

The Complex and Long-Duration 2002 April 18 M_w 6.7 Near-Trench Earthquake in the Guerrero Seismic Gap, Mexico

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Accepted 2025 June 12. Received 2025 April 24; in original form 2024 October 23

SUMMARY

The seismic behaviour of the near-trench plate interface of the Guerrero seismic gap and other segments of the Mexican subduction zone is likely to play a critical role in the seismic and tsunami hazard of the region. In this context, a detailed study of the near-trench 2002 April 18 M_w 6.7 earthquake that occurred about 55 km off the coast of Guerrero and generated a small tsunami attains particular importance. From an analysis of the teleseismic P waves and S waves, local recordings and aftershock distribution, we find that the rupture most likely began at a subducted seamount, propagated unilaterally towards NW, parallel to the trench for ~ 54 – 58 km and a duration of ~ 68 – 70 s. The moment rate function is highly rugged, with two dominant pulses separated by about 50 s. Although relatively small in magnitude, the earthquake has all the characteristics of a tsunami earthquake: the slip occurs very close to the trench, the rupture speed is slow (~ 1 km s $^{-1}$), the high-frequency radiation is deficient, and, in common with tsunami earthquakes, the moment-scaled radiated energy is low ($E_R/M_0 = 1.45 \times 10^{-6}$). We confirm that the duration of the event (~ 70 s) is extraordinarily long compared to that expected from scaling relations (~ 12.8 s), consistent with it being the most anomalous of all the events studied in the last 40 yr. Our results support a conditionally stable upper 15 km of the plate interface in the region reported from recent offshore seismic observations.

Key words: North America; Tsunamis; Earthquake source observations; Subduction zone processes.

1 INTRODUCTION

On 2002 April 18, an M_w 6.7 near-trench earthquake occurred. This event generated a small tsunami (Iglesias *et al.* 2003) and took place in a region known not to generate large earthquakes. This situation is reminiscent of the 2011 March 11 M_w 9.0 Tohoku-Oki, Japan, earthquake and the ensuing tsunami, which came as a surprise to the seismological community. Even though Japan has some of the world's most complete documentation of historical earthquakes, there was no indication of such a large earthquake occurring where the Pacific plate subducts beneath the Okhotsk plate (Ito *et al.* 2011) in the last 1000 yr. On the other hand, relatively small earthquakes of magnitudes 7.6–7.8 have been occurring regularly in the Japan subduction zone over the last decades. The rupture areas of recent events have been located at depths of 10–40 km (e.g. Seno *et al.*

1980; Yamanaka & Kikuchi 2004; Okada *et al.* 2005), with no slip near the trench, except during the 1896 Sanriku tsunami earthquake, which apparently broke only the part of the megathrust fault that is near the trench (Tanioka & Satake 1996). In contrast, during the 2011 Tohoku earthquake, large slip occurred over both deeper and shallower parts of the subduction interface, with as much as 60 m of slip near the trench (Shao *et al.* 2011; Simons *et al.* 2011; Iinuma *et al.* 2012).

A similar situation may be observed in Mexico. Fig. 1 shows rupture areas of subduction thrust earthquakes with magnitudes $M_w \geq 7.0$. These areas have been compiled from numerous sources and are based on locations of aftershocks using local, regional or teleseismic data and, in some cases, tsunami modelling (from west to east: 1932—Singh, Ponce & Nishenko 1985; 1995—Pacheco *et al.* 1997; 2003—Singh *et al.* 2003; 1973—Reyes, Brune &

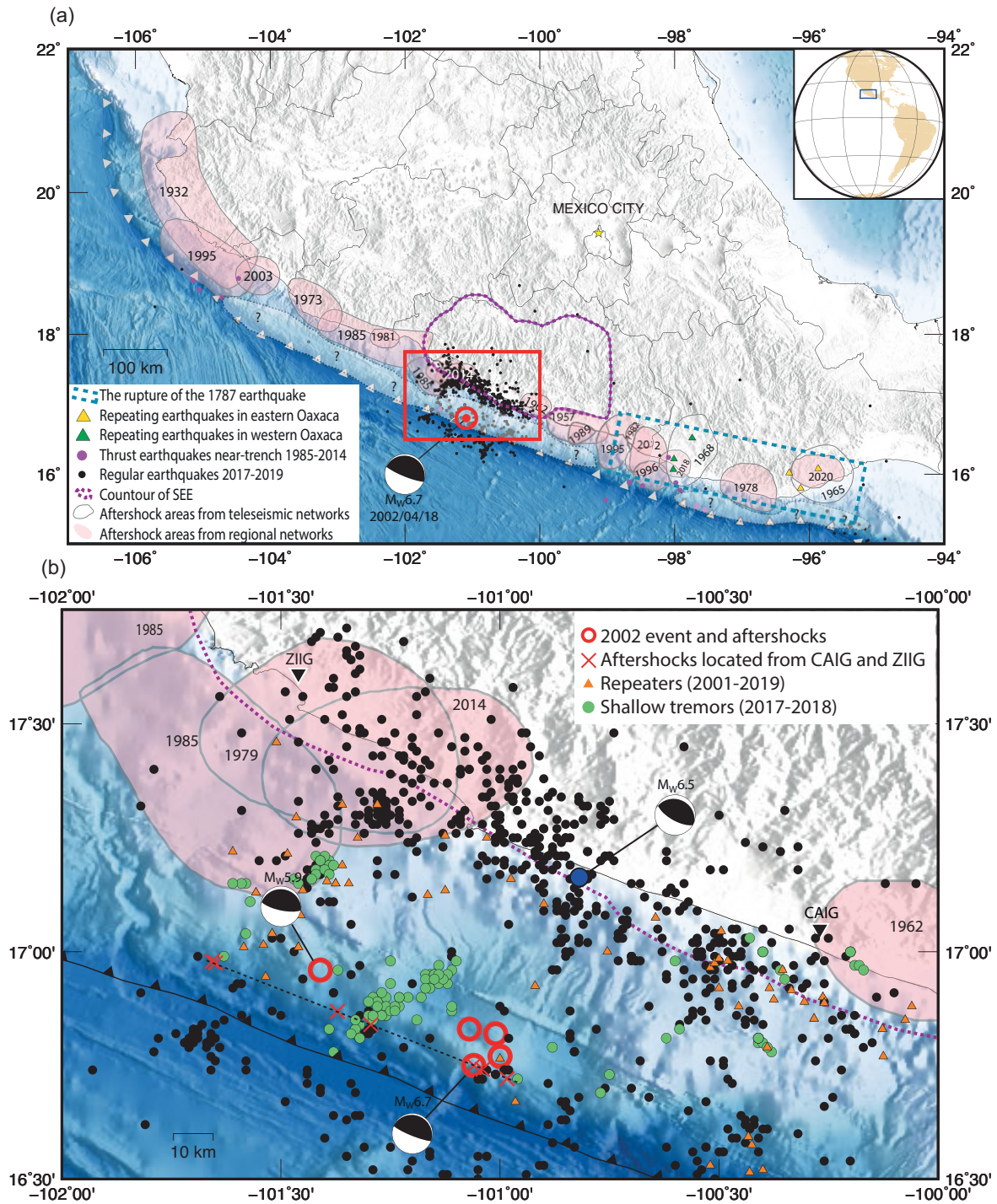


Figure 1. (a) Rupture areas of large earthquakes along the Mexican subduction zone estimated from regional and local seismograms and/or tsunami modeling (in pink) and from teleseismic seismograms (in white). Red circle shows the 2002 April 18 M_w 6.7 earthquake. Purple dots represent shallow-dipping interface thrust earthquakes near the trench, with distances less than 15 km ($5.3 \leq M_w \leq 8.0$), and characterized by deficient high-frequency radiation (Singh *et al.* 2016). Purple dashed line shows the average areal extent of 6 cm contour of slow slip events of 2001–2002, 2006, 2009–2010 and 2014 (Radiguet *et al.* 2012, 2016). Rupture areas of large earthquakes, with two exceptions, have left the upper 30 km of the plate interface unbroken (shown by quotation marks enclosed by dashed elongated contours). Black circles are locations of regular earthquakes (2017–2019) from Plata-Martínez *et al.* (2021). Yellow and green triangles are repeating large earthquakes in eastern and western Oaxaca, respectively, from Singh *et al.* (2024). The dashed blue rectangle shows the estimated rupture length of the 1787 March 28 earthquake, inferred from intensity data and tsunami reports along the coast (Suárez & Albiní 2009). An enlarged view of the rectangular area marked in red is shown in b. (b) Red circles: 2002 April 18 main shock and its aftershocks (Iglesias *et al.* 2003; Pacheco & Singh 2010). Green dots and orange triangles are locations of shallow tremors (2017–2018) and of repeaters (2001–2019), respectively, from Plata-Martínez *et al.* (2021). Red crosses: aftershocks located from the CAIG and ZIIG recordings only. Those five aftershocks were constrained to lie on the dashed black line, which passed through the hypocentre of the main shock. Blue circle: 2014 May 8 M_w 6.5 earthquake. Dashed black line: Aftershocks were located along a 111° – 291° profile; passing through the hypocentre of the main shock, which we take as the origin.

Lomnitz 1979; 1981, 1985a,b—UNAM Seismology Group 1986; 1979—Valdés *et al.* 1982; 2014—UNAM Seismology Group 2015; 1962a,b, 1957—Ortiz *et al.* 2000; 1989—Singh *et al.* 1989; 1995—Couboulex *et al.* 1997; 1982—Nava 1984; 1996—Iglesias *et al.* 2003; 2012—UNAM Seismology Group 2013; 2018—SSN 2021; 1968, 1978, 1965—Singh *et al.* 1980; 2020—SSN 2021). Typically, the rupture areas are centred on the coast, reaching about halfway towards the trench, leaving the near-trench segment, ~ 30 km wide, unbroken. There are two exceptions: the 1995 October 9 M_w 8.0 Jalisco and the 1996 February 25 M_w 7.2 Oaxaca earthquakes, both of which broke the near-trench region (Shapiro, Singh & Pacheco 1998; Iglesias *et al.* 2003; Hjörleifsdóttir *et al.* 2018). Thus, a segment of the plate interface, about 650 km long and 30 km wide near the trench between Colima and Oaxaca, as well as a 300 km long segment along the coast of Oaxaca, have not broken in a large event, at least in the last 90 yr. We will refer to these unbroken segments as the near-trench seismic gaps (N-T gaps). Since the aftershock areas often overestimate the rupture areas, the width of the N-T gaps may be greater than 30 km.

The Oaxaca segment of the Mexican subduction zone, between 95.4°W and 98.4°W , generally breaks in discrete and remarkably simple $M \sim 7.3$ – 7.6 earthquakes (Singh, Astiz & Havskov 1981; Chael & Stewart 1982; Singh & Mortera 1991). It ruptured in four large earthquakes in 1928: 22 March (M_w 7.5), 17 June (M_w 7.8), 4 August (M_w 7.3, M_s 7.4) and 9 October (M_w 7.6) (Anderson *et al.* 1989; Singh & Mortera 1991). The segment hosted earthquakes in 1965 (M_w 7.5), 1968 (M_w 7.3), 1978 (M_w 7.6) and most recently in 2020 (M_w 7.4). Some of these events are members of repeating earthquake sequences according to Singh *et al.* (2024); they show that there are two sequences of three repeating large earthquakes in eastern Oaxaca that occurred on 1928 March 22 (M_s 7.5), 1965 August 23 (M_s 7.6, M_w 7.5) and 2020 September 9 (M_s 7.4, M_w 7.4) and in western Oaxaca, on 1928 August 4 (M_s 7.4), 1968 August 2 (M_s 7.2, M_w 7.3) and 2018 February 16 (M_s 7.2, M_w 7.2) (Fig. 1a).

However, the Oaxaca segment does not always break in characteristic $M \sim 7.3$ – 7.6 earthquakes. For example, in 1787, the entire segment broke in one great $M_w \sim 8.6$ earthquake (Suárez & Albiní 2009) (Fig. 1a). The earthquake caused damage to the state of Oaxaca and produced a tsunami with an inundation distance of more than 6 km (García Acosta & Suárez Reynoso 1996). The large tsunami suggests a rupture along the near-trench plate interface, analogous to the 2011 Tohoku earthquake. This event prompts questions about whether such profound trench-reaching ruptures could also occur elsewhere along the Mexican subduction zone. Addressing this query necessitates a thorough understanding of the seismic behaviour along the near trench portion of the plate interface, that is, the N-T gap.

Along the Mexican subduction zone, between the towns of Acapulco and Zihuatanejo (from 100° to 101°W), is the western Guerrero seismic gap (Fig. 1), a segment that has not broken in a large earthquake since at least 1911 (Singh *et al.* 1981; UNAM Seismology Group 2015). Recently, Ramírez-Herrera *et al.* (2024) presented evidence of an event $M_w > 8$ around the year 1300 in the Guerrero seismic gap. A large earthquake ($M_w > 8$) in the gap poses significant seismic and tsunami hazards to towns in and near the epicentral region. Simulations show that an M_w 8.2 earthquake may cause accelerations in Mexico City twice those recorded during the disastrous 1985 Michoacán M_w 8.0 earthquake (Kanamori *et al.* 1993; Ordaz, Arboleda & Singh 1995) and, hence, may have a devastating effect on the city. These concerns have led to a relatively dense instrumentation of the region (see UNAM Seismology Group 2015

for a brief review). Consequently, much has been learned about the western Guerrero seismic gap. Large slow slip events (SSEs), some exceeding M_w 7.5, have been detected with a periodicity of about four years on the deeper part of the plate interface of the gap. These SSEs reach at least up to the coast and perhaps even further updip, releasing the stress aseismically. The coupling on the plate interface between the trench and the coast is estimated to be one-fourth of that on the adjacent segments (Kostoglodov *et al.* 2003; Radiguet *et al.* 2012, 2016). A large earthquake, M_w 7.3, ruptured the plate interface to the east of Zihuatanejo (ZIIG, Fig. 1) and to the west of the gap on 2014 April 18, leaving the gap mostly intact (UNAM Seismology Group 2015).

Since onshore instrumentation has limited sensitivity to detect processes occurring on the shallower portion of the plate interface, offshore instrumentation has been deployed in the region recently (Cruz-Atienza *et al.* 2018). The new data is beginning to provide valuable information on seismicity, shallow tremors and repeating earthquakes occurring on the shallow part of the interface (Plata-Martínez *et al.* 2021). In this context, a detailed study of the 2002 event with epicentre in the Guerrero gap (Fig. 1) is critically important as this may shed light on the seismic and tsunami potential of the gap as well as provide complementary information to the observations based on the offshore instrumentation. The 2002 event is also interesting as it is the earthquake with the longest duration relative to its magnitude recorded globally in the last 40 yr, as estimated from the difference between the hypocentral and centroid times (Duputel *et al.* 2013). This finding aligns with Bilek, Lay, & Ruff (2004), who showed that longer duration events tend to occur in the shallow part of megathrusts. A previous study of the earthquake by Iglesias *et al.* (2003) documents anomalously low accelerations produced by the event. Rodríguez Pérez, Ottemöller & Raesi (2015) present a kinematic source model and an estimate of the radiated energy based on teleseismic seismograms.

In view of the importance of the event, we, again, study the source characteristics of the 2002 earthquake. We first estimate the duration, dimension and rupture velocity by analysing the local and teleseismic data. We then invert teleseismic P and surface waves for the slip on the fault plane and calculate its radiated energy. We compare our results with those reported by Rodríguez-Pérez *et al.* (2015) and find similarities but also significant differences. Finally, we discuss the implications of the 2002 event for the Guerrero seismic gap in particular and the N-T gap in general.

2 RUPTURE LENGTH, DURATION AND SPEED

The long duration of the 2002 earthquake is suggested by the difference in the origin and centroid times listed in the Global Centroid Moment Tensor (GCMT) catalogue and is reported in a global study on source duration by Duputel *et al.* (2013). To confirm this and to obtain an approximate initial estimate of the rupture length and speed, we examine the P and S waveforms recorded at teleseismic distances (Fig. 2a). We use data downloaded from the IRIS datacenter (www.iris.edu) from the Global Seismograph Network—IRIS/DIA (II, Scripps Institution of Oceanography 1986), Global Seismograph Network—IRIS/USGS (IU, ASL/USGD 1988), the New China Digital Seismograph Network (IC, ASL/USGS 1992), the GEOFON Seismic Network (GE, GEOFON Data Centre 1993) and the GEOSCOPE network (G, IPGP & EOST 1982). The waveforms, plotted in Figs 2(b) and 2(c), are aligned at the initial break. The P waves are emergent at many stations, making it difficult to

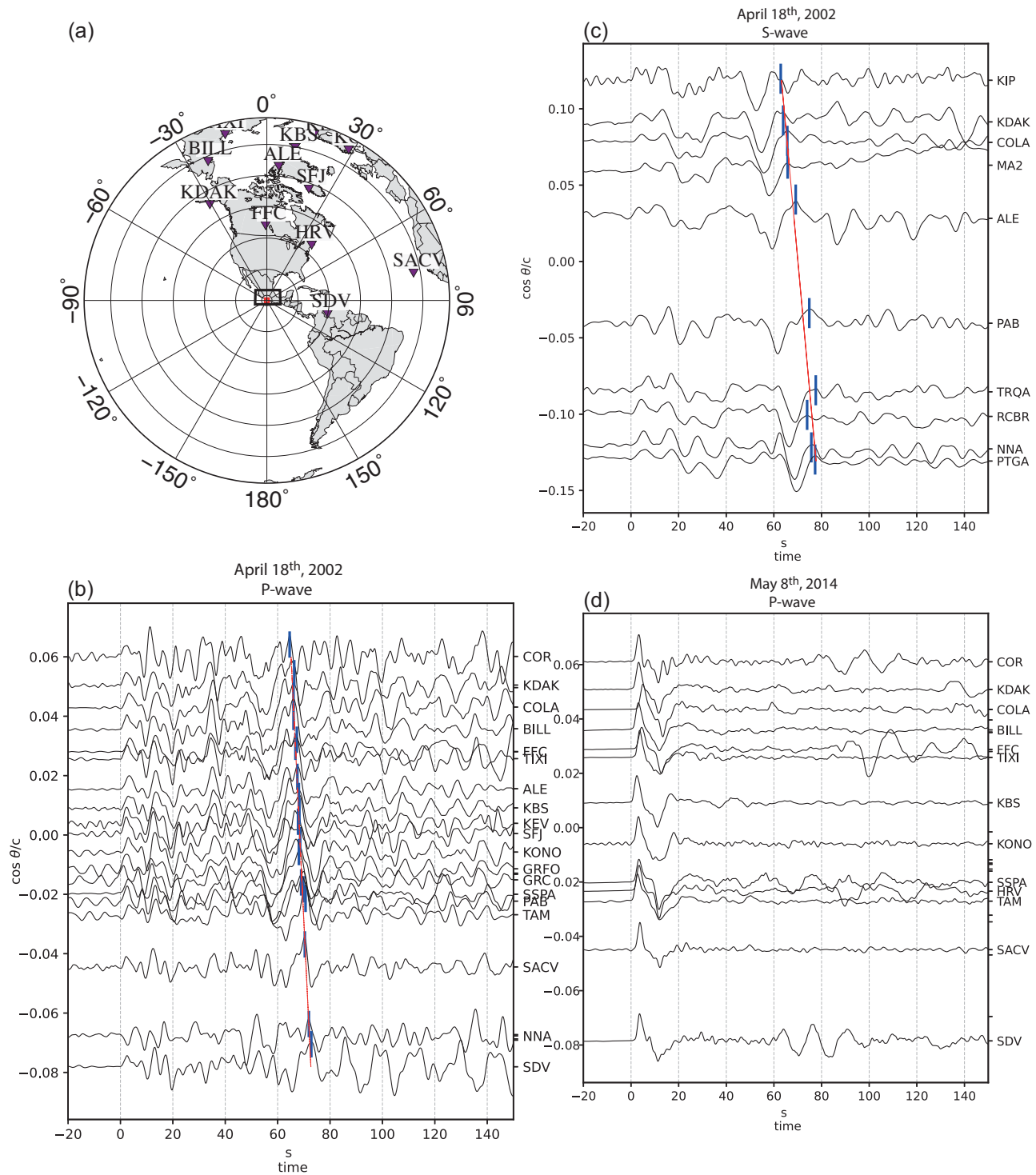


Figure 2. (a) Distribution of stations whose teleseismic earthquake records are analysed in this study. (b) *P* and (c) *S* waveforms of the 2002 event ordered by $\cos\theta/c$, where θ is the angle between the strike of the trench and the azimuth towards the stations and c is the apparent horizontal speed. Note that a selection of these traces were used in the slip distribution calculation. Blue ticks mark the last peak in the *P*- and *S*-wave train and red lines define the slopes (d) Same as (b), but for the 2014 May 8 M_w 6.5 earthquake, which occurred near the coast in the vicinity of the 2002 event.

identify the first arrival. We observe records with several coherent pulses across most stations (Figs 2b and 2c). The last and largest of these pulses arrives nearly 60 s after the first one. To confirm that these pulses are due to complexities of the source, we compare these records with those from an event with similar magnitude,

the 2014 May 8 M_w 6.5 earthquake located 55 km to the north of the 2002 event (Fig. 1b). The records of the 2014 event are very simple, with no coherent pulses arriving after the first main pulse (Fig. 2d), suggesting that the complexity in the 2002 event is due to the earthquake source, not due to structure along the path. This

also confirms that the duration of the 2002 event is at least 60 s. The difference in the waveforms of these two events, located relatively close to each other and of similar magnitude, illustrates the highly unusual character of the 2002 event.

To estimate the rupture length, we order the records by $\frac{\cos\theta}{c}$, where θ is the angle between the strike of the fault (291°) and the azimuth of each station and the apparent horizontal speed, $c = v/\sin(\delta)$, where δ is the take-off angle (using `obspy.taup` with `model='iasp91'`) and v is either v_p or v_s , the P - or S -wave velocity, respectively. We note that the last pulse at stations to the northwest (where $\frac{\cos\theta}{c}$ has positive values) arrives earlier than at stations to the southeast (where $\frac{\cos\theta}{c}$ has negative values), suggesting a rupture towards the northwest (Fig 2b and 2c). The timing between the first and last pulse at each station is given by the same equation that describes the duration of the pulse of a wave with apparent horizontal wave speed c (P or S wave) due to a unilateral rupture of length L , propagating at a rupture speed v_r (Haskell 1964):

$$\tau_c = \frac{L}{v_r} - \frac{L \cos\theta}{c}. \quad (1)$$

Using a linear regression with the time of the second large pulse observed at each station as a function of $\frac{\cos\theta}{c}$ (Fig. 2b), the intercept of the line with $\frac{\cos\theta}{c} = 0$ gives the event duration $\tau = \frac{L}{v_r} = 68$ and 70 s, with P and S waves, respectively. We then estimate L from the slope, that is, 54 and 58 km, with P and S waves, respectively. Finally, we can estimate the rupture speed $v_r = L/\tau = 0.79 \text{ km s}^{-1}$ and $v_r = 0.83 \text{ km s}^{-1}$.

The directivity and long duration of the source are also in agreement with the local data of two stations, ZIIG and CAIG, located in different azimuths (φ) and at a similar distance from the epicentre to the station location (R). Unfiltered vertical velocity traces recorded at stations ZIIG ($R = 105.4 \text{ km}$, $\varphi = 337^\circ$) and CAIG ($R = 93.4 \text{ km}$, $\varphi = 68^\circ$) are presented in the top panel of Fig. 3 (see Fig. 1b for station locations). The panels below illustrate band-pass filtered traces at the two stations. The high-frequency signals (2–4 Hz; 4–8 Hz), most likely related to bodywave radiation, have a longer duration at CAIG than at ZIIG. A large burst of high-frequency energy ~ 55 to 65 s after the P -wave arrival is clearly seen at ZIIG but not at CAIG. A complex rupture propagating NW towards ZIIG and away from CAIG and culminating in a large burst of energy after ~ 60 s agrees with both the teleseismic and local data.

3 AFTERSHOCKS

A rough estimate of the rupture dimension can also be obtained from the aftershock distribution. Unfortunately, due to the non-impulsiveness of near trench events resulting in emergent phase arrivals, the small number of seismic stations located onshore and the difficulty in picking the S phase, only four aftershocks could be located (Iglesias *et al.* 2003; Pacheco & Singh 2010). These events, including the main shock, are listed in Table 1, and their locations are shown in Fig. 1(b). Three of these aftershocks are located close to the main shock. The locations suggest a probable rupture length of about 45 to 50 km. To increase the number of aftershocks, we use more events located from regional data. We selected those that had clear P -wave arrival at the closest two stations, CAIG and ZIIG (Fig. 1b). For the main shock and each of the aftershocks, we measured T_{diff} , where $T_{\text{diff}} = (T_{P,\text{CAIG}} - T_{P,\text{ZIIG}})$ and $T_{P,\text{CAIG}}$ and $T_{P,\text{ZIIG}}$ are P -wave arrival times at CAIG and ZIIG, respectively. Since the depth of the events is poorly constrained, we assumed that they were located on the plate interface near the trench, that is, at 7 km depth. We further assumed the P -wave velocity as 5.8 km s^{-1} and that all aftershocks

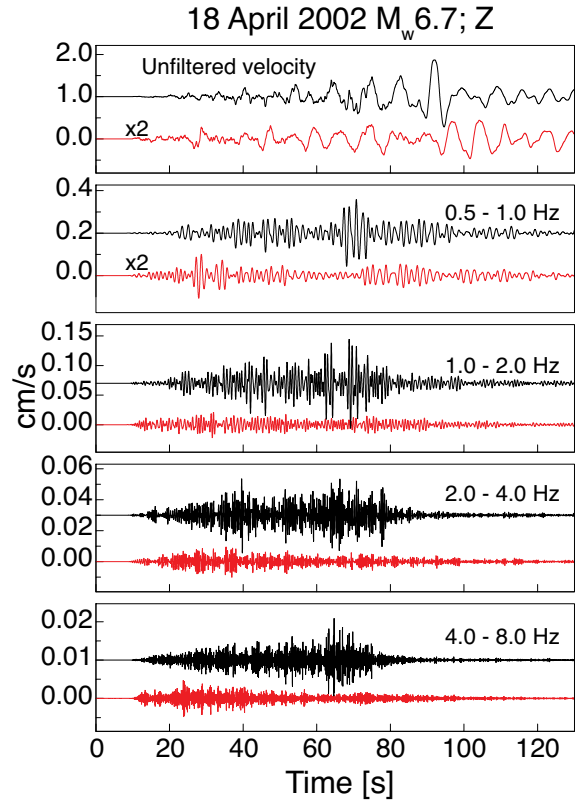


Figure 3. Velocity (Z component) recorded at stations ZIIG and CAIG (Fig. 1b), aligned by first P -wave arrival at 10 s. Black: ZIIG ($R = 105 \text{ km}$; $\varphi = 337^\circ$); Red: CAIG ($R = 93 \text{ km}$; $\varphi = 68^\circ$). Top panel: unfiltered velocity. Other panels: band-pass filtered velocity. CAIG traces in the top two panels are multiplied by 2.

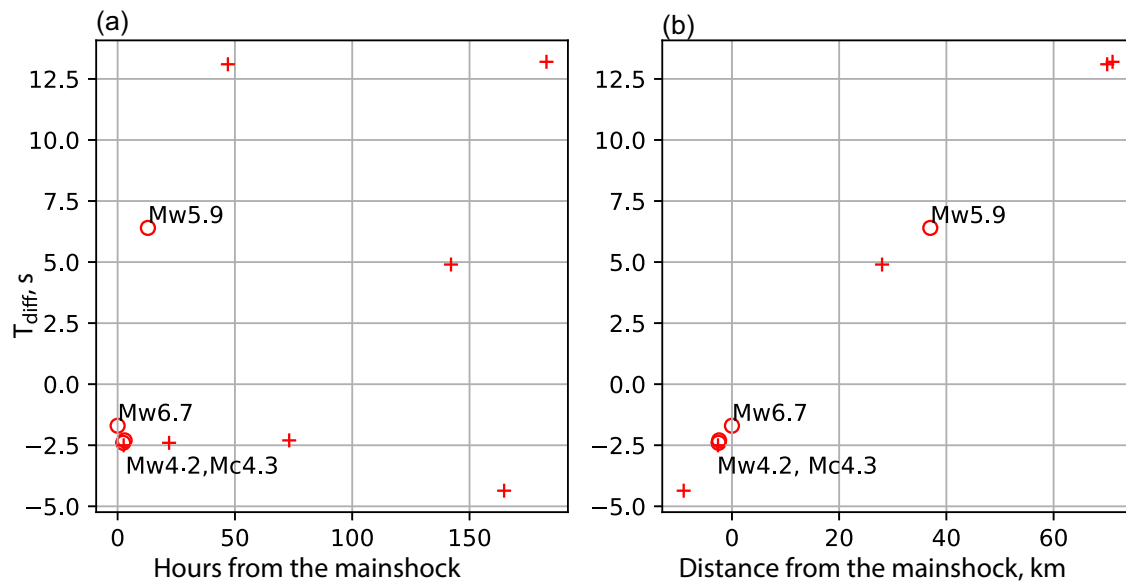
were located along a 291° profile, constrained to pass through the hypocentre of the main shock, which we take as the origin and define a positive distance from the origin towards 291° and negative to 111° . We determined the location of the events along this profile corresponding to the observed values of T_{diff} . Fig. 4(a) shows T_{diff} as a function of time. As expected, the aftershock activity is relatively high during the first 24 hr. T_{diff} of the main shock and 11 aftershocks, which occurred during the first eight days, ranged between -4.4 and 13.2 s . A plot of T_{diff} versus the distance of the aftershock measured from the main shock location is shown in Fig. 4(b). The observed T_{diff} values correspond to the distances between -10 and 70 km . The distance from the origin to the furthest aftershock in the positive direction, 70 km , represents an upper bound of the unilateral rupture length and is in rough agreement with the rupture length of $54\text{--}58 \text{ km}$, obtained in the previous section assuming a unilateral rupture.

4 DISTRIBUTION OF SLIP

Slip distribution on the fault was mapped using teleseismic P and SH waves recorded at 19 and 15 stations, respectively, with a relatively good azimuthal distribution (Fig. 5). Additionally, we used Rayleigh and Love waves registered at 5 and 6 stations, respectively (Fig. 6). The stations were located at angular distances between 30° and 110° . The P - and S -wave arrival times were picked manually, as described earlier, to remove the effect of 3-D structure along their trajectories on the traveltime of the waves. We took special care to ensure all

Table 1. Source parameters the 2002 April 18 earthquake and its aftershocks (modified from Iglesias *et al.* 2003).

Date yymmdd	Time	Latitude (deg)	Longitude (deg)	Depth (km)	Strike (deg)	Dip (deg)	Rake (deg)	M_0 (N-m)	Magnitude
020 418 ^{a,c}	05:02:43.1	16.75	-101.06	6.0	291°	9°	89°	1.5×10^{19}	6.7 M_w
020 418 ^{b,c}	07:24:27.3	16.77	-101.00	2.0	—	—	—	2.40×10^{15}	4.2 M_w
020 418 ^d	07:37:28								
020 418 ^c	08:01:34.9	16.82	-101.01	2.0	—	—	—	—	4.3 M_c
020 418 ^c	08:28:05.7	16.83	-101.07	2.0	—	—	—	—	4.3 M_c
020 418 ^{a,c}	17:57:23.4	16.96	-101.4	4.0	273°	17°	81°	8.88×10^{17}	5.9 M_w
020 419 ^d	02:58:47	—	—	—	—	—	—	—	—
020 420 ^d	04:02:35	—	—	—	—	—	—	—	—
020 421 ^d	06:09:36	—	—	—	—	—	—	—	—
020 424 ^d	03:08:20	—	—	—	—	—	—	—	—
020 425 ^d	01:47:05	—	—	—	—	—	—	—	—
020 425 ^d	19:50:14	—	—	—	—	—	—	—	—

^aMoment and focal mechanism from Global CMT catalogue.^bMoment from regional data. From Iglesias *et al.* (2003).^cEpicentre and depth from local and regional data. Depth is poorly controlled. From Iglesias *et al.* (2003).^dLocation relative using only CAIG and ZIIG stations.**Figure 4.** (a) $T_{\text{diff}} = (T_{\text{P,CAIG}} - T_{\text{P,ZIIG}})$, where $T_{\text{P,CAIG}}$ and $T_{\text{P,ZIIG}}$ are P -wave arrival times at CAIG and ZIIG, respectively, of aftershocks. T_{diff} of the main shock and the aftershocks as a function of time measured from the origin time of the main shock. Red circles: main shock and aftershocks located by Iglesias *et al.* (2003) and Pacheco & Singh (2010). Crosses are other aftershocks located only from the CAIG and ZIIG recordings. (b) T_{diff} as a function of the distance of the aftershocks from the main shock location, assuming that they occurred along a line with azimuth 291° passing through the main shock. A P -wave speed of 5.8 km s⁻¹ is assumed.

traces aligned on the same phase. The instrument response was deconvolved from the original records, resulting in displacement seismograms, which were subsequently band-passed between 1 and 80 s for the body-wave records and 170–250 s for the surface-wave seismograms.

To invert the observed records, we used the Fast Finite Fault (FFF) inversion algorithm (Ji, Wald & Helmberger 2002a, b). The algorithm uses a simulated annealing method that minimizes the weighted difference between wavelet coefficients of observed and simulated seismograms. The slip on each subfault has an asymmetric time function (Ji *et al.* 2003), and smoothing is applied both to the slip distribution and the rupture contours (Shao *et al.* 2011). The local velocity structure at the source is obtained from Crust 2.0. Other details of the procedure are given in Ji *et al.* (2002a,b).

Several subjective choices of parameters need to be made in the inversion, and each can have a significant effect on the result. We

used a fault strike of 291° and a dip of 9°, consistent with the GCMT solution. The fault strike agrees with the general strike of the trench in this region (Fig. 1). The rake was allowed to vary within $\pm 45^\circ$ from an average rake of 90°. Since the slip on the fault plane is determined relative to the hypocentre, its location affects the geographical locations of the slip patches. We used the epicentre listed in Table 1 (16.75°N, 101.06°W) reported in a study that carefully relocated events in the Guerrero segment of the Mexican subduction zone (Iglesias *et al.* 2003). The depth of the event is poorly constrained. We assigned it a depth of 7 km, consistent with the assumed plate interface. Guided by the rupture velocity obtained earlier from direct analysis of the teleseismic P and S waves, we allowed it to vary between 0.5 and 1.5 km s⁻¹, with the average constrained to 1 km s⁻¹. Observed and synthetic waveforms corresponding to our preferred model are compared in Figs 5 and 6. We draw attention to several stations located at distances less than

Body Wave

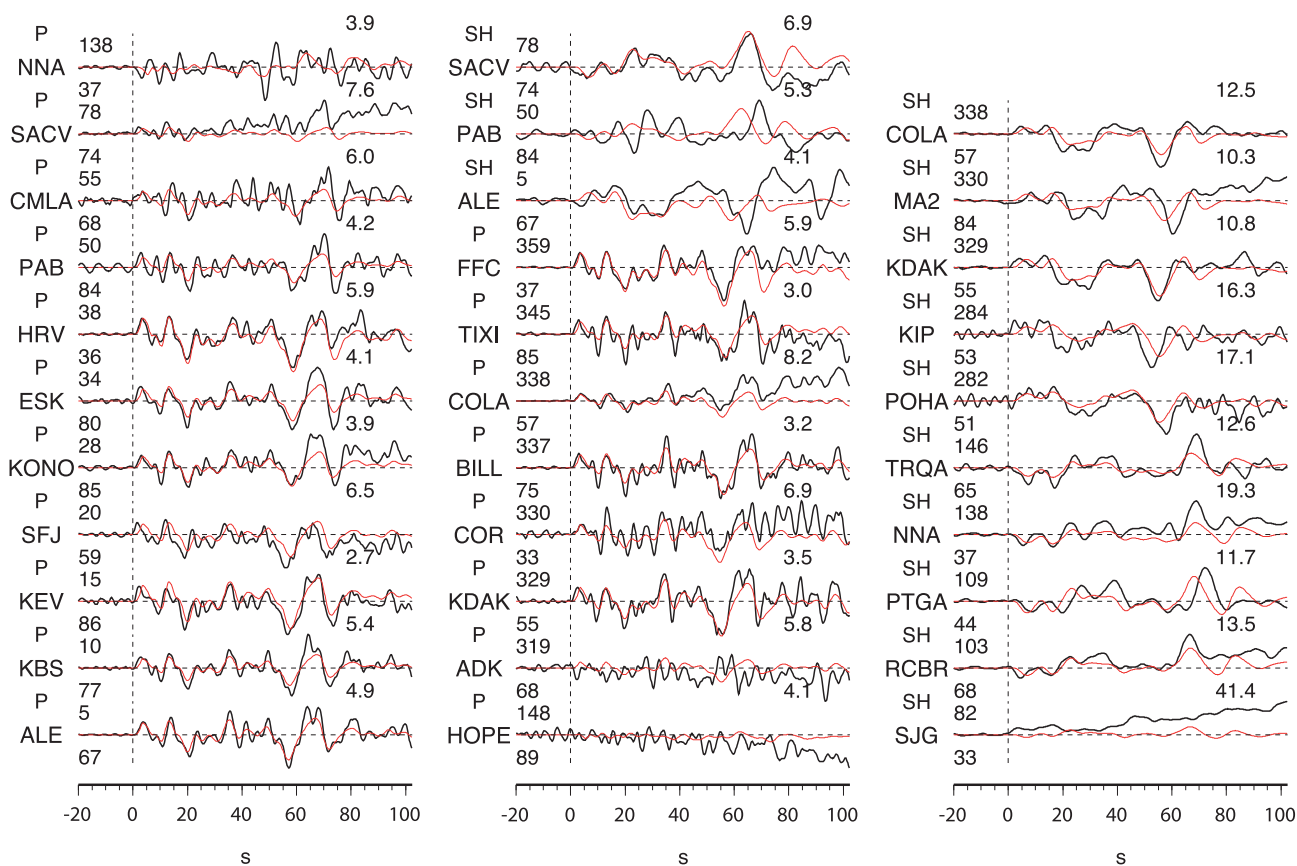


Figure 5. Comparison of observed (black) and synthetic (red) teleseismic P and SH displacement waveforms. Synthetics correspond to our preferred slip model. Labels on each trace indicate from top to bottom, left to right. Vertical (P) and transverse (SH) component, the station name, azimuth from the source, angular distance and relative amplitude of each trace.

Surface Wave

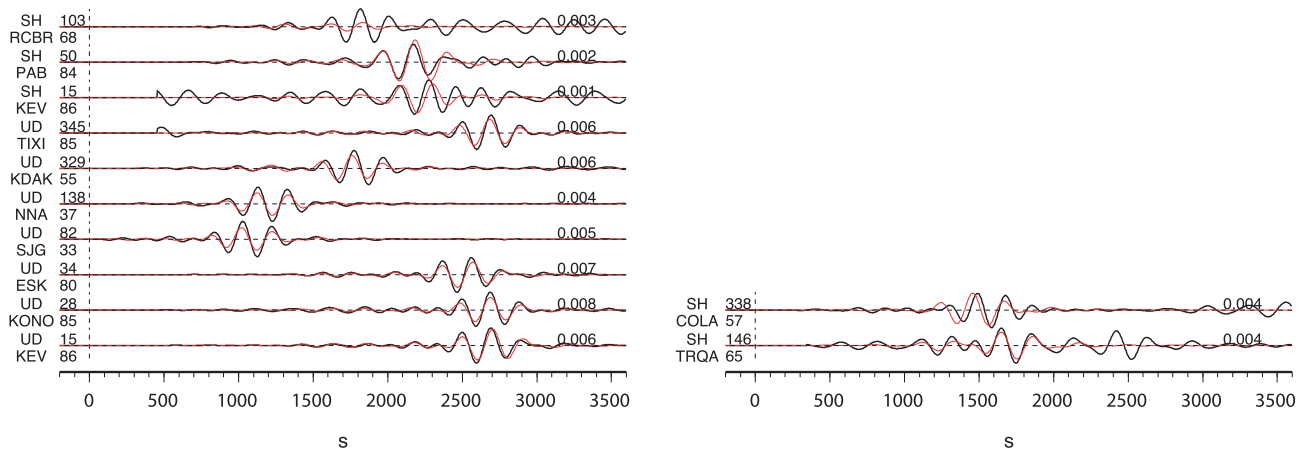


Figure 6. Comparison of observed (black) and synthetic (red) surface-wave displacement waveforms. Synthetics correspond to our preferred slip model. Labels on each trace indicate from top to bottom, left to right. Vertical (UD) and transverse (SH) component, the station name, azimuth from the source, angular distance and relative amplitude of each trace.

40 degrees (COR, FFC, HRV, NNA and SJG), which exhibit a PP arrival. This phase has a significantly different slowness than the direct P wave, and due to the wide range of epicentral distances considered, the PP arrival does not appear coherently across many traces. Therefore, although its inclusion slightly increases the uncertainty at times later than 60 s (when the PP phase arrives at ~ 33 degrees), incorporating the portion of the waveform where PP waves occur does not significantly impact the final solution.

The inverted slip, shown in Fig. 7, is distributed in a narrow strip immediately downdip from the trench. The maximum slip is 80 cm, and the slip area is approximately 10 km $\times$ 60 km, with almost all the slip occurring above 8 km depth. In this model, we see two distinct slip patches that are reflected in the two main pulses of the moment rate function (MRF) (Fig. 7b). The small patch of slip at the bottom of the slip model is poorly resolved and may be an artefact. The largest aftershock, M_w 5.9, which occurred ~ 13 hr later (Table 1 and Fig. 1b), was located near the second slip patch (Fig. 7). Even though there are only two main patches in the smoothed slip model, the overall ruggedness of the MRF, as observed directly in the multiple impulses seen in the P and S arrivals, suggests a relatively complex rupture to the smooth ruptures suggested for tsunami earthquakes, as for example the 2010 M 7.8 Mentawai earthquake (Lay *et al.* 2011; Newman *et al.* 2011). Furthermore, the waveforms from the direct P and S waves have such a long duration that they start interfering with reflected and converted phases. Hence, it is difficult to accurately determine the end of the rupture. The inversion procedure allows for a rupture duration of up to 105 s (Fig. 7b), but most of the slip has occurred before a duration τ of about 72 s, and we take that as the end of the rupture. While we cannot exclude the possibility of a longer rupture, we attribute the imaged slip occurring after this time as an artefact of the algorithm fitting noise in the data.

5 MOMENT RATE SPECTRUM AND RADIATED ENERGY

The moment rate spectrum and the seismic energy give an insight into the earthquake rupture dynamics. For that reason, we estimated the moment rate spectrum, $\hat{M}(f)$, of the 2002 event and computed the radiated seismic energy, E_R , using the expression (Venkataraman & Kanamori 2004),

$$E_R = \left[1 + \frac{3V_\alpha^5}{2V_\beta^5} \right] \frac{8\pi}{15\rho_h V_\alpha^5} \int_0^\infty f^2 |\hat{M}(f)| df, \quad (2)$$

where $V_{\alpha,\beta}$ are the P - and S -wave velocities, ρ_h is the density at the source region and $\hat{M}(f)$, the moment rate spectrum (also called the source spectrum). We took $V_\alpha = 5800$ m s $^{-1}$, $V_\beta = 3600$ m s $^{-1}$ and $\rho_h = 3000$ kg m $^{-3}$.

$\hat{M}(f)$ was constructed in a manner similar to Ye *et al.* (2016), combining the spectrum of the MRF retrieved from the finite fault inversion (Fig. 8a) at lower frequencies with the spectrum computed from teleseismic P waves at higher frequencies. To compute $\hat{M}(f)$ from the spectra of teleseismic P waves, we cut the records with a cosine window from $t_p - 15$ s to $t_p + 105$ s. This time window includes the P -wave group (P , pP and sP phases) and brackets the estimated duration of the rupture. The records were corrected for instrument response, followed by computation of the Fourier amplitude spectra, which were then corrected for attenuation and site effect following Pérez-Campos, Singh & Beroza (2003). Finally, $\hat{M}(f)$ was computed using the method of Boatwright & Choy

(1986) and modified by Pérez-Campos & Beroza (2001). Fig. 8(b) shows the logarithmic average of $\hat{M}(f)$ estimated from P waves, and $\hat{M}(f)$ from the MRF obtained from finite fault (FF) modelling. Our composite spectrum, shown in Fig. 8(c), patches $\hat{M}(f)$ computed from MRF for $f \leq 0.06$ Hz and from P waves for $(0.06 \leq f \leq 1)$ Hz. Scalloping in the spectrum is a consequence of the two distinct pulses in the MRF. For reference, Fig. 8(c) shows theoretical spectra from the Brune source model (Brune 1970) with a stress drop of 3 and 0.37 MPa. The observed spectrum is severely depleted at all frequencies compared with the theoretical spectrum, with a stress drop of 3 MPa. The observed spectrum between 0.2 and 1 Hz is consistent with the theoretical spectrum with the stress drop of 0.37 MPa, but it is still highly deficient at lower frequencies.

We assumed that the composite $\hat{M}(f)$ falls off as f^{-2} at $f > 1$ Hz and computed the total radiated seismic energy, E_R , from eq. (2) as 2.21×10^{13} J. This yields seismic moment-scaled radiated energy, E_R/M_0 , of 1.45×10^{-6} , a low value that is typical of tsunami earthquakes (Okal & Newman 2001; Venkataraman, Rivera & Kanamori 2002; Ye *et al.* 2016).

6 DISCUSSION

A consistent picture of the source process of the M_w 6.7 2002 earthquake in offshore Guerrero emerges from the analysis of teleseismic P and S waves, local recordings and aftershock distribution.

In the following discussion, we first present a comparison of our results with previous studies. We focus on the rupture length, duration and speed, highlighting similarities and differences. In the second part, we analyse the distribution of slip, MRF and energy, examining their relationship with local tectonic features.

6.1. Rupture length

In this study, we find that the rupture length is at least 60 km, as estimated from body waves when assuming unilateral rupture, and less than 80 km, as observed from aftershocks when assuming that aftershocks occur at the edges of the rupture. The finite-slip distribution suggests a rupture length close to 80 km. The slip distribution, as imaged by Rodríguez-Pérez *et al.* (2015), is somewhat smaller but consistent with the first sub-event observed in this study.

6.2. Rupture duration

The rupture duration obtained in this study, both from direct observations of coherent pulses observed on P - and S -wave records, $\tau = 68$ to 70 s, and from finite-fault inversion of slip, 72 s, is very long given the size of the earthquake.

The expected duration $\tau = 1.2 \times 10^{-8} M_0^{1/3}$ (M_0 in dyne-cm) from Duputel *et al.* (2012) for a M_w 6.7 is 12.8 s. A similar scaling law developed for the Mexican subduction zone (where the duration is $\tau = 1.18 \times 10^{-8} M_0^{1/3}$, M_0 in dyne-cm) predicts a duration of 6.1 s (Singh *et al.* 2000). Our results, therefore, indicate that the duration of this event is an order of magnitude longer than expected from scaling laws.

Other studies have noted the anomalous nature of this event. A rough estimate of the duration can be obtained from moment-tensor analysis as $\tau = 2T_c$, where T_c is the difference between the centroid time and the origin time of an event (e.g. Duputel *et al.* 2013; Ye *et al.* 2016). For this event, the GCMT and WCMT (Duputel *et al.* 2013) catalogues list T_c of 29.5 and 35 s, respectively, corresponding

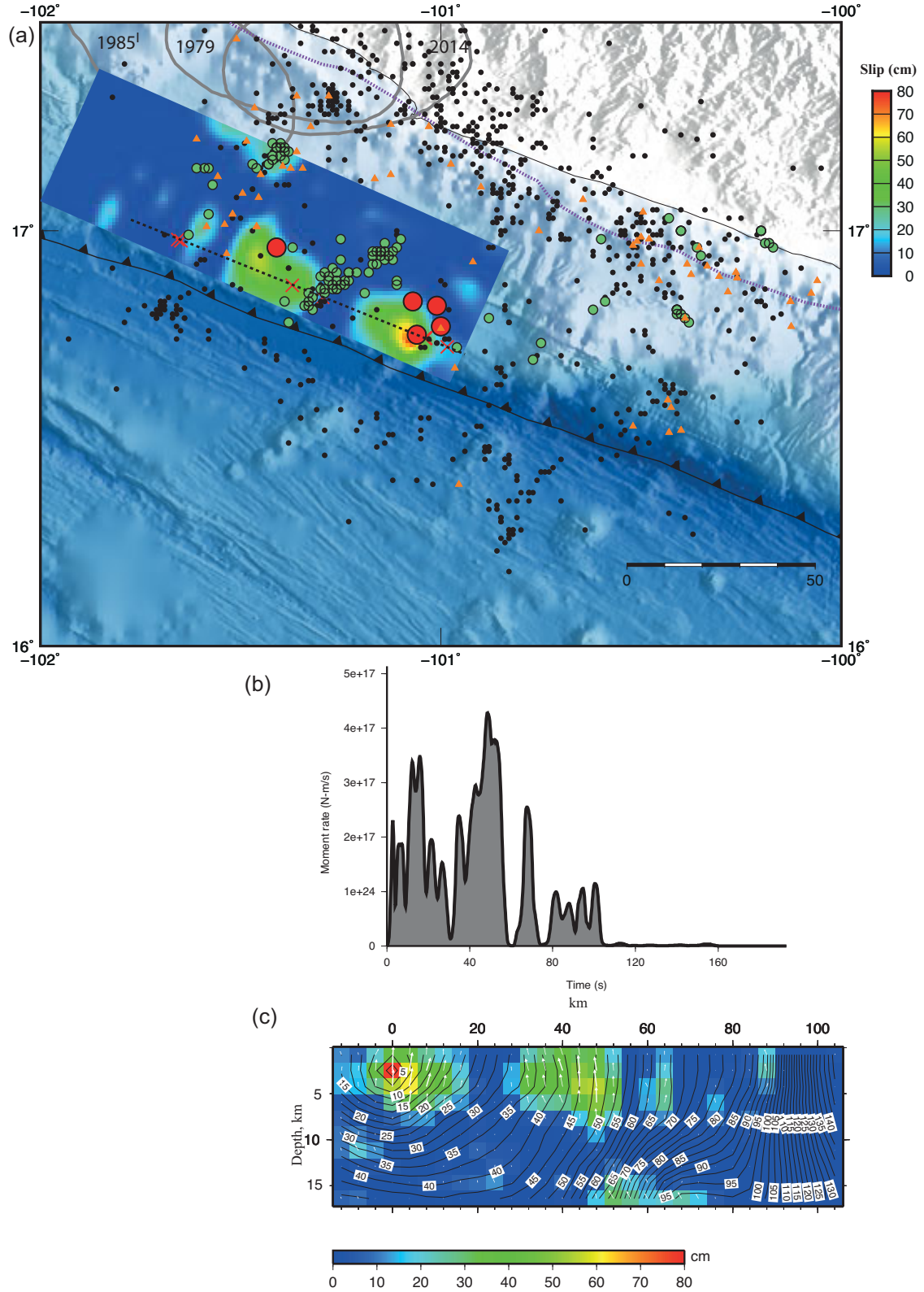


Figure 7. (a) Slip distribution in the 2002 April 18 earthquake. Colours indicate slip amplitude and direction with white arrows. Red circles show the epicentre of the main shock and its aftershocks. Blue circle is the epicentre of the 2014 May 8 M_w 6.5 earthquake, whose teleseismic records are compared with those of the 2002 event in Fig. 2. Rupture areas of previous large events are shown by grey contours. Orange triangles, green and black circles are locations of repeaters (2001–2019), shallow tremors (2017–2018) and regular earthquakes (2017–2019), respectively, from Plata-Martínez *et al.* (2021). Purple dashed line shows the average areal extent of 6 cm contour of slow slip events of 2001–2002, 2006, 2009–2010 and 2014 (Radiguet *et al.* 2012, 2016). (b) Moment rate function. (c) Slip as a function of depth with rupture front at different times (white labels).

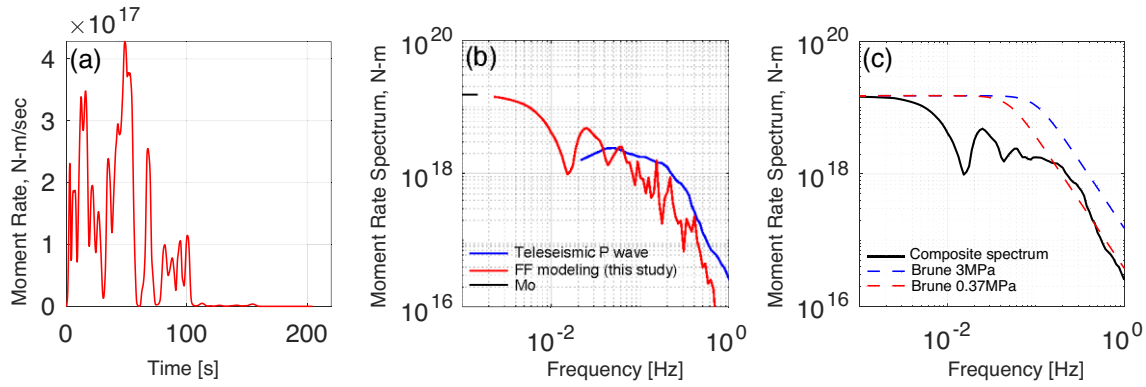


Figure 8. (a) Moment rate function (MRF) estimated from finite fault modeling. (b) Moment rate spectrum (MRS) from MRF retrieved by finite fault modelling (red). The spectrum is smoothed using 1/3 octave-band filtering. Also shown is the MRS estimated from teleseismic *P* waves (blue). (c) Composite MRS (black) and Brune source spectra for stress drops of 3 MPa (dashed blue) and 0.37 MPa (dashed red).

to a duration of 59 and 70 s, respectively, making it the event with the most anomalous value of duration in for the WCMT catalogue (Duputel *et al.* 2013).

This event was also analysed using the SCARDEC method (Vallée & Douet 2016), but the routine analysis does not permit such a long duration for a moderate-sized event, and therefore, the event was not reported in the data base. However, when relaxing this constraint, the MRF obtained for the 2002 earthquake is similar to the one obtained here: two main pulses followed by a smaller one and a duration of 70 s (Fig. 9; Martin Vallée, personal communication, 2022).

Finally, the moment-rate function reported by Rodríguez-Pérez *et al.* (2015) has a shorter duration of 38 s but with a single peak, which is consistent with the first main slip patch of our model. We suggest that their analysis does not include the second sub-event imaged in this study, hence the discrepancy. We conclude that the estimate of the unusually long rupture duration of 70 s or longer is robust.

6.3. Rupture speed

The average rupture speed can be estimated by the rupture length divided by the rupture duration or the time it takes for the rupture front to reach from one end of the fault to the other. However, the total duration of the event depends on both the rupture duration and the rise time or the time it takes for each point of the fault to slip. Therefore, while the total duration is relatively well determined, it is more difficult to separate the contribution of the rupture—and the rise times. In this study, we have attempted to do that in three ways: by direct observations of body waves, by observations of local records and by inverting for a finite-fault slip distribution. In the case of direct observations of body waves, the move-out between the initial and last pulse gives us the difference in the ‘centroid-time’ of the two pulses. We assumed that the rise time is constant along the fault, and therefore, the difference in timing between the two pulses gives the rupture duration. However, similar waveforms could be generated in a scenario where the rise time varies along the fault in such a way that the rupture travels fast, but the part of the fault that is further from the hypocentre slips slower. However, this would predict diminished high-frequency radiation from the end of the fault, which is not consistent with the local observations. Furthermore, the finite-fault slip distribution does not show a fast rupture velocity with long rise times towards the end of the fault.

The rupture speed obtained in this study for the whole rupture is similar to that obtained for the first sub-event, $v_r = 0.8 \text{ km s}^{-1}$, by Rodríguez-Pérez *et al.* (2015). We finally conclude that while there is large uncertainty in the value of the rupture speed, the main contribution to the long duration does not come from a long rise time.

6.3.1. Location of slip and relationship with local tectonics

We note that the hypocentre of the 2002 event and the location of the first slip patch coincide with residual bathymetric and gravity high (Plata-Martínez *et al.* 2021), presumably related to subducting seamounts (Nopal ridge from Černý *et al.* 2020) in the region (Fig. 7). Slow slip events (SSEs), tectonic tremors, and, in some cases, tsunami earthquakes have been attributed to subduction of seamounts (e.g. Todd *et al.* 2018).

We conclude that the rupture most probably began at a subducted seamount near the trench (Fig. 7). It propagated mostly unilaterally towards NW, parallel to the trench. The finite fault inversion reveals a complex moment rate function, with two dominant pulses separated by about 30 s. The first pulse occurs near the beginning of the earthquake, presumably related to the rupture of the seamount, and the second, a larger pulse, towards the end of the rupture. The largest aftershock (M_w 5.9) was located near the slip area corresponding to the second pulse. A linear distribution of shallow tremors, oriented NE-SW, separates the two slip patches (Plata-Martínez *et al.* 2021). However, no unusual bathymetric feature is discernible that may be related to these tremors (Fig. 7).

Several observations suggest that the 2002 event, although relatively small in magnitude, has characteristics of a tsunami earthquake: the slip occurs very close to the trench, the rupture speed is slow, the high-frequency radiation is deficient, and, in common with tsunami earthquakes, the moment-scaled radiated energy is low and the duration of the event is extraordinarily long compared to the expected duration from scaling relations. In Fig. 9, we compare the MRF of the 2002 event with those of five other known tsunami earthquakes: 1992 M_w 7.7 Nicaragua, 1996 M_w 7.4 Peru, 2010 M_w 7.7 Mentawai, 1994 M_w 7.8 Java and 2006 M_w 7.7 Java. The MRFs of the five large earthquakes are taken from Ye, Kanamori & Lay (2018). The source complexity of the 2002 earthquake bears a resemblance to that of the 2006 Java earthquake, given that both MRFs contain several pulses, which distinguish them from other MRFs.

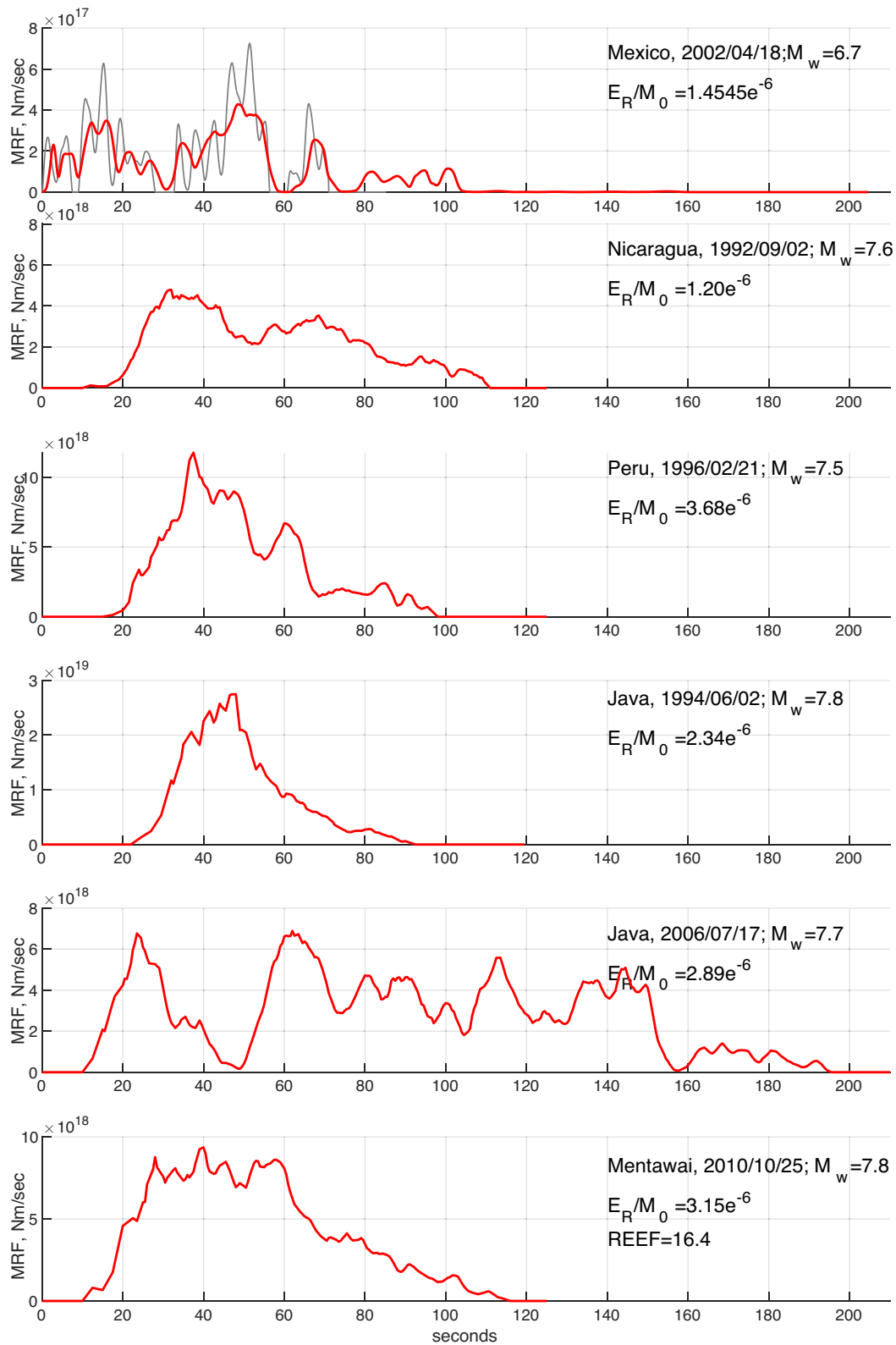


Figure 9. Comparison of moment rate function (MRF) of the 2002 tsunami earthquake of Guerrero, Mexico, with those of five large tsunami earthquakes. MRF and E_R/M_0 of the 2002 event are from this study; for other events, they are taken from Ye *et al.* (2018). Additionally, this plot shows the MRF from Martin Vallée, personal communication, 2022 (Grey line).

The 2002 magnitude was less than that of other events, consistent with the moment-scaled radiated energy.

In this study, we calculated E_R and E_R/M_0 resulting from a composite spectrum. Those values are smaller than those reported by Rodríguez-Pérez *et al.* (2015), but both results are consistent with the range of tsunami earthquakes (Okal & Newman 2001; Venkataraman & Kanamori 2004). The low E_R suggests the dissipation of a large fraction of energy during the rupture process as fracture energy with little energy to radiate seismic waves (Polet & Kanamori 2000).

Bilek, Lay, & Ruff (2004) examined earthquake source time functions for shallow subduction zone thrust earthquakes, such as the 2002 event, and identified patterns suggesting complexity in the shallow seismogenic environment. The ruggedness of the MRF of the 2002 event suggests that the uppermost segment of the plate interface in the region (width ~ 15 km) consists of relatively small but strong asperities of unstable frictional characteristics embedded in material of low frictional strength or of conditional stability. A similar conclusion has been reached from the analysis of ocean bottom seismometer recordings (Plata-Martínez *et al.* 2021). Tremors and repeaters have been observed in the shallow segment, which is likely a proxy for slow slip events and aseismic creep in this region. These shallow slow-slip events, in turn, may trigger complex earthquakes such as the 2002 event. Further downdip, seismic activity is absent over a width of about 25 km, and the seismic coupling is weak, as suggested by onshore geodetic data and the inference of aseismic slip from shallow tremors and repeaters (Radiguet *et al.* 2016; Cruz-Atienza *et al.* 2021). These observations suggest an almost freely sliding interface and a low probability of a large earthquake or long recurrence interval between large earthquakes (Plata-Martínez *et al.* 2021).

Large near-trench earthquakes along the Mexican subduction zone are infrequent. Although the subduction zone is ~ 1500 km long and the convergence rate ranges between 5 and 7 cm yr^{-1} (DeMets, Gordon & Argus 2010), only three other large ($M_w > 6.5$) earthquakes have broken the near trench area since 1990: the 1995 M_w 8.0 Jalisco earthquake (Hjörleifsdóttir *et al.* 2018), the 1996 M_w 7.2 and the 1997 M_w 6.7 Oaxaca earthquakes (Shapiro *et al.* 1998; Iglesias *et al.* 2003). It has been suggested that the apparent lack of seismicity near the trench is partially due to a ‘loading shade’ produced by the locked interface downdip of the near trench fault surface. However, most of this region has had large earthquakes in the last 100 yr, indicating that some loading should have occurred. It follows that the motion across the fault surface in the shallowest part of the plate interface near the trench is mostly accommodated by creep. However, this does not exclude the possibility that this fault surface could participate in a large seismic event when loaded with a large slip downdip, as observed in the 2011 Tohoku-Oki earthquake (Shao *et al.* 2011). This may also have occurred during the 1787 Oaxaca earthquake (Suárez & Albiní 2009). Only a wide-spread deployment of offshore seismic and geodetic instrumentation can reveal the nature of coupling on the updip, near-trench portion of the plate interface.

7 CONCLUSIONS

The Mexican subduction zone presents infrequent large near-trench earthquakes. However, this does not exclude the possibility that this area could participate in a large seismic event.

The 2002 event had a complex source process and propagated unilaterally towards the NW, parallel to the Mexican trench. Although relatively small in magnitude, it had characteristics of a

tsunami earthquake: the slip occurred very close to the trench, the rupture speed was slow, the high-frequency radiation was deficient and the moment-scaled radiated energy was low. It is possible that shallow slow-slip events may trigger earthquakes such as the 2002 event.

Finally, this opens up the possibility of tsunami earthquakes with magnitudes less than 7 near Guerrero, Mexico. Knowledge of the rupture fault could be taken into account for tsunami warnings, with the purpose of alerting the population to a potential disaster.

ACKNOWLEDGMENTS

We thank Martin Vallée for contributing the estimated source time function using his SCARDEC method. We thank UNAM/DGAPA/PAPIIT for supporting VH and KF-I. through grants IB101812, IN111316 and SKS, AI and PC through project IN109423. This work was funded by MEXT KAKENHI Grant-in-Aid for Transformative Research Areas (A) ‘Science of Slow-to-Fast Earthquakes’ (Grant Numbers JP21H05203). The views expressed herein are those of the authors and do not necessarily reflect the views of the CTBTO Preparatory Commission.

DATA AVAILABILITY

Seismograms were obtained from the IRIS Data Center, networks II, IC, IU, G and GE, as well as from SSN at www.ssn.unam.mx. We used the Python toolbox ‘ObsPy’ (Beyreuther *et al.* 2010) and Seismic Analysis Code (SAC) (Goldstein & Snoke 2005) to process the seismic data. Graphics were prepared using the Generic Mapping Tools (GMT, Wessel & Smith 1995).

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