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Active seismic deformation in the grabens of northern Central America and its relationship to the relative motion of the North America–Caribbean plate boundary

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Abstract

The active seismic deformation along the grabens of northern Central America is calculated using a method which relates relative plate velocity with the seismic strain rate tensor. The latter is, in turn, obtained from seismic moment tensors.

We calculate the typical or average moment tensor from available Centroid-Moment Tensors and fault-plane solutions. This average tensor is then used to obtain the sum of moment tensors of historical and modern earthquakes reported in the literature.

Both historical and modern earthquakes yield an average extension rate of 8 mm/yr along the grabens of northern Central America, while the relative motion between the North America and Caribbean plates expressed along the Motagua–Polochic fault system is 20 mm/yr.

This result suggests that part of the seismic deformation related to the plate boundary is taken up as extension along the grabens, which play the role of fault termination structures for the Motagua–Polochic system.

Keywords: North America Caribbean plate boundary; extension rate; seismicity; Central America

1. Introduction

Most of the North America–Caribbean plate boundary is a left-lateral transform boundary. Relative motion takes place along several faults. From East to West the main fault zones are: the Enriquillo-Plantain-Garden, Oriente, Swan Islands, and Motagua–Polochic (see Mann et al., 1990, for a review).

The complex nature of the plate boundary is particularly noticeable along its western end in northern Central America, where it approaches the triple junction with the Cocos Plate. The plate boundary offshore, the Swan Islands fault zone, becomes the Motagua and Polochic faults (Motagua–Polochic fault system, for short) inland, but their trace is lost in western Guatemala (Fig. 1). Most workers (e.g. Muehlberger and Ritchie, 1975; Plafker, 1976; Burkart, 1978; Sanchez-Barreda, 1981; Burkart, 1983; Guzmán-Speziale et al., 1989) agree that the triple junction is not a trench–trench–fault triple point in the classical view of McKenzie and Morgan (1969). Rather, it is a broad zone of deformation. To the north of Motagua–Polochic, the deformation zone includes the strike-slip faults of Southeastern Mexico tectonic province (Guzmán-Speziale et al., 1989), and

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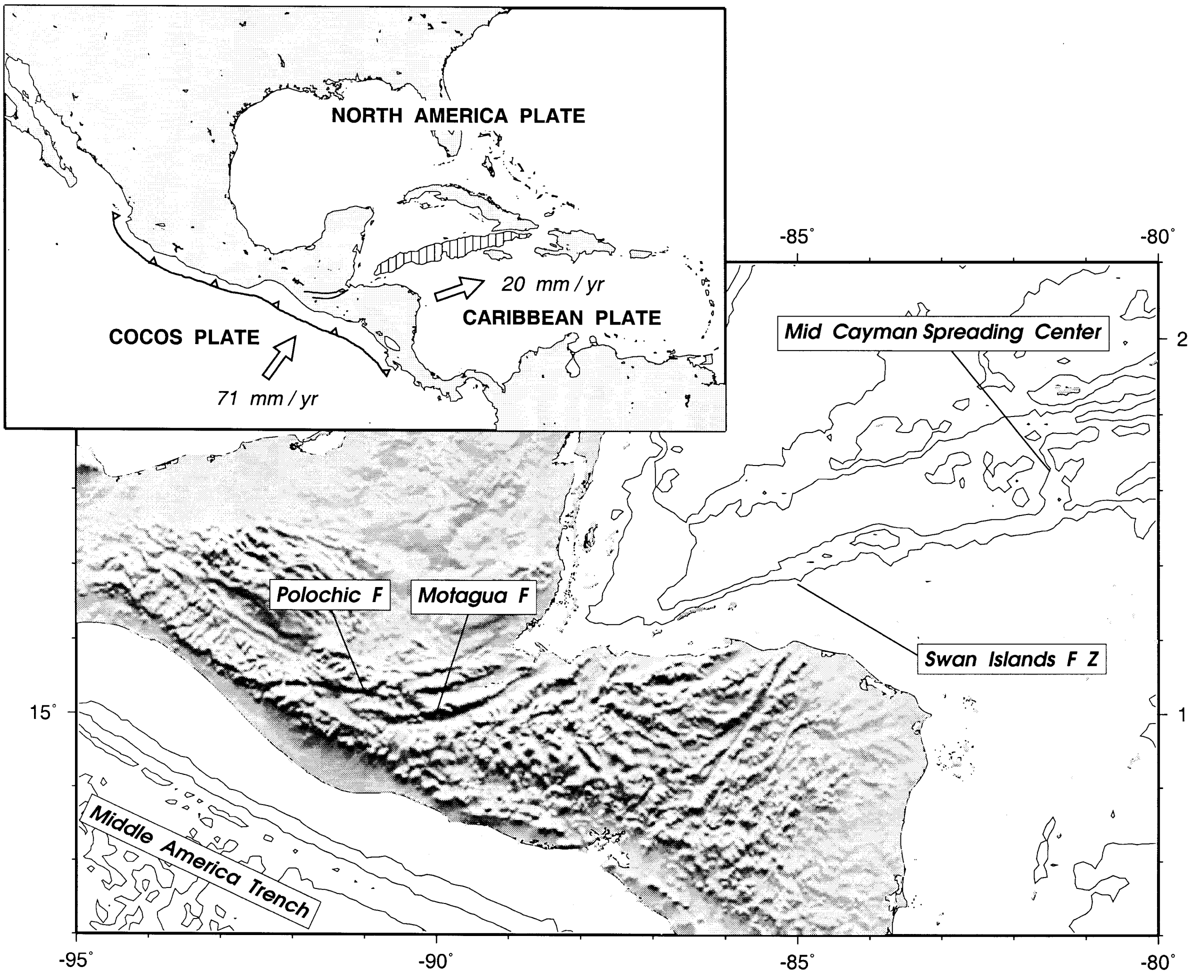

Fig. 1. Plate tectonic framework of northern Central America. Main elements of the northwestern North America–Caribbean plate boundary are shown, together with relative plate velocities with respect to North America. All figures drawn with the help of software by Wessel and Smith (1991).

the Reverse-Faults tectonic province (Guzmán-Speziale and Meneses-Rocha, 2000).

Just south of the Motagua–Polochic fault system, a broad zone of essentially E–W extension is expressed along a series of N–S-trending grabens. Several authors (e.g. Plafker, 1976; Burkart, 1983; Burkart and Self, 1985; Guzmán-Speziale et al., 1989; Gordon and Muehlberger, 1994) suggest that these structures, and the concomitant extensional stress regime, are part of the Caribbean–North America plate boundary deformation.

In this work, we estimate the extension rate of the grabens of Central America, using earthquakes in the last four centuries. We then elaborate on the contribution of the grabens to the relative motion between the North America and Caribbean plates.

2. Tectonic setting

The wedge-shaped area where the grabens of Central America are located, is bounded by the volcanic chain of Central America to the south, the Motagua–Polochic fault zone to the west and

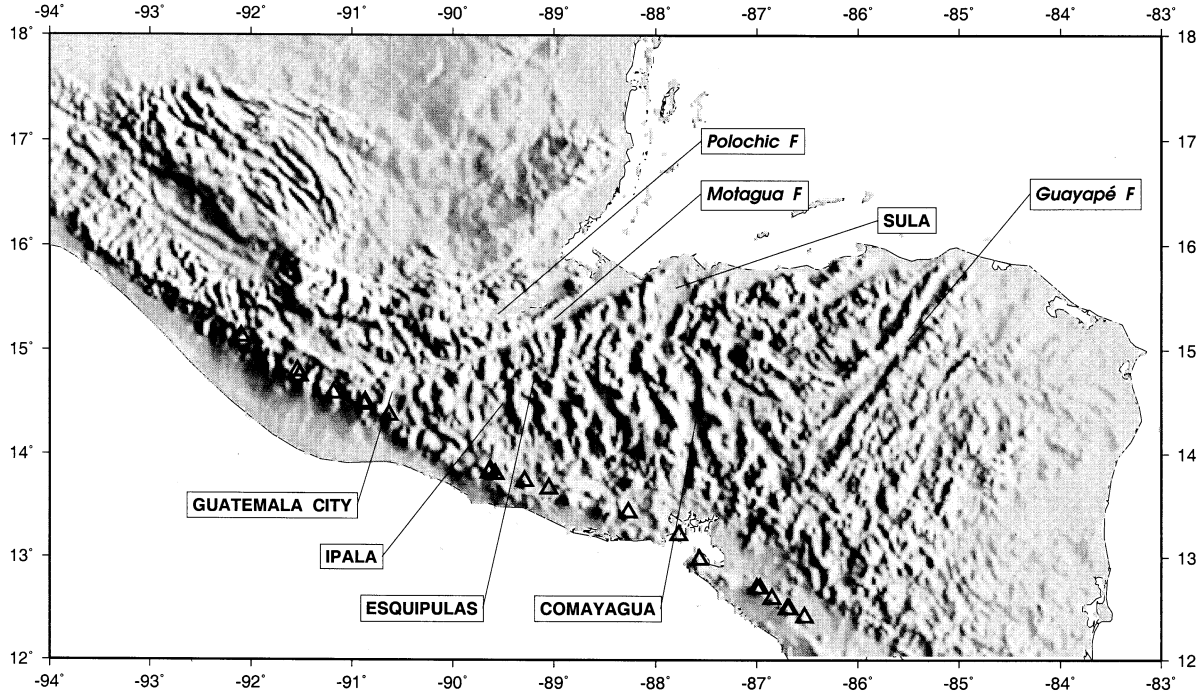

Fig. 2. Main grabens (names in upper case letters) and strike-slip faults (lower case) in northern Central America (after Weyl, 1980). Triangles represent volcanoes of the Central America volcanic arc, with documented historic eruption (Simkin et al., 1981).

Table 1
Speed and azimuth of relative plate motion of the North American Plate with respect to the Caribbean Plate at a point P (15.0°, −90.0°)

N	Pole		ω (°/my)	P	
	Lat. (°N)	Lon. (°E)		Speed (mm/yr)	Azimuth
1	50.0	116.0	0.20	21	252
2	−33.8	−70.5	0.22	19	250
3	66.0	−132.0	0.36	34	251
4	−55.2	−60.8	0.11	12	253
5	−74.3	−26.1	0.11	12	256
6	−33.9	−69.0	0.10	9	248
7	68.4	−126.3	0.25	24	255
8a	18.6	107.2	0.36	24	243
8b	26.4	109.7	0.27	21	245
9	64.9	−109.5	0.214	19	260
Average				20	251

(References: (1) Jordan (1975); (2) RM2 (Minster and Jordan, 1978); (3) Sykes et al. (1982); (4) Stein et al. (1988); (5) NUVEL-1 (DeMets et al., 1990); (6) Deng and Sykes (1995); (7) Calais and Mercier de Lépinay (1993); (8) Dixon et al. (1998); (9) DeMets et al. (2000)

northwest, the Caribbean coast to the north, and the Guayapé Fault to the east (e.g. Dengo, 1968; Mann and Burke, 1984; Gordon and Muehlberger, 1994) (Fig. 2).

The left-lateral strike-slip Motagua–Polochic fault zone marks the plate boundary between the North America and Caribbean plates. Estimates of the relative motion between the two plates along the Motagua–Polochic fault zone range from 9 to 34 mm/yr (Table 1).

The volcanic chain of Central America is an active volcanic arc, with 75 closely spaced basaltic to dacitic volcanoes which have had documented Holocene or historic eruptions (Simkin et al., 1981; Carr and Stoiber, 1990).

The strike-slip Guayapé Fault is an arcuate feature concave to the northwest, which runs across Central America in Honduras. The origin and sense of slip of the fault is not clear, although Gordon and Muehlberger (1994) suggest a right-lateral sense of slip based on regional mapping.

There are at least thirteen N–S-trending graben structures in the area. The main structures are, from west to east: Guatemala City, Ipala, Esquipulas, and Honduras Depression (e.g. Bonis et al., 1970; Burkart, 1983; Mann and Burke, 1984). The Honduras Depression is not a single structure; it is composed of the 125-km-long Sula Graben (also called the Ulúa Graben) and the Comayagua Graben (e.g. Muehlberger, 1976; Weyl, 1980; Gordon and Muehlberger, 1994). Some smaller grabens occur off the main axis of the Honduras Depression, but parallel to it, like the Morazán, Negrito, Otoro, Santa Bárbara, and Talanga basins (Muehlberger, 1976; Mann and Burke, 1984; Gordon and Muehlberger, 1994). Muehlberger (1976) indicates that the Sula Graben has subsided as much as 2000 m in the last 9 to 16 my.

There are several interpretations on how the grabens were formed. For Dengo (1968), the grabens, and particularly the Honduras Depression, formed as a result of local tensile stresses, approximately perpendicular to the regional direction of compressive stress; according to Malfait and Dinkelman (1972), the rifts formed as this portion of the Caribbean Plate became *pinned* against North America, a view shared by Plafker (1976) while Langer and Bollinger (1979) suggest that the grabens developed as fault termination structures as the Motagua Fault grew from east to west; Wadge and Burke (1983) state that the grabens represent internal deformation of the Caribbean Plate since the Miocene; Mann and Burke (1984) propose that localized extension is produced as the Caribbean Plate moves eastward about a promontory of the North American Plate in northern Central America; Burkart and Self (1985) attribute the formation of the grabens to block rotations in northern Central America; in the model of Guzmán-Speziale et al. (1989), the grabens are an extensional zone south and west of the Motagua–Polochic fault system which is part of the deformation related to the Caribbean–Cocos–North America triple junction tectonics; Heubeck and Mann (1991) view the Honduras Depression and associated N–S grabens as a transtensive zone between the western and central Caribbean blocks, which move independently. Suter (1991) believes that extension along the grabens of northern Central America is due to flexural stresses because of bending around the arcuate Motagua Fault. Finally, Gordon and Muehlberger (1994) propose that the grabens form in response to slip on the major strike-slip faults, such as the Motagua Fault.

Recent models for Caribbean–North America plate motion (e.g. Calais and Mercier de Lépinay, 1993; Deng and Sykes, 1995; Dixon et al., 1998; DeMets et al., 2000), however, do not explicitly take into account the grabens of northern Central America.

Weyl (1980) points out that slip along the graben structures is in the order of 10 km "… in the course of several million years" (p. 70). This translates into a slip of a few mm/yr.

3. Seismicity

Subduction of the Cocos Plate beneath the Caribbean Plate yields large to great interplate events, sometimes reaching magnitudes around 8.0 (e.g. McNally and Minster, 1981; Burbach et al., 1984). The Motagua–Polochic fault system is another source of seismicity. There have been at least 25 destructive earthquakes along the fault system since 1530 (White, 1984), including the $M_w = 7.5$ event of 4 February, 1976, which killed some 23 000 people. The Central American Volcanic Arc is the site of medium-sized (maximum magnitudes ~6.5), but highly destructive, earthquakes (e.g. White, 1991; White and Harlow, 1993; Peraldo and Montero, 1999). A fourth source of seismicity is that related to the grabens of northern Central America.

Carr and Stoiber (1977), Osiecki (1981), Montero-Pohly (1989), White (1991), White and Harlow (1993) and Peraldo and Montero (1999) have documented the historical seismicity of Central America, since 1500. Several of these earthquakes can be related to activity along the grabens of northern Central America because the area of largest intensity is directly located on one of the grabens, the damages reported are constrained to a small area (indicative of a shallow source), and the isoseismals are oriented in a NS direction along mapped structures. Magnitude estimates for most of the earthquakes are between 5.0 and 7.5.

Instrumental data for this century reveal an earthquake in 1915 in the Ipala Graben ($M_s = 6.4$); a sequence in December 1917 to January 1918 in the Guatemala City Graben, the largest of which reached a magnitude $M_s = 6.2$; an event ($M_s = 6.2$) in the

Ipala Graben in 1934 (White and Harlow, 1993); and two events on September, 1982 (Ipala Graben) (White, 1991). Langer and Bollinger (1979) report that many aftershocks of the great Motagua Fault earthquake of 1976 aligned along the Mixco Fault, the western flank of the Guatemala City Graben, with normal-faulting (composite) focal mechanisms. More recently, the Harvard University catalog of Centroid Moment Tensors reports four events in the grabens, with normal-faulting fault planes oriented approximately N–S and with magnitudes from 5.0 to 5.6.

Seismicity along the grabens also takes place in swarms of small earthquakes. White et al. (1980) report one that occurred in 1979–1980, in the graben between the Guatemala City and Ipala grabens, consisting of more than 100 000 earthquakes of $M \geq 1$; the epicenters were clustered in a N–S direction. The fault-plane solution of the largest of these events ($M_s = 5.0$; 9 October, 1979) is a normal-faulting mechanism along a N–S plane. Molina Cruz (1992) documented another swarm, in May–June, 1988 in the Guatemala City Graben, which consisted of about 2000 microearthquakes aligned along a N–S-trend, with several lines of evidence suggestive of normal faulting. Maps from the yearly catalogs of the Instituto de Sismología, Vulcanología, Meteorología e Hidrología (INSIVUMEH) of Guatemala consistently show epicenters of small earthquakes (usually $2 \leq M \leq 3$) trending N–S along the grabens (E. Molina, INSIVUMEH, personnal communication, 2000).

4. Method

The formulation of Jackson and McKenzie (1988) is used. It relates the strain rate tensor and the relative plate velocity within a deforming volume by:

$$\dot{\varepsilon}_{ij} = \frac{1}{2}\begin{bmatrix} \frac{2v_1^1}{l_1} & \frac{v_2^1}{l_1} + \frac{v_1^2}{l_2} & \frac{v_3^1}{l_1} + \frac{v_1^3}{l_3} \\ \frac{v_2^1}{l_1} + \frac{v_1^2}{l_2} & \frac{2v_2^2}{l_2} & \frac{v_3^2}{l_2} + \frac{v_2^3}{l_3} \\ \frac{v_3^1}{l_1} + \frac{v_1^3}{l_3} & \frac{v_3^2}{l_2} + \frac{v_2^3}{l_3} & \frac{2v_3^3}{l_3} \end{bmatrix} \tag{1}$$

Here, $\dot{\varepsilon}_{ij}$ are the elements of the strain rate tensor, $\mathbf{v}^i$ is the mean velocity vector difference along the face normal to x_i, and l_1, l_2 and l_3 are the dimensions of the volume along these directions. Jackson and McKenzie (1988) derived Eq. (1) from the relationship between the average strain rate tensor and the sum of seismic moment tensor **M** (Kostrov, 1974):

$$\dot{\varepsilon}_{ij} = \frac{1}{2\mu V\tau}\sum_{n=1}^{N} M_{ij}^n \tag{2}$$

The elements M_{ij} of each of the N seismic moment tensors are summed over the time period considered τ, within the volume of interest V. μ is the modulus of rigidity (3×10^{10} N/m^2).

In the usual right-handed coordinate system, x_1 — north, x_2 — east, x_3 — down, v_2^2 is the relative velocity in the EW direction and between the faces in the NS direction. If the grabens are indeed oriented in the NS direction, this is their extension rate.

Combining Eqs. (1) and (2):

$$v_2^2 = l_2\dot{\varepsilon}_{22} = \frac{l_2}{2\mu V\tau}\sum_{n=1}^{N} M_{22}^n \tag{3}$$

Fault-plane solutions may also be used in Eq. (3), converting them to (symmetric) seismic moment tensors with the well-known expression (e.g. Jost and Herrmann, 1989):

$$M_{ij} = M_0(u_i n_j + u_j n_i) \tag{4}$$

where **u** and **n**, the normal and slip unit vectors, are a function of the strike, dip, and rake of the focal mechanism, and M_0, the scalar seismic moment tensor (in N m), may be calculated from the surface-wave magnitude M_s by (Ekström and Dziewonski, 1988):

$$\log M_0 = 1.5M_s + 9.14 \tag{5}$$

Historic earthquakes represent a sizable portion of the strain rate in the grabens of northern Central America. Evidently, only their magnitude is known, but they can be used in Eq. (3) if we assume that their faulting mechanism is similar to an *average* seismic moment tensor of modern earthquakes, representative of the extension along the grabens.

The average tensor $\overline{\mathbf{F}}$ is given by (Kiratzi and Papazachos, 1996):

$$\overline{F}_{ij} = \frac{1}{N}\sum_{n=1}^{N} F_{ij}^n \tag{6}$$

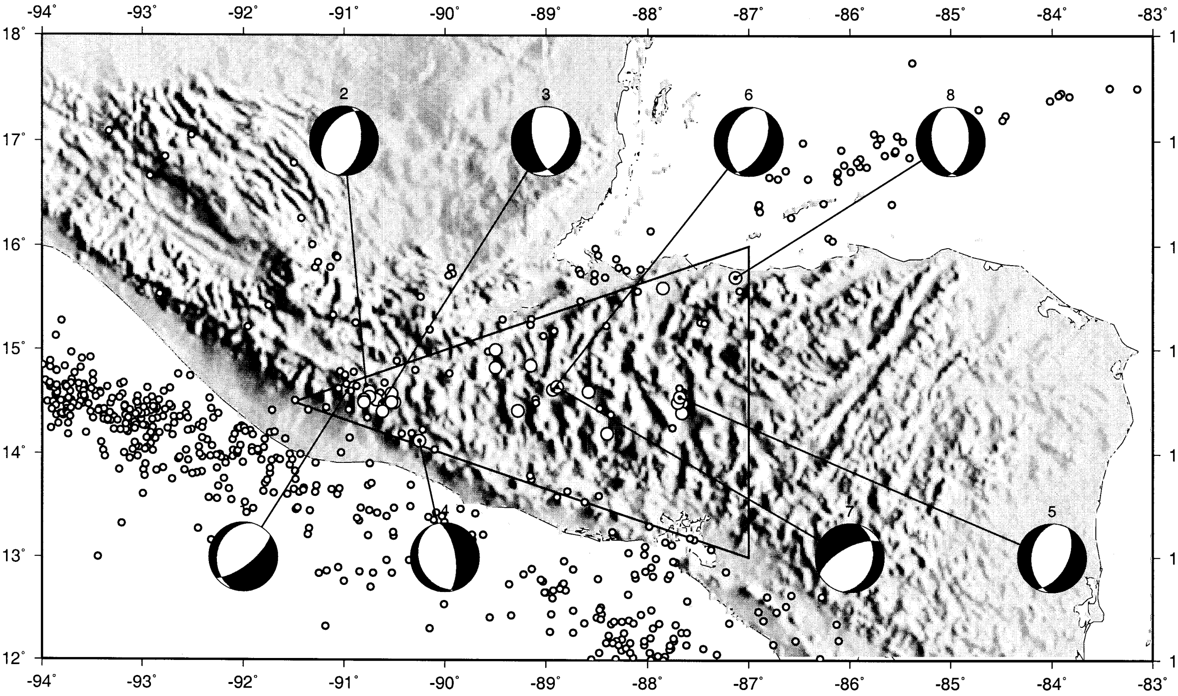

Fig. 3. Shallow seismicity in the northwestern Caribbean. Small circles are recent shallow ($z \leq 50$ km) events reported by Engdahl et al. (1998). Large circles are those earthquakes in Table 3. Also shown are shallow-focus normal-faulting focal mechanisms used in this study and keyed to Table 2. Triangle depicts area for seismic strain rate calculation.

Here, the shape tensor **F** is the normalized seismic moment tensor (Papazachos and Kiratzi, 1992):

$$F_{ij} = \frac{M_{ij}}{M_0} \qquad (7)$$

The sum of seismic moment tensors in Eq. (2) can now be expressed as:

$$\sum_{n=1}^{N} M_{ij}^n = \left(\sum_{n=1}^{N} M_0^n\right)\overline{F}_{ij} \qquad (8)$$

For historic earthquakes, M_0 may be calculated with Eq. (5).

The seismic strain rate (Eq. (2)) may now be written:

$$\dot{\varepsilon}_{ij} = \frac{1}{2\mu V\tau}\left(\sum_{n=1}^{N} M_0^n\right)\overline{F}_{ij} = \frac{1}{2\mu V}\left(\frac{1}{\tau}\sum_{n=1}^{N} M_0^n\right)\overline{F}_{ij} \qquad (9)$$

The term in the last parenthesis is the seismic moment rate for the time period considered.

In practice, we first obtain the average tensor from known seismic moment tensors and focal mechanisms and then multiply by the sum of both historic and modern scalar seismic moments.

The area of deformation was approximated by the isosceles triangle depicted in Fig. 3. It is designed so that all grabens will be included but the Motagua and Polochic fault will be excluded. The triangle runs some 300 km along its N–S base and approximately 500 km along its E–W height (also l_2). If we assume a seismogenic zone of 10 km in depth, the volume of deformation is $V = 7.5 \times 10^{14}$ m^3.

5. Data

We searched for centroid-moment tensors (CMTs) of Harvard University (e.g. Dziewonski and Woodhouse, 1983) through 1999, associated to the deformation of the grabens of northern Central America, i.e. within the triangular area of interest and with depth less than 50 km. Of the seven CMTs

Table 2
Normalized seismic moment tensors (scalar seismic moment $M_0 \times 10^{16}$ N m). References: (1) Langer and Bollnger (1979); (2) reported by White and Harlow (1993) (references therein); (3) Harvard University Centroid-Moment Tensor solution

N	Date	Lat.	Lon.	M_0	M_{11}	M_{22}	M_{33}	M_{12}	M_{13}	M_{23}	Ref.
1	197602 (a)	14.60	−90.97	[a]	0.151	0.477	−0.628	−0.361	−0.632	0.385	1
2	197602 (a)	14.66	−90.78	[a]	0.091	0.865	−0.956	−0.335	0.124	0.192	1
3	197602 (a)	14.50	−90.62	[a]	−0.134	0.919	−0.785	0.102	−0.328	−0.377	1
4	19791009	14.13	−90.25	4.37	0.046	0.722	−0.768	0.208	0.048	0.631	2
5	19820427	14.55	−87.68	12.10	0.017	0.909	−0.926	−0.264	−0.256	0.140	3
6	19820929.1	14.62	−88.93	9.41	−0.069	0.935	−0.866	−0.396	−0.170	0.165	3
7	19820929.2	14.65	−88.89	30.70	0.231	0.495	−0.726	−0.456	0.606	−0.114	3
8	19990216	15.70	−87.13	8.43	0.198	0.875	−1.074	0.026	−0.250	0.056	3
			$\sum$	65.01	0.531	6.197	−6.729	−1.476	−0.858	1.078	

[a] Composite solutions for aftershocks of the Gautemala earthquake of 19760204.

found, three are related to strike-slip faulting along the Central American volcanic arc (e.g. White and Harlow, 1993). The remaining four show a normal-faulting mechanism, with fault planes oriented in an approximate N–S direction. According to Frohlich and Apperson (1992), a focal mechanisms is normal-faulting if its P axis dips more than 60°.

Within the same triangular area, Langer and Bollinger (1979) report four additional (composite) normal-faulting mechanisms and White and Harlow (1993) report one. We use all these fault-plane solutions, converting them to seismic moment tensors by way of Eq. (4).

Parameters of all eight events used are listed in Table 2.

Historic earthquakes in Central America starting in the 16th century, are well documented by Carr and Stoiber (1977), Osiecki (1981), White (1991), White and Harlow (1993) and Peraldo and Montero (1999), among others (see above). Peraldo and Montero (1999) searched historical documents and newspapers throughout Central America and in Spain. In their work, they report original accounts as well as isoseismals and epicentral estimations for events through the 19th century. From this work, we chose earthquakes related to activity along the grabens of northern Central America through the 19th century. The selection criteria were: zone of maximum intensity located directly on one of the grabens, isoseismals elongated in a N–S direction, and rapid attenuation of intensity with distance, indicative of a shallow source (Gutenberg and Richter, 1956). It is assumed both here and by Peraldo and Montero (1999), that the epicenter is located within the isoseismal of maximum intensity and consequently within one of the grabens. Some of the events selected are also reported in the other works mentioned.

The area of study has been highly populated in the last 500 years. According to historical reports (see Peraldo and Montero, 1999, for a review), southeastern Mexico and northern Central America was inhabited by about three million people by the time of the Spanish conquest. The Spaniards established themselves in the ancient Mayan localities, so the number of settlements remained high. Hence, when an earthquake occurred, it was felt in many towns and cities and when deemed newsworthy (usually when churches were damaged or when there were casualties) reported in newspapers or to the proper authorities. Compilations of pre-instrumental earthquakes in historical documents, thus, come from reports of many towns and cities. For example, the earthquakes of 29 September, 1717 and 29 July, 1773 were reported in 21 and 12 towns and cities, respectively, with intensities ranging from V to IX (Peraldo and Montero, 1999).

White and Harlow (1993) compiled a catalog of upper-crustal earthquakes in Central America since 1900. In addition to reviewing historical documents and newspapers, they also searched several international and regional earthquake catalogs and checked available seismograms from Costa Rica and Guatemala. They discriminated between subduction-related and upper-crustal events using several criteria:

Table 3
Epicenters of major earthquakes reported for the northern Central America grabens

N	Date	Lat.	Lon.	M_s[a]	M_s[b]	M_0 (N m)[a]	M_0 (N m)[b]	Ref.
1	15861223	14.60	−90.75	5.4	6.0	1.74×10^{17}	1.38×10^{18}	1, 5
2	16071009	14.50	−90.50	5.4	6.0	1.74×10^{17}	1.38×10^{18}	1, 5
3	16510218	14.52	−90.68	5.4	6.0	1.74×10^{17}	1.38×10^{18}	1, 5
4	16890212	14.55	−90.75	6.0	6.0	1.38×10^{18}	1.38×10^{18}	1, 5
5	17170929	14.52	−90.80	6.5	6.5	7.76×10^{18}	7.76×10^{18}	5
6	173304	14.20	−88.40	4.9	5.4	3.09×10^{16}	1.74×10^{17}	3, 5
7	173305	14.42	−89.28	5.4	7.5	1.74×10^{17}	2.45×10^{20}	3, 5
8	17431015	15.00	−89.50	6.7	6.7	1.55×10^{19}	1.55×10^{19}	5
9	17650602	14.83	−89.50	6.0	7.6	1.38×10^{18}	3.47×10^{20}	1, 3, 5
10	17730729	14.50	−90.80	6.5	6.5	7.76×10^{18}	7.76×10^{18}	1, 5
11	17731214	14.50	−90.80	5.7	5.7	4.90×10^{17}	4.90×10^{17}	1, 5
12	17741014	14.50	−87.66	5.4	6.0	1.74×10^{17}	1.38×10^{18}	2, 3, 5
13	18090620	14.40	−87.66	5.0	5.7	4.37×10^{16}	4.90×10^{17}	2, 3, 5
14	18201019	15.60	−87.85	6.0	6.5	1.38×10^{18}	7.76×10^{18}	2, 5
15	18300421	14.47	−90.60	6.3	6.3	3.89×10^{18}	3.89×10^{18}	1, 5
16	18511114	14.50	−87.70	6.0	6.5	1.38×10^{18}	7.76×10^{18}	1, 2
17	18851218	14.41	−90.62	6.0	6.0	1.38×10^{18}	1.38×10^{18}	1, 5
18	19151229	14.60	−88.58	6.3	6.4	3.89×10^{18}	5.50×10^{18}	1, 3, 4
19	19171226	14.53	−90.53	5.8	5.8	6.92×10^{17}	6.92×10^{17}	1, 4
20	19171229	14.55	−90.53	5.7	5.7	4.90×10^{17}	4.90×10^{17}	1, 4
21	19180104	14.58	−90.53	6.1	6.1	1.95×10^{18}	1.95×10^{18}	4
22	19180125	14.50	−90.53	6.2	6.2	2.75×10^{18}	2.75×10^{18}	4
23	19340203	14.85	−89.15	6.2	6.2	2.75×10^{18}	2.75×10^{18}	1, 4
24	19791009	14.13	−90.25	5.0	5.0	4.37×10^{16}	4.37×10^{16}	4
25	19820427	14.55	−87.68	5.3	5.3	1.21×10^{17}	1.21×10^{17}	6
26	19820929.1	14.62	−88.93	5.1	5.1	9.41×10^{16}	9.41×10^{16}	6
27	19820929.2	14.65	−88.89	5.5	5.5	2.45×10^{17}	3.07×10^{17}	4, 6
28	19990216	15.70	−87.13	5.2	5.2	8.43×10^{16}	8.43×10^{16}	6
					Σ	5.64×10^{19}	6.66×10^{20}	

[a] Minimum value reported.

[b] Maximum value reported.

(References.: (1) Carr and Stoiber (1977); (2) Osiecki (1981); (3) White (1991); (4) White and Harlow (1993); (5) Peraldo and Montero (1999); (6) Harvard CMT).

the presence of surface rupture, indicative of a shallow focus; depth determined by a local network; epicentral location with respect to volcanic arc (more than 30 km landward of the arc would mean an upper crust event whereas more than 30 km seaward of the arc means that the event is related to subduction). They also found that upper-crustal events produced higher and more localized intensities than subduction earthquakes. Earthquake information for the period 1900–1964 is taken from this work. From the work of White and Harlow (1993), we also chose those events with maximum intensity above one of the grabens and isoseismals elongated in the N–S direction.

Considering historic events, fault-plane solutions, and CMTs, there are 28 events for which M_0 can be calculated (Table 3).

6. Results

The average tensor calculated from seismic moment tensors listed in Table 2, using Eq. (6) is:

$$\bar{\mathbf{F}} = \begin{bmatrix} 0.066 & -0.185 & -0.107 \\ -0.185 & 0.775 & 0.135 \\ -0.107 & 0.135 & -0.841 \end{bmatrix} \qquad (10)$$

The matrix $\mathbf{U}$ of eigenvectors of $\overline{\mathbf{F}}$:

$$\mathbf{U} = \begin{bmatrix} -0.245 & 0.964 & 0.100 \\ 0.965 & 0.253 & -0.071 \\ 0.093 & -0.080 & 0.992 \end{bmatrix} \tag{11}$$

shows that the eigenvector associated to the extensional axis T (the first column of $\mathbf{U}$), is practically horizontal and oriented in a S76°E direction. The compressional axis P (third column) is basically vertical, while the intermediate axis B (second column) is mainly horizontal, oriented in a N15°E direction. We rotate $\overline{\mathbf{F}}$ into a new coordinate system with x'_1 — in a N14°E direction, x'_2 — oriented S76°E, and x'_3 — down. In this new coordinate system, extension takes place in the x'_2 direction and $\overline{\mathbf{F}}$ becomes:

$$\overline{\mathbf{F}}' = \begin{bmatrix} 0.021 & 0.003 & -0.071 \\ 0.003 & 0.820 & 0.157 \\ -0.071 & 0.157 & -0.841 \end{bmatrix} \tag{12}$$

The sum of scalar seismic moments is given in Table 3, as a minimum (5.64×10^{19} N m) and a maximum (6.66×10^{20} N m) value. In the rotated coordinate system, Eq. (8) becomes:

$$\left[\sum_{n=1}^{N} M_{ij}^{n}\right]_{\min} = \begin{bmatrix} 1.164 & 0.174 & -4.014 \\ 0.174 & 46.268 & 8.848 \\ -4.014 & 8.848 & -47.432 \end{bmatrix} \times 10^{18}\ \text{N m} \tag{13a}$$

$$\left[\sum_{n=1}^{N} M_{ij}^{n}\right]_{\max} = \begin{bmatrix} 1.375 & 0.205 & -4.739 \\ 0.205 & 54.636 & 10.448 \\ -4.739 & 10.448 & -56.011 \end{bmatrix} \times 10^{19}\ \text{N m} \tag{13b}$$

Combining Eqs. (8) and (2), the minimum and maximum extensional strain rates are:

$$[\dot{\varepsilon}_{22}]_{\min} = 2.48 \times 10^{-9}\ \text{yr}^{-1} \tag{14a}$$

$$[\dot{\varepsilon}_{22}]_{\max} = 2.93 \times 10^{-8} \text{yr}^{-1} y \tag{14b}$$

whereas Eq. (3) yields the minimum and maximum value for the extension rate of the grabens of northern Central America:

$$[v_2^2]_{\min} = 1.24\ \text{mm/yr} \tag{15a}$$

$$[v_2^2]_{\max} = 14.66\ \text{mm/yr} \tag{15b}$$

with an average of 8 mm/yr.

A large portion of Eq. (14b) comes from two earthquakes in the Ipala Graben in 1733 and 1765 with reported maximum magnitudes of 7.5 and 7.6. The question is whether an intraplate, normal-faulting earthquake this large may occur. An example of such an earthquake is the Sonora, Mexico, earthquake of May 1887, an intraplate, normal-faulting event in the southern end of the Basin and Range Province with a rupture length of about 80 km and a calculated M_s of 7.4 ($M_0 = 1.27 \times 10^{20}$ N m) (Natali and Sbar, 1982). The rupture length agrees well with the relationship between magnitude M and surface rupture length (SRL) given by Wells and Coppersmith (1994) for normal-faulting earthquakes:

$$M = 4.86 + 1.32 \log(\text{SRL}) \tag{15}$$

According to this formula, the 80–100 km-long Ipala Graben is capable of producing a magnitude ~7.5 earthquake. So, strictly from this point of view, reported magnitudes of 7.5–7.6 are not unreasonable in this tectonic setting.

7. Discussion

The directions (eigenvectors) of minimum, intermediate and maximum extension rate, given by the average tensor $\overline{\mathbf{F}}$, are obtained from normalized seismic moment tensors. Consequently, no single event dominates the tensor. In fact, except for the composite solutions of Langer and Bollinger (1979), the events used in calculating $\overline{\mathbf{F}}$ are uniform in size, with seismic moments ranging from 4.37×10^{16} to 3.07×10^{17} N m (magnitudes 5.0–5.5; Table 2). Therefore, we are confident that this tensor indeed represents the average seismic moment tensor related to the grabens.

The average seismic moment tensor shows a very small non-horizontal component for T and B axes, and a practically vertical P axis (Eq. (10)). The direction of the T axis (S76°E) is essentially perpendicular to

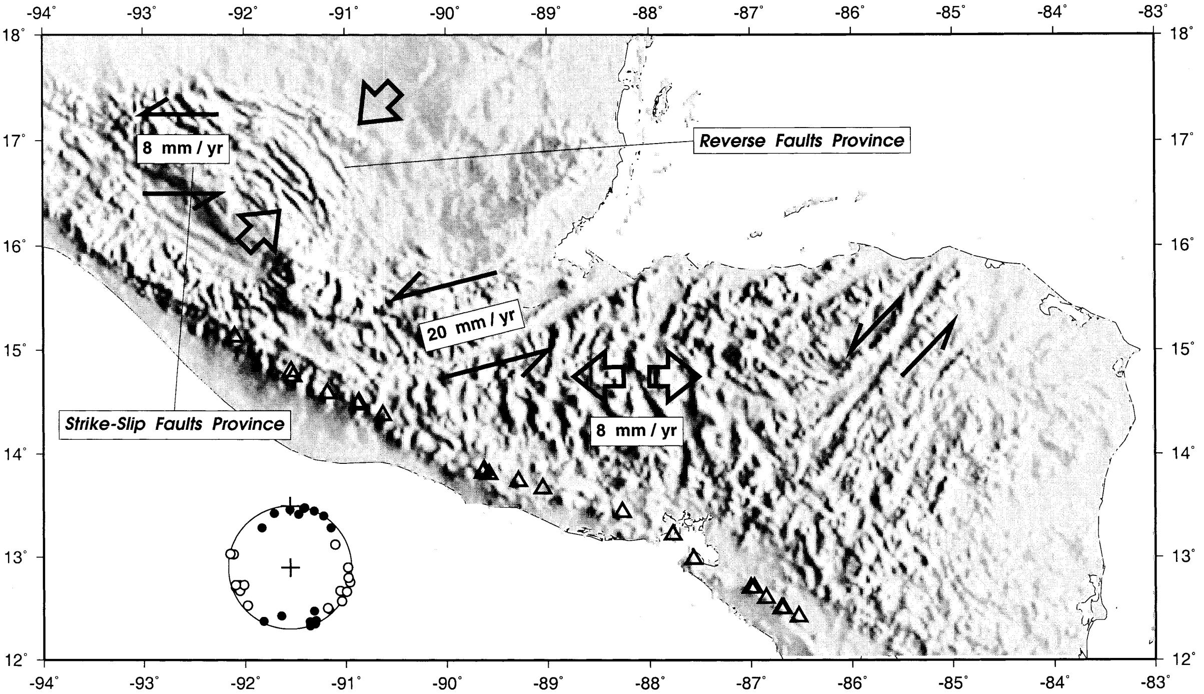

Fig. 4. Summary of deformation in the area. Areas of extension (diverging arrows), compression (converging arrows), and strike-slip faulting are shown, together with relative motion, where known. Stereographic projection inset shows *P* (solid circles) and *T* (open circles) axes for shallow-focus earthquakes along the Central American Volcanic Arc (e.g., White and Harlow, 1993). See text for details.

the N–S-trend of the grabens, which results in normal faulting along the flanks of the grabens, without a significant strike-slip component. This has also been observed in the field by Muehlberger (1976) and Gordon and Muehlberger (1994).

The seismic moment rate for the 414 years considered is 1.36×10^{17} N m/yr minimum, and 1.61×10^{18} N m/yr maximum. As a comparison, modern earthquakes in 21 years (Table 2) yield a seismic moment rate of 3.09×10^{17} N m/yr. Using Eq. (9), this moment rate corresponds to an extension rate of almost 3 mm/yr. The maximum seismic moment rate for the entire 20th century is 1.48×10^{17} N m/yr (1.3 mm/yr extension rate) whereas for the 19th century, it is 2.13×10^{17} N m/yr (1.9 mm/yr). But for the 18th century, the maximum rate is 6.25×10^{18} N m/yr (representing a 57 mm/yr extension rate!). There are two possible explanations for this change: (1) the magnitudes for the 1733 and 1765 events are overestimated; (2) seismic activity during this time period is higher than in the following two centuries. We favor a combination of both because the maximum extension rate is too high but, on the other hand, there are two other earthquakes (1743 and 1773; see Table 3) with reported magnitudes of 6.7 and 6.5, respectively. Even the minimum moment rate (for which the magnitudes of the 1733 and 1765 events are 5.4 and 6.0) is 3.33×10^{17} N m/yr, which corresponds to an extension rate of 3 mm/yr.

The average extension rate from all the data is 8 mm/yr but, considering the arguments given above, the figure might be somewhat lower, possibly closer to 3 mm/yr. This corresponds only to seismic deformation, and additional extension may be taking place aseismically. In comparison, the average relative motion between the North America and Caribbean plates is 20 mm/yr. In other words, extension along the grabens is significant with respect to relative plate motion (Fig. 4). This result suggests that the grabens of northern Central America are an important part of the deformation associated to the North

America–Caribbean plate boundary and/or the triple junction of the Cocos–North America–Caribbean plates, as proposed by several authors (Dengo, 1968; Malfait and Dinkelman, 1972; Plafker, 1976; Langer and Bollinger, 1979; Wadge and Burke, 1983; Mann and Burke, 1984; Burkart and Self, 1985; Guzmán-Speziale et al., 1989; Heubeck and Mann, 1991; Suter, 1991; Gordon and Muehlberger, 1994).

E–W extension south of the Motagua–Polochic fault system fits well into the models proposed by Malfait and Dinkelman (1972), Plafker (1976), Langer and Bollinger (1979) and Burkart and Self (1985), who suggest that the Caribbean Plate has become pinned against North America. It also agrees with the model of Guzmán-Speziale et al. (1989) who suggest that the Motagua–Polochic plate boundary has not yet reached the Middle America Trench. In both cases, continued westward motion of the North America Plate *stretches* the northwestern corner of the Caribbean Plate.

Currently active, N–S-trending normal faults are widely present in Roatán Island, off the northern coast of Honduras (Avé Lallemant and Gordon, 1999). A CMT for a shallow-focus normal-faulting earthquake (19900727; $M_0 = 1.62 \times 10^{17}$ N m) is located in the Tela Basin, south of Roatán Island. This event did not meet our selection criterion for a normal-faulting mechanism (P axis dipping at least 60°), but it shows a sub-horizontal T axis oriented in a 304° azimuth. This suggests that E–W extension continues north, off the coast of northern Central America.

According to Mann and Burke (1984), E–W extension along the northern margin of the Caribbean Plate is not constrained to the grabens of northern Central America: sixteen N–S-trending rifts can be identified along the northern edge of the Caribbean Plate, from Guatemala to Jamaica, although only the grabens of northern Central America appear to be currently active (Muehlberger, 1976; Mann and Burke, 1984; Gordon and Muehlberger, 1994).

Along the Central American volcanic arc, shallow-focus earthquakes with right-lateral, along-the-arc (or left-lateral, perpendicular to the arc), strike-slip faulting occur frequently (e.g. White and Harlow, 1993). Preliminary results from Guzmán-Speziale and White (in preparation) show that the average extensional axis of these events is horizontal and trends in an E–W direction. Furthermore, the strain rate in this direction is about $4 \times 10^{-8}\ \mathrm{yr}^{-1}$, similar to that encountered along the grabens. This suggests that the E–W extensive stress regime extends along the volcanic arc.

North of the Motagua–Polochic fault system, the *pinning* of the plate boundary produces EW compression along the Reverse-Faults tectonic province (Guzmán-Speziale and Meneses-Rocha, 2000), as North America moves westward with respect to the Caribbean Plate.

Other elements may also be taking part in the deformation related to the North America–Caribbean plate boundary (Fig. 4). The strike-slip faults of southeastern Mexico (e.g. Meneses-Rocha, 1985), north of the Motagua–Polochic fault system, are an assemblage of at least nine strike-slip left-lateral faults, with documented seismic activity which, according to Guzmán-Speziale et al. (1989), take up part of the motion between the North America and Caribbean plates. Meneses-Rocha (1985, 1991) has calculated a displacement of between 40 and 70 km since the late Miocene for all the faults, which represent an average of 6–10 mm/yr. The 290 km long Guayapé Fault is believed to be active, from geologic evidence (Gordon and Muehlberger, 1994), but there is no reported seismicty, either historical or instrumental. There is also the possibility that some of the deformation along this fault may be taking place aseismically or as large earthquakes with long return times.

8. Summary

The E–W extension rate along the grabens of northern Central America is 8 mm/yr, on average. This deformation fits well into models of the North America–Caribbean plate boundary. Neotectonic deformation associated to the plate boundary in northwestern Guatemala and southeastern Mexico, additionally consists of (Fig. 4): strike-slip left-lateral motion along the Motagua–Polochic fault system, at an average rate of 20 mm/yr; strike-slip faulting along the strike-slip faults of southeastern Mexico at a rate of 6–10 mm/yr; Compression along the Reverse-Faults tectonic province (rate unknown). E–W extension along the Central America volcanic arc, and strike-slip faulting along the Guayapé Fault, may also be part of the deformation.

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