7

Volcanic eruptions

INTRODUCTION

(Latter, 1981; Intergovernmental Oceanographic Commission, 1999)

Historically, volcanoes cause 4.6% of tsunami and 9.1% of the deaths attributable to this hazard, totaling 41,002 people. Two events caused this disproportionately large death toll: the Krakatau eruption (Figure 7.1) of August 26–27, 1883 (36,000 deaths) and the Unzen, Japanese eruption of May 21, 1792 (4,300 deaths). Tsunami account for 20%–25% of the deaths attributable to volcanic eruptions. The eruption of Santorini around 1470 BC is not included in these statistics because of a lack of written record. Santorini and the Krakatau eruption of 1883 will be discussed in more detail subsequently in this chapter. The main locations of the 65 tsunami linked to eruptions historically are plotted in Figure 7.2. The vast majority of these are restricted to the Japanese–Kuril Islands and the Philippine–Indonesian Archipelagos. Both of these regions form island arcs where one plate is being subducted beneath another. Explosive volcanism with caldera formation is a common occurrence in these regions. Other isolated cases of eruptions that have generated tsunami are associated with hot-spots beneath the Pacific Plate. Unfortunately, volcano-induced tsunami neither have been recorded well nor described except for a few events such as Krakatau in 1883.

CAUSES OF VOLCANO-INDUCED TSUNAMI

(Latter, 1981; Blong, 1984; Lockridge, 1988a, 1990)

There are ten mechanisms whereby volcanic eruptions can generate tsunami. These together with their major events are summarized in Table 7.1. Submarine landslides sloughed off from non-erupting volcanoes are not included in this table because they were dealt with in the preceding chapter. Many of the events listed in Table 7.1 were

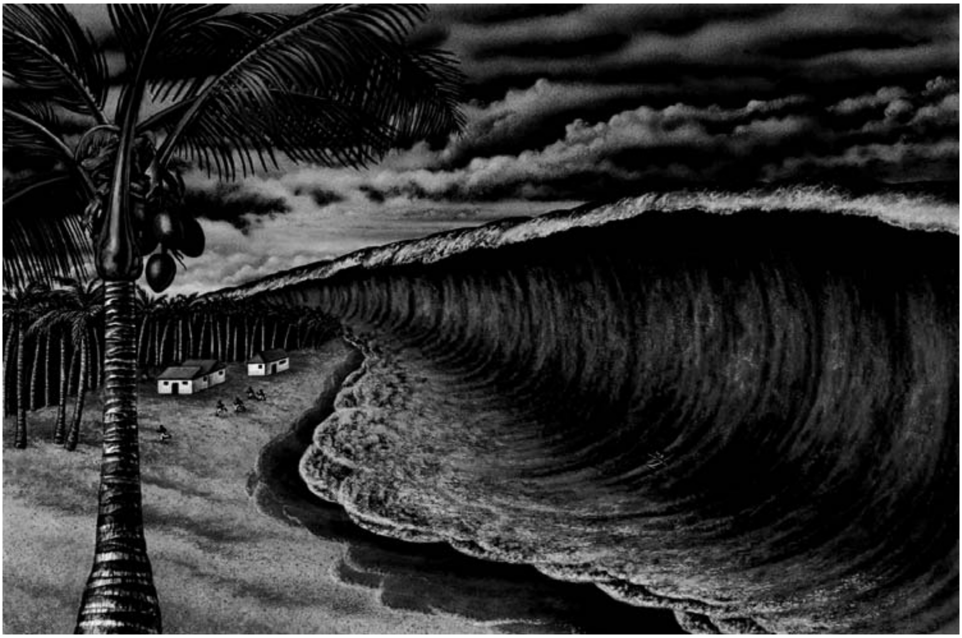

Figure 7.1. An artist's impression of the tsunami from the third explosion of Krakatau hitting the coast of Anjer Lor at about 10:30 AM on August 27, 1883. See color section. © Lynette Cook.

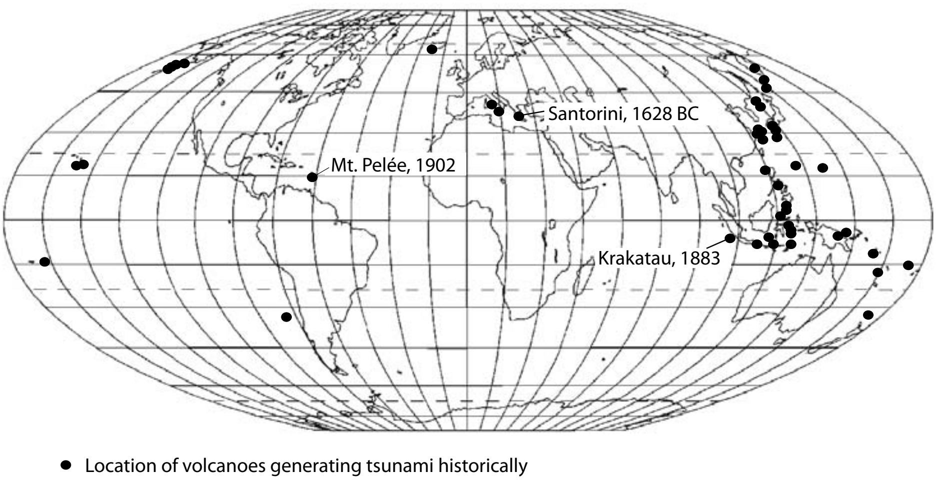

Figure 7.2. Location of volcanoes that have generated tsunami in recorded history. Based on Latter (1981), and Intergovernmental Oceanographic Commission (1999).

Table 7.1. Causes of historical tsunami induced by volcanoes.

Mechanism	*Percentage of events*	*Examples*	*Date*	*Height* (m)
Volcanic earthquakes	22.0	New Hebrides	January 10, 1878	17
Pyroclastic flows	20.0	Ruang, Indonesia Krakatau, Indonesia	March 5, 1871 August 26–27, 1883	25 >10
Submarine explosions	19.0	Krakatau, Indonesia Sakurajima, Japan	August 26–27, 1883 September 9, 1780	42 6
Caldera formation	9.0	Ritter Island Krakatau, Indonesia	March 13, 1888 August 26–27, 1883	12–15 2–10
Landslides	7.0	Unzen Volcano, Japan	May 21, 1792	6–9
Basal surges	7.0	Taal Volcano, Philippines	Numerous	?
Avalanches of hot rock	6.0	Stromboli, Italy	Numerous	?
Lahars	4.5	Mt. Pelée, Martinique	May 5, 1902	4.5
Atmospheric pressure wave	4.5	Krakatau, Indonesia	August 26–27, 1883	<0.5
Lava	1.0	Matavanu Volcano, Samoa	1906–1907	3.0–3.6

Source: Based on Latter (1981).

catastrophic. The majority of volcanic eruptions are accompanied by seismic tremors. If these are substantial enough and the volcano lies near or in the ocean, the tremors can trigger tsunami. For example, the eruption of Vesuvius in the southeast corner of the Bay of Naples on the west coast of Italy, in August AD 79, was preceded by a tsunami induced by seismic activity. Pliny the Elder, the commander of the Roman fleet at Misenum, sailed to the coast at the base of the mountain to rescue inhabitants five days before the final eruption. He could not get near the shore because of a sudden retreat of the shoreline. Two of the largest events due to seismic activity occurred on the January 10, 1878 and January 8, 1933 with the eruptions of Yasour Volcano in the New Hebrides and Severgin Volcano in the Kuril Islands, respectively. The respective tsunami reached 17 m and 9 m above sea level.

Pyroclastic flows, or *nuées ardentes*, are generated by the collapse under gravity of hot vertically ejected ash clouds. When these reach the ocean surface, they spread out rapidly as density flows that can either displace water or transfer energy to the ocean and generate a tsunami. The size of the resulting tsunami depends upon the density of the flow. If the density is less than seawater, then the ash cloud rides the surface of the ocean, generating a small wave. However, if the flow is denser than seawater, the cloud will sink to the bottom of the ocean and will displace water piston-like in front of it. In some cases, these flows can travel tens of kilometers along the seabed. Pyroclastic flows have the potential to generate devastating tsunami remote from the source of the eruption. For example, Tambora in 1815 generated a tsunami 2 m–4 m high in this manner despite lying 15 km inland. The May 7, 1902 eruption of Mt. Pelée, Martinique produced a *nuée ardente* that swept into the harbor of St. Pierre and generated a tsunami that traveled as far as Fort de France, 19 km away. Two of the largest of these types of events occurred in Indonesia during the eruption of Ruang on March 5, 1871 and Krakatau on August 26–27, 1883, producing tsunami 25 m and 10 m high, respectively.

Submarine eruptions within 500 m of the ocean surface can disturb the water column enough to generate a surface tsunami wave. Below this depth, the weight and volume of the water suppress surface wave formation. Tsunami from this cause rarely propagate more than 150 km from the site of the eruption. One of the largest such tsunami occurred during the eruption of Sakurajima, Japan, on September 9, 1780, when a 6 m high wave was generated. More significant are submarine explosions that occur when ocean water meets the magma chamber. This water is converted instantly to steam. In the process it produces a violent explosion. Krakatau during its third explosive eruption in 1883 produced a tsunami 40 m high in this manner.

Formation of a caldera during the final stages of an explosive eruption near the sea can permit water to flow rapidly into the depression. This sets up a wave train that can propagate away from the caldera within five minutes. Strato-volcanoes are particularly prone to collapse. In the *Ring-of-Fire* subduction zone around the Pacific Ocean there are hundreds of these types of volcanoes. Krakatau's numerous eruptions produced calderas that may have been responsible for some of the tsunami observed in the Sunda Strait. A comparable event occurred on Ritter Island, Papua New Guinea, on March 13, 1888. Here, the formation of a caldera 2.5 km in diameter produced a 12 m to 15 m high tsunami. While the initial wave heights radiating out from the caldera can be large, the actual volume of water displaced may be small. In addition, because the height of the wave decays inversely to the square root of the distance traveled, the effect of the tsunami diminishes rapidly away from the point source.

The slopes of a volcano are inherently unstable during eruptions because of earthquakes, inflation, or collapse. Collapsing material can form a debris avalanche that can travel at speeds of $100\,m\,s^{-1}$. Horseshoe-shaped scars are left behind as evidence of the failure. There have been 200 such avalanches over the past 2 million years. Some of these exceeded $20\,km^3$ in volume and traveled 50 km–100 km from their source. For small avalanches of $0.1\,km^3$–$1.0\,km^3$in volume, the travel distance ranges between six and eleven times the elevation of the initial avalanche. For higher volumes, the travel distance can increase to 8 to 20 times the vertical drop. For example, the collapse of a 2,000 m high volcano could generate a debris avalanche that can travel 16 km–22 km from its source. Avalanching is a significant hazard associated with volcanoes in Alaska, Kamchatka, Japan, the Philippines, Indonesia, Papua New Guinea, the West Indies, and the Mediterranean Sea. In most cases, the resulting landslide is localized enough to generate small tsunami that are highly directional in their propagation away from the volcano. However, some of the greatest death tolls have been caused by such events. For example, during the eruption on May 21, 1792 of Unzen Volcano, in Japan, 0.34 km^3 of material sloughed off its flank. The landslide traveled 6.5 km before sweeping into the Ariake Sea, where it generated a tsunami with run-ups 35 m–55 m above sea level along 77 km of coastline on the Shimabara Peninsula and along the opposite side of the sea, 15 km away. Six thousand houses were destroyed, 1,650 ships were sunk, and 14,524 people lost their lives. Similar-magnitude waves were generated by landslides on Paluweh Island, Indonesia, on August 4, 1928 (160 dead), and recently on Ili Werung Volcano, Indonesia on July 18, 1979 (500 dead). Mt. St. Augustine at the entrance to Cook Inlet, Alaska, also has the potential to generate 5 m to 7 m high tsunami through sloughing. Eleven major debris avalanches, occurring at 150 yr to 200 yr intervals, have originated from this volcano. One of the largest occurred on October 6, 1883. The avalanche swept 4 km–8 km into Cook Inlet on the north side of the volcano. Within half an hour, a 9 m high tsunami flooded settlements at English Bay, 85 km up the Inlet.

Basal surges or lateral blasts are formed when a volcano erupts sideways. Taal Volcano in Lake Bombon, Philippines, has been subject to at least five basal surge events since 1749. All of the resulting tsunami took lives. Avalanches of hot rock can generate tsunami when ejecta, piled up on the side of an erupting volcano, collapses into the sea. These events occur frequently on Stromboli volcano in Italy. The resulting tsunami are usually localized and have not been associated with any deaths. Lahars are ash deposits that fail after becoming saturated with water. Failures occur during an eruption because of the displacement of ground water, mixing of ash with water in a crater lake, or melting of snow and ice at the crest of the volcano. When the lahar reaches the ocean, it can generate significant tsunami; however, these are usually localized because the lahar tends to flow down valleys. For example, on May 5, 1902, a 35 m high lahar from Mt. Pelée swept down Rivière Blanche north of the nearby town of St. Pierre. When it reached the sea, it generated a 4.5 m high tsunami that only affected the lower part of the town, killing 100 people. Large explosive volcanoes generate a pressure pulse through the atmosphere. Krakatau, in 1883, generated tsunami in the Pacific Ocean and in Lake Taupo in the middle of the North Island of New Zealand, via this mechanism. However, nowhere did the tsunami exceed more than 0.5 m in height. The 1955–1956 eruption of Bezymianny on the Kamchatka Peninsula in Russia also generated a global pressure wave, but the resulting tsunami in the Pacific Ocean did not reach more than 0.3 m in height. Finally, if lava reaches the ocean *en masse* it can generate tsunami. These events are rare and have only been noted at one or two locations, with no widespread destruction being produced. For example, the lava from the eruption of Matavanu Volcano on Savaii, Samoa in 1907 generated a tsunami 3.0 m–3.6 m high when it poured into the sea. No deaths were reported.

KRAKATAU, AUGUST 26–27, 1883

(Verbeek, 1884; Latter, 1981; Self and Rampino, 1981; Blong, 1984; Myles, 1985; Nomanbhoy and Satake, 1995)

Krakatau was one of the largest explosive eruptions known to humanity. It is the only eruption for which detailed information exists on volcano-induced tsunami. To date, the mechanisms generating tsunami during its eruption are still debated. The volcano lies in the Sunda Strait between Sumatra and Java, Indonesia (Figure 7.3). The Javanese *Book of Kings* describes an earlier eruption, referring to Krakatau as Mount Kapi. Then, the volcano exploded and created a sea wave that inundated the land and killed many people throughout the northern part of Sunda Strait. Krakatau had last been active in 1681, and during the 1870s the volcano underwent increased earthquake activity. In May 1883, one vent became active, throwing ash 10 km into the air. By the beginning of August, a dozen Vesuvian-type eruptions had occurred across the island. On August 26, loud explosions recurred at intervals of 10 minutes, and a dense tephra cloud rose 25 km above the island. The explosions could be heard throughout the islands of Java and Sumatra. In the morning and later that evening, small tsunami waves 1 m–2 m in height swept the strait, striking the towns of Telok Betong on Sumatra's Lampong Bay, Tjaringin on the Java coast north of Pepper Bay, and Merak (Figure 7.3). On the morning of August 27, three horrific explosions occurred. The first explosion at 5:28 AM destroyed the 130 m peak of Perboewatan, forming a

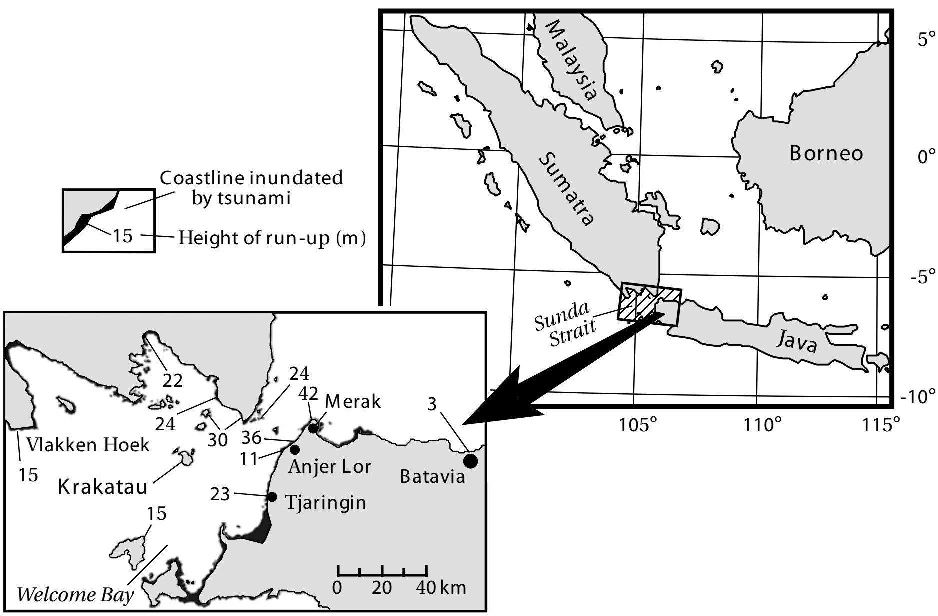

Figure 7.3. Coastline in the Sunda Strait affected by tsunami following the eruption of Krakatau August 26–27, 1883. Based on Verbeek (1884), Blong (1984), and Myles (1985).

caldera that immediately infilled with seawater and generated a tsunami. At 6:36 AM, the 500 m high peak of Danan exploded and collapsed, sending more seawater into the molten magma chamber of the eruption and producing another tsunami. The third blast, at 9:58 AM, tore the remaining island of Rakata apart. Including ejecta, 9 km^3–10 km^3 of solid rock was blown out of the volcano. About 18 km^3–21 km^3 of pyroclastic deposits spread out over 300 km^2 to an average depth of 40 meters. Fine ash spread over an area of 2.8×106 km^2, and thick pumice rafts impeded navigation in the region up to five months afterwards. A caldera 6 km in diameter and 270 m deep formed where the central island had once stood. This third blast was the largest sound ever heard by humanity, and was recorded 4,800 km away on the island of Rodriguez in the Indian Ocean and 3,200 km away at Elsey Creek, Northern Territory, Australia. Windows 150 km away were shattered. The atmospheric shock wave traveled around the world seven times. Barometers in Europe and the United States measured significant oscillations in pressure over nine days following the blast. The total energy released by the third eruption was equivalent to 200 megatons of TNT. (Kinetic energy for volcanic eruptions and asteroids exploding in the atmosphere or impacting with the ocean is expressed in megatons of TNT. One megaton of TNT is equivalent to 4.185×10^{15} joules.)

The two pre-dawn blasts each generated tsunami that drowned thousands in the Sunda Strait. The third blast-induced wave was cataclysmic and devastated the adjacent coastline of Java and Sumatra within 30 to 60 minutes (Figure 7.3). The coastline north of the eruption was struck by waves with a maximum run-up height of 42 m. The tsunami penetrated 5 km inland over low-lying areas. The largest wave struck the town of Merak. Here, the 15 m high tsunami was increased to 40 m because of the funnel-shaped nature of the bay. The town of Anjer Lor was swamped by an 11 m high wave (Figure 7.1), the town of Tjaringin by one 23 m in height, and the towns of Kelimbang and Telok Betong were each struck by a wave 22 m–24 m high. In the latter town, the Dutch warship *Berouw* was carried 2 km inland and left stranded 10 m above sea level. Coral blocks weighing up to 600 tonnes were moved onshore (Figure 7.4). Within the Strait, 11 waves rolled in over the next 15 hours, while at Batavia (now Jakarta) 14 consistently spaced waves arrived over a period of 36 hours. Between 5,000 and 6,000 boats in the strait were sunk. In total, 36,417 people died in major towns and 300 villages were destroyed because of the tsunami.

Within four hours of the final eruption, a 4 m high tsunami arrived at Northwest Cape, West Australia 2,100 km away. The wave swept through gaps in the Ningaloo Reef and penetrated 1 km inland over sand dunes. Nine hours after the blast, 300 riverboats were swamped and sunk at Kolkata (formerly Calcutta) on the Ganges River 3,800 km away. The wave was measured around the Indian Ocean at Aden on the tip of the Arabian Peninsula, Sri Lanka, Mahe in the Seychelle Islands, and on the Island of Mauritius. The farthest this tsunami wave was observed was 8,300 km away at Port Elizabeth, South Africa. Tsunami waves were measured over the next 37 hours on tide gauges in the English Channel, in the Pacific Ocean, and in Lake Taupo in the center of the North Island of New Zealand, where a 0.5 m oscillation in lake level was observed. Around the Pacific Ocean, tide gauges in Australia and Japan and at San Francisco and Kodiak Island measured changes of 0.1 m up to 20 hours after

Figure 7.4. Coral boulder moved by the tsunami generated by the eruption of Krakatau at Anyer, Indonesia. *Source:* Dr. Laura Kong, Director of the ITIC (Honolulu).

the eruption. Honolulu recorded higher oscillations of 0.24 m with a periodicity of 30 minutes. Smaller, subsequent eruptions of Krakatau generated lesser tsunami throughout the strait until October 10. The last tsunami was observed in Welcome Bay, where it surged 75 m inland beyond the high-tide mark.

The tsunami in the Pacific has been attributed to the atmospheric pressure wave because many islands that would have effectively dissipated long-wave energy obscured the passage from the Sunda Strait into this ocean. The atmospheric pressure wave also accounts for seiching that occurred in Lake Taupo, which is not connected to the ocean. Finally, it explains the long waves observed along the coast of France and England when the main tsunami had effectively dissipated its energy in the Indian Ocean. The generation of tsunami in Sunda Strait and the Indian Ocean has been attributed to four causes: lateral blast, collapse of the caldera that formed on the north side of Krakatau Island, pyroclastic flows, and a submarine explosion. Lateral blasting may have occurred to a small degree on Krakatau during the third explosion; however, its effect on tsunami generation is not known. During the third explosion Krakatau collapsed in on itself, forming a caldera about 270 m deep and with a volume of 11.5 km^3. However, modeling indicates that this mechanism underestimates tsunami wave heights by a factor of 3 within Sunda Strait. Krakatau generated massive pyroclastic flows. These flows probably generated the tsunami that preceded the final explosion. At the time of the third eruption, ash was ejected into the atmosphere toward the northeast. Theoretically, a pyroclastic flow in this direction could have generated tsunami up to 10 m in size throughout the strait; however, the mechanism does not account for measured tsunami run-ups of more than 15 m in height in the northern part of Sunda Strait. The pyroclastic flow now appears to have sunk to the bottom of the ocean and traveled 10 km–15 km along the seabed before depositing two large islands of ash. The 40 m high run-up measured near Merak to the northeast supports this hypothesis. The tsunami's wave height corresponds with the depth of water around Krakatau in this direction. As well, the third explosion of Krakatau at 9:58 AM more than likely produced a submarine explosion as ocean water encountered the magma chamber. Van Guest's description of the eruption (presented in Chapter 1) indicates that the magma chamber was visible in the strait before the third explosion at 10 o'clock in the morning. A submarine explosion could have generated tsunami 15 m high throughout the Strait. If the explosion had a lateral component northward, as indicated by the final configuration of Krakatau Island, then this blast, in conjunction with the pyroclastic flow, would account for the increase in tsunami wave heights toward the northern entrance of Sunda Strait (Figure 7.3).

SANTORINI, AROUND 1470 BC

(Yokoyama, 1978; Pichler and Friedrich, 1980; Kastens and Cita, 1981; LaMoreaux, 1995; Cita, Camerlenghi, and Rimoldi, 1996; Johnstone, 1997; Pararas-Carayannis, 1998c; Bryant, 2005)

The prehistoric eruption of Santorini around 1470 BC, off the island of Thera in the southern Aegean Sea north of Crete (Figure 7.5), is probably the biggest volcanic explosion witnessed by humans. It is also one of the most controversial because legend, myth, and archaeological fact frequently are intertwined and distorted in the interpretation of the sequence of events. The eruption has been linked to the lost city of Atlantis described by Plato in his *Critias*, to the destruction of Minoan civilization on the island of Crete 120 km to the south (Figure 7.4), and to the exodus of the Israelites from Egypt in the Bible. Certainly, Greek flood myths refer to this or similar events that generated tsunami in the Aegean. Plato's story of Atlantis is based on an Egyptian story that has similarities with Carthaginian and Phoenician legends.

The great Minoan empire was a Bronze Age maritime civilization centered on the island of Crete that flourished from 3000 BC to 1400 BC. The Minoan seafarers dominated trade in the eastern Mediterranean, and on this basis were able to accumulate great wealth and prevent the development of any other maritime power that could threaten them. The Minoans were noted for their cities and great palaces at Ayía, Knossos, Mallia, Phaestus, Triáda, and Tylissos—all decorated by detailed and lively frescoes. By far the largest and best-known palace was that of the legendary King Minos at Knossos, rivaling any other Middle Eastern structure in size. The eruption of Santorini, also known as Stronghyli—the round island—did not destroy Minoan civilization, but it certainly weakened it. The tsunami from the eruption is believed to have sunk most ships near the coast and in harbors, and to

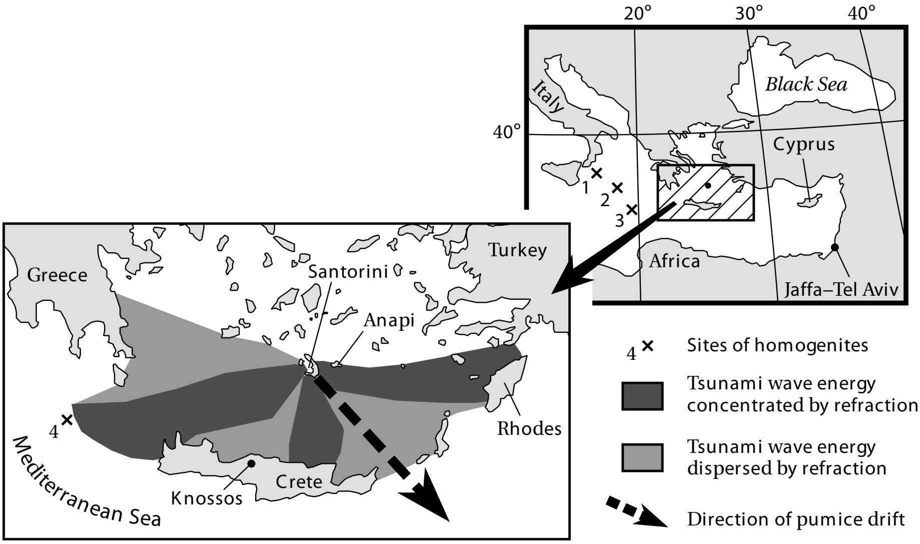

Figure 7.5. Eastern Mediterranean region affected by the Santorini eruption around 1470 BC. The numbers refer to sites where homogenites exist. (1) Calabrian Ridge, (2) Ionian Abyssal Plain, (3) Bannock Basin, and (4) Mediterranean Ridge. Refraction patterns based on Kastens and Cita (1981).

have greatly disrupted sea trade that was pivotal to the stability of the civilization. Ash falls also disrupted agriculture. Within 50 years of the eruption of Santorini, the Mycenaean Greeks, who had escaped its effects, were able to conquer the Minoans and take over their cities and palaces.

The timing of the Santorini eruption has also been linked to the plagues of Egypt (Exodus 6:28–14:31) and the exodus of the Israelites from that country. In 1 Kings 6:1 the exodus is dated as occurring 476 years before the rule of Solomon. Scholars believe that Solomon began his rule in 960 BC, putting the Exodus around 1436 BC. Other evidence indicates that the exodus occurred in 1477 BC. Both dates encompass the reign in Egypt of either Hatshepsut or her son Tuthmosis III of the 18th dynasty. The transition of rule between the two rulers is known as being a time of catastrophes. In the biblical account, the river of blood may refer to pink pumice from the Santorini eruption preceding the explosion. This pumice, after it was deposited, would easily have mixed with rainwater and flowed into any stream or river, coloring it red. The three days of darkness possibly refer to tephra clouds blowing south across Egypt at the beginning of the eruption. The darkness was described as a "darkness, which could be felt." Egyptian documents around 1470 BC refer to a time of prolonged darkness and noise, to a period of nine days that "were in violence and tempest: none . . . could see the face of his fellow," and to the destruction of towns and wasting of Upper Egypt. There is also direct reference to the collapse of trade with Crete (Keftiu). Volcanic shards have been found in soils on the Nile Delta with the same chemical composition as tephra on the Santorini Islands. The parting of the Red Sea most likely occurred in the marshes at the northern end of the sea. The Bible attributes the parting to wind (Exodus 14:21). The wind may refer to the atmospheric pressure wave produced by the explosion of the volcano. Such waves, akin to that generated by Krakatau in 1883, can generate seiching or tsunami in enclosed basins or distant oceans.

Santorini volcano is part of a volcanic island chain extending parallel to the coast of Asia Minor (Figure 7.4). The Aegean is the only zone in the eastern Mediterranean where subduction of plate boundaries is active. Of all these volcanic islands, only two, Santorini and Nisyros, have erupted in recorded times. Santorini forms a complex of overlapping shield volcanoes consisting of basaltic and andesitic lava flows. The volcano has erupted explosively at least 12 times during the last 200,000 years. Its height has been reduced over this time from a single mountain 1,500 m high to three islands less than 500 m in height surrounding a submerged caldera. The eruption around 1470 BC was the most recent of these and was one of the largest eruptions on Earth in the last 10,000 years. The timing is debatable. Acidity in Greenland ice cores suggest that the major eruption occurred in 1390 ± 50 BC, although radiocarbon dating on land suggests an age around 1450 BC or 1470 BC. Dendrochronology based on Irish bog oaks and Californian bristlecone pine puts the age of the event as old as 1628 BC. At this latter time, Chinese records report a dim Sun and failure of cereal crops because of frost. Large volcanic events cool temperatures globally by as much as 1°C over the space of several years. The range of dates may not be contradictory because there is evidence that Thera may have erupted several times over a timespan of 200 years.

The eruption around 1470 BC had four distinct phases. The first was a Plinian phase with massive pumice falls. This was followed by a series of basal surges producing profuse quantities of pumice up to 30 m thick on Santorini. The third phase was associated with the collapse of the caldera and production of pyroclastic flows. About 4.5 km^3 of dense magma was ejected from the volcano, producing 10 km^3 of ash. The volume of ejecta is similar in magnitude to that produced by the Krakatau eruption in 1883. The ash drifted to the east–southeast, but did not exceed 5 mm thickness in deposits on any of the adjacent islands, including Crete. The largest thickness of ash measured in marine cores appears to originate from pumice that floated into the eastern Mediterranean. It is possible at this stage that ocean water made contact with the magma chamber and produced large explosions, which generated tsunami in the same way that the eruption of Krakatau did. The final phase of the eruption was associated with the collapse of the caldera in its southwest corner. The volcano sank over an area of 83 km^2 and to a depth of between 600 m and 800 m. According to the Krakatau model, this final event produced the largest tsunami, directing most of its energy westward (Figure 7.4). It is estimated that the original height of the tsunami was 46 m–68 m, and maybe as high as 90 m. The average period between the dozen or more peaks in the wave train was 15 minutes.

Evidence of the tsunami is found in deposits close to Santorini. On the island of Anapi to the east, sea-borne pumice was deposited to an altitude of 40 m–50 m above present sea level. Considering that sea levels at the time of the eruption may have been 10 m lower, this represents run-up heights greater than those produced by Krakatau in the Sunda Strait. On the island of Crete, the wave arrived within 30 minutes with a height of approximately 11 m. Refraction focused wave energy on the northeast corner of Crete, where run-up heights reached 40 m above sea level. In the region of Knossos, the tsunami swept across a 3 km wide coastal plain reaching the mountains behind. The backwash concentrated in valleys and watercourses, and was highly erosive. Evidence for the tsunami is also found in the eastern Mediterranean on the western side of Cyprus, and farther away at Jaffa–Tel Aviv in Israel. At the latter location, pumice has been found on a terrace lying 7 m above sea level at the time of the eruption. However, the tsunami wave here had already undergone substantial defocusing because of wave refraction as it passed between the islands of Crete and Rhodes. The greatest tsunami wave heights occurred west of Santorini. Based upon linear wave theory, the wave in the central Mediterranean Sea was 17 m high. Closer to Italy over the submarine Calabrian Ridge, it was 7 m high. Bottom current velocities under the wave crest in these regions ranged between 20 cm s^{-1} and 50 cm s^{-1}—great enough to entrain clay-to-gravel-sized particles. The maximum pressure pulse produced on the seabed by the passage of the wave ranged between 350 kdyne cm^{-2} and 850 kdyne cm^{-2}. Spontaneous liquefaction and flow of water-saturated muds is known to occur under pressure pulses of 280 kdyne cm^{-2} and greater.

Some of the evidence for a large tsunami comes from the discovery of unusual deposits on the seabed of the central Mediterranean Sea, where wave heights were highest. These deposits—labeled homogenites—formed in the deep sea as the result of settling from suspension of densely concentrated, fine-grained sediment. This process produced homogeneous units up to 25 m thick with a sharp basal contact. Homogenites can be linked hydrodynamically to the passage of a tsunami wave. As sediment fails via liquefaction due to the pressure pulse, oscillatory flow under the wave suspends finer particles, creating turbulent clouds of sediment. It is estimated that the slurries exceeded concentrations of 16,000 mg L^{-1}. In comparison, the highest measured sediment concentrations on the ocean seabed and in muddy tidal estuaries rarely exceed 12 mg L^{-1} and 300 mg L^{-1}, respectively. Gravity-sorting occurred under this extreme concentration. Sand-sized particles settled first to the bottom and were deposited at the erosional contact with the seabed as a fining-upward unit, whose thickness ranged from a few centimeters to several meters. Finer clay-sized sediment was deposited over the next few days as a massive undifferentiated clay deposit that was up to 20 m or more thick. Homogenites differ from turbidites described in Chapter 3 by their greater thickness, lack of laminations, and undifferentiated particle size. Homogenites differ from debris flows by the absence of large clasts or rock pieces derived from continental sediments.

Four types of homogenites can be differentiated. In the western Mediterranean, on the Ionian abyssal plain, a 10 m to 20 m thick deposit, with an estimated volume of 11 km^3, was laid down on the seabed over an area of 1,100 km^2. It appears that the tsunami wave slammed into the continental shelf of North Africa and either directly or indirectly triggered a mega-turbidity current. This current carried terrigenous and shelf sediment into the deep Mediterranean Sea, eroding flanks of undersea ridges and depositing homogenites with an erosional base on upslopes. In one location, this turbidity current rode up a ridge 223 m above the abyssal plain and deposited sediment. In the eastern part of the Mediterranean, bottom velocities and the related powerful pressure pulse liquefied sand into depressions, forming uniform deposits several meters thick with a sandy base overlying an erosional contact. These deposits form in cobblestone-shaped basins with a vertical relief of 200 m. Finally, in the Bannock Basin, the passage of the wave destabilized evaporites. The resulting deposits are 12 m thick and consist of 3 m of sand overlain by 9 m of graded mud deposited from suspension in high-density brines trapped at the bottom of 100 m deep depressions in the seabed. All of the homogenites found in the Mediterranean are derived from a single event and date around the time of the Santorini eruption. Homogenites are not found in the eastern Mediterranean Sea, where tsunami wave heights were insufficient to cause resuspension or liquefaction of bottom sediment.

Since the Santorini eruption around 1470 BC, there have been many others. Since 197 BC at least 11 eruptions have formed the two islands that presently exist in the center of the caldera. Eruptions in 1650, 1866, and 1956 have given rise to tsunami with damaging consequences. An earthquake preceded the 1650 eruption and generated a 50 m high tsunami that swept 4 km inland in places. The 1866 event generated two tsunami that had run-up heights of 8 m along nearby coasts. Earthquakes associated with the latest eruption on July 9, 1956 produced a tsunami that had a run-up height of 24 m and killed 53 people. The Santorini volcano remains one of the most dangerous in terms of tsunami in the world today.

The last three chapters have summarized how geophysical processes originating from the Earth generate tsunami. When reviewed, the magnitudes of tsunami associated with these processes are indeed impressive. Tsunami have dispersed across the Pacific after numerous historical earthquakes. Five events since 1600 have produced run-up heights of 51 m–115 m in elevation. The Indian Ocean event of 2004 and the Lisbon event of 1755 indicate that catastrophic tsunami can occur in any ocean. Volcanic eruptions, while rarer in terms of tsunami, have generated similar-magnitude run-ups, but these have been localized. The Santorini eruption of around 1470 BC may hold the record for the biggest volcano-induced tsunami with an initial wave height of 90 m. Tsunami generated by submarine landslides may be bigger yet. The Lituya Bay landslide of July 9, 1958 generated a wave that achieved a run-up height of 524 m above sea level. Whether or not this wave consisted of a mixture of water and air is a moot point. In terms of area affected, the Storegga slide of 7,950 ± 190 years ago may have been the biggest—considering that some suspicion hangs over some of the evidence attributed to the Lanai slide in Hawaii. Many of the tsunami induced by these processes produced some of the signatures of tsunami outlined in Chapters 3 and 4. However, only the Storegga event can be linked to the full range of signatures that includes bedrock-sculpturing features. A dichotomy thus exists in that observable tsunami have not commonly generated bedrock-sculpturing features that exist so widely along rocky coasts, especially those in Australia. One mechanism, asteroid/comet impact with the ocean, is capable of generating tsunami equivalent to or bigger than the largest tsunami produced by other mechanisms. The nature of cosmogenically induced tsunami will be discussed in the next chapter.