Rupture length of the October 9, 1995 Colima-Jalisco earthquake
($M_w$ 8) estimated from tsunami data

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Abstract. We analyze tsunami data of the great 1995 Colima-Jalisco, Mexico earthquake, recorded in Cabo San Lucas, Baja California Sur, to estimate its rupture length, $L$. To model the tsunami arrival time, we assume a rectangular source area, oriented parallel to the trench, whose SE limit is fixed at the point of rupture initiation. The NW limit of the source area, i.e., $L$, is varied between 120 and 200 km. The comparison between synthetic and observed data strongly suggests that $L$ of the earthquake was 160±20 km. This length agrees with those reported in various other studies of the earthquake. It, however, disagrees with a previous study, based on the same tsunami data, which suggested that the rupture may have extended 250 km NW from the epicenter [Tanioka and Ruff, 1996]. The cause of this discrepancy is most likely an error in the timing of the records used by Tanioka and Ruff [1996]. We conclude that the earthquake only partially ruptured the Rivera-North America plate interface. A 120 km-long segment in the NW extreme of this interface, which apparently ruptured in 1932, remains presently unbroken.

Introduction

The October 9, 1995 Colima-Jalisco earthquake ($M_w$ 8.0) was a shallow, thrust event which occurred near the junction of Rivera (RIVE), Cocos (COCOS), and North American (NOAM) plates (Figure 1). The seismotectonics of the region, however, is controversial. The location and orientation of the boundary between the RIVE and COCOS plates, as well as the relative convergence rate between RIVE and NOAM plates are subject to debate [e.g., Rando et al., 1995; Kostoglodov and Rando, 1995; DeMets and Wilson, 1996] (Figure 1). DeMets and Wilson [1996] suggest a diffused RIVE-COCOS plate boundary, which is shown in Figure 1. The aftershock area of the 1995 earthquake suggests that this event may have ruptured the interplate between the subducted diffused plate boundary and the NOAM plate (Figure 1, modified from Pacheco et al., 1997).

Previous large/great earthquakes in the region occurred on June 3 and 18, 1932 ($M_S$ 8.1, 7.8). From the length of the aftershock areas, ~280 km (Figure 1), Singh et al. [1985] concluded that the 1932 earthquakes broke the RIVE-NOAM plate boundary, and that the lack of knowledge of the precise location of RIVE-COCOS boundary did not affect this conclusion.

The source characteristics of the October 9, 1995 earthquake have been discussed in several papers [Courboulex et al., 1997; Pacheco et al., 1997; Melbourne et al., 1997; Escobedo et al., 1998]. Local and regional data show that the rupture initiated about 24 km S of Manzanillo. The length of the aftershock area is 170 km. It is rectangular in shape, lies almost entirely offshore, and is parallel to the trench [Pacheco et al., 1997]. The epicenter lies near the SE end of the aftershock area, suggesting that the rupture propagated 130 km towards the NW (Figure 1). The deconvolution of surface waves of the mainshock with the corresponding waves from empirical Green's functions indicates an almost unilateral rupture propagation for 150 km towards N70°W [Courboulex et al., 1997]. Teleseismic body-wave

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Figure 1. Tectonic map of the region. Note that the location of the Rivera-Cocos plate boundary is uncertain; the diffused boundary shown is taken from DeMets and Wilson [1996]. Aftershock areas of the June 3 and 18, 1932 earthquakes are outlined by continuous line. Also shown is the aftershock area of the 1995 earthquake (dashed contour) and its epicenter (star). Triangles: Volcanoes; EPR: East Pacific Rise; EGG: El Gordo graben.
modeling gives a similar result: a unilateral rupture towards the NW for 130 to 150 km [M. Kikuchi, personal communication, 1995; Tanioka and Ruff, 1996; Escobedo et al., 1998]. The inversion of GPS deformation data shows the 1-m slip contour extending about 150 km parallel to the trench [Melbourne et al., 1996]. All of these estimates are fairly consistent and suggest a rupture length $L$ of about 150 km. However, based on modeling of tsunami data recorded at Cabo San Lucas, Baja California, Tanioka and Ruff [1996] suggested that the rupture length may have reached 250 km. If so, it would mean that the 1995 event ruptured the entire area broken in 1932. It also raises the question: if the true $L$ was 250 km why do seismic and GPS analyses give $L$ ~ 150 km? While it may be possible to miss a very slow rupture process from the analysis of seismograms, such a slip should have been detected in the GPS data. There are three possible explanations for the discrepancy. (1) $L$ was ~250 km, but, since there were only two GPS sites near the NW portion of the rupture (Chamela and Puerto Vallarta, Figure 1), which were about 100 km apart, this slip could not be detected. (2) $L$ was ~250 km but the slow seismic slip in the NW part was recuperated within one week following the earthquake when the GPS sites were recovered. This is a very unlikely explanation because the post-seismic afterslip would almost certainly increase rather than decrease the coseismic slip. (3) $L$ was ~150 km and the estimate of $L$ ~ 250 km resulted from some error in the tsunami data or in its analysis.

We note that a definitive knowledge of the rupture length of the 1995 earthquake is important for understanding the seismotectonics and the seismic hazard of the region. For this reason we reanalyze the tsunami data from Cabo San Lucas.

**Tsunami Record from the Tide Gauge of Cabo San Lucas**

Two analog, flotation-type tide gauges are in operation in Cabo San Lucas. The analog data of one tide gauge is digitized at an interval of two minutes and is transmitted via the GOES satellite to the Pacific Tsunami Warning Center, Hawaii. The analog data from the second gauge is sampled every 6 minutes and is transmitted via satellite to the Tropical Global Atmosphere Project, Hawaii. The time of both instruments is kept by a single Transmitter Platform. Data from the gauges are also recovered by Centro de Investigación Científica y de Educación Superior de Ensenada (CICESE), Baja California, Mexico, via telephone modem to the GOES receiver in USA. We checked the time stamp of the tide recording of the 1995 earthquake, sampled every two minutes and kept at CICESE. This was done by comparing the predicted tide (using harmonic constants obtained from the tide data for the period 1978-1990) with the observed tide. We found that the fit required a time correction of +16 min to the observed data. As a further check, we compared the corrected time series with the data from the second tide gauge (available from internet site http://tides.watson.hawaii.edu/cihs/). With the correction of +16 min to the tide data available at CICESE, the traces were almost identical. In further analysis, we used the corrected tide data, kept at CICESE.

Figure 2 illustrates the tide data (top) and the residual tide (bottom), where the residual tide is obtained by subtracting the predicted tide from the observed one. The time shown is relative to the centroid origin time of the earthquake, 15:36:29.4, given in the Harvard CMT solution (PDE origin time +38.4 sec).

In both traces shown in Figure 2, the first tsunami arrival can be identified. Although the noise masks the beginning of the tsunami, from an enlarged view of the residual tide, shown in Figure 3, it is clear that it occur at 57.5±2 min after the earthquake.

**Numerical Modeling of the Tsunami**

The vertical movement of the sea floor was computed using the expressions given by Mansinha and Smylie [1971]. The earthquake was modeled by a buried, rectangular fault of width 70 km, with its shallow edge located at a depth of 10 km, oriented parallel to the trench with an azimuth of 314°, dip of 12°, rake of 90°. The SE limit of the fault was taken to coincide with the point.
of rupture initiation. As an initial condition the sea-level change due to the rupture was taken to be same as the sea-floor uplift calculated from the dislocation model. This change was assumed to occur instantaneously at the centroid origin time of the earthquake, 15:36:29.4, given in the Harvard CMT solution. In prescribing the initial condition, we only took the area of sea-level change of 4 cm or greater. In the simulations the length of the fault varied from 120 km to 200 km, in steps of 10 km, while the slip was computed for a constant seismic moment, $M_w$, of $1 \times 10^{28}$ dyne-cm. The offshore four-cm contours are 20 km from the trench. Figure 4 illustrates the four-cm contour of the initial condition for rupture lengths of 140, 160, and 180 km. From Figure 4 it can be noted that the tsunami travel time to Cabo San Lucas decreases as $l$ increases.

The propagation of the tsunami was simulated by the following linearized shallow-water equations in a spherical coordinate system [Pedlosky, 1979]:

$$\frac{\partial \eta}{\partial t} + \frac{1}{R \cos \theta} \left[ \frac{\partial U}{\partial \lambda} + \frac{\partial}{\partial \theta} \left( \frac{v \cos \theta}{R \cos \theta} \right) \right] = 0$$

$$\frac{\partial U}{\partial t} + \frac{\partial}{\partial \lambda} \left( gh \frac{\partial \eta}{\partial \lambda} \right) = fV$$

$$\frac{\partial V}{\partial t} + \frac{\partial}{\partial \theta} \left( gh \frac{\partial \eta}{\partial \theta} \right) = \frac{1}{R} \frac{\partial \eta}{\partial \theta}$$

(1)

In equations (1), $t$ is time, $\eta$ is the vertical displacement of the water surface above the still water level; $R$ is the Earth's radius; $g$ is the gravitational acceleration; $h$ is the still water depth; $f = 2 \Omega \sin \theta$ represents the Coriolis acceleration, $\Omega$ being the angular velocity of the Earth; $U$ and $V$ are the discharge fluxes in longitudinal ($\lambda$) and latitudinal ($\theta$) directions. The shallow water approximation is valid when $\delta$, the ratio of the water depth to the wavelength, is small. In the present case, $\delta = (3 \text{km}/50 \text{km})$, which is much smaller than 1. Thus, the shallow-water approximation is valid.

Equations (1) were solved using an explicit central finite difference scheme [Nagano et al., 1991]. In the computation, the grid spacing was 2 min arc, time step was 10 sec, and the bathymetry was taken from the ETOPO-2 data bank [Smith and Sandwell, 1997]. The time series was computed at point $P_1$, ($22.81^\circ$N, $109.88^\circ$W) which is 9 km from the location of the tide gauge. The propagation time from $P_1$ to the tide gauge, $T_p$, was estimated from the harbor navigational chart [S.M., 1994], using the relation $v = (gh)^{1/2}$, where $v$ is the phase velocity. The value $T_p$ is 4.1 min.

The tsunami time series at the tide gauge, corresponding to different values of $L$, are shown in Figure 5. In this figure, the time $T_p$ has been added in the time series computed at point $P_1$, so that the computed arrival time of the tsunami can be directly compared with the residual tide shown in Figure 3. We note that the time of first arrival at the tide gauge of 57.5±2 min agrees best with the time series corresponding to $L = 160$ km for which the arrival time is 57.5 min.

Figure 5. The computed time series at the tide gauge corresponding to various values of rupture length $L$. Observed time of first arrival best fits the time series corresponding to $L = 160$ km.

**Uncertainty in the Results**

The speed of tsunami propagation critically depends on the bathymetry. To estimate the error in the computation of the time of arrival of the tsunami due to the uncertainty in the bathymetry, we compared depths from the ETOPO-2 data bank with the navigation chart [S.M., 1996] in the vicinity of the tsunami trajectory, S, shown in Figure 4. The figure also shows sites where the depths were obtained from the navigation chart and were compared with ETOPO-2. We find that the difference between the two sources of bathymetry is normally distributed with a mean of 0.6% and a standard deviation of the difference of 6.8%.

Figure 4. Four-cm elevation contours of the initial tsunami profile corresponding to rupture lengths, $L$, of 140, 160, and 180 km, used as initial condition in the modelling of tsunami propagation. The circle with a cross is the epicenter of the earthquake. The thick line from the four-cm elevation contour ($l = 160$ km) to point $P_1$ shows the tsunami trajectory. The crosses show the great circle path between $P_1$ and $P_2$. The points in the vicinity of the trajectory are sites where the depths were taken from navigational chart [S.M., 1996]. Arrow heads indicate the trench.
The propagation time along the trajectory $S$ can be computed from the equation:

$$\tau = \frac{dS}{\sqrt{gh(S)}} \tag{2}$$

From equation (2) we find that a difference of ±4% in the bathymetry results in a difference in $\tau$ of ±1.1 min. Since the error in the bathymetry is less than ±4%, the expected error in $\tau$ is also less than ±1 min. Thus the uncertainty in $\tau$ is dominated by the sampling rate of the observed time series and not by the uncertainty in the bathymetry. From Figure 5, an uncertainty in $\tau$ of ±1 min roughly translates into an uncertainty of ±10 km in $L$. This gives the uncertainty in the rupture length of ±20 km.

Another source of uncertainty in the estimation of $L$ comes from the assumption that the rupture occurred instantaneously at the centroid origin time, which, as mentioned before, is 38.4 sec later than the time of rupture initiation. The total estimated rupture time is about 62 sec [Courbois et al., 1997]. If we take the time of the stoppage of the rupture as the origin time of the tsunami then its travel time to Cabo San Lucas decreases by about 24 sec. This would increase the estimated $L$ by about 5 km.

**Discussion and Conclusions**

An analysis of the tide gauge data from Cabo San Lucas, Baja California Sur, Mexico, shows that the rupture length of the October 9, 1995 Colima-Jalisco earthquake was about 160 km. This result is consistent with those obtained from seismological studies and inversion of GPS deformation data. Our result contradicts the rupture length of 250 km reported by Tanikawa and Ruff [1996] from the modeling of the same tide gauge data. The reason appears to be an inadvertent error of 16 min in the timing of the data, which was made available to Tanikawa and Ruff [1996]. This error of 16 min, however, does not completely explain the discrepancy. Our calculations, using either equations (1) or (2), show that the rupture length corresponding to the tsunami travel time of 41.5 min (the value that Tanikawa and Ruff [1996] should have estimated from the data available to them), is not 250 km but 320 km. It is not clear why Tanikawa and Ruff [1996] put the upper limit as 250 km. Whatever the cause of the discrepancy, we may now definitely conclude that, unlike the 1992 earthquakes, the 1995 earthquake did not rupture the NW end of the Middle America trench. The unbroken segment extends from Chamela to offshore Puerto Vallarta, for a length of about 120 km (Figure 1).

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**References**


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