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History of Earthquakes and Tsunamis Along the Eastern Aleutian-Alaska Megathrust, with Implications for Tsunami Hazards in the California Continental Borderland

Professional Paper 1795–A

U.S. Department of the Interior
U.S. Geological Survey

COVER
View of Southwest Anchorage beach on Chirikof Island, Alaska, where the U.S. Geological Survey (USGS) conducted paleotsunami and paleoseismology studies in 2010. The USGS base camp is visible on the far shore of the lake in the middle distance. USGS photo by Alan Nelson.

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Abstract

During the past several years, devastating tsunamis were generated along subduction zones in Indonesia, Chile, and most recently Japan. Both the Chile and Japan tsunamis traveled across the Pacific Ocean and caused localized damage at several coastal areas in California. The question remains as to whether coastal California, in particular the California Continental Borderland, is vulnerable to more extensive damage from a far-field tsunami sourced along a Pacific subduction zone. Assuming that the coast of California is at risk from a far-field tsunami, its coastline is most exposed to a trans-Pacific tsunami generated along the eastern Aleutian-Alaska subduction zone. We present the background geologic constraints that could control a possible giant (M_w ~9) earthquake sourced along the eastern Aleutian-Alaska megathrust. Previous great earthquakes (M_w ~8) in 1788, 1938, and 1946 ruptured single segments of the eastern Aleutian-Alaska megathrust. However, in order to generate a giant earthquake, it is necessary to rupture through multiple segments of the megathrust. Potential barriers to a through-going rupture, such as high-relief fracture zones or ridges, are absent on the subducting Pacific Plate between the Fox and Semidi Islands. Possible asperities (areas on the megathrust that are locked and therefore subject to infrequent but large slip) are identified by patches of high moment release observed in the historical earthquake record, geodetic studies, and the location of forearc basin gravity lows. Global Positioning System (GPS) data indicate that some areas of the eastern Aleutian-Alaska megathrust, such as that beneath Sanak Island, are weakly coupled. We suggest that although these areas will have reduced slip during a giant earthquake, they are not really large enough to form a barrier to rupture. A key aspect in defining an earthquake source for tsunami generation is determining the possibility of significant slip on the updip end of the megathrust near the trench. Large slip on the updip part of the eastern Aleutian-Alaska megathrust is a viable possibility owing to the small frontal accretionary prism and the presence of arc basement relatively close to the trench along most of the megathrust.

Introduction

The primary goal of the U. S. Geological Survey (USGS) Multi-Hazards Demonstration Project (MHDP) is to use the best science available to reduce losses from natural hazards in the southern California community. In order to facilitate the application of scientific research to decisionmaking, the USGS partners with county and city government agencies, academic researchers, the Federal Emergency Management Agency (FEMA), the National Oceanic and Atmospheric Administration (NOAA), and local emergency response agencies to reduce losses. The natural hazard scenarios developed by MHDP include earthquakes (Jones and others, 2008), severe floods, coastal erosion, landslides, and wildfires. The final scenario being planned for emergency managers will involve a tsunami that is sourced along the eastern Aleutian-Alaska megathrust and will travel across the Pacific Ocean to strike the coast of southern California. A tsunami is generated by the vertical displacement of the seafloor, in response to an earthquake or a landslide, with that displacement transferred to the water column. This results in the formation of a series of long-period waves that can cause destruction when they strike both nearby and distant coasts. Within the past several years, the devastating impact to both life and property of tsunami waves generated during giant earthquakes in Sumatra (2004), Chile (2010), and Japan (2011) has provided increased impetus to better prepare southern California for future tsunamis. Although not rigorously defined, a “giant earthquake” is one that has a magnitude (M) ~9.0 or larger, and a “great earthquake” is larger than an M~8. When available, we prefer to use M_w (moment magnitude), which is based on the surface area and slip of the actual rupture during an earthquake, directly related to total seismic energy release. For many of the older historical earthquakes, the only available magnitude is a surface wave magnitude (M_s).

We first provide a history of tsunami hazards in the California Continental Borderland, with an emphasis on tsunamis generated by local earthquakes and landslides. Most of the paper is focused on the possibility of the rupture of a giant earthquake beneath several eastern sections of the

Aleutian-Alaska megathrust, which could generate a trans-Pacific tsunami affecting the southern California coast. For the eastern Aleutian-Alaska megathrust, we summarize available information about both historical and prehistoric seismicity as it relates to prior great earthquakes. This is followed by an analysis of geologic factors indicating that parts of the Aleutian-Alaska megathrust are at least partially locked and accumulating strain, including (1) Global Positioning System (GPS) data, (2) the presence and location of forearc basins, (3) the subduction of bathymetric features such as fracture zones, ridges, and seamounts, and (4) the amount of sediment in the trench, which could smooth roughness on the subducting plate and promote coupling. Finally we conclude with a discussion of recent great subuction-zone earthquakes (2004 Sumatra-Andaman, 2010 Maule, Chile, and 2011 Tohoku-Oki, Japan) and how those events have affected our understanding of seismic hazards along the Aleutian-Alaska megathrust.

Previous Tsunamis Affecting the California Continental Borderland

The evidence for tsunamis affecting southern California during both historical and prehistoric times is sparse. McCulloch (1985) provides a comprehensive overview of the potential impact of a tsunami in southern California. For a summary of significant tsunamis that were recorded in California (in addition to the rest of the United States and its possessions), see Lander and Lockridge (1989). During historical times in southern California—dating from the formation of a mission in San Diego in 1769—only two tsunamis caused any damage and both of them were locally generated: 1927 M_w 7.3, off Point Arguello, and 1812, off Santa Barbara (McCulloch, 1985). In 1812 a moderate tsunami, on the order of 2 m, was generated following a local earthquake of magnitude greater than *M* 7 located most likely beneath the western part of Santa Barbara Channel (Lander and Lockridge, 1989). The 1927 *M* 7.3 earthquake occurred on an offshore reverse fault that has not been definitively identified. It generated a tsunami that was observed to have a runup of about 2 m just north of Point Arguello (Lander and Lockridge, 1989; Satake and Somerville, 1992).

Few studies have been conducted to identify tsunami deposits preserved in coastal sediments in southern California. However, urbanization likely has destroyed most geologic evidence for prehistoric tsunamis. The only interpreted paleotsunami deposit in southern California was documented near Carlsbad, north of San Diego (Kuhn, 2005). Near Santa Barbara, possible sand layers consistent with tsunami deposition were found beneath the Carpenteria salt marsh, although that study is ongoing and results are as yet inconclusive (R. Peters, unpublished report, 2009).

Both locally sourced and remotely sourced tsunamis have the potential to affect the southern California coast. Sources of locally generated tsunamis include both landslides and earthquakes. Lee and others (2009) summarize landslide hazards in the California Continental Borderland, an offshore area of islands and channels south of Point Arguello that lies between the shoreline and the deep ocean basin to the west. Most submarine landslides that have been imaged using high-resolution bathymetry in the California Continental Borderland are relatively small. The largest known landsides are the Goleta slide in Santa Barbara Channel and the Palos Verdes debris avalanche south of the Palos Verdes Peninsula (fig. 1). The Goleta landslide is composed of three main lobes and is one of the few deep-seated landslides in the borderland. It shows evidence for numerous separate mass wasting events and is spatially associated with a major fault system (Fisher and others, 2005). A numerical model of a tsunami generated by movement of all three lobes of the Goleta slide concurrently shows that it would result in a peak runup of as much as 20 m (Borrero and others, 2001). If only an individual lobe failed, the peak tsunami runup would be on the order of 10 m (Greene and others, 2006). The three main lobes are dated as having failed at 10 ka, 8 ka, and most recently (and least constrained) at ~6 ka (Fisher and others, 2005). Using the estimated ages of the past seven paleolandslide events as imaged on seismic reflection profiles, Geist and Parsons (2010) approximate the probable recurrence interval of a Goleta landslide to have a mean of 31 k.y.

The other major potential landslide source for a tsunami is the Palos Verdes debris avalanche. It is composed of blocky landslide debris that extends over a distance of 10 km and has a volume of between 0.3 and 0.7 km^3 (Bohannon and Gardner, 2004). Borrero and others (2004) modeled a tsunami generated by the Palos Verdes slide and determined a maximum runup of more than 5 m in the area of steep cliffs along the Palos Verdes Peninsula. The most recent movement of the debris avalanche was at about 7.5 ka (Normark and others, 2004). Additional local landslide sources include large-scale basement failures along the flanks of high relief bathymetry, such as the Thirtymile and Fortymile Banks (Legg and Kamerling, 2003). The ages of these failures are not well constrained and are most likely Pleistocene and older (Legg and Kamerling, 2003).

Locally sourced tsunamis in the California Continental Borderland also can be generated by vertical motion on offshore reverse or thrust faults and restraining bends within strike-slip fault zones. Fisher and others (2009) and Ryan and others (2009) summarize evidence for active faults between Point Arguello and the Mexican border. Thrust faults with primarily dip-slip displacement in the inner California Continental Borderland occur beneath the Santa Barbara Channel, the Palos Verdes Peninsula, and northern San Diego County. In particular, thrust faults in deeper waters, such as the Channel Islands Thrust beneath Santa Barbara Channel and the San Mateo fold and thrust belt offshore San Diego County, have the potential to cause coastal damage immediately landward of the fault. The Channel Islands Thrust is a blind thrust that dips to the north at ~23° and is capable of generating an M_w 7.3 earthquake (Borrero and others, 2001). A tsunami modeled from such an event would produce local runup on the

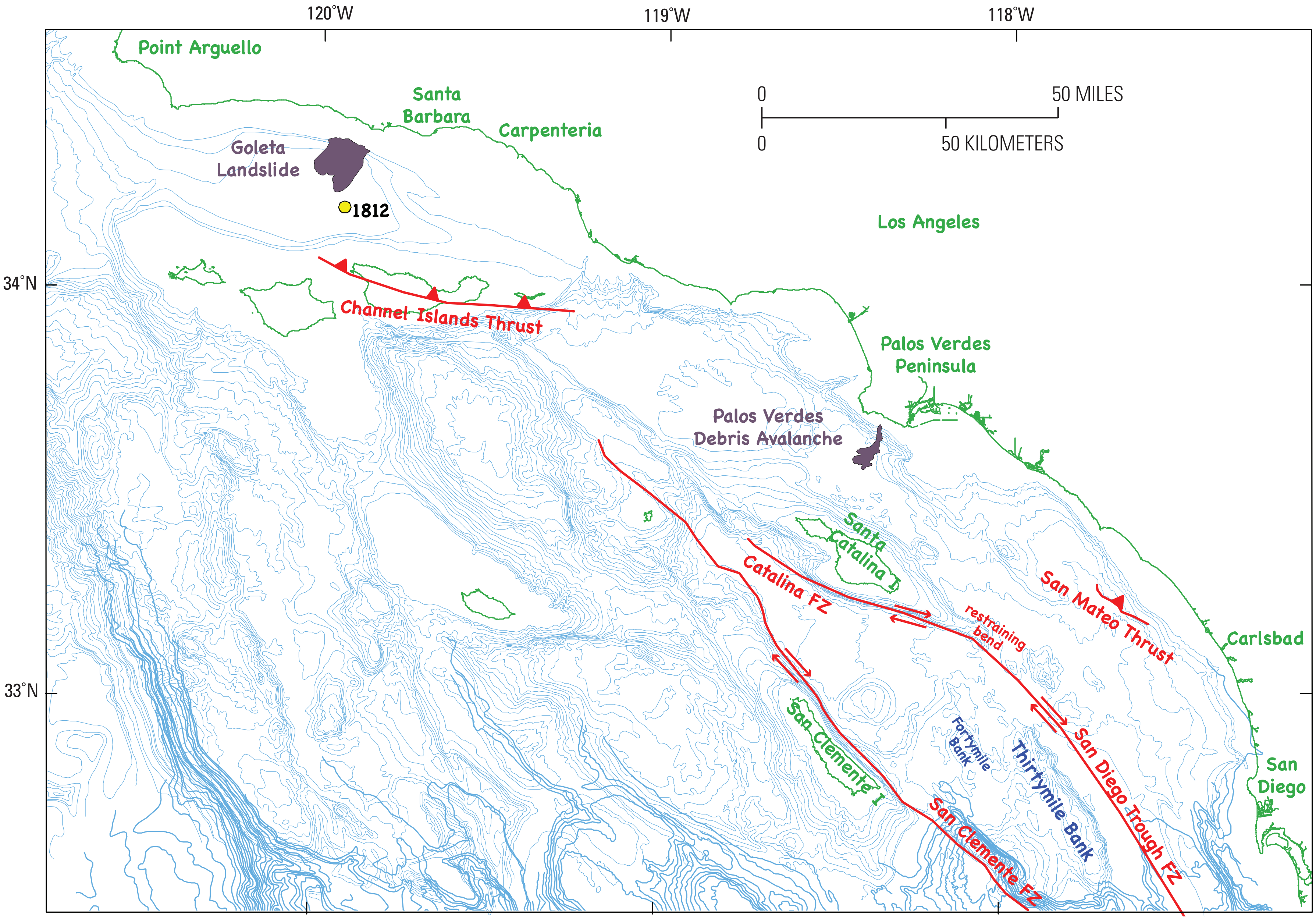

Figure 1. Examples of local earthquake and landslide sources that have been modeled for tsunami generation in the Southern California Borderland. The yellow dot is located at the inferred location of an earthquake that generated a tsunami near Santa Barbara in 1812. Bathymetric contour interval is 100 m. Bathymetric data compiled by Pete Dartnell (U.S. Geological Survey) from multibeam and other sources (see Lee and others, 2009). FZ, fault zone.

order of 2 m (Borrero and others, 2001). Less is known about the San Mateo Thrust. It has been modeled as capable of an M_w 7.0 earthquake on a 45° dipping fault plane, which would produce concentrated local runup of more than 5 m (Borrero and others, 2004).

The majority of offshore faults in the borderland are strike-slip faults showing primarily horizontal displacement and thus do not generally pose a tsunami threat, except in areas of releasing and restraining bends with significant dip-slip displacement. Restraining bends along the San Clemente and San Diego Trough-Catalina Fault Zones were evaluated for tsunami hazards (Borrero and others, 2004; Legg and Borrero, 2001; Legg and others, 2004). Legg and Borrero (2001) used elastic dislocation modeling to determine the amount of dip slip predicted within a left bend along the right-lateral San Clemente Fault Zone, southeast of San Clemente Island (fig. 1). Their model results showed only 40 cm of seafloor offset, which the authors consider to be most likely too low for the generation of a significant tsunami. Legg and others (2004) modeled the restraining bend between San Diego Trough Fault and Catalina Fault that is proposed to uplift Catalina Island. However, more recent studies indicate that the Catalina Fault is not active and that Catalina Island is presently subsiding as evidenced by submerged marine terraces (J. Conrad, oral commun., 2010).

The two large landslides in the borderland that pose the greatest tsunami threat, the Goleta slide and Palos Verdes debris avalanche, last failed several thousand years ago and have recurrence intervals on the order of tens of thousands of years. This is consistent with the observation that many landslides are temporally tied to the glacial cycle and date to before 5 ka (Lee, 2009). For locally generated earthquake tsunamis, it is more difficult to evaluate the probability of a tsunami, because slip rates and recurrence intervals on most offshore faults are not well known. However, in general, these faults likely have low slip rates, because most of the Pacific-North American Plate motion can be accommodated by faults on land, such as the San Andreas and San Jacinto Faults. Low slip rates imply long repeat times between events, with therefore a relatively low annual probability of occurring. With the exception of the 1927 Arguello tsunami and possibly the 1812 Santa Barbara event, all damaging historical tsunamis in the state of California (a total of five between 1812 and 1988) were remotely generated by distant earthquake sources; the most damaging was the tsunami from the 1964 Alaskan earthquake (McCulloch, 1985). Thus, the most likely tsunami source to cause damage to coastal California would be a giant earthquake along a Pacific-rim subduction zone.

Subduction zones pose one of the greatest hazards to communities situated along continental margins. McCaffrey (2007) wrote, “For policy purposes, one lesson we should take away from the Sumatra-Andaman earthquake is that every subduction zone is potentially locked, loaded, and dangerous.” Because subduction zone earthquakes rupture elongate areas parallel to the zones, the wave amplitude of any resultant tsunami is greatest in the direction perpendicular to the subduction zone. If the earthquake rupture is long, it creates a beaming (directivity) effect that, together with the bathymetry along the tsunami path, will control the locations where a transoceanic tsunami will have its greatest impact. For Los Angeles, the highest danger for a distant tsunami from any reasonable source in the Pacific is from a tsunami source region along the eastern Aleutian-Alaska megathrust, based on a probabilistic tsunami hazard assessment (Thio and others, 2010; fig. 2). A trans-Pacific tsunami spawned along the Aleutian-Alaska megathrust, however, will have potential impacts in areas far beyond California, including Hawaii, Alaska, and the rest of the Pacific Basin (McCulloch, 1985; Lander and Lockridge, 1989).

Aleutian-Alaska Subduction Zone

Background

The Aleutian-Alaska subduction zone (AASZ) stretches 3,000 km from the Near Islands in the far west to Middleton Island in the Gulf of Alaska (figs. 3, 4). Its physiography is composed of a deep trench (5–6 km), the Aleutian Ridge and Alaska Peninsula volcanic arc, and a complexly structured forearc that includes the Aleutian Terrace at about 4-km depth west of Unimak Pass (near long. 166°W) (fig. 4). At least 29 active volcanoes have erupted along this arc during historical times dating back to 1760 (Miller and others, 1998). Along the AASZ, the Pacific Plate subducts beneath the North American Plate at velocities that range from 54 to 76 mm/yr, with subduction varying from roughly perpendicular to the trench in the east to increasingly oblique in the west (Demets and others, 2010) (fig. 3).

The geologic framework of the Aleutian arc is summarized in Vallier and others (1994) and references therein. At Unimak Pass, there is a transition from subduction beneath the intraoceanic Aleutian Ridge to the west and the continental margin of Alaska to the east. This transition is marked by a westward decrease both in the distance between the trench and arc and in the width of the shelf (von Huene, 1989). West of Unimak Pass, the intraoceanic Aleutian Ridge formed as an active volcanic arc in the Eocene (Scholl and others, 1987) and is now mostly submerged. East of Unimak Pass, the continental margin, including the Alaska Peninsula and Kodiak, Shumagin, and Sanak Islands (fig. 4*B*), is composed primarily of Mesozoic through Holocene igneous rocks and a Mesozoic to early Tertiary accretionary complex (Plafker and others, 1994). Since the early Eocene, the entire AASZ has had a similar tectonic history.

The geometry of the surface of the subducting plate along the Aleutian Trench is depicted by the USGS Slab1.0 model, which is based on a variety of datasets, including earthquake catalogs, seismic profiles, sediment thickness, and bathymetry (Hayes and others, 2012; fig. 5). Contours of the subducting slab are relatively smooth, with the exception of offshore Kodiak and particularly beneath the Shumagin Islands, where the slab surface is distorted (fig. 5). The dip of the slab shallows beneath Kodiak Island; it locally steepens in the vicinity of the Shumagin Islands (fig. 6). West of Sanak Island, the slab is slightly steeper, with little along-strike variation in dip as far west as the Fox Islands.

The age of the subducting Pacific Plate varies from about 33 Ma in the Gulf of Alaska to ~90 Ma in the far western Aleutians; oceanic crust increases in age southward from the trench (Atwater, 1989). Because the Pacific Plate increases in age to the west, the older and colder lithosphere results in an increase in the depth of the trench to the west. Major fracture zones within the subducting Pacific Plate that intersect the trench include the Amlia, Rat, Adak, Aja, and 58 degree Fracture Zones (fig. 7). In addition to the fracture zones, two low ridges, the Kodiak-Bowie and Patton-Murray, extend across the Gulf of Alaska and subduct near the Kodiak margin (fig. 8). The crests of both ridges are marked by chains of seamounts, including the prominent Kodiak Seamount located just seaward of the trench at the north end of the Kodiak-Bowie seamount chain.

Convergence between the Pacific and North American Plates is close to orthogonal along the eastern Aleutian Trench, but it becomes increasingly oblique to the west (fig. 3). The difference in obliquity of plate convergence results in segmentation of the AASZ. This is most pronounced in the central and western Aleutians, which are composed of rotating and westward-translating blocks of various sizes separated by transverse, fault-controlled canyons (Geist and others, 1988). The easternmost of these rotating blocks is the Andreanof block that extends from Adak to near Seguam Island (fig. 3). East of the Andreanof block, the AASZ is also considered segmented as the result of differences in volcanism and crustal composition controlled by magma rupture through fractured or less fractured lithosphere (Kay and others, 1982; Shillington and others, 2004). A segment boundary occurs at Unimak Pass, near the intersection between the Bering shelf and the arc, which marks the transition between continental and oceanic crust (Lizarralde and others, 2002). The Unimak Pass area is characterized by an unusually wide and diffuse zone of volcanism. The lowest crustal velocities found along the east-central Aleutians occur at the segment boundaries near Seguam Island and Unimak Pass (see figure 4 for locations; Shillington and others, 2004).

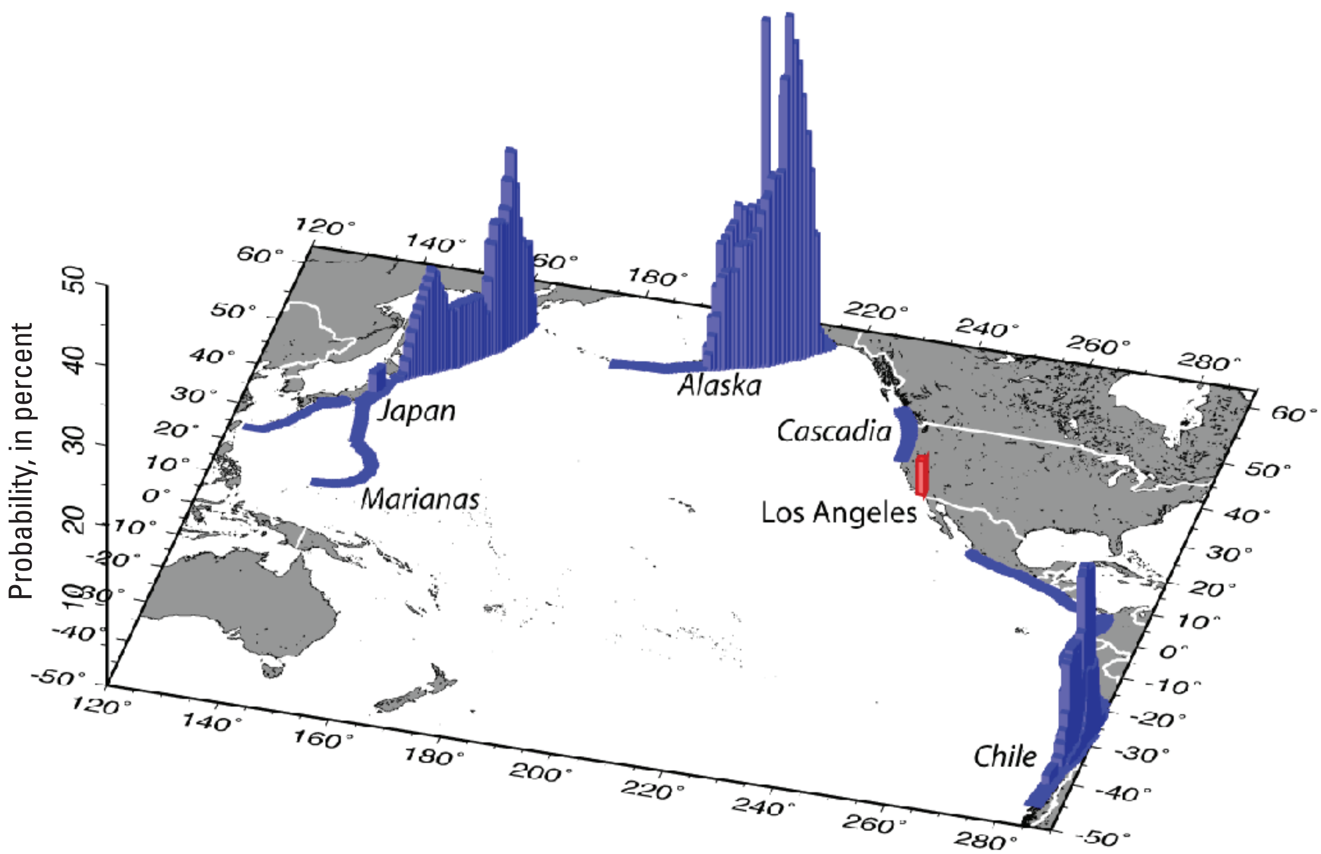

Figure 2. Tsunami source disaggregation for peak wave height at 475-yr return period for Los Angeles (Thio and others, 2010). A tsunami hazard curve is an aggregation of tsunami sources as used by engineers, based on the probability that a runup will be as large or greater than a certain value. A disaggregation plot indicates the location of the predominant source area. The height of the blue bars indicate the percent probability of such an event.

Historical Seismicity

Along with those off Chile and Indonesia, the Aleutian-Alaska megathrust is one of the three most seismically active subduction zone systems in the world, and it is the most seismically active region in the United States. The historical instrumental seismicity recorded in Alaska since 1898 for magnitude > 4 is shown in figure 9 (Alaska Earthquake Information Center, 2010). Ruppert and others (2008) provide a summary of historical Aleutian-Alaska seismicity. As the result of variations in the obliquity of convergence, the maximum depth of seismicity varies from about 250 km in the east to 50 km in the west. Normal faulting occurs beneath the trench and outer rise, with shallow thrust faulting resulting from slip between the plates at depths less than about 60 km. Earthquakes at intermediate depths (70–300 km) that define the Wadati-Benioff zone occur as the result of deformation within the slab. Double seismic zones with two layers of intermediate depth earthquakes were resolved beneath the central Aleutians (Engdahl and Scholz, 1977) and the Shumagin Islands (Hudnut and Taber, 1987), indicating different stresses in those areas.

Most of the strain generated by plate convergence is released in the main thrust zone, which is located from just arcward from the trench to a maximum depth of 40–50 km (Boyd and others, 1995). Of the 11 largest earthquakes recorded worldwide during historical times (with $M_w > 8.5$), 4 occurred along the Aleutian-Alaska megathrust, with 3 of them (1946, 1957, and 1964) spawning large, destructive trans-Pacific tsunamis. During the mid 20th century, most of the Aleutian-Alaska megathrust ruptured in a sequence of $M >$ 8 events: 1938, 1946, 1957, 1964, and 1965; the rupture zones of these earthquakes are shown in figure 10 (Rhea and others, 2010). For earthquakes that occurred since the introduction of digital broadband seismometers in the early 1960s, reliable finite slip models were determined. For older events, the distribution of moment release is generally determined by tsunami waveform modeling from tide gauge data. Areas that did not rupture in a great earthquake in the past century include the "Shumagin gap," which lies between the 1938 and 1946 rupture zones, and possibly the "Unalaska gap" within the eastern part of the 1957 rupture zone (Sykes, 1971; Davies and others, 1981; House and others, 1981).

We briefly summarize the recurrence history of significant earthquakes that ruptured the main thrust zone of the Aleutian-Alaska arc, as described primarily by Nishenko and Jacob (1990), which updates Davies and others (1981). The historical seismicity is discussed from east to west with respect to the great megathrust earthquakes of the 20th century that occurred in 1938, 1946, and 1957. However, the preinstrumental record of seismicity only goes back to 1788 and is

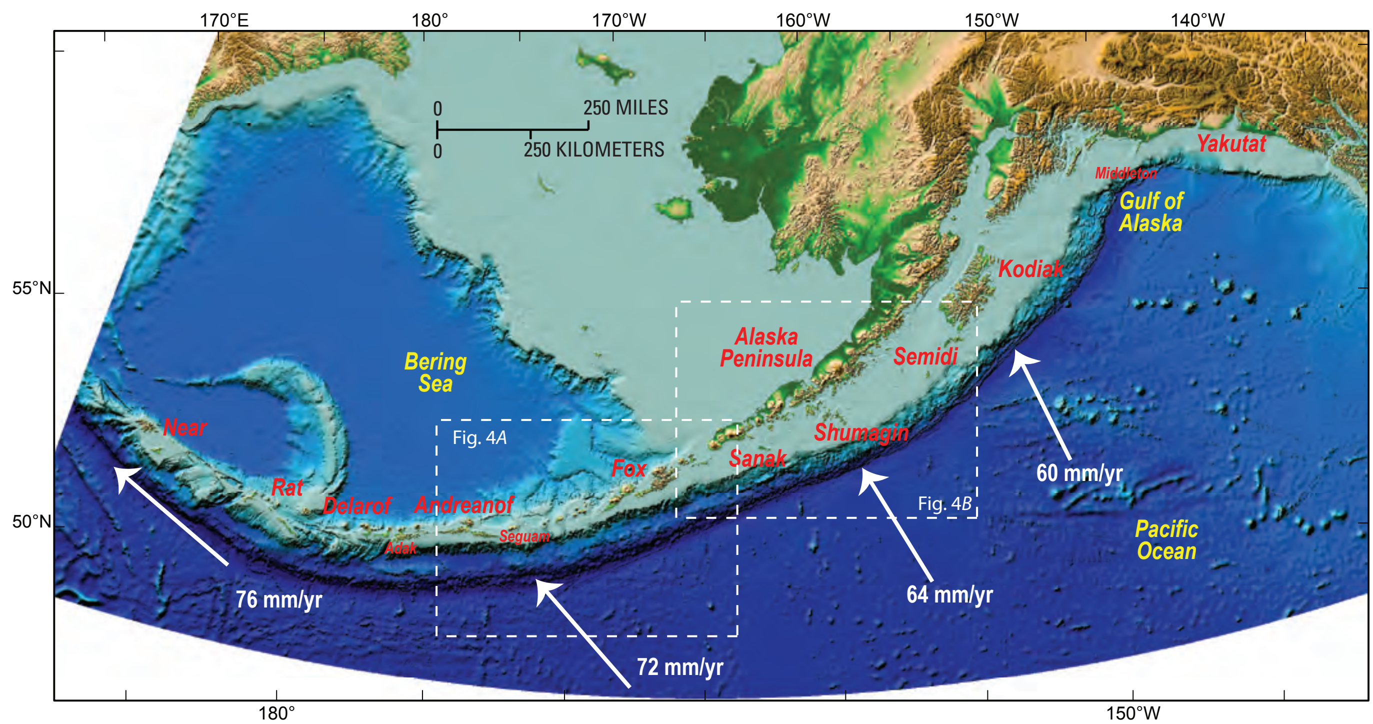

Figure 3. Map of the Aleutian Islands and southern Alaska mainland, showing locations of major features, including islands and island groups discussed in the text. Numbers and arrows show relative motion between the Pacific and North American Plates at the location of each arrow's tip (angular velocity and pole of rotation from DeMets and others, 2010). Bathymetry for this and other figures is from National Oceanic and Atmospheric Administration's southern Alaska coastal relief model with a range in values from about 8 km below sea level to 6 km above (Lim and others, 2009). Locations of figs. 4*A* and 4*B* shown by dashed white lines.

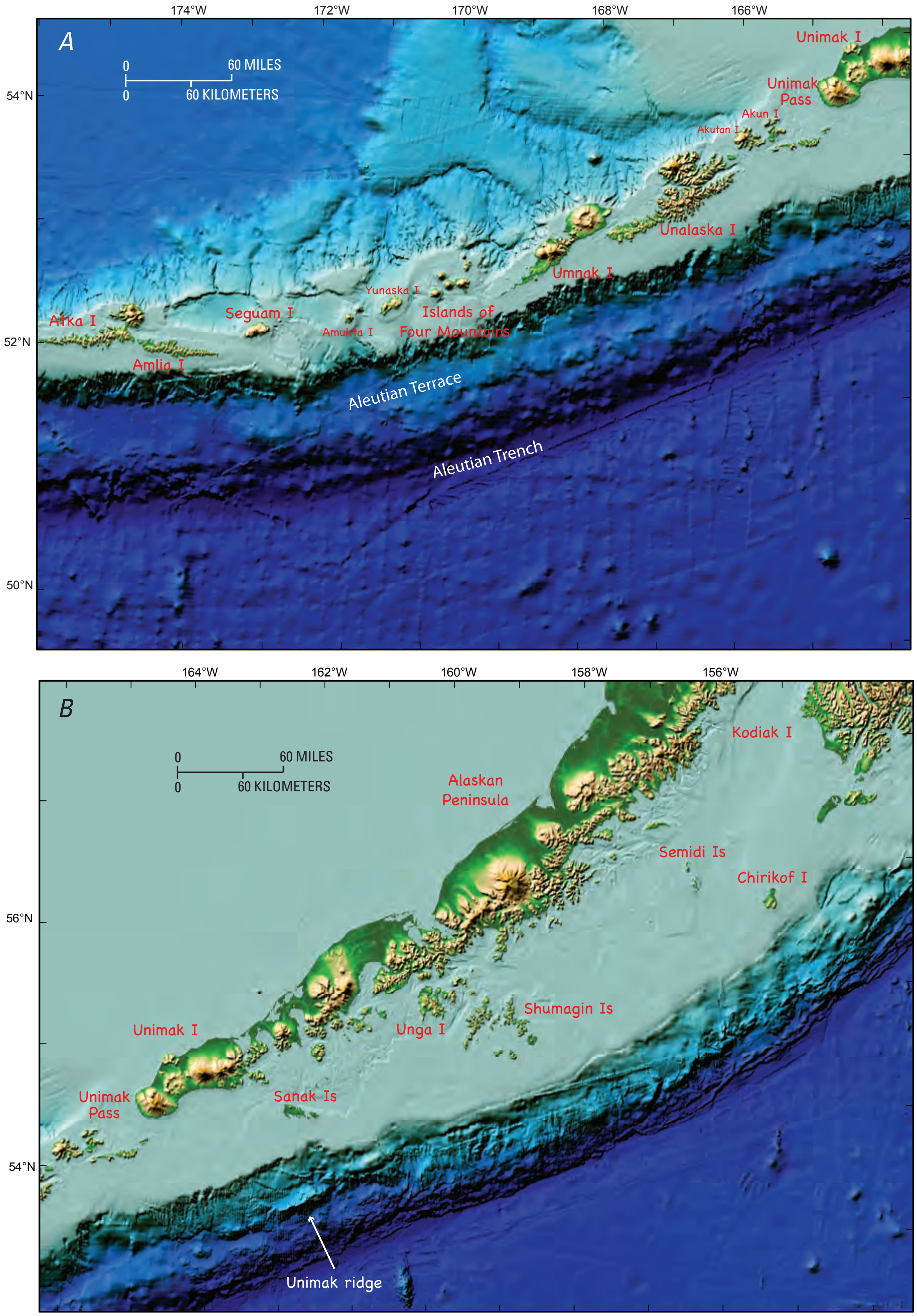

Figure 4. Location maps of (*A*) the oceanic sector of the eastern Aleutian-Alaska arc from Atka Island to Unimak Pass and (*B*) the continental sector of the eastern Aleutian-Alaska arc from Unimak Pass to Kodiak Island.

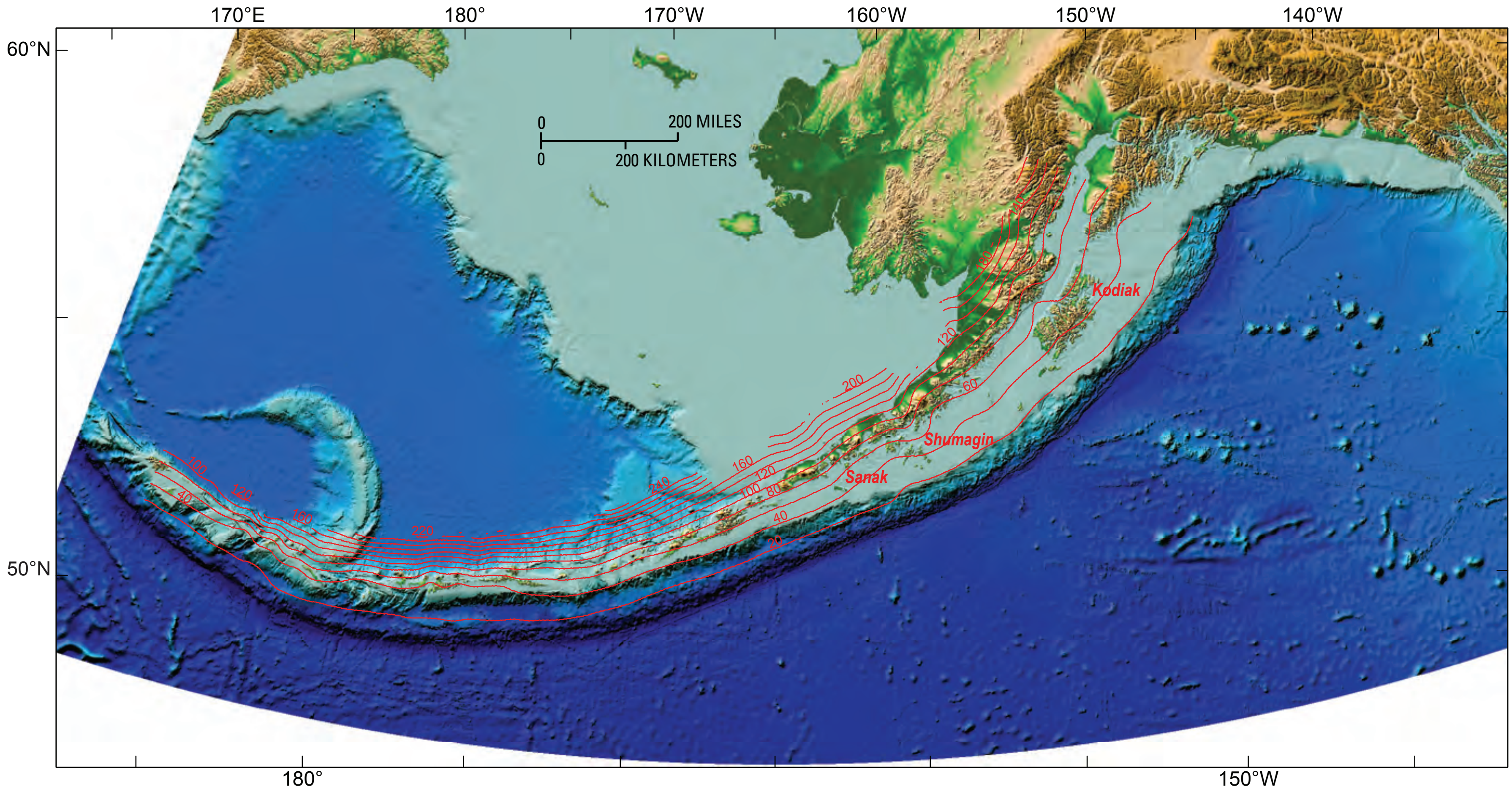

Figure 5. Contour map of the depth to the top of the subducting Pacific Plate from the Alaska-Aleutians Slab1.0 model (Hayes and others, 2012). Depths given in kilometers below mean sea level; contour interval varies. The slab model is constructed from a variety of datasets, including historical earthquake catalogs, centroid moment tensor (CMT) solutions, seismic reflection profiles, bathymetry, and sediment thickness information.

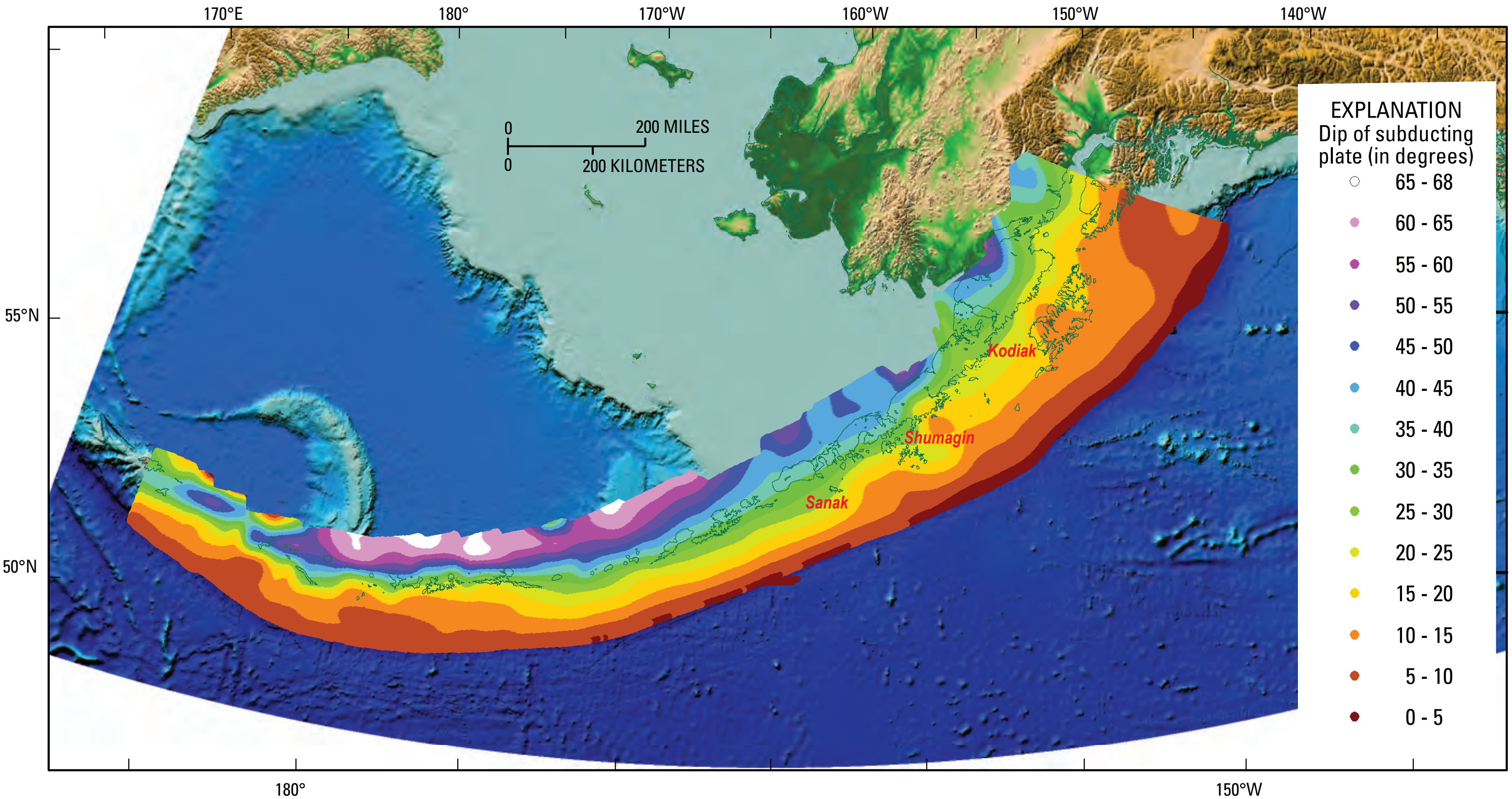

Figure 6. Map showing variation of the dip (in degrees) of the subducting Pacific Plate, from the Alaska-Aleutians Slab1.0 model (Hayes and others, 2012). Beneath Kodiak Island, the slab dips gently; it locally steepens in the vicinity of the Shumagin Islands. West of Sanak Island, the slab slightly steepens, with the dip profile remaining relatively constant along strike to the west.

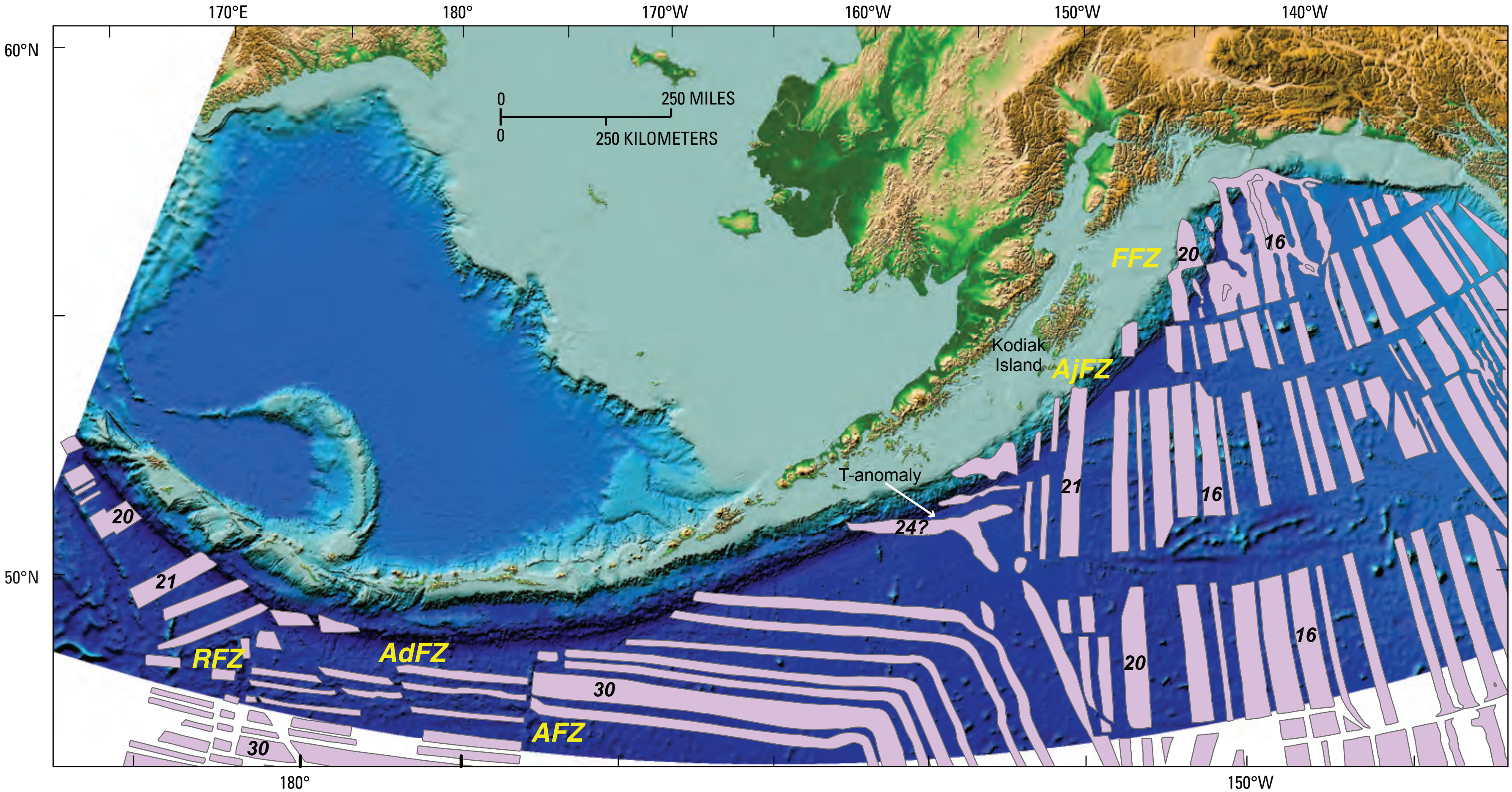

Figure 7. Marine magnetic anomalies in the north Pacific Ocean digitized from Atwater (1989); the numbers of some of the magnetic anomalies are labeled on the map. The age of the subducting Pacific Plate varies from about 33 Ma in the Gulf of Alaska to ~90 Ma in the far western Aleutians. The Zodiac fan intersects the trench between the T-anomaly and Kodiak Island (Atwater, 1989). AFZ, Amlia Fracture Zone; AdFZ, Adak Fracture Zone; AjFZ, Aja Fracture Zone; FFZ, 58° fracture zone; RFZ, Rat Fracture Zone.

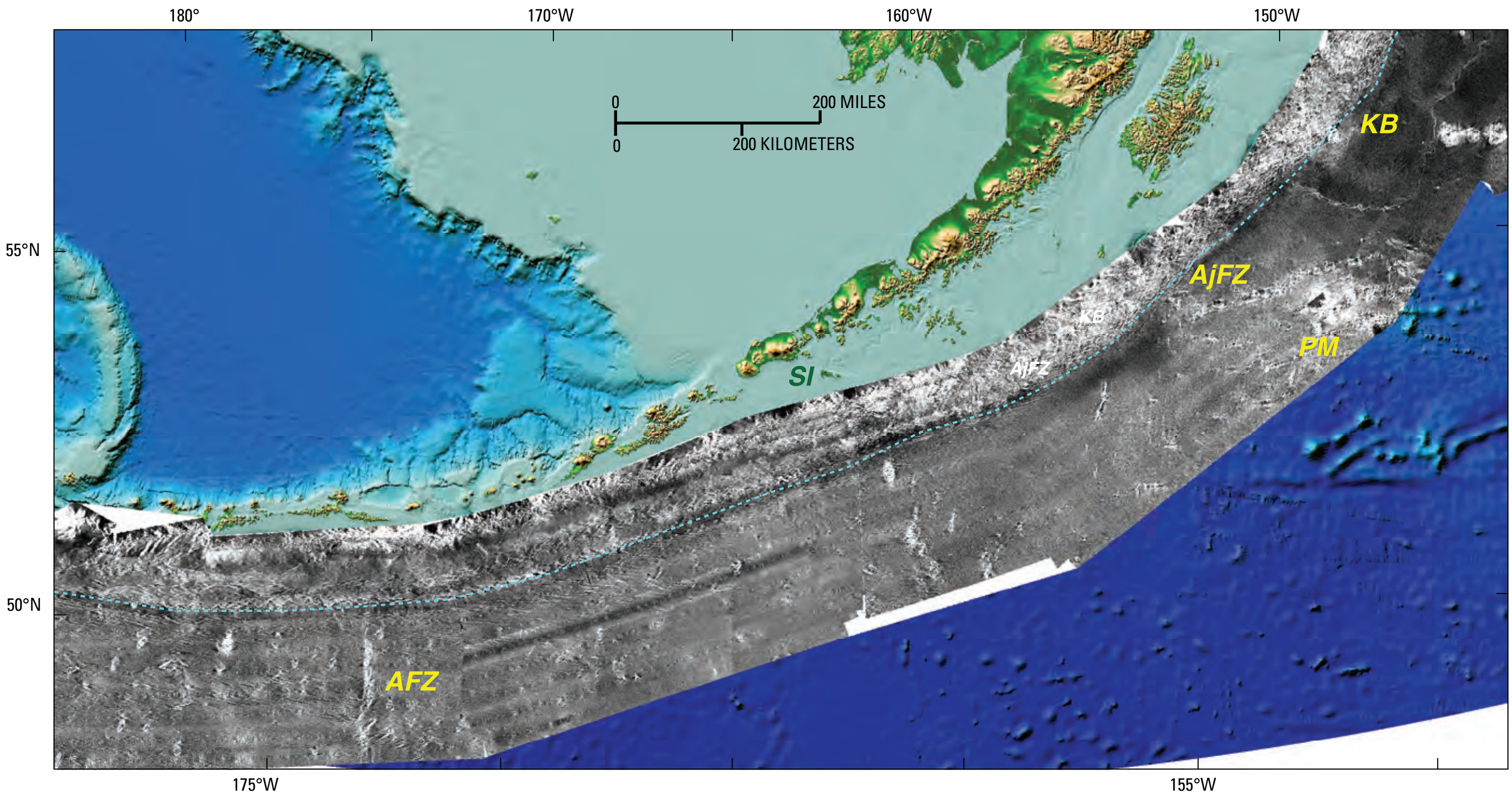

Figure 8. Sidescan sonar imagery (grey; from Paskevich and others, 2010) taken by GLORIA (Geological LOng-Range Inclined Asdic) in a broad swath straddling the Aleutian Trench and showing relief on the Pacific Plate adjacent to the trench. GLORIA transmits acoustic energy at a nominal frequency of 6 kHz and receives energy backscattered from ocean bottom features; high backscatter (light areas) indicates steep slopes. Dashed blue line denotes arcward edge of the trench. Major fracture zones within the subducting Pacific Plate that intersect the trench include the Amlia (AFZ) and Aja (AjFZ) Fracture Zones. The Kodiak-Bowie (KB) and Patton-Murray (PM) seamount chains extend across the Gulf of Alaska and subduct beneath Aleutian-Alaska Trench. Although there are no prominent fracture zones or seamount chains that might form a barrier to a throughgoing rupture between PM and AFZ, a seamount is located near the trench offshore of Sanak Island (SI).

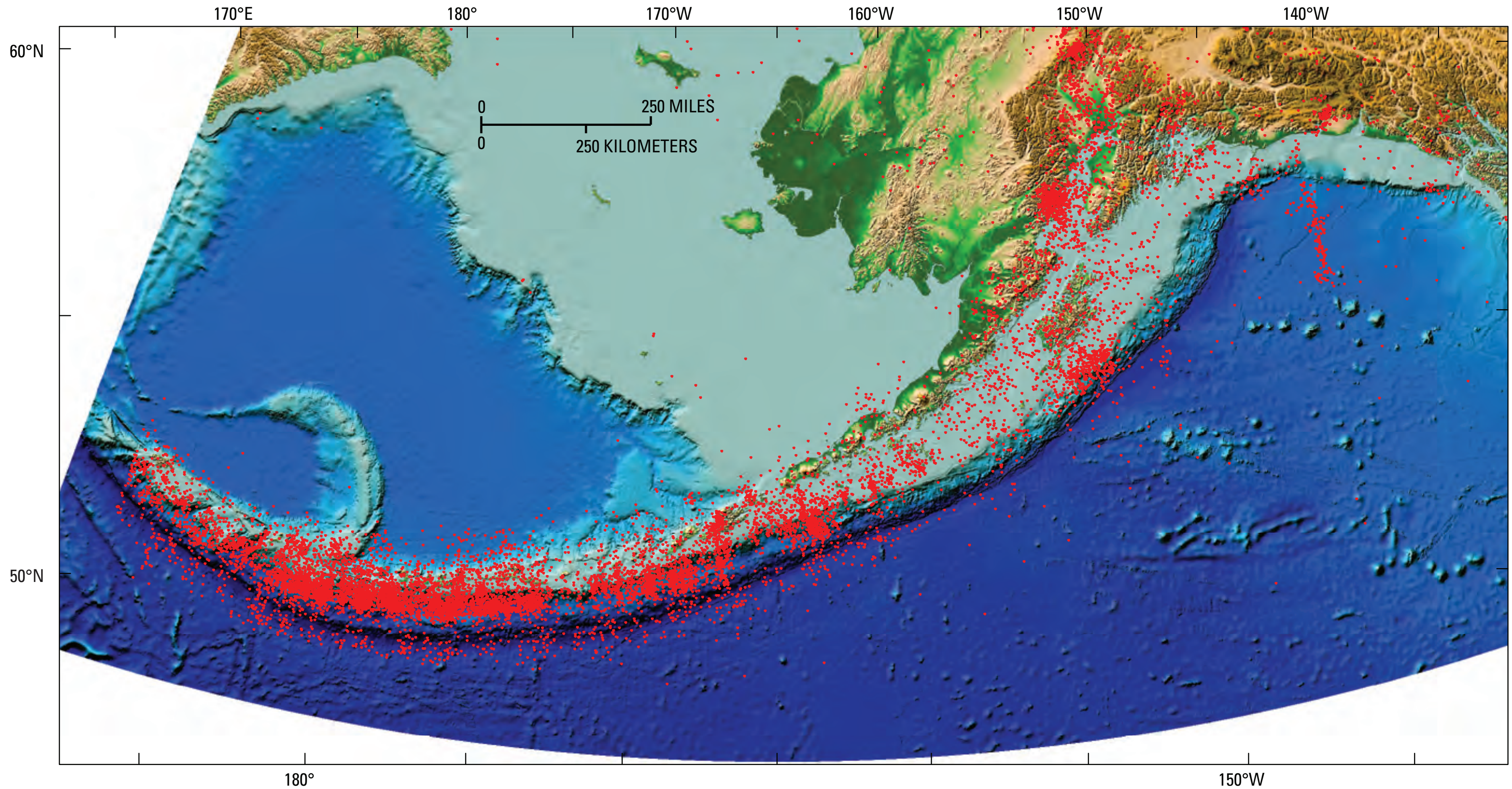

Figure 9. Distribution of earthquakes (red dots) in Alaska with magnitudes greater than 4 that occurred from 1898 to 2010 (Alaska Earthquake Information Center, 2010).

determined primarily from sparse historical intensity and tsunami observations (Jacob, 1984). Thus, the earthquake record for the Aleutians is relatively short (just over 200 yrs) and may not contain a complete seismic cycle (McCaffrey, 2008).

In 1938, an *M*8.3 earthquake ruptured the Aleutian-Alaska megathrust from 156°W to 158.5°W offshore of the Alaskan Peninsula in the area of the Semidi Islands (fig. 10). The event occurred beneath the broad shelf from east of the Shumagin Islands to Chirikof and the Trinity Islands, rupturing a 300 km by 100 km section of the plate boundary, although the aftershock zone was difficult to define (Sykes, 1971). The epicenter was along the northern edge of the rupture zone near Shumagin Basin; however, most of the moment released during the 1938 earthquake was in the eastern third of the rupture zone, with a maximum slip of 3.3 m (Johnson and Satake, 1994). Additional slip occurred in the western third of the rupture zone (0.8 m), with little slip resolved along the updip (deeper water) portion of the megathrust. The repeat time for a similar rupture is estimated to be 50–90 yrs (Davies and others, 1981). A minor local tsunami was observed associated with the 1938 earthquake and caused minor flooding on Unga Island (fig. 4*B*) (Lander and Lockridge, 1989); there was also a weak tsunami observed at Santa Monica in southern California (McCulloch, 1985).

Although earthquakes before the 20th century are poorly constrained, several prior historical events are considered to have ruptured all or part of the 1938 rupture zone. An earthquake in 1890 occurred near Chirikof Island that resulted in an abrupt tilt of the island, although the along-strike extent of this event is somewhat uncertain. Great earthquakes that broke both the entire 1938 rupture zone and at least part of the "Shumagin gap" to the west occurred in 1847 and 1788 (Davies and others, 1981) (see further discussion below).

Between the rupture zones of the great 1938 and 1946 earthquakes lies the "Shumagin gap," a section of the Aleutian-Alaska megathrust that did not experience an *M*>8 earthquake during the 20th century, as has most of the rest of the megathrust. The length of the Shumagin gap is 250 km, which would be capable of producing an *M*8.2 quake if a rupture occurred along the full length of the gap (Davies and others, 1981). Since at least the early 20th century, the entire Shumagin seismic gap has not ruptured in a great earthquake. An *M*7.4 earthquake ruptured the eastern edge of the Shumagin gap in 1917, but it did not rupture as far west as Sanak Island; no historical large earthquakes ruptured the western segment of the Shumagin gap (Estabrook and Boyd, 1992). In 1947, an earthquake with an M_s 7.5 ruptured the deepest part of the main thrust zone in the central part of the Shumagin gap (Davies and others, 1981). Since 1947, no accumulated slip has been released in the eastern part of the Shumagin Islands (Estabrook and Boyd, 1992). Wesson and others (2007) suggested that the megathrust beneath the Shumagin Islands has more frequent, but moderate earthquakes than along the rest of the arc.

Earthquakes in 1788 and 1847 that broke the megathrust in the area of the 1938 event are thought to have simultaneously ruptured into, at a minimum, the eastern half of the Shumagin gap as great earthquakes (Davis and others, 1981). At least 500 km of the megathrust ruptured in 1847, and at least 600 km ruptured in 1788. The event of July 22, 1788, provides the best evidence that the Shumagin Islands section of the megathrust has ruptured in a great earthquake in the past. The magnitude of this earthquake was *M*8 or greater, rupturing the megathrust from Kodiak Island through the Shumagin Islands and perhaps to Sanak Island (fig. 4*B*) (Davis and others, 1981; Lander and Lockridge, 1989). There may have been two events in 1788 separated by a few weeks, with the second event on August 16, 1788. Historical descriptions of the two 1788 events suggest that the July event ruptured the megathrust from Kodiak to Unga Island and the August event from Unga to Sanak Island. Thus these two events may have between them ruptured most of the Shumagin gap (Davies and others, 1981). On Unga Island, water levels rose as much as 90 m (Lander and Lockridge, 1989). The occupation of Unga Island by native Shumagin people ended in 1788 as a result of the destruction and abandonment of the last native village on the island, most likely from tsunami inundation (Winslow and Johnson, 1989).

The great Scotch Cap earthquake of April 1, 1946 (fig. 10), ruptured an area west of the Shumagin gap and offshore of Unimak Island (fig. 4*B*) (Kanamori, 1972). The tsunami generated by that event was one of the most destructive tsunamis recorded in the Pacific Basin—it was in its aftermath that the Pacific Tsunami Warning Center was formed. The tsunami generated by the Scotch Cap earthquake traveled for more than 13,000 km from its source in the Aleutians to the Antarctic Peninsula, where damage from the tsunami was observed (Fryer and others, 2004; Okal and Hebert, 2007). Hawaii was particularly affected by this tsunami, with more than 150 deaths and millions of dollars in property damage. In central California, sea level was elevated 2.6 m above the tides in Half Moon Bay, with several boats washed across the coastal highway; one fatality occurred in Santa Cruz (Lander and Lockridge, 1989). In southern California, cargo vessels were pulled from their moorings in San Pedro (Lander and Lockridge, 1989).

The epicenter of the 1946 Scotch Cap earthquake was located near the deep-water trench (fig. 10). It was an archetypical "tsunami earthquake," meaning that it was impoverished in high frequencies (Kanamori, 1972). Thus the tsunami that was generated was much larger than expected from the surface-wave magnitude alone (M_s 7.4), which is based on the higher frequency components of the earthquake. Lopez and Okal (2006) relocated the aftershocks and determined that the rupture area was 180 by 115 km, with an average slip of 6–8 m. This rupture area estimate resulted in a recalculation of the magnitude to an M_w of 8.6, much higher than the surface-wave magnitude. The rupture was relatively slow, which may have been the result of slip either through a sedimentary wedge or a "corrugated slab interface" in a sediment-starved environment (Lopez and Okal, 2006). In addition to the far-field impact from the tsunami, the near-field destruction of the lighthouse

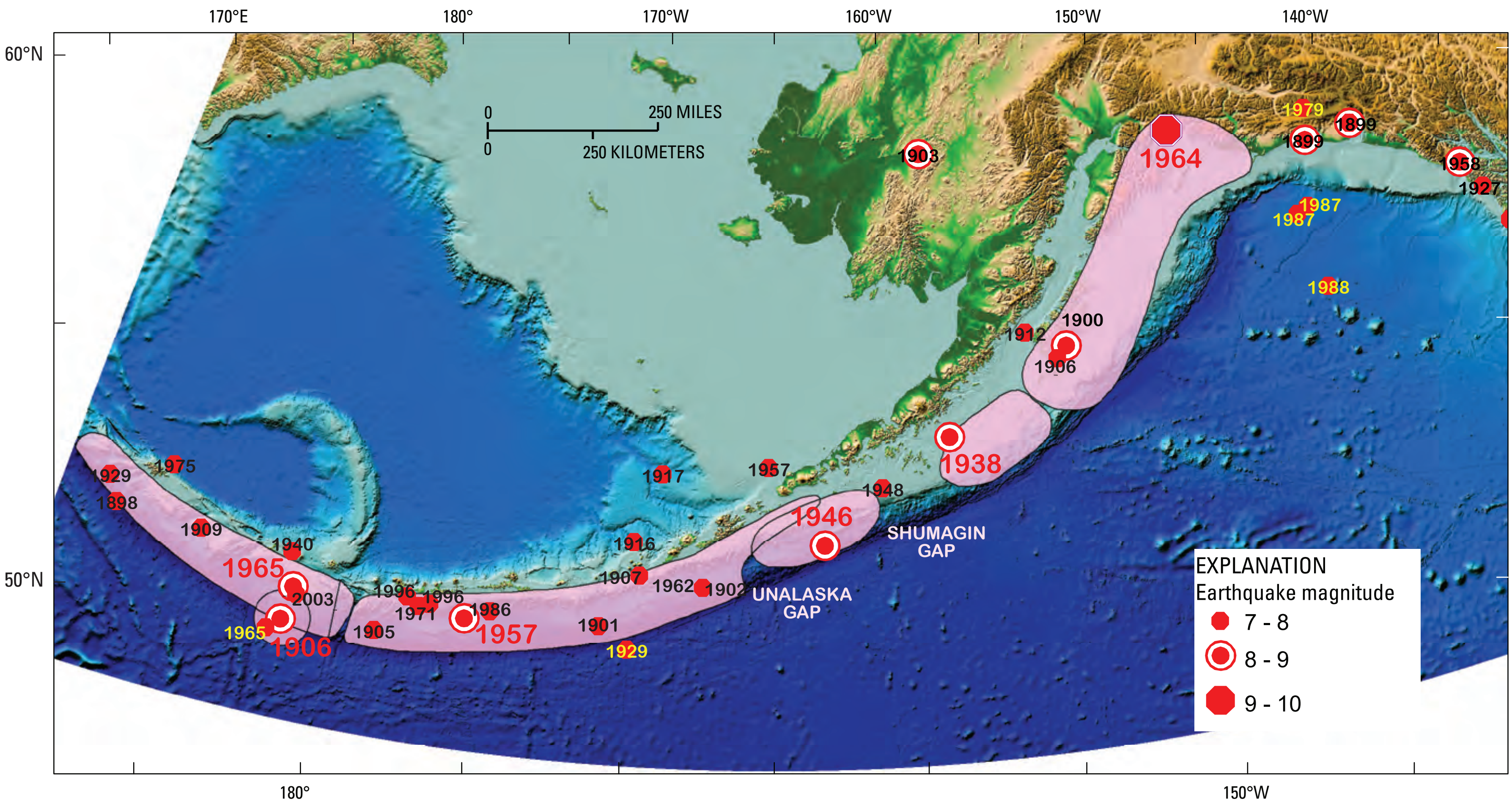

Figure 10. Map showing rupture zones (in pink) and epicenter locations of great earthquakes (M>8) on the Aleutian megathrust during the 20th century. From west to east, they are: 1906 and1965 Rat Islands earthquakes; 1957 Andreanof earthquake; 1946 Scotch Cap earthquake; 1938 earthquake; and 1964 great Alaskan earthquake. Rupture zones are from Rhea and others (2010). Epicenters of significant earthquakes with magnitudes >7.5, and their years of occurrence, are also shown with red symbols (from the database of the National Oceanic and Atmospheric Administration, 2010). That database contains information on destructive earthquakes from 2150 B.C.E. to the present. In order to be included in the database, the event must meet one of the following criteria: moderate damage, 10 or more deaths, magnitude 7.5 or greater, Modified Mercalli Intensity X or greater, or a tsunami generated.

at Scotch Cap by a 35-m-high wave can only be explained by a local source such as a landslide (Fryer and others, 2004) or splay fault. Some aspects of this earthquake and tsunami are still not well understood.

The March 9, 1957, M_w 8.6 Andreanof earthquake ruptured an area 1,147 by 324 km from the Delarof Islands (fig. 3) to near the edge of the1946 rupture zone (Johnson and Satake, 1993; Johnson and others, 1994); it is the 4th largest earthquake to occur within the United States in historical times. The trans-Pacific tsunami from this earthquake caused extensive damage in Hawaii. On the West Coast of the United States, it caused minor damage in San Diego Bay. This earthquake had one of the longest rupture zones ever measured, based on the lateral distribution of aftershocks (Johnson and others, 1994). However, the eastern end of the rupture zone was defined by few aftershocks. If slip were evenly distributed over the entire aftershock zone, the rupture zone would have only slipped on the order of 1 m, using the relation between magnitude and rupture area.

The 1957 earthquake happened before the establishment of the World Wide Standardized Seismographic Network (WWSSN), and thus its source parameters are not well constrained. The slip distribution for the earthquake was calculated by inverting tsunami waveforms, which were resolved into three major slip patches with a maximum slip of 7 m (Johnson and Satake, 1993). Most of the moment release was near the western end of the aftershock zone, between 174°W and 177°W longitude. There was a major change in the rupture characteristics of the 1957 event east of the Amlia Fracture Zone. Tidal gauge data indicate that the rupture zone of the 1957 event terminated near Unalaska Island (Fig. 4*A*); however, little moment was released between that island and the Amlia Fracture Zone to the west. A small asperity occurred near the downdip end of the megathrust near 168° to 169°W (Johnson and Satake, 1993). It has been proposed that the Unalaska section of the megathrust (between the Amlia Fracture Zone and Unalaska Island) is a seismic gap, because little strain was released along that section of the arc during the 1957 Andreanof earthquake (House and others, 1981).

There is scant evidence for previous great earthquakes in the vicinity of the Fox Islands (fig. 3). Based on tsunami arrivals at Hawaii, California, and Oregon, the earliest known earthquake in the Fox Islands was on August 23, 1872, although the location of its epicenter is not well constrained (Nishenko and Jacob, 1990). The 1872 event was considered to be an “earthquake of considerable magnitude” and, on the basis of tsunami arrival times probably ruptured from the west end of Umnak Island into the Four Mountain Islands area (fig. 4*A*) (Cox, 1984). A thrust event near Unalaska Island may have occurred in 1878, but this event is also poorly constrained (Nishenko and Jacob, 1990). In 1929, an *M*8.1 normal-fault earthquake occurred near the trench offshore of Umnak Island (Kanamori, 1972). During the 20th century, there were numerous events in the *M*6.5 to *M*7.4 range in the Fox Islands.

Prehistoric Seismicity

Little is known about the rupture history of the Aleutian-Alaska megathrust west of Kodiak Island for times prior to the historical record, which dates from the time of Russian settlements in the Aleutians in 1741. Paleoseismology studies documenting subduction zone earthquakes were conducted farther east in Cook Inlet, Prince William Sound, and Kodiak Island (Carver and Plafker, 2008). The best paleoseismology record for the Prince William Sound area is recorded in the Copper River delta sediment, where 11 megathrust earthquakes over the past 6,000 yrs were documented (Carver and Plafker, 2008). Based on this record, estimates of the recurrence interval for this part of the Aleutian-Alaska megathrust vary from 333 to 875 yrs. In the Shumagin Islands, Winslow and Johnson (1989) show evidence for uplifted Holocene terraces that yield a clustering of radiocarbon ages that correlate inversely with native habitation in the Shumagin Islands. This suggests that following times of terrace uplift (in other words, earthquakes), the natives abandoned the islands. Few other studies have been conducted on the Aleutian Islands. Beginning in the summer of 2010, the USGS has been conducting paleoseismology and paleotsunami studies on islands southwest of Kodiak, including Chirikof, Semidi, and the Shumagin Islands (fig. 4*B*) to better constrain the rupture history of the eastern Aleutian-Alaska megathrust (A. Nelson, oral commun., 2010).

Alongstrike Variations in Plate Coupling

Global Positioning System Data

Geodetic data are often used to study interseismic strain accumulation and to estimate the slip deficit at a plate interface. A fully coupled or locked plate interface will show a slip deficit that is equal to the convergence rate times the time elapsed since the last event. If the slip deficit is less than this, the plate interface is either only partially locked (only discrete patches of the plate interface are locked) or is creeping at less than the convergence rate. Particularly in the Aleutian Islands, Global Positioning System (GPS) data are limited by the unavailability of stations near the trench (most islands are at least 150 km from the trench), harsh weather conditions, remoteness of sites that make reoccupation difficult, and the presence of active volcanoes near many of the sites, which can obscure the signal at the plate interface. Despite these difficulties, Freymueller and others (2008) provide constraints on the magnitude of slip deficit along much of the Aleutian-Alaska megathrust.

We summarize the geodetic results as outlined in Freymueller and others (2008) for the Aleutian-Alaska megathrust between Kodiak Island and the Andreanof Islands (fig. 11). The GPS data were collected for the time periods 1993–1996 and 2003–2005. GPS data show considerable alongstrike variability in the degree of plate coupling from the Semidi Islands in the east to the Andreanof Islands in the west (fig. 11). In the

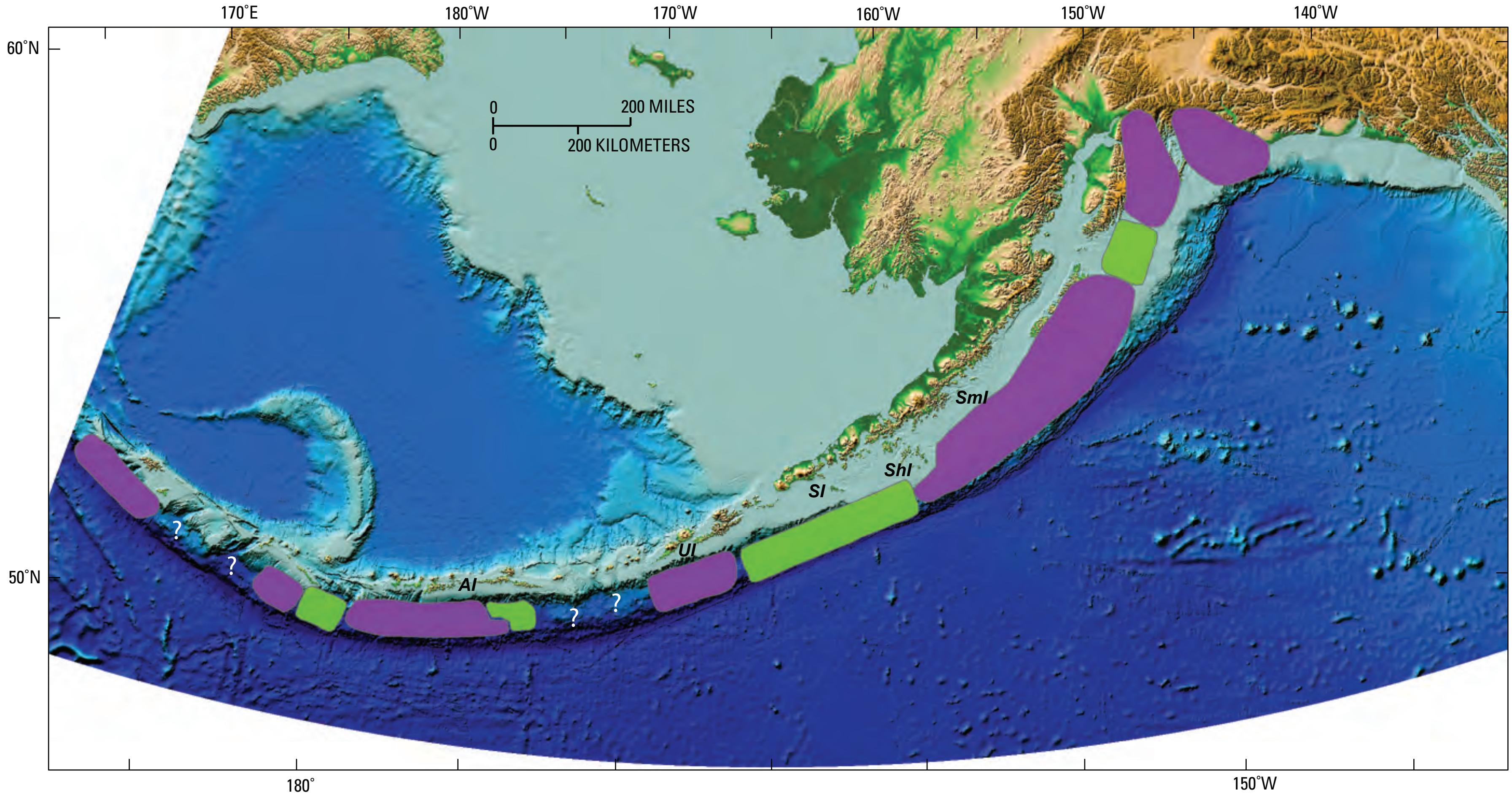

Figure 11. Areas of locked (purple) versus creeping (green) regions of the plate interface along the Aleutian-Alaska arc, based on Global Positioning System (GPS) and seismic rupture data (modified from Freymueller and others, 2008). Question marks denote those areas where data are insufficient to assess the degree of plate coupling. From the Semidi Islands (SmI) (about 80 percent coupled) west to Sanak Island (SI) (nearly 0 percent coupled), GPS data show an overall decrease in the slip deficit, with a transition from locked to creeping sections near the Shumagin Islands (ShI).

area of the Semidi Islands, which includes the 1938 rupture area, the plate interface is 70 to 90 percent coupled. From the Semidi Islands west to Sanak Island, GPS data show an overall decrease in the slip deficit. There is a transition from a highly coupled zone to a mainly creeping zone occurring over a small area around the Shumagin Islands (Fournier and Freymueller, 2007). East of the Shumagin Islands, the coupling is on the order of 70 percent, with coupling in the area of the Shumagin Islands themselves at about 30 percent (Cross and Freymueller, 2008). These more recent studies confirm earlier studies in the Shumagin Islands that showed a lack of measurable strain accumulation (Savage and Lisowski, 1986; Lisowski and others, 1988; Freymueller and Beavan, 1999). Leveling data acquired between 1977 and 1988 did show a time-variable ground surface tilt in the Shumagins (Beavan and others, 1983, 1984).

There is little trench-perpendicular strain measured at Sanak Island, with the only possibility of a locked zone at very shallow depths (~11 km) near the trench (Fournier and Freymueller, 2007). West of Sanak Island, most of the GPS stations are located on the volcanic arc; however, most stations located away from the influence of volcanism show little slip deficit. GPS data indicate that Unimak Island, in the area of the 1946 M_w 8.6 earthquake, is poorly coupled, with the exception of the station located at the western end of Unimak Island at Scotch Cap (Mann and Freymueller, 2003). However, it is difficult to determine the degree of coupling from GPS data collected on the islands if the width of the coupling zone is relatively narrow (48 km or less). This is because for a narrow width, a completely unlocked interface cannot be differentiated from one that is completely locked using GPS data (Mann and Freymueller, 2003). No geodetic data are available for the section of the arc from Unimak Island to 150 km to the west, an area large enough to potentially include a locked plate interface.

The amount of strain accumulation in the Fox Islands section of the AASZ is not well documented, because most GPS stations are located to monitor volcanoes and not the megathrust boundary. There are two areas, however, where GPS data suggest that the plates are at least partially locked beneath the Fox Islands. A station located at the western end of Umnak Island indicates partial coupling along the megathrust at mid (14–30 km) to deep (30–47 km) depths (Cross and Freymueller, 2008). In addition, offshore of Unalaska Island, the plate interface could be coupled at the shallowest (7–14 km) updip end of the megathrust (Cross and Freymueller, 2008). There are no GPS stations located between Atka and Umnak Islands, and thus the state of coupling of the plate interface in this area is unknown.

In the eastern Andreanof Islands, Atka Island stations show low slip deficit, indicating that the plate interface is weakly coupled. Conversely, the rupture zone of the 1986 Andreanof earthquake, which coincides with the area of highest moment release during the 1957 Andreanof earthquake offshore of Adak Island (Hwang and Kanamori, 1986) is now nearly 100 percent locked (Cross and Freymueller, 2007). Therefore, the eastern section of the 1957 rupture zone appears to be creeping, with strain accumulation occurring in the west.

Forearc Basins

Areas of high moment release during great subduction zone earthquakes have a strong tendency to occur beneath gravity lows centered on forearc basins, with the arcward edge of the high-slip zone aligned with a strong gravity gradient (Wells and others, 2003; Song and Simons, 2003; Llenos and McGuire, 2007). Forearc basins are long-lived features that record the average stress on the plate interface over time scales on the order of a million years. If interseismic subsidence is not completely recovered during an earthquake, a basin could form above the region of strong coupling (Wells and others, 2003). Alternatively, the location of a forearc basin may reflect an increase in shear traction and subduction erosion at locations where interplate coupling is strongest (Wells and others, 2003; Song and Simons, 2003). Thus, the locations of forearc basins along a subducting margin may be coincident with areas of high strain release during earthquakes.

Wells and others (2003) use satellite gravity data to correlate the locations of forearc basin depocenters with areas of high moment release during great earthquakes along circum-Pacific subduction zones. Along the Aleutian-Alaska subduction zone (AASZ), the locations of forearc basins vary depending on whether a basin is built upon older continental crust (east of Unimak Pass) or upon oceanic crust. Forearc basins in the west underlie the Aleutian Terrace (water depth 3,500–4,200 m); in the east they occur in shallow water beneath the broad Alaskan continental shelf (fig. 12). The major forearc basin depocenters located in the Alaska Peninsula-eastern Aleutian forearc include Tugidak, Shumagin, Sanak, Unimak, and Unalaska Basins. Atka Basin, a relatively large forearc basin, is located in the central Aleutians (fig. 12).

Tugidak and Shumagin Basins are the largest of the forearc basins beneath the shelf west of Kodiak Island (fig. 12). A transverse structural high associated with the Semidi Islands separates the two basins. During the great 1938 earthquake, the maximum moment release occurred beneath Tugidak Basin, with additional moment released beneath Shumagin Basin (Wells and others, 2003). Between the Shumagin Islands and Unimak Pass at the continental-oceanic crust transition, the forearc basin configuration is complex and includes three depocenters: Sanak Basin, west Sanak Basin, and Unimak Basin (fig. 12). Sanak Basin is a graben oriented transverse to the margin and contains as much as 8 km of sediment (Bruns and others, 1987; von Huene, 1989). According to von Huene (1989), these margin-oblique basins formed along structures inherited from the old Beringian margin, although Bruns and others (1987) suggest that Unimak Basin formed as the result of subduction erosion. Moment released during the great 1946 earthquake, which was located at the east end of the Aleutian Terrace, was at least partially coincident with Unimak Basin (Wells and others, 2003).

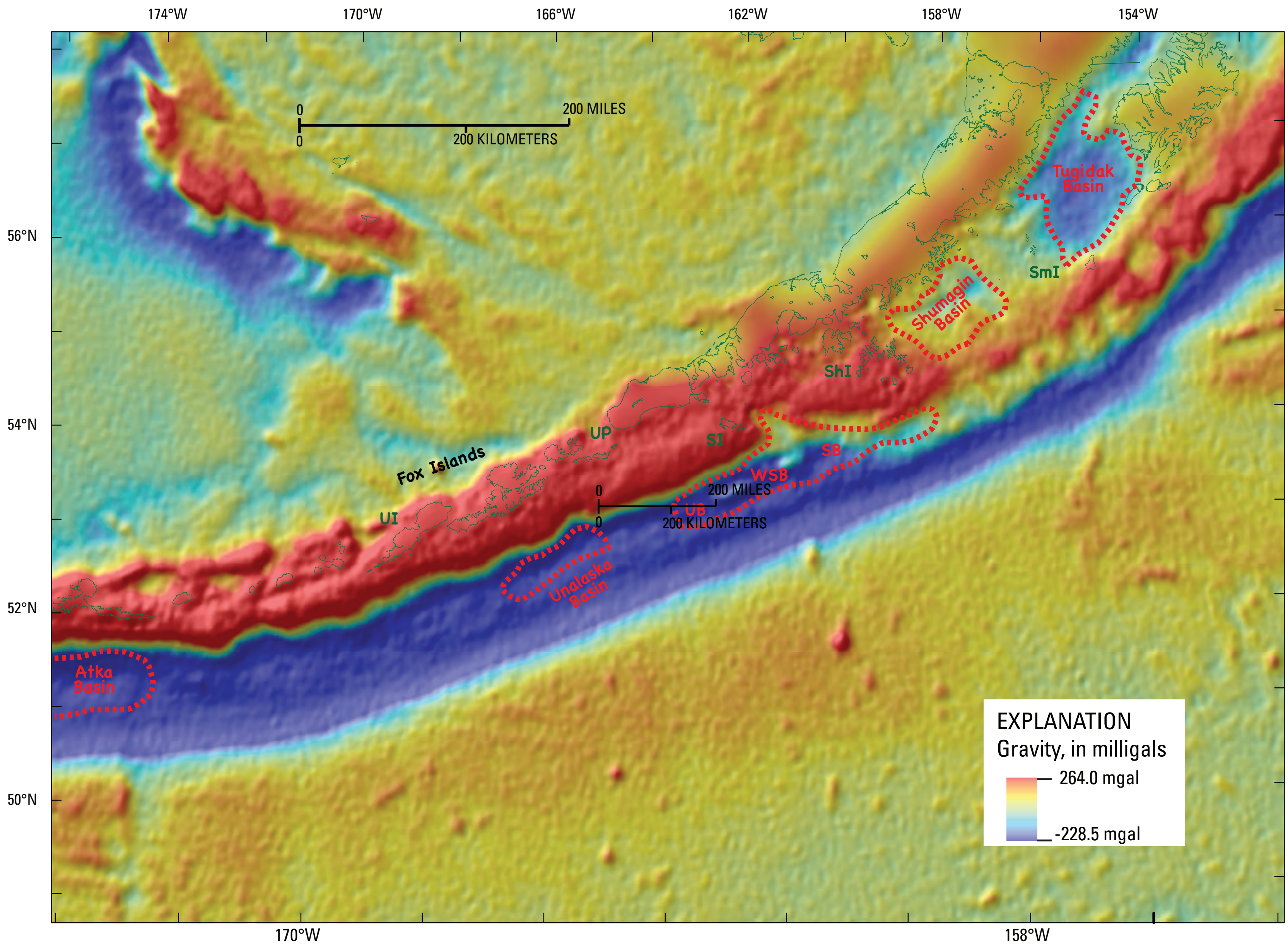

Figure 12. Satellite gravity data for the region of the eastern Aleutian-Alaska arc, showing locations (red dashed outlines) of main forearc basin depocenters, as indicated by gravity lows. These locations are modified from Wells and others (2003). UI, Umnak Island; UP, Unimak Pass; SI, Sanak Island; ShI, Shumagin Islands; SmI, Semidi Islands; UB, Unimak Basin; WSB, West Sanak Basin; SB, Sanak Basin. Gravity data are from Sandwell and Smith (1997) and were downloaded from version 16 (March 2007) at http://topex.ucsd.edu/WWW_html/mar_grav.html.

West of Unimak Pass, the Aleutian Terrace forms a broad gravity low with several forearc basin depocenters located beneath the terrace (fig. 12). Geophysical data are sparse offshore of the Fox Islands, and thus not much is known about the size and location of forearc basin depocenters in this area. Wells and others (2003) show localized gravity lows associated with the 175-km-long Unalaska Basin (fig. 12) and with a small basin off of Umnak Island (not shown in fig. 12). The megathrust beneath Unalaska Basin did not rupture during the 1957 Andreanof earthquake and thus may be within a seismic gap (Boyd and Jacob, 1986). In the Andreanof Islands, Atka Basin is located east of where most of the moment was released during the 1957 earthquake; GPS data indicate that the megathrust beneath Atka Basin is only partially coupled (Freymueller and others, 2008).

Fracture Zones, Ridges, Seamounts

An understanding of what controls the alongstrike limits of earthquakes is important to determining rupture length. Features on the subducting plate such as fracture zones, ridges, and seamount chains have been both barriers to rupture and locations of rupture nucleation. In some instances, features on the subducting plate correlate with patterns of aftershocks associated with great earthquakes. Major fracture zones within the subducting Pacific Plate that intersect the Aleutian Trench include the Amlia and Aja Fracture Zones (fig. 7). In addition to the fracture zones, the Kodiak-Bowie (KB) and Patton-Murray (PM) seamount chains are prominent bathymetric features that extend across the Gulf of Alaska and are subducted at the trench near Kodiak Island (figs. 7 and 8). Seamounts are also observed near the trench offshore of Sanak Island (fig. 8).

Although the Amlia Fracture Zone has only moderate bathymetric expression (fig. 4*A*), it is considered to be a major segment boundary along the Aleutian-Alaska subduction zone (Lu and Wyss, 1996). Focal-mechanism stress inversions indicate that there are major changes in stress directions across the fracture zone at the trench (Lu and Wyss, 1996). The fracture zone separates a rougher, slightly shallower Pacific Plate with more numerous bathymetric features to the west from a somewhat smoother Pacific Plate to the east (fig. 8). Magnetic anomalies are offset more than 4200 km in a left-lateral sense across the fracture zone (fig. 7). Near the Amlia Fracture Zone, there is an offset in the volcanic arc, a paucity of moderate earthquakes, and a 40-km right-lateral offset of the Wadati-Benioff zone by either a tear or a flexure in the subducting plate (House and Jacob, 1983). Although the Amlia Fracture Zone apparently did not form a barrier to the 1957 Andreanof earthquake, few aftershocks occurred and little moment was released east of the fracture zone. The fracture zone sweeps along the arc at a rate of more than 2 cm/yr in response to oblique convergence between the Pacific and North American Plates (DeMets and others, 2010).

The Aja Fracture Zone intersects the trench near the southern end of Kodiak Island (fig. 7). Although not as pronounced as the Amlia Fracture Zone, the Aja Fracture Zone is well expressed in both magnetic anomalies and bathymetry. Where the fracture zone intersects the continental margin, there are huge slide scars coincident with an old structural boundary in the upper plate, and a broad zone of disruption is suspected from satellite bathymetry (von Huene and others, 1999). Just south of the Aja Fracture Zone, the Patton–Murray Ridge intersects the trench axis. Although no pronounced seamounts mark the crest of the ridge here, subduction of the Patton–Murray hot spot swell corresponds with a change in axial gradient of the trench and a thinning of sediment in the trench (von Huene and others, 1999). The Aja Fracture Zone, together with the Patton-Murray Ridge, formed the southwestern barrier to rupture during the 1964 Alaskan earthquake and the northeastern boundary of the 1938 earthquake (von Huene and others, 1999). The southwestern barrier of the 1964 rupture is comparable to the Juan Fernandez barrier that ended the rupture of the 2010 M_w 8.8 Chilean earthquake (von Huene and others, 2012). Other features may influence seismicity at the plate interface. One of these is Unimak Ridge, a ~100-km-long ridge along the outer edge of the Aleutian Terrace offshore of Unimak Island (fig. 4*B*) that is composed of igneous arc basement. Although it is unclear how the ridge might influence seismicity, the aftershocks of the 1946 great earthquake tended to occur along that ridge (Bruns and others, 1987; von Huene and others, 2012).

Thickness of Trench Sediment

A commonality of great subduction zone earthquakes is that many are generated along convergent margins where a blanket of trench sediment smooths the roughness of the subducting plate (Ruff, 1989). Reducing the roughness of the subducting plate could remove possible barriers to earthquake propagation and thus increase the magnitude of an earthquake. As the result of late Cenozoic mountain building and Quaternary glaciation, thick sedimentary deposits occur on the northeast Pacific Plate (Gulick and others, 2004). These deposits include a large sediment load that was transported westward down the west-dipping axis of the Aleutian Trench, reaching a maximum sediment thickness at the Amlia Fracture Zone (fig. 13). Because of the relatively steeper westward dipping bathymetric gradient along the trench axis between the Shumagin and Fox Islands, turbidity currents bringing sediment down the trench axis are less likely to deposit their load in this section of the trench, resulting in little to no trench sediment accumulation (fig. 13).

In addition to sediment transported down the axis of the Aleutian Trench, large deep-sea submarine fans have also been deposited on the deep seafloor adjacent to the Gulf of Alaska. Along the eastern Aleutian arc, this includes the Zodiac deep-sea fan (Stevenson and Embley, 1987). The Zodiac fan is as much as 1.6 km thick and contains a volume of sediment greater than that of the Amazon fan (Stevenson and Embley, 1987); it enters the trench at a point just to the northeast of the

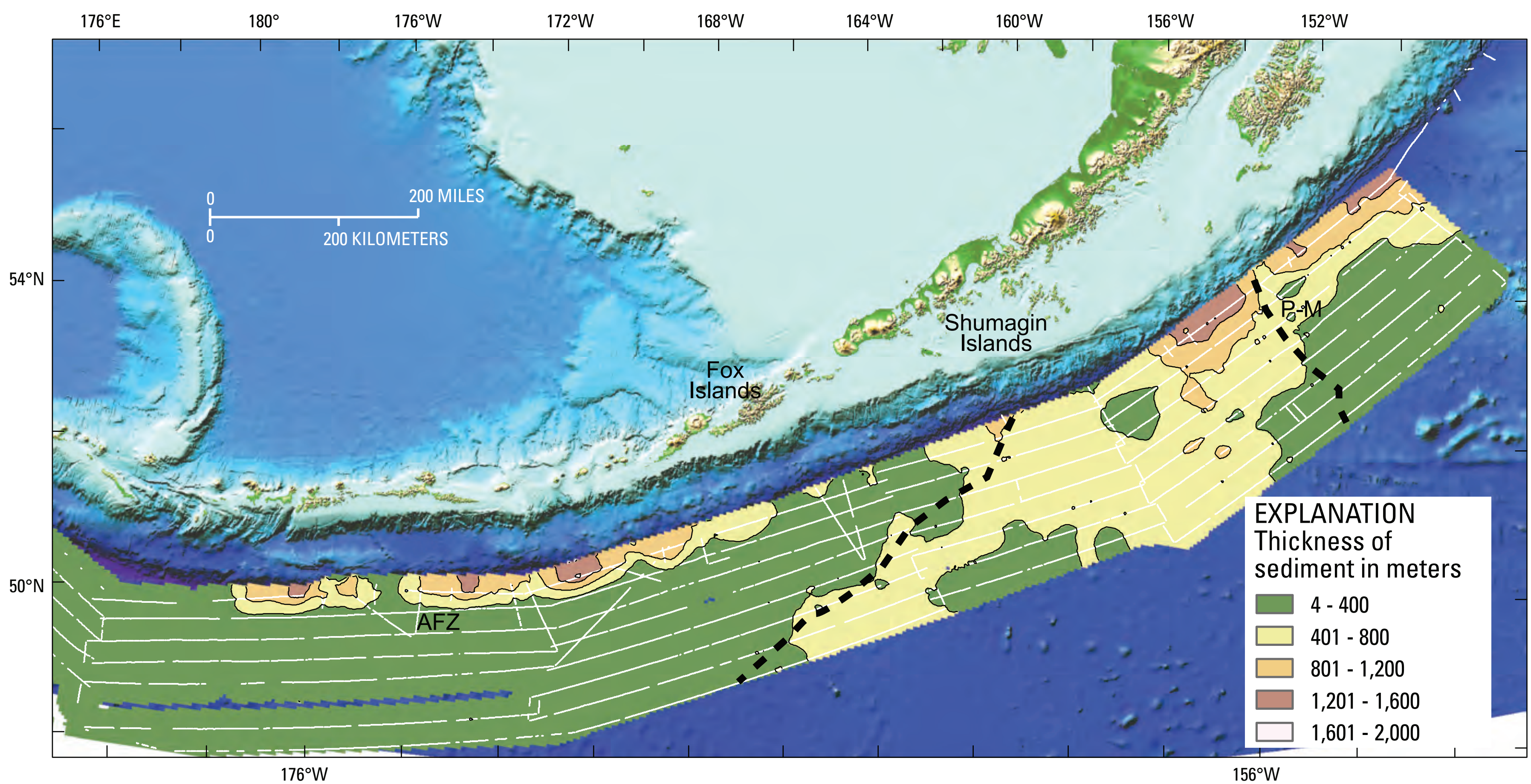

Figure 13. Thickness of sediment on subducting Pacific Plate adjacent to the Aleutian Trench. Thicknesses are in meters, assuming a seismic velocity in the sediment of 2,000 m/s. Interpretations are based on multichannel and single-channel seismic reflection profiles available at http://walrus.wr.usgs.gov/NAMSS/. Lines of white dots show location of data; the area where there are no white dots (shows as jagged blue line) is the result of missing data. P-M, Patton–Murray ridge near its intersection with the trench; AFZ, approximate location of Amlia Fracture Zone. Dashed lines show approximate margins of the Zodiac fan, modified from von Huene and others (2012).

Shumagin Islands (fig. 13). Zodiac fan sediment can be traced beneath the landward trench slope, where it underlies the area from northeast of the Shumagin Islands to Chirikof Island (fig. 14; von Huene and others, 2012). The aftershock zone of the great 1938 earthquake corresponds to the area of the subducting Zodiac fan, which has been named the Semidi asperity (von Huene and others, 2012).

Although the presence of roughness-smoothing sediment in the trench is conducive to the generation of great earthquakes, sediment accreted to form a frontal prism immediately seaward of the trench is generally composed of relatively weak, porous, and water-retaining sediment, which can impede fault rupture propagation (Byrne and others, 1988). At temperatures of about 150°C, diagenetic and low-grade metamorphic changes, including the phase change of clays from stable sliding smectite to illite and chlorite, can result in velocity weakening, a process whereby increasing displacement velocity results in decreasing friction along the fault surface, which promotes stick-slip behavior (earthquakes) within the frontal prism (Vrojlik, 1990; Moore and Saffer, 2001). Thus, along a subduction zone with a large frontal prism, the updip limit of the seismogenic zone generally occurs where temperatures reach about 150°C.

In order to estimate the updip location of a potential rupture in the eastern Aleutians, we identified the arcward limit of the frontal prism using available single and multichannel seismic reflection profiles (fig. 14). We mapped the updip limit at the pronounced break in slope that occurs at the contact between the frontal prism and the rocks of the arc basement, which are able to maintain a steeper slope. In addition, we often were able to image a velocity pullup, in which the two-way travel time makes the top of the downgoing plate appear to be shallower than expected as it subducts beneath higher velocity framework rocks. As a result of the sediment bypass along this part of the trench between the Fox and Shumagin Islands (fig. 14), the width of the frontal accretionary prism is relatively narrow, with arc basement rock generally located less than 15 km from the trench deformation front. The lack of a significant frontal prism, combined with evidence for collapse of the outer Aleutian shelf, indicates that the eastern Aleutian-Alaska subduction zone is not an accretionary margin, but is primarily erosive (von Huene and others, 2012). This has important implications for the generation of tsunami earthquakes, which are shown to preferentially occur along erosive margins where competent bedrock occurs relatively close to the trench (Bilek, 2010).

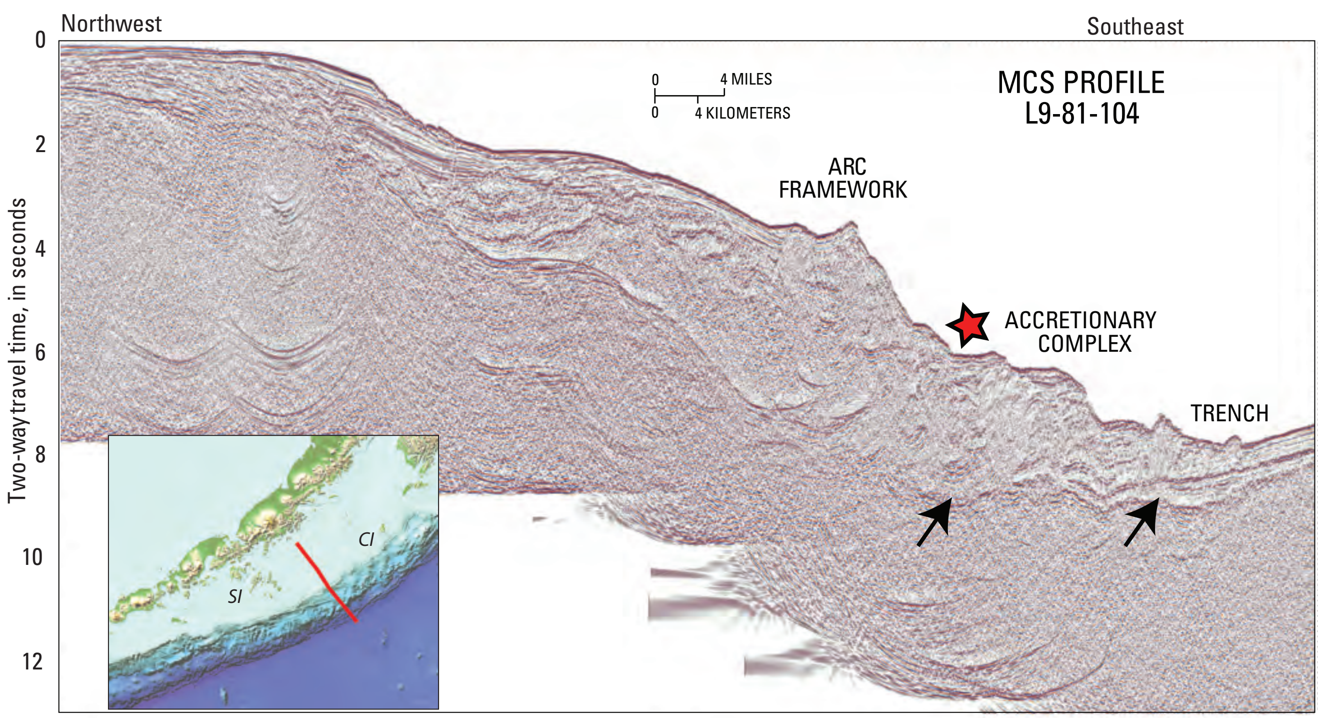

Figure 14. Example of a multichannel seismic (MCS) reflection profile (L9-81-104) that shows the relation between the frontal accretionary prism and the arc basement at the trench. The updip limit of a megathrust rupture zone is generally placed at the contact between an accretionary frontal prism, composed of relatively weak, porous, and water-retaining sediment, and stronger arc basement rock. The arcward limit of the frontal prism (denoted by the red star in this profile) was determined using both single-channel and multichannel seismic reflection profiles. A pronounced break in slope (at star) is at the contact between the weak sediment of the accretionary prism and the arc framework rocks, which are able to maintain a steeper slope. Arrowtips mark the top of subducting plate. The vertical exaggeration is about 6.7:1. Inset shows location of profile between Shumagin Islands (SI) and Chirikof Island (CI).

Discussion

A rupture length along a subduction margin on the order of 800 km is necessary to produce an *M*9 or greater earthquake (McCaffrey, 2007). It is these long ruptures that potentially generate tsunamis capable of traveling great distances with little loss of amplitude. For the eastern Aleutian-Alaska subduction zone, McCaffrey (2008) suggests that the maximum possible earthquake is M_w 9.3, with a possible slip range of 16–38 m. In order to generate an earthquake and tsunami in the eastern Aleutians of such a high magnitude, multiple strongly coupled patches (asperities) of the megathrust need to be ruptured. In addition, the earthquake would have to rupture across multiple locked patches of the plate interface separated by creeping or partially coupled patches. The absence of barriers large enough to inhibit rupture between strongly coupled sections of the plate interface in the Aleutians enhances the likelihood of generating such an earthquake there.

In order to generate a giant earthquake (M_w~9) along the eastern Aleutian-Alaska megathrust, the rupture area must be greater than what has been observed during 20th century earthquakes that occurred west of Kodiak Island. In a "characteristic earthquake" model, seismic moment is released in repeated occurrences of earthquakes of similar magnitudes, and that magnitude is the largest that would be expected on that section of the fault. Subduction zones, however, tend to have rupture histories that are distinctly noncharacteristic (Schwartz, 1999). In particular, characteristic earthquake models are clearly questionable for the Aleutian-Alaska subduction zone (Wesson and others, 2007). An oft-cited example for noncharacteristic behavior of ruptures in the Aleutians is the comparison of the 1957 *M*8.6 and 1986 *M*7.9 earthquakes along the megathrust offshore the Andreanof Islands (Boyd and others, 1995; Schwartz, 1999). Although the epicenters of the 1957 and 1986 events were nearly coincident, there were pronounced differences in the distribution of slip and the size of the aftershock zone, suggesting that earthquakes do not always rupture in similar patterns from cycle to cycle.

Part of the problem in identifying how the plate interface ruptures from seismic cycle to seismic cycle is the lack of historical observations. Paleoseismology studies can provide information that allows extension of that record much further back. Although few paleoseismology studies have been conducted west of Kodiak Island, evidence from paleoseismology studies along the Aleutian-Alaska megathrust east of Kodiak show variability in the timing of past ruptures. The 1964 Alaskan earthquake is the 2nd largest recorded worldwide in historical times. However, paleoseismology studies show that the previous great earthquake in this area, which occurred about 900 yrs ago, was approximately 15 percent larger, rupturing the portion of the megathrust from Kodiak Island east to the Yakutat sector in the far eastern Gulf of Alaska (Shannan and others, 2009). This illustrates the variation in rupture mode from earthquake cycle to earthquake cycle, with some earthquakes involving the rupture of single segments and some involving multiple segments. Future paleoseismology studies in the Fox, Shumagin, and Semidi Islands will be critical for deciphering the variability of past earthquake ruptures and their magnitudes to a greater extent than has been possible from historical observations.

One possible scenario for generating a giant earthquake on the eastern Aleutian-Alaska megathrust is to include rupture beneath the Shumagin and Sanak Islands. The rupture of this section of the megathrust is problematic because GPS data indicate that it has a low degree of plate coupling and hence might not be expected to rupture (Freymueller and others, 2008, and references therein). GPS stations that are closest to the trench and thus most likely to record evidence for strain accumulation on the megathrust include sites on Chirikof, Shumagin, and Sanak Islands; most of the other stations are located of necessity on the volcanic arc, where the signal can be contaminated by deformation related to volcanism. As a result, the degree of coupling at the updip end of the megathrust is often poorly determined, with a wide range of coupling allowable within the 95-percent confidence limits (Cross and Freymueller, 2008). In addition, the GPS record may not be long enough to record temporal variability in the locking plate interface (that is, the low coupling may be transient) (Freymueller, 2010).

Although the GPS record indicates that the Sanak and (or) Shumagin Islands sections of the megathrust are not presently accumulating strain, the historical earthquake record indicates that in the past, earthquakes of *M*>8 occurred along these areas. Earthquakes of *M*>8 that occurred in 1788 and 1946 ruptured parts of the megathrust that are not currently accumulating any strain, although the rupture zone of the 1788 earthquake is not well constrained. The earthquake in 1788 is inferred to have ruptured through the plate interface beneath the Shumagin Islands, which is evidence that this section of the subduction zone has been at least partially coupled in the past. However, since the main evidence for the 1788 earthquake is derived from the tsunami that it generated, the tsunami could have been caused by a local event such as a landslide or offset along an upper plate splay fault and not necessarily by the rupture of an *M*>8 earthquake on the megathrust (Savage and Lisowski, 1986). Although the mechanism for generating the tsunami in the 1946 earthquake is still not fully resolved, the earthquake is considered to have been a slow "tsunami" earthquake that ruptured the updip portion of the megathrust (Lopez and Okal, 2006). The updip portion of the megathrust is generally not well resolved using GPS data, and thus there could be a locked portion of the megathrust near the trench.

Geodetic studies along other subduction zones have also documented a variation of locking at a plate interface with time (Arnadottir and others, 1999; Burgmann and others, 2005; Prawirodirdjo and others, 2010). As Burgmann and others (2005) noted for their studies in Kamchatka, geodetic data spanning only a few years do not necessarily predict the eventual degree of seismic coupling. Geodetic data collected before and after the M_w 9.1 Sumatra-Andaman earthquake document a time-transient GPS signal along the Sumatra megathrust (Prawirodirdjo and others, 2010). Before the 2004 earthquake, geodetic studies showed that the area off northern Sumatra that ruptured in 2004 was only partially coupled; this area was not expected to be the likely site for a giant earthquake (Prawirodirdjo and others, 1997). Conversely, one of the most strongly coupled areas located further south along the Sumatra megathrust, which was thought to be the more likely site of an earthquake before 2004, has since been observed to be uncoupled and slipping freely.

An alternate possibility that would allow the incorporation of the areas of low strain accumulation into a possible earthquake rupture area is to consider that the creeping sections of the plate interface are not large enough to inhibit stress transfer. Areas of low slip deficit do not necessarily form barriers to rupture. Perfettini and others (2010) show that the distance between the locked patches is the critical factor in determining whether an interplate earthquake will propagate through an intervening aseismic area. Creeping patches along the main plate boundary will not necessarily inhibit the alongstrike propagation of an earthquake rupture as long as stress can be transferred across the creeping zone (the creeping zone being relatively small). Some larger areas, however, always slip aseismically, and these would form a barrier to any earthquake rupture propagation.

Considering the GPS record along the Aleutian-Alaska megathrust (Freymueller and others, 2008), there are two freely slipping sections of the megathrust that did not inhibit the rupture of a great earthquake, assuming that the current state of stress is similar to what it was at the time of the earthquake. One of these freely slipping sections, which did not form a barrier to the 1957 Andreanof earthquake, is off Atka Island; the other section, which did not form a barrier to rupture during the giant 1964 Alaska earthquake, lies between Kodiak Island and Prince William Sound (fig. 11). These creeping sections have alongstrike lengths on the order of 100–150 km. It is important to note that those areas that showed the most moment release during the 1957 and 1964 earthquakes are also presently nearly fully coupled (Freymueller and others, 2008). Therefore, it is possible that an earthquake could rupture through relatively narrow (<150 km) poorly coupled or uncoupled sections of the megathrust.

Nischenko and Jacob (1990) suggest that although a future great earthquake may not necessarily nucleate in the Shumagin "gap" region, it could rupture through that area, with most of the moment released in adjacent areas. If sections of the megathrust adjacent to the Shumagin section are locked and near failure, an earthquake of a more moderate magnitude located beneath the Shumagins could trigger rupture in adjacent areas. The Shumagin Islands region displays a pronounced gravity high that extends out toward the trench. This gravity high is associated with trench-normal anticlines that might be more resistant to bending during elastic loading (Wells and others, 2003). In addition, the subducting plate is warped, as evidenced by changes in dip (fig. 6), and there is a discontinuity in moderate-depth seismicity (Hudnut and Taber, 1987). Thus the complexities inherent to the megathrust in the Shumagin Islands region may form a local asperity capable of nucleating an earthquake that could then propagate into adjacent segments of the arc, resulting in a giant earthquake.

Lessons Learned from Recent Subduction Zone Earthquakes

If anything has been learned from the recent giant (M_w>9) earthquakes and devastating tsunamis in the Sumatra–Andaman Islands (2004), Chile (2010), and Japan (2011), it is that we still do not fully understand the conditions that lead to these unusually large and devastating events. In particular, the Sumatra-Andaman and Japan earthquakes each occurred along segments of a megathrust that were not considered to be likely to rupture in earthquakes of that magnitude. This conclusion was based on both the physical parameters that characterize these subduction zones and the historical earthquake record (Subarya and others, 2006; Kerr, 2011). Past studies have linked earthquake magnitude with such factors as plate age, convergence rate, and heterogeneities on the subducting plate (Ruff and Kanamori, 1983). In general, strong seismic coupling is presumed to occur on margins where the subducting plate is young and is converging at a fast rate, and where sediment has blanketed and smoothed bathymetric features on the subducting plate (Ruff, 1989). Many of these physical controls that were thought to contribute to earthquake magnitude, however, were not present along the rupture zones of these most recent giant earthquakes.

2004 Sumatra–Andaman Earthquake

No one could have foreseen the magnitude or complexity of the devastating M_w 9.1 earthquake and deadly tsunami that were generated along the Sumatra-Andaman trench on December 23, 2004 (Bilham, 2005). This is the 3rd largest earthquake recorded during modern times, despite the fact that there were no previous historical records of giant earthquakes in the area that ruptured between Sumatra and Myanmar (Subarya and others, 2006). In terms of the Sumatra-Andaman subduction zone, this sector of the arc was considered the least likely to rupture (Subarya and others, 2006; McCaffrey, 2007). That plate boundary is experiencing slow subduction of an old plate accompanied by back-arc extension, physical conditions considered unlikely to lead to the generation of great earthquakes, according to the seismic coupling model of Ruff and Kanamori (1983). Thus, the earthquake was larger than predicted on the basis of this prior understanding of the physical parameters indicative of the degree of seismic coupling (Stein and Okal, 2007).

The Sumatra-Andaman earthquake's rupture area was unusually long (~1,300 km) but relatively narrow

(100–150 km), when compared to subduction zone earthquakes along continental margins (Geist and others, 2007). The rupture initiated near a bend in the subducting plate and was multisegmented. Slip was concentrated at three main asperities, with slips as high as 30 m, separated by areas that slipped aseismically; most of the slip occurred updip of the locations of previous earthquakes (Chlieh and others, 2007). Although the presence of weak sediment accreted near the trench is generally thought to impede updip rupture in heavily sedimented subduction zones, Gulick and others (2011) provide evidence that the sediment on this subducting plate was dewatered and compacted and thus strong enough for stick-slip behavior unusually close to the trench in 2004.

2010 Maule, Chile, Earthquake

On February 27, 2010, an M_w 8.8 earthquake struck south-central Chile—the 6th largest earthquake in the world in the past hundred years. The earthquake ruptured an area about 650 by 180 km, with a maximum slip on the order of 15–20 m (Delouis and others, 2010, Pollitz and others, 2011; Vigny and others, 2011). The earthquake nucleated near the coast and ruptured bilaterally, with two main asperities separated by little slip in the area of the epicenter. Little slip occurred at the updip end of the megathrust near the trench, resulting in a far-field tsunami that was not as large as might be expected for an earthquake of this magnitude.

The 2010 Maule earthquake has called into question the validity of the seismic gap model and also the assumption that fault segments that have built up the most stress over the longest time periods will be the next to rupture (Lorito and others, 2011). Although the Maule earthquake nucleated within a seismic gap (the Darwin gap), little strain was released within the gap, and the area with the highest pre-earthquake coupling remained unbroken. Furthermore, the area of maximum slip occurred where stress had been released by an earthquake just 80 yrs ago in 1928 (Lorito and others, 2011). Thus, the slip distribution of the Maule earthquake did not match what was expected on the basis of the areas of high pre-earthquake coupling. The seismic gap hypothesis was also violated, because the 2010 earthquake rupture zone overlapped the rupture zones of both the 1960 M_w 9.5 and 1906 M_w 8.5 earthquakes—segments of the megathrust that had already ruptured in the not too distant past (Vigny and others, 2011).

2011 Tohoku-Oki, Japan, Earthquake

We are just beginning to learn about the rupture dynamics of the devastating M_w 9.0 earthquake and tsunami generated offshore northern Japan on March 11, 2011. Because there is a denser network of seismometers in Japan than anywhere else in the world, this event is expected to be one of the best studied giant earthquakes in the modern instrument era. At this point, what is known about this earthquake is that it ruptured an area that was expected to produce large, but not giant earthquakes (Kerr, 2011). Along the northern Japan margin, old (140 Ma) oceanic crust with little sediment cover is subducting at the trench; these physical conditions had been thought not to be conducive to generating giant earthquakes at subduction zones. As a result, the apparent lack of seismicity in the area of high slip before the earthquake was thought to be the result of slow, steady slip without the accumulation of strain (Kerr, 2011).

What is perhaps the most surprising aspect of the Japan earthquake is the enormous amount of slip that occurred during the earthquake—slip that may have been more than 50 m at some locations (Simons and others, 2011; Lay and others, 2011). Geodetic stations deployed on the seafloor near the epicenter of the earthquake recorded a maximum slip of 24 m, with horizontal movements more than four times that recorded on land (Sato and others, 2011). Before the earthquake, onland geodetic data had indicated partial to no coupling of the megathrust near the trench, illustrating the difficulty in using land-based stations to evaluate the degree of plate coupling at the updip end of the megathrust (Simons and others, 2011). A key aspect that contributed to the large tsunami that struck Japan was that when the updip end of the rupture propagated towards the seafloor, an unusually large amount of slip occurred on the megathrust (Ide and others, 2011). The 2011 Japan earthquake is thought to have involved a compound frequency-dependent rupture that included a conventional rupture at the downdip end of the megathrust and slow rupture (a tsunami earthquake) at the updip end (Koper and others, 2011).

The subduction zones off Chile and Japan are two of best studied in the world. Yet, in these areas and in the Sumatra-Andaman area, devastating earthquakes and tsunamis did not occur either along the segment or at the magnitude that was predicted by previous studies. Although an understanding of the characteristics of a subduction zone that are conducive to the rupture of a giant earthquake may indicate which subduction zones are more likely to spawn a teletsunami, capable of traveling long distances across oceans, such understanding may not be indicative of where on the subduction zone it is possible to generate a teletsunami. Stein and Okal (2007) suggest that the historical earthquake record is much too short to sample the variability of slip at a given subduction zone. Thus, one should not dismiss any subduction zone segment as not being able to generate a potentially devastating tsunami. As a case in point, recent studies in New Zealand suggest that "there could be any number of potential rupture segments and sizes that rupture individually or cascade into larger ruptures depending on initial stress conditions, nucleation and dynamic rupture properties" (Wallace and others, 2009).

While there is no evidence that a rupture of the Aleutian-Alaska megathrust from the Fox Islands to west of Kodiak Island occurred within the short historical records available, we consider that this is a viable event that could happen in the future. Paleotsunami and paleoseismology studies of the megathrust that could extend the historical earthquake record back in time are needed in order to be able to evaluate the rupture scenario. These studies are lacking for most of the

Aleutian-Alaska subduction zone. In addition, few studies to identify evidence for prehistoric far-field tsunamis in California have been conducted. Such studies are critical to finding evidence for rare giant events, as illustrated by paleotsunami studies in both Thailand (Jankaew and others, 2008) and Japan (Minoura and others, 2001). Both those studies showed evidence for tsunami inundation much greater than was expected on the basis of the historical record. In Thailand, paleotsunami evidence indicates a previous event, 550–700 yrs ago of similar magnitude to the 2004 Indian Ocean tsunami (Jankaew and others, 2008). This time-variable behavior of tsunami-generating earthquakes is confirmed by coral uplift patterns along the Sumatra-Andaman subduction zone that indicate that rates of interseismic coupling over the past 50 yrs differ from those of the past 300 yrs (Meltzner and others, 2010). Although the studies in Sumatra were conducted after the 2004 Sumatra-Andaman earthquake, paleotsunami studies in Japan prior to the 2011 event suggested that the area affected by the 2011 earthquake experienced a large tsunamigenic earthquake in 869 C.E. (Minoura and others, 2001).

Conclusions

In the aftermath of both the Sumatra-Andaman and Japanese devastating earthquakes and tsunamis, we review existing studies to evaluate whether a similar tsunami could strike the coast of southern California. The historical record does not support the likelihood for a tsunami to be generated locally from either an earthquake or landslide in the California Continental Borderland. However, a tsunami triggered by an earthquake along the eastern Aleutian-Alaska megathrust would cross the Pacific Ocean and could cause damage to southern California coasts, with ports being particularly vulnerable. Such a tsunami would impact not only the coast of California, but also Hawaii and the Pacific Basin in general, and it would cause considerable damage to the southern coasts of Alaska. An M_w9 earthquake sourced in the eastern Aleutians would most likely involve a complex, multiple-segment rupture, perhaps similar to the 1957 Andreanof earthquake or the 2004 Sumatra-Andaman earthquake. A multiple-segment rupture in the eastern Aleutians is possible, in that there are no major fracture zones along the Fox-Shumagin-Semidi Islands sectors of the megathrust that might form a barrier to a throughgoing rupture. The lack of a large accretionary complex along the eastern Aleutian-Alaska megathrust allows for large slip at the updip (deep water) end of the megathrust, contributing to the size of the tsunami generated. A tsunami generated in the eastern Aleutians could create a wave amplitude as high as 2 m, with currents as high as 4 m/s in Los Angeles (Uslu and others, 2010).

Although we suggest that the Fox-Shumagin-Semidi Islands section of the Aleutian-Alaska megathrust is potentially capable of generating a destructive trans-Pacific tsunami, the Aleutians overall are not a well-studied subduction zone. In particular, few geophysical data have been collected offshore of the Fox Islands, and thus little is known about this section of the subduction zone. In order to determine the most likely source in the eastern Aleutians for a damaging tsunami-generating earthquake, we will need additional paleoseismology and paleotsunami studies, more GPS stations, multibeam bathymetry, and geophysical profiling to better image the seismogenic zone. Until more is understood about this remote, most seismically active part of the United States, it is prudent for California to be prepared for a possible damaging tsunami sourced along the Aleutian-Alaska megathrust.

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