, both could be caused by earthquakes such as the 1929 Grand Banks earthquake. This topic is included because it is considered part of the tsunami literature. Turbidity currents and tsunamis could also be generated by landslides whether or not an earthquake occurred.

Most submarine telegraph cables from North America to Europe pass south of Newfoundland (Heezen and Ewing 1952). At about 2032 h GCT, Nov. 18, 1929, an earthquake of magnitude 7.2 occurred on the continental slope southeast of the Cabot Trench. This generated a tsunami that caused considerable property damage and loss of life along the shores of Placentia Bay.

Another important consequence occurred during the 13 h and 17 min following the earthquake. An orderly sequence of breaks occurred in the telegraph cables to 483 km south of the epicenter. According to Heezen and Ewing (1952), although all cables along the continental slope and on the floor of the ocean south of the epicenter were broken, none on the continental shelf were disturbed. The exact times and locations of the cable breaks were known, respectively, from the telegraph records and resistance measurements.

Bucher (1940) hypothesized that erosion in the submarine canyons caused by the tsunami left the cables unsupported, thus leading to breakage. Heezen and Ewing (1952) criticized Bucher's explanation on the grounds that the cable breaks were too regular in time and the cables were of different ages and breaking strengths. In turn, they offered the following explanation (Heezen and Ewing, p. 864):

> "A severe shock jarred the continental slope and shelf, setting landslides and slumps in motion. These virtually instantaneous movements affected an area 129 by 241 km along the continental slope. These mass movements, starting on the relatively steep continental slope, raced downward, and by the incorporation of water, the moving sediment was transformed from sliding masses into turbidity currents. Undoubtedly the pattern of canyons and tributaries caused initial concentrations of the flow, but as the canyons joined the currents grew larger, quickly becoming so large that they were not restricted by the submarine canyons and eventually covered the bottom of the 322-km wide bight which lies between the southern Grand Banks and the continental slope off Sable Island. The currents had many times the force necessary to break the cables, and snapped each cable shortly after reaching it."

Plapp and Mitchell (1960, p. 983) defined a turbidity current as follows:

> "A turbidity current occurs along a sloping bottom in a large body of liquid when liquid adjacent to the bottom contains suspended sediment that causes the average density of the mixture to be greater than the density of the surrounding clear water. The current flows down the slope and may continue to flow for a long distance after it reaches a relatively level section of the bottom."

Calculations of the progress of the turbidity current following the 1929 Grand Banks earthquake was based on the orderly breaks in the cables. The first cables to break were on the continental slope. In the literature, many more details are given of this particular turbidity current and the cable breaks.
A schematic representation of a turbidity current with low and high velocity is shown in Fig. 2.11.

![Diagram of turbidity currents](image)

**Fig. 2.11.** Diagrammatic section of turbidity currents. (A) current with relatively coarse suspended load (or small velocity), (B) current with relatively fine-grained suspended load (or high velocity). (Kuenen 1956)

Heezen and Ewing (1952) showed rather convincingly that the successive series of breaks in the telegraph cables following the Grand Banks earthquake of 1929 were caused by the turbidity current generated. Heezen et al. (1954) presented additional evidence.

Kuenen (1952) agreed that turbidity currents broke the cables, nevertheless, he wondered whether velocities of up to 100 km/h are possible over hundreds of kilometers of level ground. The following estimate made by him shows that these velocities are not unreasonable.

He estimated the size, $V$, of the turbidity current from the relation:

$$ V = C \sqrt{msd} $$  \hspace{1cm} (2.34)

where $C$ is a constant, $s$ is the slope of the ground on which the current is flowing, $d$ is the effective density, and $m$ is a hydraulic mean depth. Actually this formula is normally used to calculate the velocity of flow in a river. As is obvious, the constant, $C$, which depends on friction, must be larger for the turbidity current than for river flow.

According to Kuenen, $C$ could have maximum values to 700 or 800 cgs. He gave tables of the velocity of the turbidity current for different slopes with $C = 400$ and $d = 0.6$. He also varied $C$ from 200 to 600 and $d$ from 0.3 to 0.8. The maximum velocity of the turbidity current was calculated to be 101 km/h.

Heezen and Ewing (1952) estimated that the velocity of the turbidity current at the first breaking point was about 101 km/h whereas Kuenen’s (1952) estimate gave an average velocity of 143 km/h. The initial slide was estimated to be 50 m thick, but the thickness of the turbidity current was estimated at 270 m, and the
average thickness of the deposited bed was 1 m. According to Kuenen, these values are reasonable and the hypothesis of turbidity current causing the breaks is sound.

Heezen and Ewing (1955) and Heezen (1963) provided some data on the break of submarine cables in the Mediterranean Sea following an earthquake in Orleansville, Algeria. The cable breaks occurred mainly in the Balearic abyssal plain.

Plapp and Mitchell (1960) developed a steady state theory for turbidity currents based on the concepts of boundary-layer flow. The driving force for the turbidity current is the excess density of the fluid in motion over that of the surroundings. In this situation the Reynolds Number will be such that the flow would be turbulent and this turbulence keeps the sediment particles suspended.

Plapp and Mitchell identified the following as the important parameters of the problem: ocean bottom slope, thickness, velocity, and density of the turbidity current.

Benjamin (1968) presented a theory of gravity currents under rather ideal conditions. He presented schematically a gravity current (also called a density current) consisting of a head wave followed by a turbulent zone where a heavy fluid flows along a horizontal bottom (Fig. 2.12A). As the density of the surrounding fluid is less than the density of the current, the piezometric pressure gradient provides the motive force.

Benjamin referred to the work of Prandtl (1952), who assumed that the velocity of the mean flow is greater than that of the front so that a portion of the stream is deflected up near the front. Benjamin also referred to the work of Von Kármán (1940) (see Fig. 2.12B) where the coordinate system is moving with the denser fluid (density, \( \rho_1 \)). Von Kármán used Bernouilli's theorem for steady irrotational flows to get the following results: (a) at the forward stagnation point, the angle between the bottom and the interface is 60°, and (b) the flow velocity is given by:

\[
c_1 = \sqrt{\frac{2gH(\rho_1 - \rho_2)}{\rho_2}}
\]  

(2.35)

where \( H \) is the asymptotic height of the interface above the bottom as shown in Fig. 2.12B.

Benjamin argued that although (2.35) is essentially correct, the arguments of Von Kármán leading to it are not, because energy conservation is not possible in this case. A more appropriate condition is the overall balance of momentum fluxes against the fluid forces.

Benjamin also made an analogy to the flow past a cavity shown in Fig. 2.12C. In this setup, a long rectangular box is filled with liquid, then one end is opened; as the liquid flows out the air cavity travels along the length of the box.

Benjamin gave the following expression for \( c_1 \), to compare the theory with Keulegan's (1958) experimental measurements on density currents involving the movement of saline water into fresh water:

\[
c_1 = C_1 \sqrt{\frac{gH\Delta\rho}{\rho_2}}
\]  

(2.36)

where \( \rho_2 \) is the density of the lighter fluid. Benjamin's theoretical estimate for \( C_1 \),
is 1.414 and Keulegan's value is 1.20. Benjamin attributed the influence of the upper boundary to this discrepancy. He also mentioned that Keulegan gave a universal value of 1.07 for \( C_1 \) applicable for Reynolds numbers larger than 500; the Reynolds Number being defined as:

\[
R_e = \frac{c_1 H}{\nu}
\]  

(2.37)

where \( \nu \) is the kinematic viscosity of the stream.

**WATER WAVES GENERATED BY LANDSLIDES**

There have been numerous instances of very large water waves generated by huge rock falls and landslides. On July 9, 1958, following an earthquake, a portion of a mountain fell into Lituya Bay, Alaska, causing an upsurge of water over 518 m (Miller 1960). Submarine landslides following the 1964 Alaskan earthquake caused
huge waves with amplitudes to 30.5 m in several places on the coast of Alaska (Kachadoorian 1965). On Oct. 9, 1963, a great rock slide fell 175 m into a reservoir in the Vaiont Valley in Italy and crested a wave that wiped out a town and killed 3000 people (Wiegel et al. 1970). In Norwegian fiords there have been numerous instances of landslides followed by huge water waves (see Miller 1960; Muller 1964).

Noda (1970) developed a theory for water waves generated by landslides. He made a distinction between vertical and horizontal landslides: the vertical landslide could be thought of as a two-dimensional box falling to the bottom at one end of a semi-infinite channel whereas the horizontal landslide could be modeled by making a two-dimensional wall move into the fluid.

Noda established the region of validity of his asymptotic solution by comparison with the solution from a direct numerical integration of the governing equations. He compared his theoretical results with results from the experiments of Wiegel et al. (1970). The theoretical results deviate more and more from experimental values with increasing B/D (where B is thickness of the slide and D is water depth). This means the linearization used by him is not satisfactory. Based on the analysis of the experimental data, Noda identified the different regions of wave characteristics as shown in Fig. 2.13.

Noda applied this theory to waves generated by the vertical landslide in Lituya Bay, Alaska, on July 9, 1958, and Miller (1960) gave the following data for this incident: thickness of slide = \( \lambda \approx 91.5 \) m, height of fall \( \approx 610 \) m, depth of water \( \approx 122 \) m, height of largest wave generated \( \approx 61 \) m, calculated velocity of free fall 109 m/s, and B/D equal to 0.75. The relevant value of V for this case is \( V \approx 3.16 \) and this puts the wave in the solitary wave region, D, shown in Fig. 2.13. The solution for this case gives a maximum wave height of 91.5 m.

If friction is included and the geometry of the problem might make \( V = 1.0 \), the same solution will be valid and \( \eta_{\text{max}} \) would be 76.9 m.

Noda essentially proceeded along analogous lines for the horizontal slide problem, and used the method of stationary phase to approximate the velocity time-history of the wall, by means of N straight lines. However, there were no experimental data in the linear range to compare with Noda’s theory, and in the nonlinear range he compared his theory with the experimental results of Miller and White (1966) discussed in Section 2.5.

2.4 Inverse Tsunami Problem

Reid and Knowles (1970) and Knowles and Reid (1970) are probably the first to use the terminology “inverse tsunami problem” to mean the determination of the deepwater signature of a tsunami based on records made at or near an island station. The significance of this terminology will be broadened to include determinations in the deeper water of such parameters as source region, tsunami signature, and tsunami energy based on recordings made near land, whether a mainland coast or an island.
FIG. 2.13. (A) Box-drop problem; regions of wave characteristics, (B) wave characteristics for box-drop problem. (Noda 1970)
Van Dorn (1970a) determined the source mechanism for the 1964 Alaskan earthquake tsunami based on the water-level record at Wake Island (Fig. 2.14). This diagram shows the first crest is the highest of the whole wave train. This observation led Van Dorn to assume that the displacement caused by the earthquake was a long-line source. The tsunami wave train generated was distorted subsequently by refraction.

Van Dorn theoretically (using Airy integral, see Section 1.2) determined the wave form for three different types of sources: (1) abrupt monopolar uplift, (2) dipolar motions with zero net displacement, and (3) impulsive sources initiated from rest. The wave form shown in Fig. 2.14 resembles more the wave form obtained from source one. However, at the outset this result appears to be in disagreement with other deductions. (Van Dorn (1964) suggested that dipolar movement occurred.) Van Dorn resolved this by arguing that “...Because most of the uplifted region was under water, while a large fraction of the depressed sector was on land, the net effect of the dipolar ground motion was to add water to the ocean. Thus the oceanic effect was essentially a modular uplift.”

Reid and Knowles (1970) determined the deepwater signature of a tsunami based on its record near an island and some of their assumptions are: the tsunami in deep water can be represented by a plane wave and the distance between the epicenter and the island is much larger than the dimension of the island. Processes such as scattering, diffraction, refraction, and resonance were ignored and the linear long-wave theory was used.

Let P be the location near an island where the tsunami is recorded. Reid and Knowles defined a transfer function as the ratio of the Fourier Transform of the response at P to the Fourier Transform of the input, the deepwater signature of the tsunami. If this transfer function (generally a complex function of frequency and direction of incoming waves) is determined, then from the mareograms the deepwater signature of the tsunami can be determined.

This transfer function, P, can be determined from either observational studies, laboratory experiments, or numerical models; the topography and other conditions duplicated as closely as possible in the latter two cases. Reid and Knowles used a simple geometry for the island so that analytical solutions of the problem could.
be obtained. Then they used a numerical model to incorporate several improvements such as broad-band input spectrum, real topography of the island, and different wave directions.

With reference to Fig. 2.15 let $R_m$ be the radius of the circle surrounding an island such that, for radii greater than $R_m$, the region is deep and has uniform depth. Let $X(t)$ and $Y(t)$ be the water-level records at some point slightly outside $R_m$ and at the point, $P$, near the island.

Assume that at the source, $Q$, the wave input is of plane progressive form and that the response at $P$ is linear, then $X$ and $Y$ can be expressed as convolutions of each other:

$$Y(t) = \int_{-\infty}^{\infty} K(\lambda) X(t-\lambda) d\lambda \quad (2.38)$$

and

$$X(t) = \int_{-\infty}^{\infty} G(\lambda) Y(t-\lambda) d\lambda \quad (2.39)$$

where $\lambda$ is the wavelength and $K(\lambda)$ and $G(\lambda)$ are the kernel functions that depend on the bathymetry of the island, the wave propagation direction, and the location of $P$.

Let $F_X(f)$ and $F_Y(f)$, where $f$ is the frequency, be the Fourier transforms of $X(t)$ and $Y(t)$ and let $R(f)$ be the Fourier Transform of $K(t)$. Then the Fourier Transform of $G(t)$ is $R(f)^{-1}$. By definition, then:

$$F_X(f) = \int_{-\infty}^{\infty} X(t) e^{i2\pi ft} dt$$

$$F_Y(f) = \int_{-\infty}^{\infty} Y(t) e^{i2\pi ft} dt \quad (2.40)$$

$$R(f) = \int_{-\infty}^{\infty} K(t) e^{i2\pi ft} dt$$
In general these will be complex. The Fourier Transform of (2.38) gives:

\[ F_j(f) = R(f) F_j(f) \]  \hspace{1cm} (2.41)

If \( X(t) \) and \( Y(t) \) are known then from Equation (2.41) the transfer function \( R(f) \) can be determined.

Reid and Knowles (1970) used only simple island geometry (a circular cylinder) so that analytical solutions for the transfer function were known. However, they claimed that the results were encouraging enough to warrant further work involving real topography.

Aida (1972) used the "impulse response method" and the "method of characteristics," under the one-dimensional propagation assumption, to obtain the offshore form of tsunami signature based on the tide-gage record at the head of a bay. At Enoshima, he obtained satisfactory comparisons between the theoretically determined signature and observed record for the 1963 Iturup earthquake tsunami, for the 1964 Alaskan earthquake tsunami, and the 1965 Aleutian earthquake tsunami.

The work of Ben-Menahem and Rosenman (1972) is a very significant contribution to the problem of the relation between the seismic and tsunamigenic source

![Calculated tsunami radiation patterns for 1964 Alaskan earthquake (Ben-Menahem and Rosenman 1972)](image)
parameters. The technique they developed is substantially different from the inverse refraction technique considered so far in this section. Indeed, their technique gave smaller tsunamigenic areas compared to those by other authors for the same earthquakes. These authors used the mantle Rayleigh and Love waves to deduce the source parameters for certain Kuril Island earthquakes and the Rat Island earthquake of Feb. 4, 1965. They derived an expression relating the tsunami amplitudes in deep water to the seismic source parameters and the topography of the propagation path. They also used the Kranzer-Keller solution.

Ben-Menahem and Rosenman used their theory to predict an amplitude ratio of 1:13 between the amplitude in the direction of the fault (i.e. $\phi = 0^\circ$), and the amplitude normal to the fault (i.e. $\phi = 90^\circ$). This deduction is confirmed by tide-gage records, e.g. the wave amplitudes at Crescent City and Avila Beach, Calif., and at Arica, Valparaiso, Talcahuano, and Corral, Chile; all those corresponding to $\phi = 90^\circ$ were about 15 times larger than those at Wake Island and Guam which corresponded to $\phi = 0^\circ$ (Fig. 2.16).

The same principle can be used to account for the high directivity of the Grand Banks earthquake tsunami of 1929, toward Newfoundland (in a northerly direction) as compared to the progression of the tsunami in a westerly direction (toward Nova Scotia).

Some important results of the paper by Ben-Menahem and Rosenman are: (1) the tsunamigenic area obtained was considerably smaller than those obtained by other authors; (2) the inverse refraction technique of determining the tsunamigenic area based on coastal run up was inaccurate; and (3) there need not be any simple relation between the tsunamigenic area and the area of the aftershocks as assumed by Hatori (1970), because the mechanisms were different for the main shock and aftershocks.

### 2.5 Laboratory Experiments

**Experiments with Generating Source at the Bottom**

These experiments are relevant for tsunamis generated by earthquakes. Takahasi (1934) appears to be one of the first to conduct laboratory experiments on tsunami generation and he was concerned with the form of free surface and the nature of wave motion due to the dislocation of a portion of the sea bottom. He also used the Law of Similitude to relate his laboratory experiments to the natural situations.

Assume that a circular area, $\pi r^2$, of the sea bottom is dislocated through a vertical extent, $S$, in a time interval, $T$, in a linear fashion and let $s$ denote the dislocation at time, $t$. Then

$$s = SF \left( \frac{t}{T} \right)$$

(2.42)

Let $D$ be the ocean depth at the position of dislocation and let $\lambda, \eta, \nu$ be the length, height, and velocity of the waves following the dislocation. Let $\mu$ be the dynamic viscosity of water, $\rho$ its density, $g$ its gravity, and $R$ the distance of the point of observation (where $\lambda, \eta, \nu$ are measured) from the origin. The wave motion
is determined by the following nondimensional parameters.

\[
\frac{\lambda}{D}, \frac{\eta}{D}, \frac{\nu}{\sqrt{gD}}, \frac{r}{D}, \frac{s}{D}, \frac{\sqrt{g}}{\sqrt{D}}, \frac{T}{D}, \frac{R}{D}, \frac{\sqrt{g}}{\sqrt{D}}, \frac{t}{\rho \frac{T}{D^2}} \tag{2.43}
\]

According to the Law of Similitude, these nondimensional parameters must have the same values in the model as well as in nature for the model to be truly representative of nature. An examination of (2.43) reveals the difficulties in satisfying the Law of Similitude. For example, if the space dimension is scaled down to \(1/\xi\), then the time scale in the model must be \(1/\sqrt{\xi}\) of that in nature. This requires the bottom dislocation to occur with a greater speed in the model than in nature.

In addition, if the viscous condition has to be satisfied (\(\frac{\mu}{\rho \frac{T}{D^2}}\) must be the same in the model and in nature), then a fluid with a small kinematic viscosity, \(\nu/\sigma\), must be used. Indeed, the kinematic viscosity of the fluid must be \(1/\xi^{3/2}\) that of sea water. Silicone oils probably will be suitable for this purpose, but when Takahasi performed his experiments in 1933, these oils were not commercially available and the only fluid he could use was mercury. However, use of mercury requires the ratio of the natural scale to the model scale 2.58, thus requiring an impossibly large model. For this reason, Takahasi abandoned the idea of satisfying the viscous condition.

In these experiments, Takahasi (1934) used a wooden tank 200 \(\times\) 150 \(\times\) 30 cm. To simulate upheaval of a portion of the sea bottom, a circular piston was fitted to the center of the tank bottom. A fine resin powder sprinkled on the water surface and photographed by an optical system provided a record of the wave forms. A total of 45 experiments were performed with different combinations of water depth, \(D\), length, \(S\), of the piston stroke, and velocity of the piston, with the following results: although a series of progressive waves was formed, the first appeared to be of maximum height. The front portion of the wave train (wave front) had a propagation velocity much greater than that of the first wave crest. Hence, the first wave gradually lengthened and the slope of the front part became less. Although a qualitatively similar change of form occurred with other parts of the wave train, dispersion was not as pronounced. The length of the wave train as a whole increased with travel distance. The first wave crest traveled with a velocity considerably greater than \(\sqrt{gD}\) in the immediate vicinity of the piston but approached \(\sqrt{gD}\) asymptotically with increasing distance from the piston. Above the piston, only stationary waves occurred and progressive waves started to form only at a distance of twice the piston's radius. The experiments also showed that heights of the circular progressive waves decreased as \(1/R^{0.6}\).

Measurements on the wavelength were less certain, but indications were that the wavelength was approximately constant, except when the piston moved very slowly. In this case, the wavelength was greater.

In 1957, Takahasi performed further experiments at the University of California, and reported the results in his 1963 paper. He stated that these experiments were conducted to clarify the following features of waves produced by movement of the bottom of the sea (Takahasi 1963, p. 235): (1) wave shapes and propagation
velocity of the frontal part of the wave train in the neighborhood of the origin, 
(2) attenuation law of wave heights in the neighborhood of the origin, (3) directionality in the energy radiation when the original area of dislocation is extended in one direction; for this purpose, two or more circular origins were lined up very close to each other, (4) phenomena observable in the neighborhood of the origin when the origin consists of two circular areas, and (5) coupling coefficient of the bottom movement and water movement.

These experiments were conducted in a basin 19.5 × 45.7 × 0.85 m. An iron box 2.4 × 1.2 × 0.2 m was used as the wave generator. It had a lid with 10 circular openings each 0.3 m in diameter, with the idea that, by a suitable combination closing and opening, various shapes of submarine dislocations could be achieved in the experiments.

Four types of experiments were performed: in the first set a hydraulic cylinder was used in the generating mechanism; in the second, one air operated piston was used; in the third and fourth sets, two and six pistons were used, respectively. While maintaining a constant stroke of 4.4 cm, 22 runs were made with one piston, 33 with two pistons, and 14 with six pistons. Each experimental run consisted of a series of upward and downward strokes, each producing a wave train.

Based on these experiments Takahasi arrived at the following conclusions:

a) One piston
   i) The front of the wave train is propagated with the velocity \( \sqrt{gD} \) except within one wavelength of the origin, where the velocity is much greater.
   ii) A wave train produced by the upward motion of the piston (designated upwave, hereafter) is quite different in shape from the downwave, when the water depth is small, but, as the depth becomes greater, both waves gradually become similar, except that the sense is reversed.
   iii) The first trough or first crest has a velocity very close to \( \sqrt{gD} \).
   iv) The wave form seems to be composed of a dispersive wave train superposed on a nondispersive one.
   v) The amplitude of the initial crest or trough seems to decrease roughly as \( r^{-\frac{3}{8}} \).

b) Two pistons
   Thirty-three runs were made with two pistons moving with the same sense and with opposing sense. E is the direction connecting the center of the pistons, and S the direction perpendicular to E. It seems that energy of the wave was emitted much more (16 times in the case of upwave and 4 times in the case of downwave) in the direction of S than in the direction of E. There seems to be a remarkable difference also in the period of the waves emitted into the two perpendicular directions.

c) Six pistons
   Pistons were lined up in the direction, E. In this case, a nearly plane wave is propagated toward direction, S. The decline of the crest or trough with the distance differed with the azimuth. It was \( r^{-\frac{3}{4}} \) toward S and \( r^{-\frac{5}{4}} \) toward E. Superposition of the elementary waves produced by an individual piston
does not seem to be equal to the wave height produced by all pistons moving simultaneously in water 10.2 cm deep.

Finally, Takahasi estimated the tsunami energy (in the model) based on the following considerations: the energy, $E$, of a circular wave train at distance, $r$, from the origin moving with a velocity, $V$, with amplitude, $A$, and period, $T$, is (see also Section 2.1):

$$E = \pi \rho g r \sum A^2 TV$$

(2.44)

where $g$ is gravity and $\rho$ is the density of sea water. Takahasi assumed that the amplitude and periods in a wave train varied uniformly from one wave to another and wrote:

$$A = A_0 a^n; \quad T = T_0 p^n; \quad V = V_0 c^n$$

(2.45)

From these two relations and taking:

$$a = \frac{1}{1.3} = 0.77 \quad V_0 = \sqrt{gD} \quad T = 0.8 \text{ s}$$

$$p = \frac{1}{1.1} = 0.91 \quad r = 1.83 \text{ m} \quad V_0 = 70 \text{ cm}$$

$$c = \frac{1}{1.05} = 0.95 \quad D = 5.1 \text{ cm} \quad A_0 = 0.27 \text{ cm}$$

he estimated $E = 4.5 \times 10^6$ erg.

The work done by the piston on the water is:

$$W_p = S_p h(\rho g D + KCV\rho)$$

(2.46)

where $D$ is the water depth, $S_p$ is the area of the piston, $V$ is the velocity with which the piston moves, $h$ is the piston stroke, $K$ is the coupling coefficient, and $C$ is the speed of sound in water. Part of the work done by the piston on water increases the potential energy of water, $S_p h \rho g (D - \frac{h}{2})$. The energy, $W_w$, that could be transformed into wave energy is:

$$W_w = Sh(\frac{1}{2} \rho gh + KCV\rho)$$

(2.47)

In the laboratory experiment, the coupling coefficient, $K$, is very small and the second term on the right side of (2.47) can be ignored. The reason $K$ is small is because the piston diameter is extremely small compared to the length of the sound wave emitted by the piston. Because of the large bottom displacements involved in actual tsunamis, sound waves might be transmitted significantly to cause sea shocks strong enough to rock nearby ships. However, this sound wave energy may not be efficiently transformed into tsunami energy.

Coming back to the laboratory situation, Takahasi estimated that $W_w = 4.6 \times 10^6$ ergs which agreed favorably with the wave energy estimated from the experiments. Based on this agreement, Takahasi concluded that the potential energy of sea water upheaved at the origin imparts energy to the tsunami waves.
For the Sanriku earthquake tsunami of 1933, wave energy was estimated at $1.6 \times 10^{23}$ ergs. The earthquake had a magnitude of 8.3 and an estimated energy of $2.8 \times 10^{24}$ ergs. This means about $\frac{1}{17.5}$ of the earthquake energy had been converted into tsunami energy (see also Section 2.1).

Improved technology has made better and more efficient tsunami generators for laboratory experiments possible since Takahasi's (1934) experiments (see Iwagaki et al. 1970). Hammack and Raichlen (1972) performed theoretical calculations as well as laboratory experiments on tsunami generation due to bottom motion and near field propagation. The experiments dealt with only two-dimensional situations whereas the theory included three-dimensional cases. The authors identified a parameter called a "time-size ratio" which is nondimensional and includes the duration of the bottom movement, the water depth, gravity, and the extent of the disturbance in the direction of wave propagation. Some features of the tsunami, such as maximum amplitude and duration of the leading wave, can be expressed as a function of this time-size ratio.

The time-size ratio has three regions: an impulsive region, where details of the bottom movement have no significant influence on the generated forms of the waves; a creeping region, which corresponds to very slow movement of the bottom when wave forms are strongly dependent on the details of the bottom motion; and an intermediate region between the two.

Hwang and Tuck (1970b) also performed laboratory experiments where a hinged plate capable of moving upward simulated ground motion. They recognized the comparison was only qualitative because of limitations of their theory as well as the experiments.

**Experiments of Garcia**

Reference has been made to the numerical model studies of Garcia (1972) for simulation of tsunamis caused by motion on the eastern end of the Mendocino Escarpment near California. He also performed four series of laboratory experiments and verified his numerical calculations.

The eastern end of the Mendocino Escarpment along the San Andreas Fault was modeled as the source and the experiments represented either displacement of the ocean floor or landslides. Two different sections along the eastern end of this escarpment, at 125°30′W and 128°W, were considered in detail.

In the experiments, the wall displacements were such that the boundary Froude Number $N_{F,avg} = \frac{V_{avg}}{\sqrt{gd}}$ had a moderate range (Fig. 2.17). Here $g$ is gravity, $d$ is the water depth, and $V_{avg}$ is the average speed with which the wall moves. The four series of experiments consisted of studies of waves produced by a moving vertical wall, moving slope walls with two different slopes of 1:1 and 1:2 vertical to horizontal, and a moving underwater step.

Figure 2.17 shows experimental results for the vertical wall. Here $\lambda$ is the maximum boundary displacement and $\eta$ is the height of the crest of the first wave. The diagram also includes results from the numerical calculations of Garcia (1972) and experimental results of Das and Wiegel (1972).
EXPLOSION-GENERATED WAVES

Prins (1956, 1958a, b) studied experimentally the characteristics of waves generated by a local disturbance at or near the water surface. Usually, experiments aimed at such problems consider simulating one or more of three initial states: (a) initial elevation or depression of the surface with zero initial velocity field, (b) undisturbed surface with an initial distribution of surface impulse, or (c) undisturbed surface with an initial distribution of a submerged impulse (underwater explosion). In his 1958b paper, he dealt with condition (a).

His experiments were performed in a flume 18.3 m × 0.3 m. At one end an elevation or depression in the water could be created without disturbing the rest of the liquid; at the opposite end was a wave absorber. Vertical motion of the water was monitored at five different locations along the tank length. The important parameters in these experiments were the undisturbed water depth, D, and the height, η, of the elevations or depression. Experiments were carried out with the following values for these parameters: D = 6.1, 10.7, 15.2, and 70.1 cm; η = ± 3.0, ± 6.1, and ± 9.1 cm; and L = 10, 30.5, and 61 cm.

Experimental results were compared with values calculated from the theories of Unoki and Nakano (1953a, b) and Kranzer and Keller (1955). Although Unoki
and Nakano's two-dimensional theory is for an initial elevation of a finite area in water of infinite depth, Kranzer and Keller's three-dimensional theory is applicable for an initial elevation or depression of a finite area in water of finite depth.

Prins (1956; 1958a, b) distinguished the types of leading parts of the generated wave train as (1) leading wave with oscillatory characteristics as part of the dispersive wave pattern, (2) leading wave with solitary wave characteristics with respect to its velocity of propagation, followed by a trough connecting it with the dispersive wave pattern, (3) leading wave a single wave with solitary wave characteristics separated from the dispersive wave pattern by a more or less flat part, and (4) leading part a complex form, which, while traveling outward, breaks up into a few waves with solitary wave characteristics separated from the dispersive wave pattern.

The experiments with \( \eta/D < 0.18 \) and \( L/D < 0.9 \) gave the following results: (1) generated wave pattern was of a dispersive character, (2) variation of height and extent of the initial disturbance did not affect the phase velocities, (3) phase periods agreed with the theory of Kranzer and Keller (1955) and for deep water with the theory of Unoki and Nakano (1953a, b), (4) wave patterns showed an interference phenomenon, (5) leading part of the wave pattern for an initial elevation showed exactly the negative performance of the waves generated by an initial depression \( (\eta_{\text{elev}} = -\eta_{\text{depr}}) \). Probably this can account for the initial withdrawal of water that has been reported associated with several tsunamis, (6) amplitudes of the leading waves were directly proportional to the height of the initial disturbance, (7) wave amplitudes did not show a satisfactory agreement with the theoretical values of Unoki and Nakano. In general, the measured values were found to be smaller than the theoretical values. This is not surprising because the theory of Unoki and Nakano is for deep water.

Miller (1970, 1971) performed experiments on impulsively generated waves in a wave tank of uniform width with a water layer of uniform depth, to simulate waves generated by a single impulse. The nondimensional variables have been identified by Miller as the relevant parameters of the problem: (1) ratio of the piston displacement to undisturbed water depth, \( \ell/D \), and (2) piston Froude Number \( F = V^2/D \) where \( V \) is the piston speed. Several experiments were run with various combinations of \( \ell/D \) and \( F \).

Miller identified four possible wave modes in these experiments: sinusoid, solitary, undular bore, and fully developed bore. His results showed that: (1) the fully developed bore decays rapidly to the unbroken undular bore form, (2) the lead wave of the undular form takes on the "solitary" mode leaving behind the rest of the "undular" wave. In several cases, the second and even the third undulations in turn take on the "solitary" mode, and (3) the "sinusoid" mode generated at short piston displacements transforms gradually as the trailing trough rises to the undisturbed water level, and finally enters the solitary mode.

Williams and Jordaan (1970) constructed a laboratory model to study so-called double-humped waves, the theory developed by Butler (1967). These double-humped waves or waves with two crests could be represented by:

\[
\eta = 1024 \varepsilon x^2 (a + 8x)^2 e^{-16px} 
\]

(2.48)
where the variables $x$ and $\eta$ denote the negative distance from the beach ($x=0$) and the free surface elevation, both in nondimensional form. The two parameters $a$ and $p$ determine the relative height and spacing of the two crests, and $\epsilon$ is a vertical scale magnification parameter which could depend on wave breaking.

The parameters $a$ and $p$ and the positions of the crests determined a characteristic length, $\ell_0$, used to nondimensionalize $x$, whereas $\alpha \ell_0$ (where $\alpha$ is the slope of the beach) has been used to nondimensionalize $\eta$. Butler calculated the elevation as a function of time at the beach for the double-humped waves, where their leading edges were at the beach and had zero particle velocity initially.

Williams and Jordaan performed laboratory experiments to check Butler's calculations. Their wave tank, made of lucite, was 7.6 m $\times$ 0.3 m with a maximum depth of 0.46 m and a beach slope of 1/7. A wave generator constructed from several strips of lucite was dropped vertically into the water and the wave profiles thus created were recorded by resistance probes. The surface tension effects near the beach were reduced by coating the beach with a mixture of water and detergent. The experimental results did not agree well with theoretically predicted wave forms: probably the theory is inadequate to represent double-humped waves satisfactorily. Further theoretical developments and experiments on double-humped waves could be useful to interpret coastal tsunami records.

LeMéhauté (1971) achieved some simplification of the solutions of Kajiura (1963) and Kranzer and Keller (1959) by identifying two cavity parameters, $\eta_{0,\text{max}}$ and $R$, which determine the wave envelope amplitude, $\mathcal{A}$. Correlations between his simplified theory and the results from field observations involving a 4363-kgm TNT explosion, showed that, except for irregularities caused by noise (wind waves), wave refraction, and interference patterns due to wave reflection from boundaries, good agreement is indeed possible. However, there are some limitations to the theoretical model. The bulk of data on which these correlations were based is from explosions of 0.23 to a few hundred kgm of TNT. It is not clear how reliable the extrapolation from this would be to large yields in the nuclear range (e.g. $10^{10}$ kgm TNT). In fact, LeMéhauté (1971, p. 18) cautioned that:

"...similarity is never achieved so that no simple scaling law can be expected to cover the entire range of yields of interest. In particular, the transition from chemical to nuclear sources with increasing yield may be a point of similarity violation; indeed, it would be surprising if such were not the case. Extrapolation then, to large yield from small chemical yield data must be accepted with a considerable measure of caution."

Van Dorn (1966, 1970a) performed laboratory experiments to simulate the tsunami generated by the Alaskan earthquake of Mar. 28, 1964.

The experiments were performed in a wave tank 27 m long. On one end of the tank was a shallow region 5 m long and 3.3 cm deep. Over the next 3-m length the depth increased from 3.3 cm to the full depth of 31.5 cm and in the remaining 19 m the depth was uniformly 31.5 cm. According to Van Dorn (1970a, p. 40) the relevance of this model is:

"On a scale of 1:17,000, this model represents a section of the continental shelf normal to the fault axis 90 km in half-breadth and 565 m deep, descending at the appropriate
slope to the uniform depth of 5400 meters corresponding to the effective depth computed for the travel path to Wake Island..."

The wave-generating mechanism consisted of a long plunger hinged at one end, which, when raised to its extreme position, just cleared the surface of the water, and when lowered dipped to a maximum depth of 1 cm. The 6.4-cm width of the plunger was relatively small compared to the 40-cm width of the tank. According to Van Dorn, by rapidly lowering the plunger a positive half-wave of 0.25 cm maximum amplitude and 5 m length could be produced. A displacement transducer was used to measure the motion of the plunger and pressure transducers were employed to measure wave heights at four different positions; \( T_1 \) at the shallow end of the tank, \( T_2 \) at 5 m where depth starts to increase, \( T_3 \) at 8 m where depth increased to the full value, and \( T_4 \) near the deepwater end of the tank.

Many experimental runs were made under different initial conditions and plunger speeds. Van Dorn (1970a) selected two particular runs for inclusion in his paper on the basis of close resemblance to the Wake Island record. Figure 2.18 shows the plunger (trace \( T_0 \)) started from its bottom center position and cycled approximately sinusoidally to its initial position during a time interval of about 8 s, which simulated an initial depression of the sea floor accompanied by an uplift of some magnitude.

Van Dorn (1970a, p. 42) interpreted Fig. 2.18 as:

"The other traces give the time-history of surface elevation at the transducer stations, including reflections from the far end of the channel. Trace \( T_2 \) shows that the tsunami at the edge of the continental shelf consisted of a leading trough, followed by a double pulse \( (A) \). The reflected train at the same station \( (B) \) exhibits a shallower trough and a dispersive train rather like the Wake record."

Experiments by Williams and Jordaan (1970) on the so-called double-humped waves have been described. These are somewhat similar to the double-pulse waves
FIG. 2.19. Model experiment recording for single rapid plunger depression corresponding to a simulated shelf uplift. (Van Dorn 1970a)

that appeared in the experiments of Van Dorn. He also estimated the half-breadth of the source as follows: in Fig. 2.19, let $t_1$ and $t_2$ be the arrival times of the first crest at the shelf edge, $(T_2)$, and at some other station, $T_4$ (say). The duration of the pulse at $t = t_1$ will be $2t_1$, because it consists of its direct and reflected wave from the end of the tank at station $T_1$. The pulse duration, $\Delta t$, at station $T_4$ can be represented by the following relation, using the long-wave approximation:

$$\Delta t = 2t_1 \left( \frac{t_2}{t_1} \right)^{\frac{1}{2}}$$

(2.49)

If $a$ is the half-breadth of the shelf between $T_1$ and $T_4$, then:

$$t_1 = \frac{a}{\sqrt{gD_1}}$$

(2.50)

From (2.49) and (2.50) then:

$$a = \left( \frac{\Delta t}{2} \right)^{\frac{1}{2}} \sqrt{\frac{gD_1}{t_2}}$$

(2.51)

Van Dorn used the following representative values based on Fig. 2.18: $\Delta t = 24.5$ s, $t_1 = 8.5$ s, $t_2 = 21.5$ s, and $D_1 = 3.3$ cm. He used these in (2.51) to obtain: $a = 525$ cm which agreed favorably with the value of 500 cm, the shelf width in the model. For the Wake record itself, he took: $\Delta t = 80$ min, $t_2 = 417$ min, and $D_1 = 100-200$ m, which gave $a = 23$ km for $D_1 = 100$ m, and 33 km for $D_1 = 200$ m. These values bracketed the width of 30 km between the 200-m contour and the assumed axis of the maximum uplift.
EXPERIMENTS ON UNDERWATER SLIDES, LANDSLIDES INTO WATER, AND TURBIDITY CURRENTS

Wiegel (1955) studied through laboratory experiments the system of gravity waves generated by the movement of a submerged body, to test whether a tsunami could be caused by a dipolar disturbance such as an underwater landslide (which is unlikely according to Shepard et al. (1950)). Wiegel remarked that his experiments applied equally well to the situation when the vertical movement of a block is compensated by equal movement in the opposite direction.

The earlier work of Sauer and Wiegel (unpublished data) consisted of piling coarse gravel over a sheet metal piece at one end of a laboratory channel and suddenly pulling the sheet. However, this did not create a slide as was planned but a slump occurred. Hence, in the later experiments they used a box sliding down an inclined plane to simulate an underwater slide.

Later Wiegel (1955) performed tests where a submerged body was made to fall vertically through water of different depths from several different heights. Also, experiments were rerun with different slopes of the inclined plane for the slide experiments. Most experiments were done in a wave tank 18.3 m × 0.3 m × 0.9 m, and water depth of 0.76 m. An inclined beach was arranged at the far end of the channel to minimize reflection but some reflection occurred, and was apparent after the second wave crest.

Results of these experiments can be summarized as follows: In general, within the range of experimental conditions, dispersive waves were generated by a body either falling vertically or sliding down an incline. An exception was a flat plate falling in quite shallow water (3.0-6.1 cm) when an Airy-type wave was generated, with dispersive waves forming the tail. Surface-time histories of the disturbance measured a short distance from the source showed that a crest always formed first, followed by a trough from one to three times the amplitude of the first crest (depending on the slope of the incline primarily), followed by a crest with about the same amplitude as the trough. Due to the dispersive qualities of the waves, additional crests and troughs continued to form with increasing distance from the origin, while at the same time the amplitudes of the initial crests and trough decreased.

Magnitudes of the amplitudes depended primarily on the submerged weight of the body, but also on the depth of submergence, the water depth, and other characteristics of the generation.

The “period” associated with gravity waves of the disturbance was found to be independent of water depth, initial depth of submergence, weight of body, or time of fall. It was, however, found to be related to length of the body, with the period increasing with increasing length, and to slope of the incline, with the smaller the incline, the greater the period.

Computations of the approximate energy of the wave disturbance generated by the vertically falling body indicated that about 1% of the initial potential energy (net, submerged) of the body was transformed into wave energy. Within the band of results it was seen that for water of constant depth, the less the initial depth of submergence of the body the greater the percentage of energy transformed into
wave energy; and for constant initial depth of submergence, the smaller the water depth the greater the percentage of energy transformed into wave energy.

A landslide in Lituya Bay, Alaska, that caused a giant wave has been discussed. The landslide caused the water on the opposite side of the bay to surge upward 366 m above its mean level, and a wave almost solitary in form traveled from the bay into the ocean. Wiegel (1963a) modeled this phenomenon in the laboratory and was successful to the extent of creating a solitary wave but it was followed by “a complex tail.” One interesting result of these experiments was that the crest moved with almost uniform speed across the bay even though it was much shallower near its entrance than at its center.

Earlier discussions described how turbidity currents generated by the Grand Banks earthquake of 1929 broke transatlantic cables and caused heavy damage. The work of Benjamin (1968) on turbidity currents has been briefly discussed. Some details of Middleton's (1966b) experiments will be described. According to Middleton, a turbidity current is a density current formed by sediment in suspended form. Middleton performed these experiments for a dual purpose: first, to infer the hydraulic behavior of turbidity currents; and second, to determine differences in the hydraulic behavior between turbidity currents of interest to geologists (in this case, the turbidity current carries coarse sediment in suspension) and turbidity currents of interest to engineers. By the latter, Middleton meant clay suspension turbidity currents and saltwater density currents. Harleman (1961) and Middleton (1966a) reviewed these types of currents.

Middleton (1966b) used a transparent plastic flume $500 \times 15.4 \times 50$ cm that could be tilted when necessary. In the first set of experiments Middleton created saline density currents by pumping salt solution into the tilted flume. In the second set, he simulated turbidity currents by suddenly releasing plastic beads at one end of the flume which was kept horizontal. The results of both experiments were somewhat similar in the sense that the saline density current and the turbidity current formed a characteristic head.

Middleton's experiments showed that for slopes less than 4%, the velocity of the head of the density current more or less agrees with (2.36).