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Evaluation of Tsunami Risk from Regional Earthquakes at Pisco, Peru

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Abstract We evaluate tsunami risk for the port city of Pisco, Peru, where major liquefied natural gas facilities are planned. We use a compilation of instrumental and historical seismicity data to quantify the sources of six earthquakes that generated tsunamis resulting in minor inundation (1974) to catastrophic destruction (1687, 1746, 1868) in Pisco. For each of these case scenarios, the seismic models are validated through hydrodynamic simulations using the MOST code, which compute both flow depth on virtual offshore gauges located in Pisco harbor and the distribution of runup in the port and along the nearby beach. Space-time histories of major earthquakes along central and southern Peru are used to estimate recurrence times of tsunamigenic earthquakes. We conclude that Pisco can expect a metric tsunami, capable of inflicting substantial damage every ~53 years, and a dekametric tsunami resulting in catastrophic destruction of infrastructures every ~140 years. The last such event occurred 138 years ago.

An important result of our study is that total destruction of the city of Pisco during the famous 1868 Arica tsunami requires an earthquake rupture straddling the Nazca Ridge, which thus constitutes at best an imperfect "barrier" for the propagation of rupture during megathrust events. This gives a truly gigantic size to the 1868 Arica earthquake, with a probable seismic moment reaching 10^{30} dyne cm.

Introduction and Background

We present a pilot study quantifying tsunami risk along a portion of South American coastline including the port of Pisco, Peru (Fig. 1), where major liquefied natural gas (LNG) facilities are planned. The coastline of South America, located at the boundary between the Nazca and South American plates, features exceptionally large earthquakes, which have characteristically triggered major tsunamis inflicting severe destruction in both the near and far fields. This warrants critical assessment of tsunami risk in the context of the development of facilities such as the proposed LNG terminal at Playa Loberia.

The city of Pisco, Peru, is located at 13.7° S, 76.2° W, 200 km southeast of Lima along the Peruvian shore, at the northern termination of a bay limited to the south by the Paracas peninsula (Fig. 1). The peninsula separates two segments of coastline with slightly different azimuths, and different histories of seismic activity: to the north, the central shore extends from Pisco to Chimbote (9° S) at an azimuth of N330°E, with seismic rupture recently expressed through moderately large earthquakes; to the south, the Peruvian southern shore, trending N305°E to the Arica Bight at 19° S, was the site of gigantic historical earthquakes, such as the famous 1868 event. A further important tectonic feature is the aseismic Nazca Ridge, in general, interpreted as a fossil hotspot track (Pilger and Handschumacher, 1981), which subducts along a 175-km segment of coastline, between latitudes 14.5° S and 15.5° S.

We address the question of the tsunami risk at Pisco by examining the historical record of tsunami damage along the coast of central and southern Peru, from 9° S (Chimbote) to 19° S (Arica; now in Chile), building seismic models of the principal events involved and conducting numerical simulations of the run-up at Pisco (and other coastal locations) that these models predict. We then evaluate possible return periods for the main seismic events under consideration. These individual simulations make our approach significantly different from the recent work of Kulikov *et al.* (2005), based on the empirical concept of so-called "tsunami magnitudes" (Iida, 1963; Abe, 1981), which consists of compiling and averaging tsunami run-up heights at various sites, regardless of the often nonlinear response of specific coastal bathymetry and harbors. By conducting full-scale run-up simulations with specific earthquake-source scenarios, we can provide more realistic estimates of the actual tsunami hazard at a site such as Pisco.

Our sources include the tsunami catalogs of Solov'ev and Go (1984) and Solov'ev *et al.* (1986), and the seismological compilations of Silgado (1992) and Dorbath *et al.* (1990). Other catalogs of South American tsunamis have been published, notably by Lomnitz (1970) and Lockridge

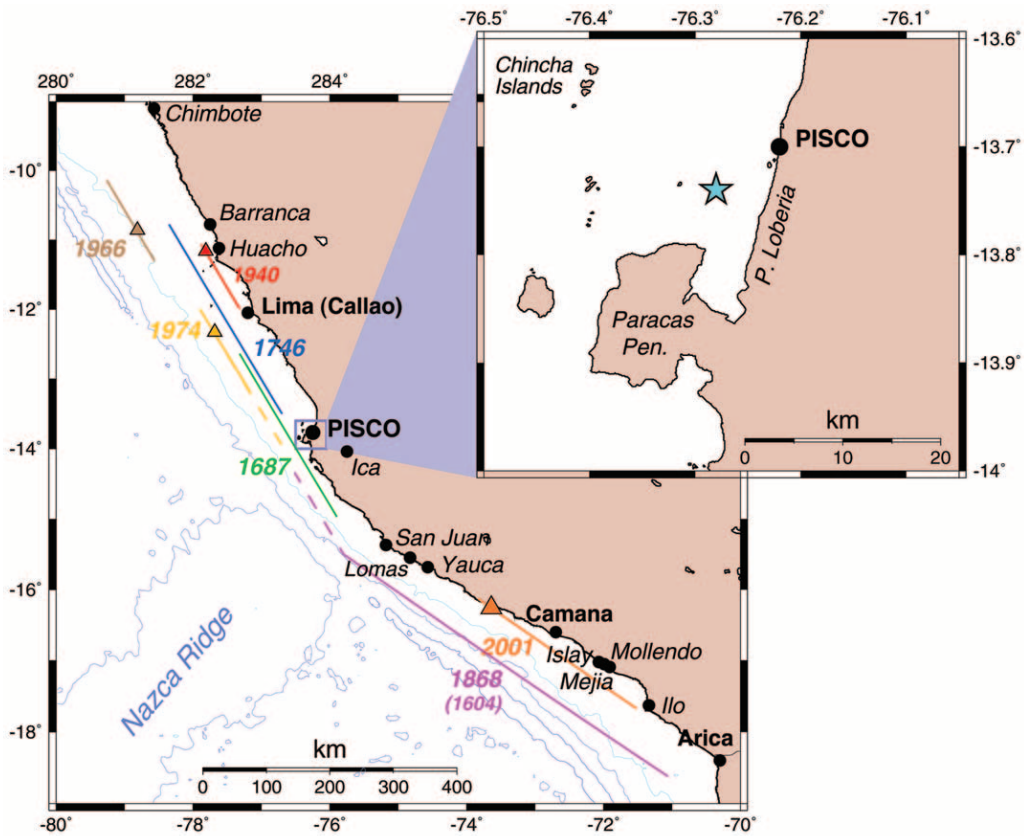

Figure 1. Map of central Peru detailing (inset) the location of the port of Pisco, Playa Loberia, and the proposed LNG terminal facility (star). On the main map, the color-keyed segments are models of rupture for historical tsunamigenic earthquakes. Triangles are epicenters of modern events from the instrumental era. For 1974 and 1868, dashed lines are alternate models involving a proposed longer rupture. To improve clarity, fault lines are traced at variable distances from the coast, but this is not meant to express differences in epicentral distance from the shoreline or in possible depth extent of the rupture.

(1985), but all such compilations tend to be based on interpretations of the same original reports, and for this reason we refer, for each event and whenever possible, to those individual studies.

In very general terms, we can classify tsunamis according to the amplitude of their run-up on the coast: decimetric tsunamis (with run-up between 0.1 m and 1 m) are mostly recorded by tide gauges and do not carry a specific hazard over and beyond that presented by storm waves; an example in Peru would be the Nazca earthquake of 12 November 1996 (Swenson and Beck, 1999). Metric tsunamis (with run-up of a few meters) can inflict substantial damage to coastal and harbor communities, especially because they can result in inundation distances of up to 1 or 2 km inland, as demonstrated recently by the Camaná tsunami of 23 June 2001 (Okal *et al.*, 2002). Finally, dekametric tsunamis (with run-up greater than 10 m) are catastrophic events leading to the total destruction of coastal communities, with inundation reaching several kilometers inland. Examples would include the overflow of the saddle between Banda Aceh and Lhoknga over a total distance of 15 km during the recent 2004 Sumatra tsunami (Borrero, 2005), and the 1868 Arica, Peru (now Chile), tsunami (Billings, 1915), which deposited the *U.S.S. Wateree* 3.5 km from the shoreline. For the purpose of the present study, the first class of events are not considered to contribute to tsunami risk. Note that our ranking shares the philosophy of Solov'ev's (1970) tsunami intensity scale while using a simplified three-tier classification.

From the standpoint of the seismological investigation of the parent earthquakes, we can distinguish three periods: After 1976 (in the era of digital instrumentation), the Centroid Moment Tensor (CMT) project at Harvard University (Dziewonski *et al.*, 1987 and subsequent quarterly updates) routinely computes a coherent set of earthquake source parameters, and the size of the rupturing fault can be inferred from source tomography and the relocation of aftershocks (e.g., Bilek and Ruff, 2002) During the period of instrumented historical seismicity (roughly from 1900 to 1976), seismic records available from several archival sites can be used to apply techniques similar to (or derived from) the CMT algorithm (Kanamori, 1970; Okal and Reymond, 2003) to retrieve the earthquake source characteristics; most large earthquakes with tsunami potential have been the focus of individual studies (e.g., Swenson and Beck, 1996). In addition, epicentral locations are available, and precise relocations can be carried out (Wysession *et al.*, 1991; Engdahl *et al.*, 1998), using datasets of arrival times listed by the International Seismological Centre (ISC; earlier the International Seismological Summary or ISS). For events predating 1900, when no seismic records are available (preinstrumental era), the size of the largest earthquakes can often be inferred through the use of macroseismic techniques such as the interpretation of isoseismal lines, which in turn can be reconstructed from detailed reports of destruction by civilian authorities and clergy, for example, and compiled in Peru by Silgado (1992). For all practical purposes, and to our best knowledge, no quantifiable records exist for any seismic and/or tsunami disasters in Peru predating the Spanish colonization in 1531.

In addition to their classical generation through the deformation of the seafloor under the coseismic dislocation, tsunamis damaging in the near field can also result from underwater landslides, themselves triggered during or shortly after a seismic event. This scenario, which was responsible for the disastrous 1998 tsunami in Papua New Guinea (Synolakis *et al.*, 2002), should be kept in mind when assessing tsunami hazard, even though the quantification of its risk is made immensely difficult by the fundamentally nonlinear nature of the triggering phenomenon, and by our general ignorance of the population statistics of underwater landslides.

Review of Historical Tsunamis Having Affected the Central and Southern Coasts of Peru

In this section, we review systematically the largest earthquakes documented along the central and southern coasts of Peru over the past 350 years. We focus on the tsunamis they generated, especially regarding their impact on the city of Pisco, either as reported in various sources, or in the case of older events, as inferred from reports at other locations along the coast. For each significant tsunami, we either retrieve from the literature, or otherwise estimate, a source model that is later used in our numerical simulations. Events are listed chronologically, starting with the most recent.

23 June 2001, Camaná, $M_0 = 4.9 \times 10^{28}$ dyne cm (Not Observed at Pisco)

This is the most recent large tsunamigenic earthquake, with a human toll attributable to the tsunami of 22 fatalities and 52 missing. Reports from the International Tsunami Survey Team (Okal *et al.*, 2002) indicate a maximum run-up of 7 m at Camaná; the tsunami was not observed above the high-water mark outside the Yauca-Ilo segment of the coast. We do not model this tsunami.

12 November 1996, Nazca, $M_0 = 4.6 \times 10^{27}$ dyne cm (Not Reported at Pisco)

This low-angle thrust event involved the subduction of the prominent Nazca Ridge, at a location recognized earlier as a seismic "gap" between the rupture zones of the 1942 and 1974 events (Beck and Nishenko, 1990). The tsunami was decimetric, reaching only 35 cm (peak-to-trough) at Arica and 24 cm at Callao. We do not model this tsunami.

21 February 1996, Chimbote "Tsunami Earthquake," $M_0 = 2.2 \times 10^{27}$ dyne cm (Not Reported at Pisco)

This was a typical "tsunami earthquake" (Kanamori, 1972; Newman and Okal, 1998), characterized by a slow rupture at the source, resulting in enhanced tsunami excitation relative to its moment inferred from seismic mantle waves. The run-up reached 5.1 m at Chimbote, resulting in 12 deaths and considerable damage to more than 150 houses and beach huts (Bourgeois *et al.*, 1999). However, it was not recorded above the high-water mark south of 11° S. We do not model this tsunami.

03 October 1974, Lima, $M_0 = 1.5 \times 10^{28}$ dyne cm (Okal, 1992) (Mild Destruction at Pisco)

The tsunami was recorded by maregraphs with peak-to-trough amplitudes of 1.83 m at Callao and 1.2 m at San Juan. It reportedly inundated houses on the waterfront at Pisco, which would suggest amplitudes of at least 2 m, but less than 4 m, which would probably have led to more systematic destruction.

This earthquake was studied by several authors, including G. S. Stewart (reported by Kanamori, 1977a), Dewey and Spence (1979), and in detail by Beck and Ruff (1989). Okal and Newman (2001) showed that its source was marginally slow. Based on the aftershock distribution, Dewey and Spence favored a 250-km-long fault, whereas the source tomography by Beck and Ruff suggested a shorter length. Consequently, we test two models, a short fault (model a) extending 150 km with a slip of 5 m, and a long fault (model b) extending 250 km, but with a slip of only 3 m, the latter shown as a dashed line on Figure 1. We use a hypocentral depth of 22 km, and Stewart's focal mechanism ($\phi = 340°$; $\delta = 17°$; $\lambda = 90°$).

03 September 1967, $M_0 = 6.3 \times 10^{26}$ dyne cm (Okal and Newman, 2001) (Not Reported at Pisco)

This earthquake generated only a minor tsunami with decimetric amplitudes at Callao and Chimbote. It is clearly too small to bear significantly on the tsunami risk along the coast, and we do not model it.

17 October 1966, Barranca, $M_0 = 1.95 \times 10^{28}$ dyne cm (Abe, 1972) (Not Reported at Pisco)

The tsunami reached 6 m at Tortuga (presumably on the northern coast) and a peak-to-trough amplitude of 3.5 m at

Callao, and was recorded at San Juan (no amplitude reported). Thus, it would be feasible that a metric oscillation took place at Pisco.

The earthquake was studied by Abe (1972), Dewey and Spence (1979), Beck and Ruff (1989), and Okal and Newman (2001). We model the rupture as a 150-km-long, 50-km-wide fault with a slip of 4 m. The focal mechanism ($\phi = 330°$; $\delta = 12°$; $\lambda = 90°$) is taken from Abe's work, the hypocentral depth (38 km) from the ISC *pP* determination.

24 August 1942, San Juan, $M_0 = 1.3 \times 10^{28}$ dyne cm (Okal, 1992) (Not Reported at Pisco)

This large earthquake was centered east of the Nazca Ridge, but a relocation by Okal and Newman (2001) suggests that its epicenter was on land. This could explain that despite the large seismic moment, tsunami damage was limited, with a lone report at Lomas; indeed Solov'ev and Go (1984) suggested that the tsunami may have been generated by a local landslide. We do not model this event.

24 May 1940, Huacho, $M_0 = 2.5 \times 10^{28}$ dyne cm (Kanamori, 1977a; Okal, 1992) (Possibly Recorded at Pisco)

The tsunami is poorly documented, but reports quoted by Solov'ev and Go (1984) describe it as stronger than in 1966. The earthquake was studied by Beck and Ruff (1989) and Okal and Newman (2001). We model the fault as 50-km-wide, 120-km-long fault with a slip of 5 m. The hypocentral depth is taken as 40 km, and the focal mechanism after Beck and Ruff ($\phi = 340°$; $\delta = 17°$; $\lambda = 90°$).

13 August 1868, Arica, $M_0 = 7$ to 10×10^{29} dyne cm (Estimated) (Total Destruction at Pisco)

This is the great Arica tsunami, probably the largest to affect Peru in historical times. Although no run-up amplitude is available for Pisco, Solov'ev and Go (1984) reported that "[the tsunami] destroy[ed] everything in its path. In particular, a stone breakwater was utterly flattened." Catastrophic destruction was also reported in the nearby Chincha Islands, where a down-draw of at least 20 m was observed. Run-up heights of 15 m or more were reported along the Southern Coast, for example, in Mejia.

The tsunami was recorded all over the Pacific, including in Japan and New Zealand with amplitudes of several meters, a situation repeated in modern times only by the 1960 Chilean earthquake in the Pacific Basin and the 2004 Sumatra earthquake in the Indian Ocean. Thus, it seems warranted to give this event an exceptionally large size. The detailed investigation of isoseismals by Dorbath *et al.* (1990) suggests a length of rupture of 600 km, which would be surpassed among modern events only by Chile (1960), Alaska (1964), and Sumatra (2004). In this framework, we tentatively assign the Arica earthquake a moment of 7×10^{29} dyne cm (i.e., slightly smaller than the estimate for the 1964 Alaska earthquake using conventional mantle waves [Kanamori, 1970]), and model its rupture as a 600-km-long, 150-km-wide fault with a slip of 15 m. Our chosen focal mechanism ($\phi = 305°$; $\delta = 20°$; $\lambda = 90°$) expresses the local geometry of subduction. As discussed under "Hydrodynamic Simulations," we also consider an even larger source, by prolonging the rupture 300 km to the northwest in the azimuth 330°. This would imply rupturing through the Nazca Ridge segment and boost the seismic moment to 10^{30} dyne cm.

13 May 1784, Camaná

This earthquake was not listed by Solov'ev and Go (1984), but Dorbath *et al.* [1990] reported a tsunami "observed at Camaná, Mollendo and Ilo, but [. . .] not produc[ing] any damage." Their macroseismic investigation, as well as their estimate of its moment, could make it comparable to the 2001 earthquake, albeit with a slightly more eastward location. The stronger tsunami damage in 2001 could be due to the recent development of the beach area at Camaná, where most of the 2001 damage occurred.

28 October 1746, Lima, $M_0 = 2$–3×10^{29} dyne cm (Estimated by Swenson and Beck, 1996) (Tsunami Waves Destroyed Pisco)

This is the last of three major earthquakes that repeatedly destroyed Lima in the 211 years following its foundation. A catastrophic tsunami followed, which devastated the port of Callao, where run-up heights of 24 m are suggested in historical reports; the tsunami is reported to have destroyed the city of Pisco. Note that the latter had been rebuilt farther inland after the 1687 tsunami (see following), so that a direct comparison of the two disasters is unwarranted.

Intensity reports compiled by Dorbath *et al.* (1990) suggest a rupture length of 350 km, extending northwest from the Chincha Islands area. We use a width of 100 km, and a slip of 15 m, which fit the moment estimated by Swenson and Beck (1996) and express the much larger size of the earthquake, relative to the events of 1966 or 1940. The focal mechanism ($\phi = 330°$; $\delta = 20°$; $\lambda = 90°$) expresses normal convergence between the Nazca and South American plates.

20 October 1687, Pisco, $M_0 = 2$–3×10^{29} dyne cm (Estimated by Swenson and Beck, 1996) (Catastrophic Destruction at Pisco)

This is probably the event that most directly affected Pisco, which was reported to have been totally destroyed by the waves, the latter carrying at least three ships "over what used to be the town" (Silgado, 1992). The tsunami was reported in Japan with decimetric amplitudes (Solov'ev and Go, 1984).

Intensity reports compiled by Dorbath *et al.* (1990) suggest a fault length of 300 km. This geometry is interesting in that the earthquake presumably ruptured the segment to the south of Pisco where the Nazca Ridge is subducted. We use a width of 100 km, and a slip of 12 m, with the following focal mechanism: $\phi = 315°$; $\delta = 20°$; $\lambda = 90°$.

As discussed in detail by Dorbath *et al.* (1990), this earthquake was almost certainly followed by a significant event on the next day on the southern shore, suggesting a process of triggering by stress transfer (Stein *et al.*, 1997; McCloskey *et al.*, 2005). There is no mention of a tsunami for the latter shock.

22 May 1664, Ica, $M_0 = 6 \times 10^{27}$ dyne cm (Estimated) (Tsunami Drowned 60 People at Pisco)

Based on the macroseismic investigation by Dorbath *et al.* (1990), this event most probably ruptured the Nazca Ridge segment (which was again involved in 1687). It may have had a 100-km-long fault, which would suggest a moment slightly larger than that of the 1996 earthquake. There is a report of 60 people drowned at Pisco, suggesting a destructive tsunami.

23 November 1604, Southern Shore, $M_0 = 7 \times 10^{29}$ dyne cm (Estimated) (Probable Damage at Pisco)

This catastrophic earthquake is interpreted by many authors as comparable to the 1868 Arica event (Dorbath *et al.*, 1990; Swenson and Beck, 1996), based on the similarity between their isoseismal maps and on the strength of its tsunami, which affected at least 1200 km of South American coastline, and possibly even 2800 km, from Lima to Concepción (Solov'ev and Go, 1984). In the case of Pisco, however, conflicting reports suggest that the waves may have been locally deflected, with destruction limited to certain sections of town (Perrey, 1857; Solov'ev and Go, 1984). In this framework, we regard the 1604 event as equivalent to the short fault scenario for the 1868 earthquake, and model only the latter.

09 July 1586, Huacho

This earthquake was the first one to destroy Lima after its foundation in 1535. Solov'ev and Go (1984) reported a catastrophic local tsunami with run-up amplitudes of 24 m at Callao, which would be corroborated by reports of a transoceanic tsunami observed in Japan. Yet, there is no mention of flooding at other Peruvian sites. Additionally, Dorbath *et al.*'s (1990) macroseismic investigation leads to a relatively short rupture of only 175 km, and hence to a presumably moderate seismic moment, not exceeding that of the 1974 event. The disparity in tsunami amplitude may be due either to a very slow component of the source, not expressed in the acceleration field constituting the source of the macroseismic dataset (but then the absence of local tsunami reports outside Lima is unexplained), or to the triggering of an underwater landslide causing the local tsunami (but this would not explain the wave amplitudes in Japan). Based on the available information, a satisfactory model of this tsunami cannot be derived.

22 January 1582, Mollendo

This earthquake affected the southern coast of Peru from Camaná to Ilo. Dorbath *et al.* (1990) report a probable tsunami at Islay. Based on their macroseismic investigation, the earthquake appears to be substantially smaller than the 2001 event.

09 May 1877, Iquique, 3-m Wave at Pisco

In addition to the preceding Peruvian earthquakes, we also consider the case of the Tarapacá (Iquique) earthquake, which ruptured the Chilean coast (then part of Peru and Bolivia) on 09 May 1877. This very large event unleashed a destructive tsunami throughout the Pacific, with 3-m waves reported at Pisco. It is generally interpreted (e.g., Lomnitz, 1970) as having ruptured a segment of coast extending from the Arica Bight to the Mejillones peninsula near Antofagasta, a length of 450 km, which would suggest a moment of 5×10^{29} dyne cm.

In concluding this section, note that at least five of the historical earthquakes considered (1604, 1687, 1746, 1868, and 1877) have moments estimated between 2 and 10 times 10^{29} dyne cm, that is, surpassed only, in the instrumental record, by the events in Chile, 1960 ($2–5 \times 10^{30}$ dyne cm [Cifuentes and Silver, 1989]); Alaska, 1964 (1.2×10^{30} dyne cm [Nettles *et al.*, 2005]); Sumatra, 2004 (1×10^{30} dyne cm [Stein and Okal, 2005]); and for the "smaller events" of 1687 and 1746, by the 1952 Kamchatka earthquake (3.5×10^{29} dyne cm [Kanamori, 1976]). The region is clearly host to some of the largest earthquakes in the world, and the seismic record during the instrumental era obviously underestimates its true potential. The city of Pisco was completely destroyed by tsunamis in 1868, 1746, and 1687, and also probably strongly affected in 1664.

Hydrodynamic Simulations

Simulations were performed for each of the six earthquakes listed in Table 1, using the MOST code (Titov and Synolakis, 1997), which solves the nonlinear shallow-water wave equations using a variable staggered grid with the method of fractional steps; a full description is given in Synolakis (2002). In contrast to early-generation models that would stop the wave-evolution calculations at some threshold depth (e.g., 5 or 10 m), and essentially treat the shoreline as a rigid, fully absorbing vertical wall, MOST is a so-called "two + one" inundation code, that is, it includes the full calculation of run-up as the wave propagates up the beach over initially dry land.

The validity of this technique was tested through the successful numerical modeling both of laboratory data, obtained notably by Briggs *et al.* (1995a, 1995b) at the USACE

Table 1
Results of Hydrodynamic Simulations

Number	Year	Area	Moment 10^{28} dyn cm	Fault Parameters Length (km)	Width (km)	Slip (m)	Run-Up at Pisco Computed (m)	Reported
1	1687	Pisco	20–30	300	150	12	13	catastrophic destruction
2	1746	Lima	20–30	350	100	15	22	city destroyed
3	1868a	Arica	70	600	150	15	4	total destruction
	1868b	Arica	100	1000	150	15	15	total destruction
4	1940	Huacho	2.5	125	90	5.5	0.8	possibly recorded
5	1966	Barranca	1.95	160	60	4	1.5	not reported
6	1974a	Lima	1.5	150	40	5	2	mild inundation
	1974b	Lima	1.5	250	40	3	3	mild inundation

Coastal Engineering Research Center, and of posttsunami datasets obtained during field surveys conducted since 1992 in the wake of several major tsunamis (Synolakis and Okal, 2005). The validation of the MOST code (earlier known as VTCS-3) is exemplified in the 1993 Hokkaido-Nansei-Oki tsunami (Shuto and Matsutomi, 1995) which struck Okushiri Island in the Sea of Japan with extreme run-up heights reaching 30 m and currents on the order of 10–18 m/sec. MOST was the only initial code which presented and published an accurate simulation of these extreme heights and currents (Titov and Synolakis, 1997)

For each of the six case studies, the rupture model proposed in "Review of Historical Tsunamis Having Affected the Central and Southern Coasts of Peru" was used to compute the static vertical deformation of the ocean floor through the algorithm of Okada (1985). Two scenarios (a "short" and a "long" fault) were considered for the 1868 and 1974 events. An example of the resulting displacement field for event 1 is contoured on Figure 2a. The deformation field was then used as an initial condition of the hydrodynamic computation. This constitutes a legitimate approximation since the deformation of the ocean floor always takes place much more rapidly than the tsunami waves can propagate the sea surface deformation away from the source region. Even in the so-called "tsunami earthquakes" characterized by very slow rupture velocities, the latter remain on the order of at least 1 km/sec (Polet and Kanamori, 2000), and thus hypersonic with respect to tsunami velocities, even in the deepest oceanic basins.

For the purpose of simulations at Pisco, we used the ETOPO-2 two-minute bathymetric dataset (Smith and Sandwell, 1997), complemented by local bathymetric charts. We built a staggered grid, with an initial spacing of 2 km offshore, refined near the coast to 100 m in the north–south direction, that is, mostly along shore, and 40 m in the east–west direction, or mostly across shore.

In this context, our simulations have two classes of products: we first compute time series of water surface elevation (e.g., Fig. 2c) at a profile of virtual gauges located offshore from the port of Pisco, at water depths of 15, 10, 5, 0.4, and −1.3 m (i.e., the last one being on initially dry land). We also keep track of the maximum amplitude of currents simulated in the various models. Second, we consider the full run-up along the beach (Fig. 2b), calculated on a coarse grid along a stretch of coastline extending 200 km northwest of Pisco, and on a refined grid in the most relevant 25-km shore segment around Pisco.

Results

Results are compiled in Table 1 and selected cases are illustrated in Figures 2,3,4, where we present profiles of the maximum value of run-up along the coast line. For the four tsunamis of 1687, 1746, 1940, and 1966, we obtain an excellent agreement between run-up heights computed in our hydrodynamic simulations and the reported level of destruction for the city of Pisco. The total destruction of the city in 1687 and 1746 is well predicted by run-up heights of 13 and 22 m, respectively, while run-ups simulated for 1940 (0.8 m) and 1966 (1.5 m) agree well with the lower levels of destruction reported for these modern events. For the 1974 event, Figure 3 shows that the long fault scenario (Fig. 3b) leads to marginally larger run-up (3 m) in Pisco than the shorter model (Fig. 3a; 2 m). However, the difference is not resolvable from the reported observation of "houses inundated on the waterfront" (Dorbath *et al.*, 1990), which is reasonably fit under either model.

The 1868 Arica Earthquake: Reassessing a True Giant

This leaves the intriguing case of the 1868 Arica tsunami for which our first model (shown on Fig. 4a) predicts a run-up at Pisco of at most 3 m, which is difficult to reconcile with the utter destruction described in Solov'ev and Go (1984). We recall that the source model was derived from Dorbath *et al.*'s (1990) study and featured a fault length of 600 km, bounded in the west by the Nazca Ridge. By contrast, Figure 4b shows that an enlarged source, rupturing through the Nazca Ridge, for a total fault length of about 900 km, can lead to run-up approaching 15 m at Pisco, which is more in line with a tsunami having "destroyed everything

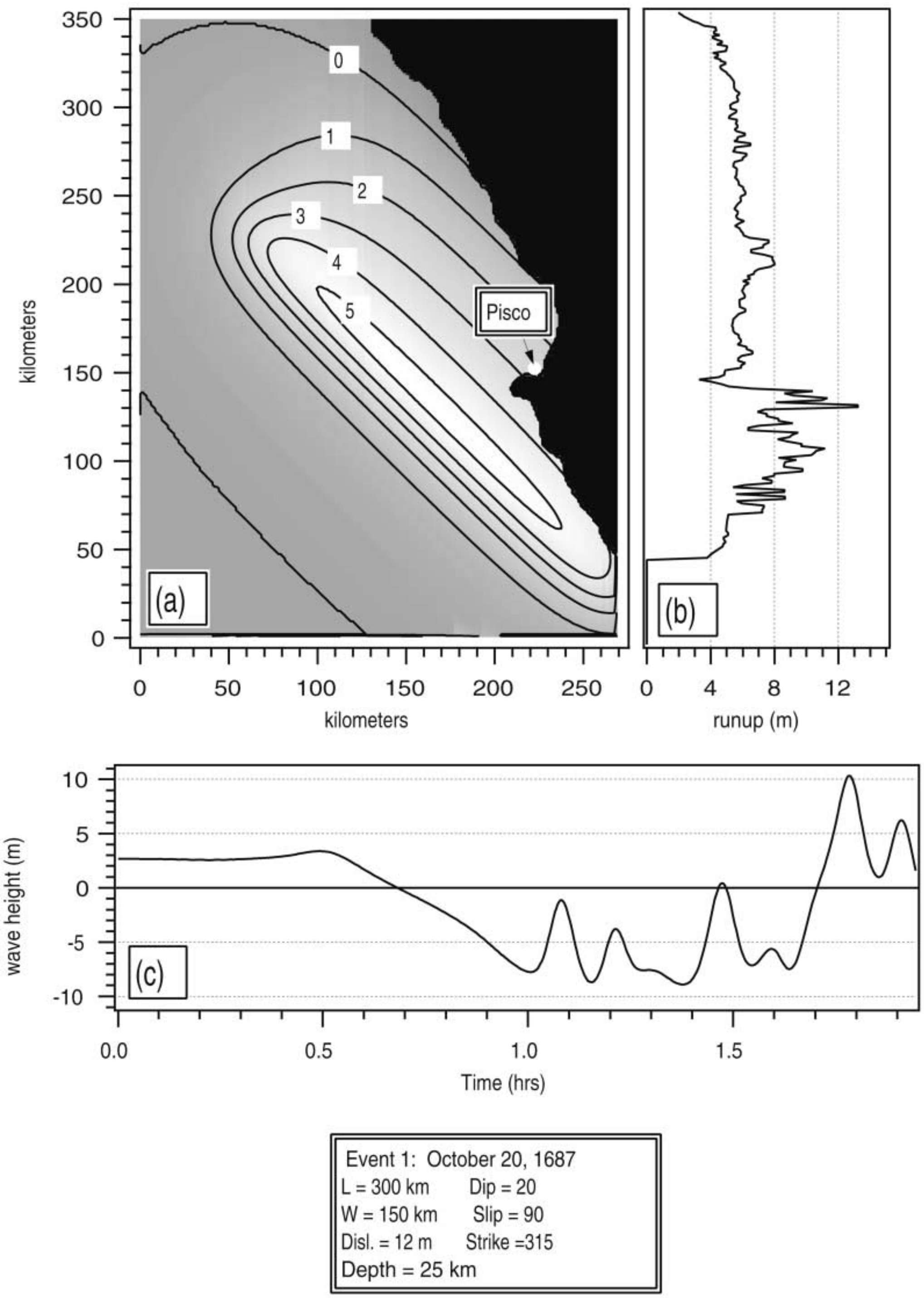

Figure 2. Simulation of event 1. (a) Map of the static deformation computed from the earthquake source model using Okada's (1985) algorithm, contoured at intervals of 1 m, and used as an initial condition of the hydrodynamic computation. (b) Run-up along the coastline of Peru. Note the value of 13 m in agreement with the total destruction reported by Silgado (1992). (c) Time series of a virtual gauge located in Pisco harbor at a water depth of 10 m.

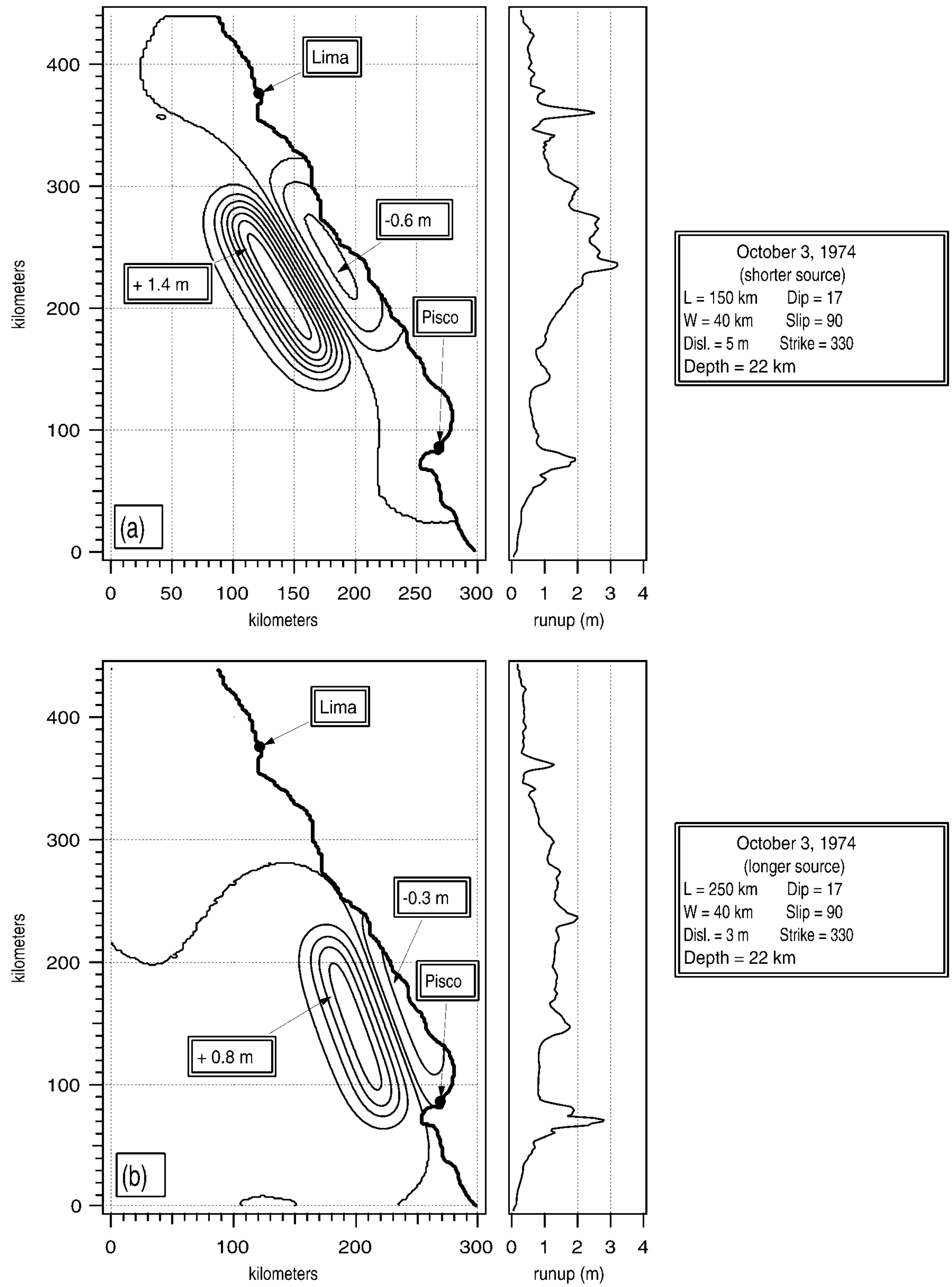

Figure 3. Static deformation and run-up for event 6, using both a short fault model (a) according to Beck and Ruff (1989), and a longer fault model (b) according to Dewey and Spence (1979). Although the former results in marginally smaller run-up at Pisco, the difference is not resolvable from the observations reported by Dorbath *et al.* (1990).

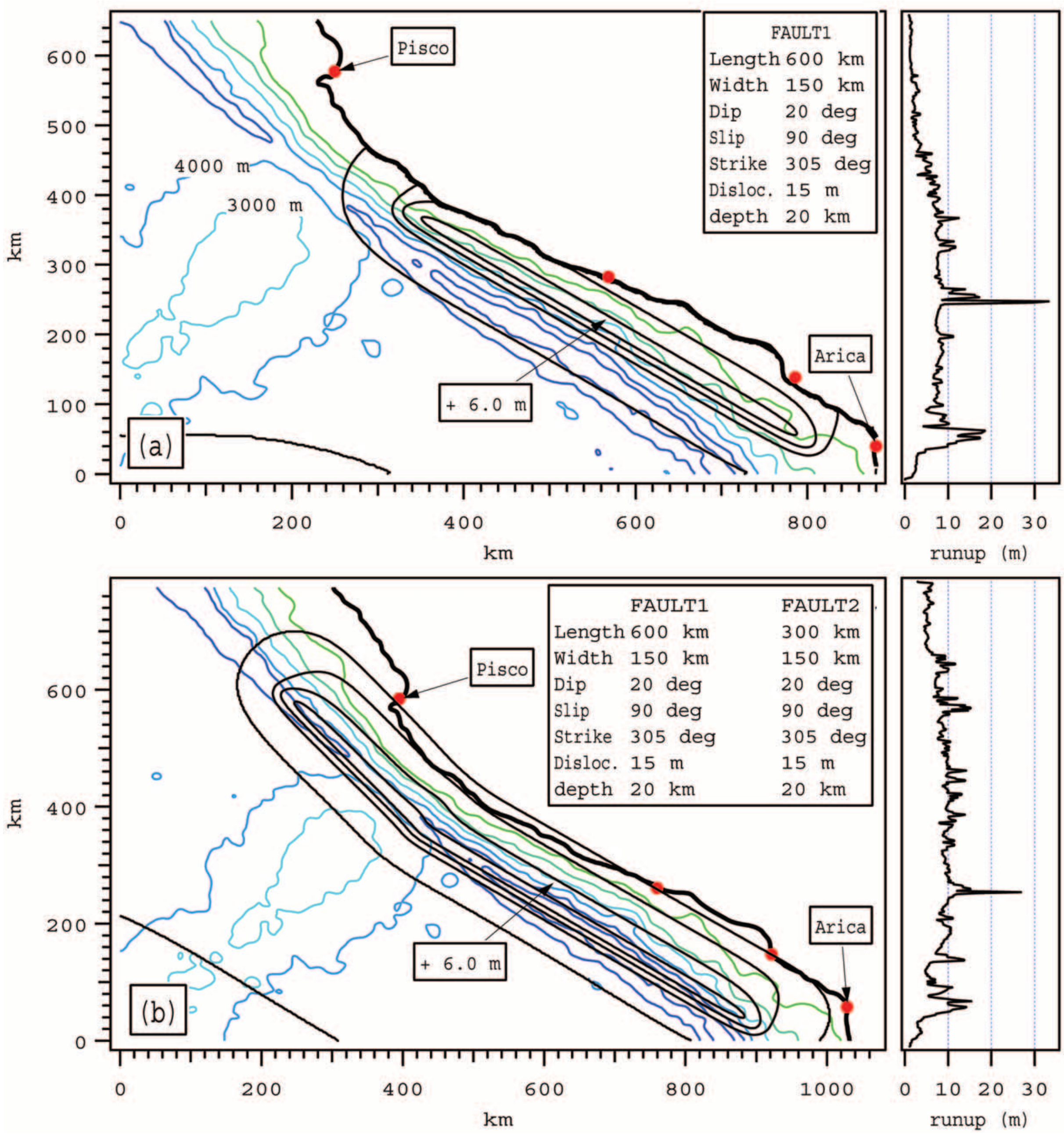

Figure 4. Simulation of the great 1868 Arica tsunami. In model a, we use a relatively short fault line (600 km), inspired from Dorbath *et al.* (1990). While it correctly predicts a catastrophic run-up in excess of 15 m at Arica, the tsunami reaches no more than 3 m in Pisco, which cannot account for the total destruction reported in the city and on the Chincha Islands. By contrast, model b, which allows the rupture to propagate across and beyond the Nazca ridge predicts a run-up of up to 15 m at Pisco.

in its path" (Solov'ev and Go, 1984). Accordingly, we retain this second scenario as the most probable model of the great 1868 Arica earthquake and tsunami; its 900-km rupture is comparable to that of the 1960 Chilean earthquake, and intermediate between those of the 1964 Alaska and 2004 Sumatra events (Plafker, 1965; Plafker and Savage, 1970; Stein and Okal, 2005).

An alternative model would have consisted of keeping the "shorter" fault (600 km), but increasing the slip. In the simplified model of a homogeneous fault considered here, the slip needs to be at least doubled, to 30 m, to reach catastrophic levels of run-up at Pisco. We regard this as extremely unlikely, since this level of average slip has not been reported even for megathrust earthquakes, because it would lead to unrealistic down-dip components of the strain-release tensor. Thus, we prefer the "long fault" model for the 1868 event.

This result has fundamental geophysical implications for our understanding of the fragmentation of rupture along subduction zones. In particular, it invalidates the concept of the Nazca Ridge functioning as a robust "barrier" limiting the propagation of the rupture to the north. This scenario may have applied in 1604, but our results clearly require a more northerly component to the rupture to explain the total destruction at Pisco and in the Chincha Islands in 1868. In our review of tsunamis that historically have affected the central and southern coasts of Peru, we argued that the smaller event of 1687 to the north may also have ruptured through the Nazca Ridge. The Nazca Ridge then appears as more of a hurdle than a barrier, serving as a rupture terminator for certain events (1604 from the south, 1746 and 1974 from the north), but being "jumped" by a few large ones (1868 from the south, 1687 from the north), while featuring its own contained local earthquakes (1664, 1942, 1996).

In this framework, one can only speculate as to the validity of the concept in the case of other barriers that have been described in the literature as providing an intuitive "natural" boundary to the rupture process of recent major earthquakes, and it becomes only legitimate to question such supposedly recognized barriers as the Amchitka corner in the Aleutians (Johnson *et al.* 1994; Okal, 2005), the Martinique passage in the Lesser Antilles (which maps the diffuse boundary between the North and South American plates), or even the Louisville Ridge at the Tonga-Kermadec boundary (which features a strong change of dip and thus a tear in the downgoing slab). In all such cases, our perception of the maximum expectable earthquake, based on the concept of a robust barrier, should be revised upward, as the latter could be jumped during a superlative megathrust event.

At any rate, our results provide a quantification of the 1868 Arica earthquake from its tsunami, revealing its truly gigantic proportion, in a league with the 1964 Alaska and 2004 Sumatra events, and surpassed only by the 1960 Chilean earthquake.

Summary

The previous extensive compilation and simulation of tsunamigenic earthquakes in central and southern Peru define tsunami hazard at Pisco in the following terms:

- The city of Pisco was destroyed by tsunamis at least three times in its history (by what our simulations suggest were waves of dekametric amplitude), in 1687, 1746, and 1868. It was also strongly affected in 1604, with reports conflicting as to the exact extent of destruction. The earthquakes involved were of gigantic proportions, with macroseismic studies suggesting rupture lengths of between 300 and 600 km (probably 900 km in 1868), and seismic moments greater than 10^{29} dyne cm. These earthquakes could originate either along the southern shore (1604, 1868), or along the central coast (1687, 1746).
- In addition, the city is at risk of less severe but substantial inundation by tsunamis of metric amplitude, as happened at least three, and possibly five, times in the period studied: in 1664, 1877, and 1974, and probably in 1940 and 1966. Such tsunamis are typically generated along the central shore (i.e., northwest of Pisco), as in 1940, 1966, and 1974, but can also originate in the Ica province where the Nazca Ridge subducts (1664), or from catastrophic earthquakes at the Chilean subduction zones (as in 1877).
- Finally, we should stress that Pisco, like any other shore location in the Pacific, is at risk from transpacific tsunamis originating from gigantic earthquakes in distant Pacific subduction zones. Although we could find no such record at the specific location of Pisco, the great earthquakes of 1946 and 1957 in the Aleutians, 1952 in Kamchatka, 1960 in southern Chile, and 1964 in Alaska, all generated tsunamis reaching 1-m amplitude at Callao, and thus presumably a similar height at Pisco.

Recurrence Rates of Tsunamigenic Earthquakes

In attempting to define a repeat time for the various events identified previously as posing a tsunami hazard to the city of Pisco, we considered several approaches.

First, we investigated the frequency-magnitude characteristics (Gutenberg and Richter, 1941) of coastal earthquakes in Peru. Figure 5 shows the result of regressing a dataset of 249 earthquakes having occurred along the central and southern segments of the Peruvian subduction zone during the years 1965–2000, thus excluding the population of aftershocks of the large 2001 earthquake. The magnitude M used in the regression is the largest magnitude (m_b, M_s, M_L) published for each event. The number of earthquakes N with a magnitude of at least M is well fit by the relation

$$\log_{10} N = 6.88 - 0.96\, M \tag{1}$$

in the magnitude window 4.8–6.6, but the extrapolation of this function to larger magnitudes clearly fails, and the Gutenberg-Richter relation is not an adequate predictor of the recurrence of the major events of interest for tsunami risk.

This situation can be explained on several accounts. First, it is well known that source-scaling laws governing the frequency-magnitude relationships break down when the earthquake fault width becomes comparable to the thickness of the brittle layer. In particular, considerable debate remains regarding the exact mode of growth of earthquake sources beyond this point of width saturation, with conflicting models as to whether a growth or a decay in b-value is expected at large magnitudes (Scholz, 1982; Pacheco *et al.*, 1992; Romanowicz and Rundle, 1993). Second, as investigated by

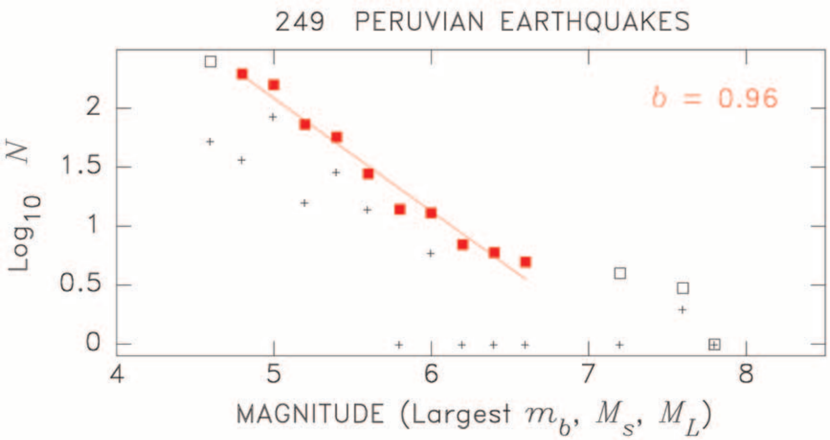

Figure 5. Frequency-magnitude relationship for a population of 249 central and southern Peru earthquakes. The dataset is binned into windows of 0.2 units of magnitude *M*, and the corresponding populations are shown using + signs on a logarithmic scale. The larger square symbols illustrate the cumulative number *N* of events with a magnitude of at least *M*. The dataset is regressed as the straight line (1) for $4.8 \le M \le 6.6$. The open squares represent values not used in the regression.

Okal and Romanowicz (1994), the saturation of magnitude scales (Geller, 1976) further degrades the linear fit of $\log_{10} N$ to *M*. Thus, the extrapolation of any population statistics to the domain of large earthquakes is at best difficult, at worst unwarranted, especially given that any statistical method is expected to grow large uncertainties in the domain of the very small population samples characteristic of the large earthquakes involving tsunami risk.

An alternative model has proposed that the largest event in a given region may occur in the form of a so-called "characteristic earthquake" (Wesnousky, 1994), which may take place more frequently than otherwise suggested by the Gutenberg-Richter law, because no moment would be released at even larger earthquake sizes. However, this model is put in doubt, for example, by the recent work of Cisternas *et al.* (2005), who showed that megathrust events of the past 300 years in central Chile were not repeat earthquakes (the 1960 event being by far larger than its immediate predecessors, as already hinted by Stein *et al.* [1986]); in our context, the disparity between the 1604 and 1868 shocks along the southern shore would be another example. This idea has fueled attempts to amend the Gutenberg-Richter distribution for large earthquakes using the concept of a "maximum" event (M_{xg} in Kagan's [1999] formalism; M_∞ in Kulikov *et al.* [2005]), which in particular allows for the convergence of the total seismic-moment release integrated over all earthquake sizes. While Kagan (1999) stressed that M_{xg} is not necessarily an absolute bound to earthquake size (as opposed to M_∞ in Kulikov *et al.*'s [2005] formalism), in practice, the choice of such a parameter relies strongly on our perception of the population of large earthquakes in a given subduction zone (and thus of the largest event known at the time of study) and of such parameters as the seismic efficiency coefficient, which describes the fraction of tectonic deformation expressed through seismic events. The former is largely subjective, being subject to revision upon occurrence (Sumatra, 2004) or reassessment (Arica, 1868) of an occasional monstrous event. As for the latter, it remains a poorly constrained parameter, despite the simplicity of its concept. As a result, it is not clear that such improved Gutenberg-Richter statistics have the potential to significantly enhance our ability to assess future megathrust events in a given subduction province.

In this framework, we turn to an alternate, more empirical, approach consisting of investigating the space-time relationship of the rupture involved in all documented historical earthquakes. This approach (which leads to the concept of seismic gap) has been used extensively, notably by Dewey and Spence (1979), Nishenko (1985), Beck and Ruff (1989), Dorbath *et al.* (1990), and Rabinovich *et al.* (2001) in central and southern Peru. On Figure 6, we adapt Dorbath *et al.*'s (1990) figure 6, and update it to include the 2001 earthquake. We then attempt to define a return time for the major earthquakes in the region, based on patterns of spaciotemporal similarity between sequences of major seismic events, in the framework of models of plate kinematics that predict a rate of convergence between the Nazca and South American plates varying from 8.8 cm/yr at Lima to 9.25 cm/yr at Antofagasta, Chile (DeMets *et al.*, 1990). We wish at this point to emphasize the tentative nature of any estimate of the return period of major subduction zone earthquakes, because it has long been known that the segmentation of a given subduction province into blocks rupturing during individual earthquakes is not necessarily repetitive, as demonstrated, for example, by Ando (1975) in the Nankai trough of Japan, and more recently by Cisternas *et al.* (2005) in central Chile. In addition, written history documents, at best, one earthquake cycle along some of the relevant segments of the subduction system. Under the circumstances, it is not practical to attempt any correction to the perceived recurrence times, as discussed, for example, by Ogata (1999) to account for documented quiet-time intervals before the first recorded event and after the last one.

A major feature of Figure 6 is the subduction of the Nazca Ridge under the Ica province at about 14° S, which tends to separate individual ruptures along either the central or southern segments of the Peruvian shoreline, even though the large 1687 event, and most probably the 1868 Arica earthquake, do propagate through the ridge, making it an imperfect barrier. This reservation notwithstanding, we will

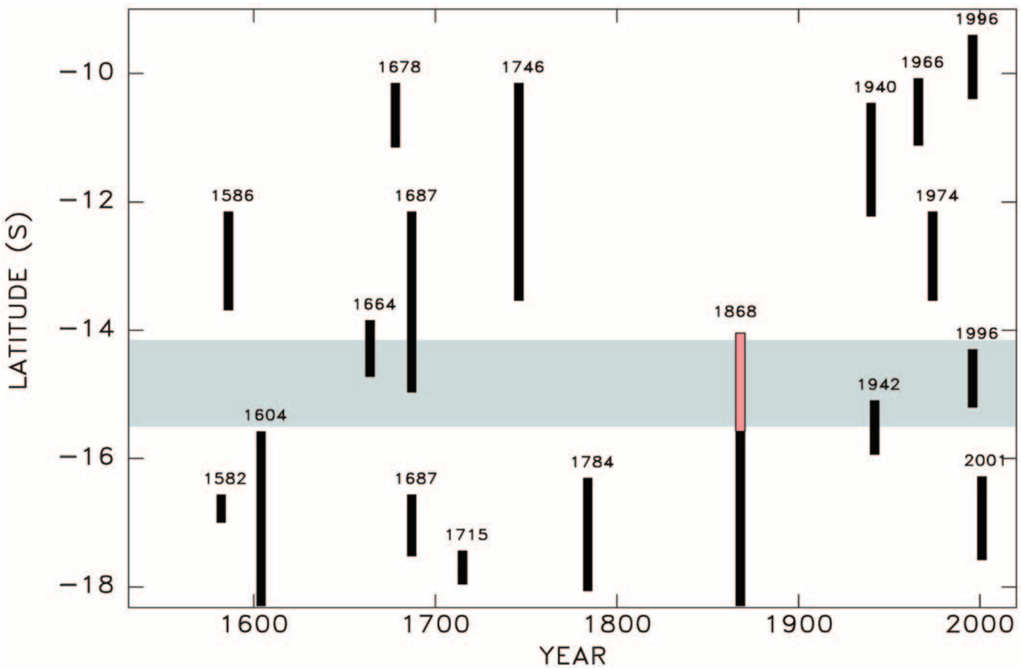

Figure 6. Spatiotemporal distribution of rupture along the central and southern segments of the Peruvian subduction zone. This figure is adapted from Dorbath *et al.*'s (1990) figure 6, and updated to 2001. The vertical bars indicate the extent of rupture involved in each of the major events; the shaded bar for 1868 corresponds to the longer fault (model b in Fig. 4). Note that, because of the variable azimuth of the shoreline, the vertical axis, scaled in latitude, does not directly express rupture length. The horizontal shaded bar, between latitudes 14.2° S and 15.5° S, symbolizes the Nazca Ridge.

study the recurrence of major earthquakes separately in the two provinces to the north and south of the Nazca Ridge.

Along the southern shore, the similarity between the macroseismic effects of the 1604 and 1868 earthquakes and tsunamis suggests that they are indeed repeat events, in the sense that their time separation (264 years) could be representative of the earthquake cycle, even if the 1868 earthquake, rupturing through the Nazca Ridge was larger than its 1604 counterpart. When compared with the plate kinematics rate of 9.1 cm/yr, this would suggest a slip of 24 m during characteristic events, which is probably too large for the 1604 and 1868 earthquakes (since that would make them equal to, or even larger than, the 1960 Chilean earthquake). This apparent deficiency in seismic slip could be explained either (1) by a shorter return period of 1868-type earthquakes, the lone documented interval (264 years) being anomalously long; or (2) by a significant contribution of smaller earthquakes to the plate kinematic budget; or (3) by slip partitioning, implying that the hanging Andean wall of the fault is not rigidly attached to the stable South American plate; or (4) by the presence of a substantial contribution of aseismic slip. On the basis of the local tectonic context (Klosko *et al.*, 2002) and of the known historical seismicity, we favor a combination of explanations 3 and 4, the latter being expected based on the age of the oceanic plate and the convergence rate (Kanamori, 1977b), a conclusion also reached by Nishenko (1985). Note also that the interval between the two comparable "Camaná" earthquakes of 1784 and 2001 is 217 years, which is on the same order as that between the gigantic events. For an order-of-magnitude computation, we thus use a repeat time of 250 years for the tsunami risk at Pisco due to catastrophic earthquakes along the southern shore. We further assume that one half of such events will break across the Nazca Ridge, resulting in catastrophic dekametric tsunamis at Pisco, whereas the other half, remaining more distant, will only generate metric tsunamis.

Along the central shore, the historical record provides no repeat event comparable to either the 1746 or the 1687 shocks. Rather, we note a long interval (194 years) with no major seismic event reported. The sequence of events in the preceding two centuries could be interpreted as a buildup of smaller events, culminating in one (or possibly several) gigantic shocks. Under this interpretation, the renewal of activity in 1940 could signal the beginning of a new cycle of seismic events, with a possible repeat time of ≈290 years. This number is again comparable to that obtained along the southern shore and could suggest a major event comparable to the 1746 earthquake in about 2035. Note, however, that in this scenario, the 1974 earthquake would be interpreted as a repeat of the 1687 one, despite being clearly much smaller. This may illustrate the relative randomness of the segmentation of the subduction zone, the 1687 event having broken the 14° S barrier and ruptured into the Nazca gap, whereas the 1974 shock was contained north of Pisco, this pattern being reminiscent, for example, of the variability of earthquake sequences in the Nankai Trough (Ando, 1975).

Regarding events filling the so-called Nazca gap proposed by Beck and Nishenko (1990), we note the general macroseismic similarity between the 1664 and 1996 events, which would suggest a 332-year repeat time, but recall the latter had only a small tsunami, but the former resulted in drownings at Pisco.

Finally, regarding the return time of a 1877-type earthquake along the northern coast of Chile, and in the absence of a comparable earthquake, we note Nishenko's (1985) estimate of 148 to 444 years, and will use its lower bound (150 years in round numbers), suggesting a possible megathrust event in about 20 years.

We can then combine the preceding estimates for the return times of the historical earthquakes known to have inflicted tsunami damage on the city of Pisco in the following fashion.

- The frequency of a catastrophic, dekametric tsunami at Pisco is the sum of those for a Southern shore, 1868-type event (0.5/250 yr^{-1}, taking into account a 50% probability that an event jumps the Nazca Ridge); for a Northern shore, 1746-type event (1/290 yr^{-1}); and for a northern shore, 1687-type event (estimated at 0.5/290 yr^{-1} on account of the latter's apparently exceptional rupture of the Nazca gap; hence, the factor 0.5). These figures combine to an average repeat time of 140 years.
- The frequency of a significant tsunami in the metric range is the sum of those (1/290 yr^{-1}) for typical 1940- or 1966-type events, of which there appear to be an average of 2.5 per cycle; for a Nazca-gap event of the 1664 type (0.5/332 yr^{-1}, the factor 0.5 expressing a provision for a tsunamiless shock in 1996); for a southern shore megaevent not breaking the Nazca Ridge (1604 type; 0.5/250 yr^{-1}); and for a distant 1877-type event (1/150 yr^{-1}). These figures combine to an average repeat time of 53 years.

In summary, we estimate that a scenario of metric run-up, on the order of a few meters, which could impose substantial damage, in particular, on port facilities, and lead to several fatalities in the absence of mitigation and warning, may have a repeat time of about 50 years. A catastrophic tsunami of dekametric amplitude, capable of totally destroying harbor infrastructures, may have a repeat time of about 140 years. This result is also consistent with the "back-of-the-envelope" observation that the city was destroyed three and perhaps four times during the past 400 years. The last such tsunami took place 138 years ago.

Finally, we stress once again that these numbers include neither the possibility of a transpacific tsunami originating in the Western Pacific or Alaska, nor that of a tsunami generated by an underwater landslide, which could conceivably be triggered by an earthquake no larger than magnitude 6. Based on the historical record at Callao, the former may not give rise to oscillations much larger than storm waves. As documented in many recent studies (e.g., Borrero *et al.*, 2001), the impact of the latter could be locally catastrophic.

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