

# Geodetic determination of crustal deformation across the Strait of Gibraltar

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## SUMMARY

In 1982/83 the Moroccan and Spanish authorities jointly established a geodetic network spanning the Strait of Gibraltar. In 1984 the network, comprising four stations on the Moroccan and four on the Spanish shore, was observed for horizontal and vertical angles, distances by EDM, levelling, and for astronomical latitudes, longitudes, and azimuths. In 1989 the network was expanded to include a further six stations in Spain, and one in Ceuta, and the whole was observed by GPS using a set of eight receivers.

The observational data used proved adequate to determine significant horizontal homogeneous strain, but less apt for vertical deformation. The strain results are given with accuracies of 1 standard deviation ('tensor' values for shear): dilatation rate  $-(0.07 \pm 0.04) 10^{-6} \text{ yr}^{-1}$  (i.e. diminution of surface area; only determinable if the EDM and GPS length-scales are held to be comparable); shear strain rate  $(0.12 \pm 0.04) 10^{-6} \text{ rad yr}^{-1}$ ; and axis of maximum relative shortening  $87^\circ/267^\circ \pm 8^\circ$  (determinable from changes in shape of the network, and largely independent of the comparability of length-scales).

With only eight stations involved in the repeated observations, the results could be influenced by local effects at one or two stations. However, they demonstrate the capability of EDM and GPS observations to detect crustal strain within a period of 5 yr to an accuracy of the order of  $0.05 \times 10^{-6} \text{ rad yr}^{-1}$ .

**Key words:** crust, deformation, geodesy, Gibraltar, strain, tectonics.

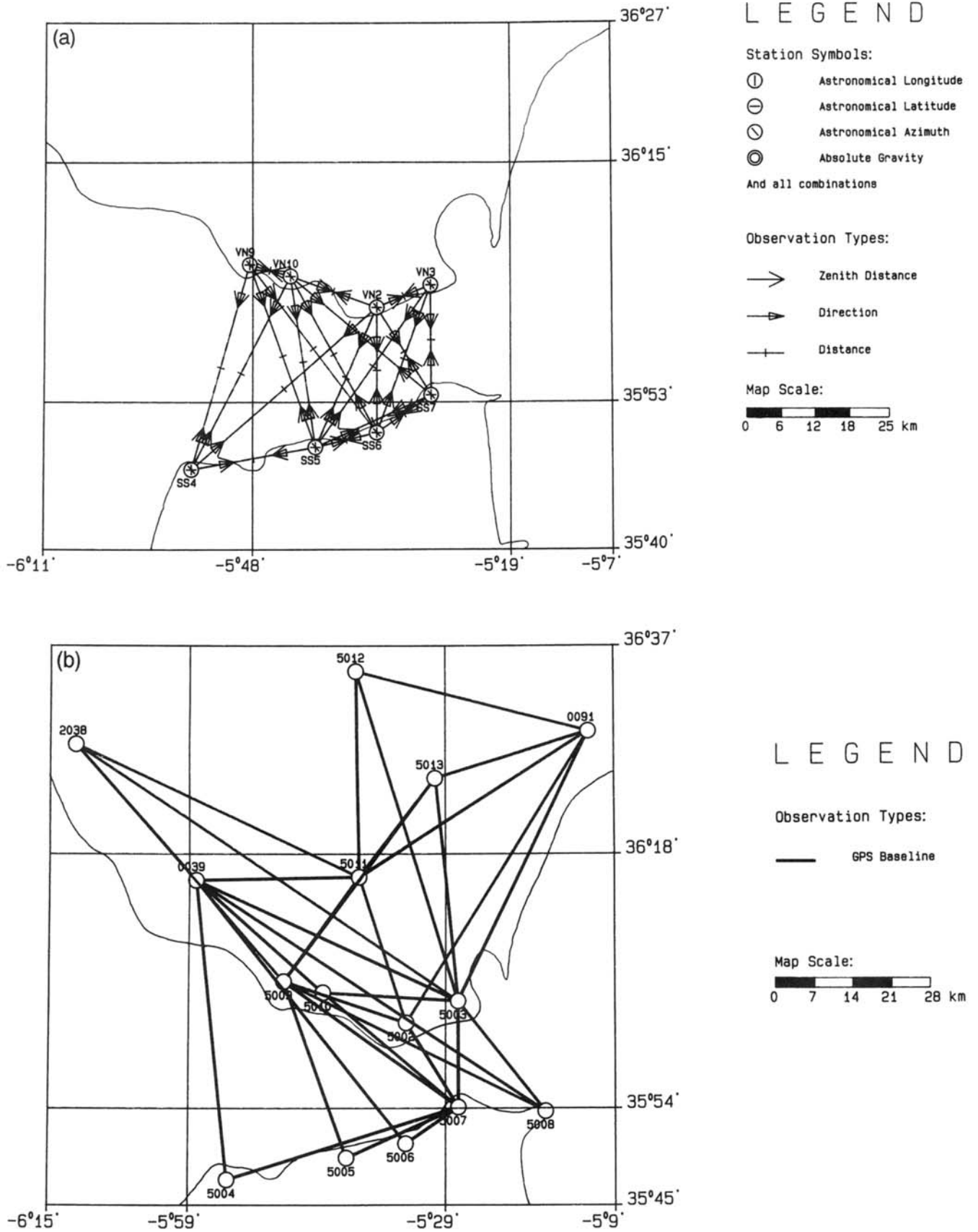
## INTRODUCTION

In 1982/83 the governments of Spain and Morocco established a geodetic network spanning the Strait of Gibraltar, as part of a joint project to improve communication links between their two countries, and thus between the European and African continents. This network—*Red Geodésica para Observaciones Geodinámicas del Estrecho de Gibraltar* (RGOG)—initially comprised of four stations on the Moroccan and four on the Spanish shore, and in 1984 was measured by EDM for distances, by theodolite for horizontal and vertical angles, and by levelling, as well as for astronomical latitude, longitude and azimuth (Fig. 1a). Following an international workshop in Madrid (SNED & SECEG 1990), it was decided to enlarge the network and to make GPS measurements. In 1989, therefore, the network was enlarged (to become RGOGA—

*RGOG Ampliada*) by the addition of a further six stations in Spain, and one at Ceuta on the African shore (Fig. 1b); and a GPS campaign was undertaken to link the stations of the expanded network.

This work was divided amongst several organizations: *Sociedad Española de Estudios para la Comunicación fija a través del Estrecho de Gibraltar* (SECEG), *Société Nationale d'Études du Déroit* (SNED), *Instituto Geográfico Nacional de España* (IGNE), and *Division de la Cartographie de Maroc* (DCM). SECEG and SNED were responsible for coordination and establishment of the network. IGNE and DCM carried out the geodetic measurements for both epochs, and results have been given by IGNE & SECEG (1987), and by Caturla *et al.* (1990). The IAPG collaborated in the processing of the GPS data, and these results have been further discussed by Fredrich & Hein (1991).

As was implicit in its establishment, and in its title,



**Figure 1.** Geodetic networks spanning the Strait of Gibraltar. (a) Network observed in 1984 by EDM distance measurements, horizontal and vertical angles, levelling, and for astronomical latitude, longitude, and azimuth. (b) Expanded network observed in 1989 by satellite (GPS) methods.

RGOG was designed to study the geodynamical aspects of the European–African link, and the availability of geodetic observations repeated after an interval of 5 yr makes it possible to carry out an analysis of deformation, using the software system GEONET at the IAPG (Reilly 1990a, b).

## DEFORMATION ANALYSIS

Evidence for crustal deformation can be sought in the data of repeated geodetic observations. Analysis of deformation within the space–time window of the observations may depend on assumptions as to the geophysical environment. In the case of observations that span, in time and space, an earthquake with a clearly defined fault plane, for instance, the introduction of a dislocation model may be appropriate. In the absence of such specific indications, however, the most general and unbiased approach is to use a deformation model that is continuous in both space and time, and to combine it in a simultaneous reduction of all relevant geodetic measurements. Such a method, introduced by Bibby (1973, 1982) for the comparisons of triangulation data, has been extended to the general three-dimensional case by Reilly (1982, 1990a) and incorporated in the software program system GEONET (Reilly 1990b). The continuum model is based on the assumption of a field of velocity (of the particles of the Earth's crust) that is (usually) held constant in time, and which has a spatial variation that can be approximated by a Taylor's expansion in three dimensions about a suitable local origin point. The method of simultaneous reduction means that it is not necessary for observations to have been repeated on an identical set of points, only that there be points in common between successive surveys; in this way, all relevant observations can be used, and none need be arbitrarily discarded.

In the present case, with in effect two surveys at an interval of 5 yr, and only eight stations in common, the only continuum model justified by the data is that of homogeneous strain under constant velocity; and since there was only weak comparability between the observations for determining height changes (vertical angles in the first epoch, GPS in the second), the continuum model can be restricted to that of deformation in a horizontal plane. In this case, the homogenous deformation model can be expressed in terms of four parameters:

- $\Delta$  the rate of dilatation (positive for extension);
- $\gamma$  the magnitude of the shear strain rate;
- $A$  the azimuth of the direction of relative maximum shortening; and
- $\Omega$  the rate of equivalent rigid-body rotation (positive anticlockwise).

Using these symbols, we can express the rate of extensional strain of a line in the direction of azimuth  $\theta$  as

$$e(\theta) = \Delta - \gamma \cos 2(\theta - A), \quad (1)$$

and of azimuth  $(\theta + \pi/2)$  as

$$e(\theta + \pi/2) = \Delta + \gamma \cos 2(\theta - A), \quad (2)$$

whence the rate of dilatation is seen to be

$$\Delta = (1/2)[e(\theta + \pi/2) + e(\theta)], \quad (3)$$

and the rate of areal dilatation is  $2\Delta$ . The difference in extensional strain rates gives

$$(1/2)[e(\theta + \pi/2) - e(\theta)] = \gamma \cos 2(\theta - A). \quad (4)$$

The rate of rotation of a line in the direction of azimuth  $\theta$  is

$$r(\theta) = \Omega + \gamma \sin 2(\theta - A) \quad (5)$$

(positive anticlockwise), and of azimuth  $(\theta + \pi/2)$  is

$$r(\theta + \pi/2) = \Omega - \gamma \sin 2(\theta - A), \quad (6)$$

whence the mean rate of rotation is

$$\Omega = (1/2)[r(\theta + \pi/2) + r(\theta)], \quad (7)$$

and the rate of shear on an axis of azimuth  $\theta$  can be defined as

$$\gamma(\theta) = -(1/2)[r(\theta + \pi/2) - r(\theta)] = \gamma \sin 2(\theta - A), \quad (8)$$

in which the sign is positive for right-lateral (dextral) shear, and negative for left-lateral (sinistral) shear.

In the general case of two observation lines at azimuths  $\theta_1$  and  $\theta_2$ , we have from (1) the difference in extensional strain rate of

$$(1/2)[e(\theta_2) - e(\theta_1)] = \gamma \sin(\theta_2 - \theta_1) \sin[(\theta_2 + \theta_1) - 2A], \quad (9)$$

and in rotation

$$(1/2)[r(\theta_2) - r(\theta_1)] = \gamma \sin(\theta_2 - \theta_1) \cos[(\theta_2 + \theta_1) - 2A]. \quad (10)$$

These last two equations (9 and 10) show how the parameters ( $\gamma, A$ ) of the shear strain rate can be deduced from the changes in the lengths of observation lines in different directions, or, alternatively (and equivalently), from the changes in the angles between them. Details of the formation and solution of geodetic observation equations are given by Reilly (1982, 1987, 1990a).

## OBSERVATIONAL DATA

The observational data fall into the following two groups.

(a) The campaign of 1984, comprising distance measurements, horizontal and vertical angles, and astronomic azimuth, latitude and longitude, involving the eight stations of RGOG, together with levelling between the four stations on the European (Spanish) shore of the Strait, and between the four stations on the African (Moroccan) shore.

(b) The campaign of 1989, comprising GPS observations on 14 observing days at the eight original stations, plus a further six on the European and one on the African side, a total of 15 stations of RGOGA. The data are summarized by Fredrich & Hein (1991) from sources in IGNE & SECEG (1987) and Caturla *et al.* (1990).

The types and numbers of observations are as follows:

Distance (EDM)	21	(or)
Distance ratio	14	(alternatively)
Horizontal angle	33	
Vertical angle		
(zenith distance)	41	
Astronomic azimuth	42	

Astronomic latitude	8
Astronomic longitude	8
GPS difference vectors	70.

(The levelling data have been excluded, as the levelling does not span the Strait).

The observations of the 1984 campaign were made with Rangemaster III laser geodimeters, Wild T3 geodetic theodolites, and Kern DKM3-A astrogeodetic theodolites; the campaign of 1989 employed six Trimble 4000 SLD GPS receivers, and two Trimble 4000 ST (Caturla *et al.* 1990). The data were introduced as mean directions (for horizontal angles); mean observed angles for zenith distances, and astronomical observations; and for EDM observations, as slope distances corrected for meteorological effects. The GPS observations of the 1989 campaign were processed at IAPG by the TOPAS software package (Landau 1991); this multistation adjustment yields a set of difference vectors for each day and frequency combination, together with the corresponding error covariance matrix, which are input as such to the GEONET program. The L1 carrier phase was used for baselines shorter than 40 km, and an ionospherically corrected dual-frequency combination for those longer than 40 km. This procedure was adopted following a detailed investigation of the ionospheric effects for this network. Comparisons with solutions using precise GPS ephemeris as well as orbit improvement techniques show that the contribution of orbital errors lies within the GPS noise level. Day-to-day GPS repeatability revealed that the discrepancies in baseline components were less than their standard errors. The data of both campaigns has also been adjusted separately using the OPERA software, and the results have been discussed by Fredrich & Hein (1991).

## METHODS OF SOLUTION

The above observations were all introduced into a simultaneous adjustment using the program system GEONET (Reilly 1990a). Three different solutions were calculated, the differences depending on the treatment of the distance measurements, both EDM and GPS.

(1) The EDM distances are accepted as internally consistent, and comparable with the GPS distance scales. Under this assumption there are comparable distance measurements at both epochs, and the dilatation rate  $\Delta$  is therefore determinable.

(2) The EDM observations are considered to have possible residual errors in scale determination, and are introduced to the solution as distance ratios in seven groups; there are thus no distance observations to be compared between the two epochs, and no dilatation rate is determinable.

(3) In addition, the GPS observations are considered to have scale variation between observation days, and a scale correction factor is introduced as an unknown for each observation day after the first: distance-scale is thus set by the first day's observations, and no dilatation rate is determinable.

In each case, there are sufficient and suitable observations from which to derive changes of shape, and thus shear strain. In no case, however, are there direction observations

in terms of unequivocally comparable reference frames which would permit the determination of the mean rate of rotation  $\Omega$ .

In addition to the strain rate unknowns, each solution yields at each network station a triplet of geocentric position coordinates, and the astronomic latitude and longitude (i.e. the direction of the vertical). The reduction takes account of the local configuration of the gravity field (i.e. geoid) by using the available observations of astronomical latitude, longitude and azimuth.

## RESULTS

### Horizontal deformation

The adjustment was carried out using the GEONET program system for the three options discussed above. As a guide to the quality of the data used, Table 1 gives the rms adjusted residual for the major observational groups. The corresponding results for the homogeneous strain rates are given in Table 2.

From Table 2, the results of solution (1) would lead to values of extensional strain rate in the principal axes of strain as follows:

$$\begin{aligned} &\text{axis of maximum relative shortening} \\ &\quad -(0.19 \pm 0.06) 10^{-6} \text{ yr}^{-1} \text{ at } 87^\circ/267^\circ \pm 9^\circ; \text{ and} \\ &\text{axis of maximum relative lengthening} \\ &\quad (0.05 \pm 0.05) 10^{-6} \text{ yr}^{-1} \text{ at } 177^\circ/357^\circ \pm 9^\circ. \end{aligned}$$

In this case, it is seen that the value of the extensional strain rate on the N-S (lengthening) axis is insignificant at the accuracy of measurement ( $2 \text{ SD} \approx 0.1 \times 10^{-6} \text{ yr}^{-1}$ ).

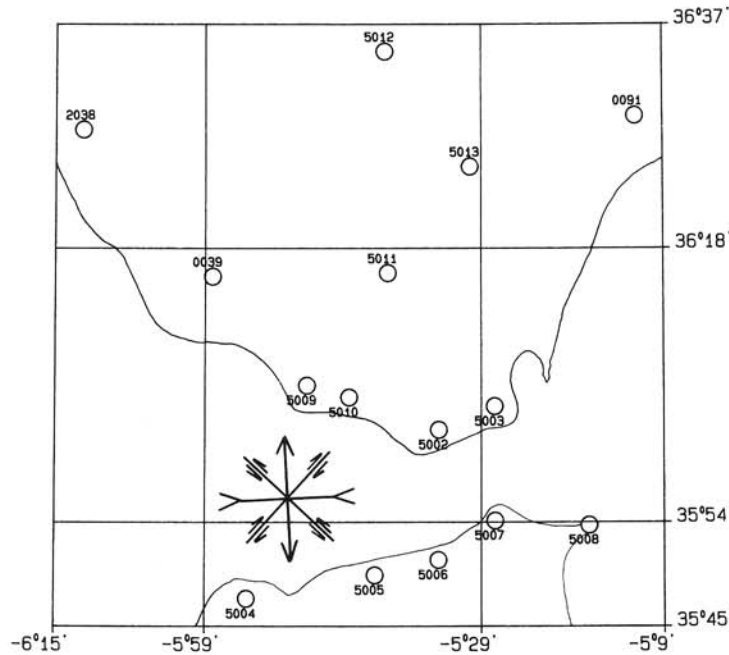
The results for the shear strain rate of solution (3) are shown in Fig. 2. For comparison, we illustrate in Fig. 3 the point displacement solution by Fredrich & Hein (1991). Here, the co-ordinates resulting from the separate adjustment (by OPERA software; Landau *et al.* 1988) of the data of each campaign were related by a Helmert Transformation, i.e. the network of one epoch was subjected to a translation, rotation and scaling with respect to that of the other so as to minimize the sum of squared lengths of the displacement vectors. The length-scales of EDM and GPS measurements are thus assumed not to be comparable.

**Table 1.** Numbers of observations ( $N$ ), and rms adjusted residuals for the major observational groups used in the three different solutions.

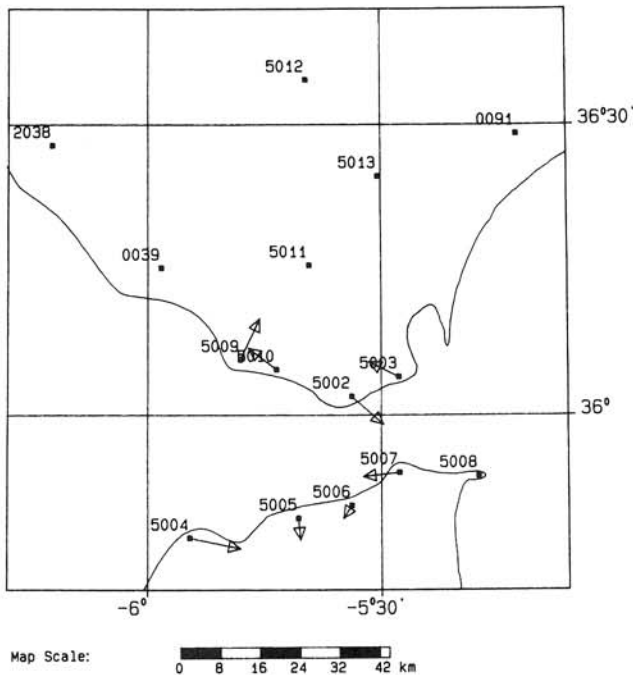
	N	(1)	(2)	(3)
Distance	21	8.5 mm		
Distance ratio	14		0.21 mm/km	0.24 mm/km
Horizontal angle	33	0.28 sec	0.28 sec	0.28 sec
Zenith distance	41	6.4 sec	6.4 sec	6.3 sec
GPS vector (magnitude)	70	36.4 mm	36.5 mm	31.5 mm
east	70	17.9 mm	18.1 mm	16.4 mm
north	70	17.7 mm	17.9 mm	13.8 mm
vertical	70	33.6 mm	33.7 mm	30.2 mm

**Table 2.** Homogeneous strain rates (in units of  $10^{-6} \text{ yr}^{-1}$ ) for the three different solutions.

	(1)	(2)	(3)
Dilatation rate	$-(0.07 \pm 0.04)$		
Magnitude of shear strain rate	$(0.12 \pm 0.04)$	$(0.13 \pm 0.04)$	$(0.14 \pm 0.04)$
Axis of maximum relative shortening	$87^\circ/267^\circ \pm 9^\circ$	$81^\circ/261^\circ \pm 8^\circ$	$87^\circ/267^\circ \pm 8^\circ$



**Figure 2.** Principal axes of homogeneous strain rate determined for the Strait of Gibraltar. The axis of maximum relative shortening is nearly E–W ( $87^\circ/267^\circ \pm 8^\circ$ ), the axis of maximum relative lengthening N–S, with the axes of maximum shear lying between, as indicated. The magnitude of the (tensor) shear strain rate is  $(0.14 \pm 0.04) \times 10^{-6} \text{ rad yr}^{-1}$  (Solution 3). If the EDM and GPS scales are taken as comparable (Solution 1), then the extensional strain rate on the shortening (E–W) axis is  $-(0.19 \pm 0.06) \times 10^{-6} \text{ yr}^{-1}$ , and on the lengthening (N–S) axis is  $(0.05 \pm 0.05) \times 10^{-6} \text{ yr}^{-1}$ .



**Figure 3.** The point displacement solution for the Strait of Gibraltar as determined by Fredrich & Hein (1991).

**Vertical deformation**

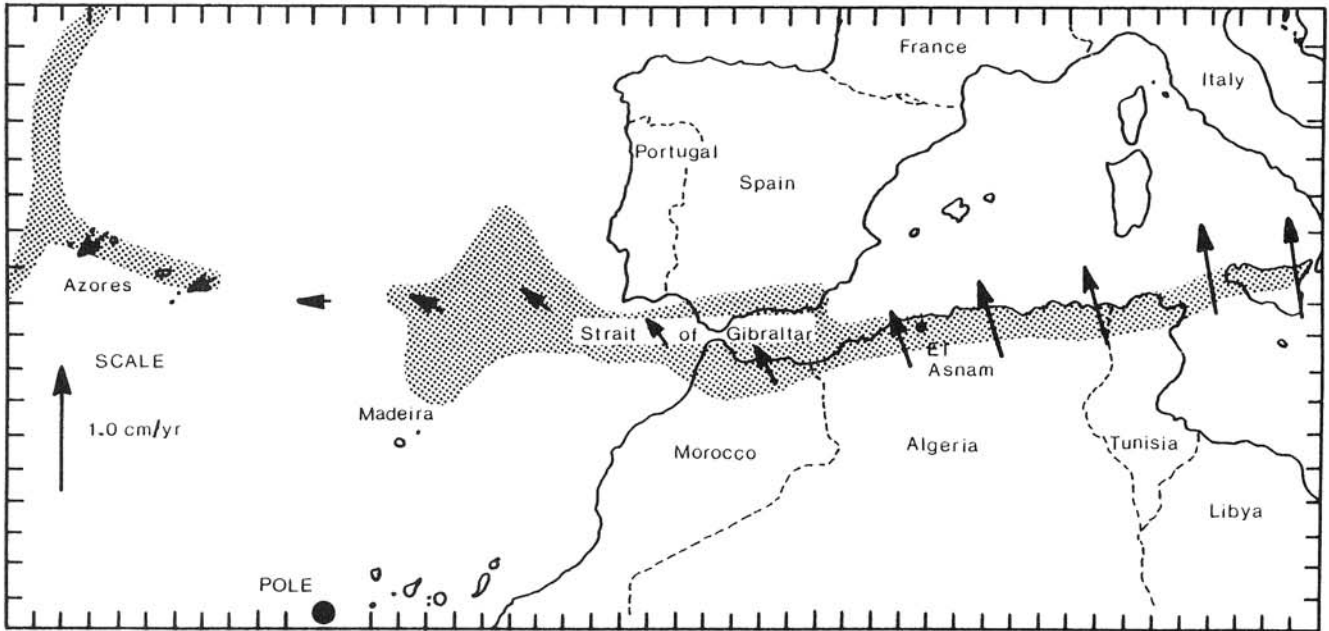
An attempt was made to determine the tilting of the network: this depended essentially on the results of vertical angle (zenith distance) observations in the first epoch, and

GPS in the second. The solution yielded a standard deviation of  $1.3 \times 10^{-6} \text{ rad yr}^{-1}$  of the homogeneous tilt vector, almost two orders of magnitude greater than that for horizontal strain components. This was achieved by introducing the zenith distance observations individually, and determining single coefficient of refraction for the whole set. If the zenith distance observations were treated, as they were observed, as simultaneous reciprocal observations, and differenced to eliminate the first-order refraction effect, then the normal equation matrix became ill-conditioned, and the iterative solution for the non-linear observation equations diverged. It appears that the ability of each observation line to assume an unrestricted curvature in the latter solution leaves insufficient strength in the network, in the vertical sense, to permit the determination of both vertical position and tilt—a ‘mirage’ effect. Introduction of the levelling data, running along both the Spanish and the Moroccan shores, strengthens the network in the E–W direction; but the tilt determination is then highly anisotropic, with the error variance in the N–S direction 16 times that in the E–W direction, and the resultant tilt vector inevitably aligned nearly N–S.

**DISCUSSION**

**Plate tectonic models**

The boundary between the Eurasian and the African lithospheric plates is usually assumed to extend from the Azores triple junction along the Azores–Gibraltar Ridge, through the Strait of Gibraltar, and into the Mediterranean Sea (e.g. Anderson & Jackson 1987, Figs 1 and 2; Argus



**Figure 4.** The Eurasian–African plate boundary zone from the Azores to Sicily. Shaded areas denote the zones of most intense seismicity. The arrows denote the relative velocity of the African with respect to the Eurasian plate, with respect to a pole of rotation at 27.6°N, 19.7°W (shown at lower left). (After Anderson & Jackson 1987, Fig. 2).

*et al.* 1989, Fig. 1). The seismicity of the boundary zone is in part diffuse, as shown in Fig. 4 (taken from Anderson & Jackson 1987, Fig. 2), and particularly so in the region of the Strait of Gibraltar. Evidence for the relative motion between the Eurasian and African plates is summarized by these authors, and incorporated in the NUVEL-1 model of DeMets *et al.* (1990); at the Strait of Gibraltar (36°N, 5°W) the velocity of the African with respect to the Eurasian plate is predicted as  $5 \text{ mm yr}^{-1}$  at an azimuth of 316°. If there were in fact an active plate margin trending E–W through the Strait of Gibraltar, it would be obliquely convergent. The existence of an active margin through the Strait, the slip on which would be characterized by a persistent and concentrated band of seismicity, does not appear to be supported by the available evidence, as presented by Purdy (1975), Grimison & Chen (1988) and Argus *et al.* (1989). A recent report summarizing the tectonic activity of the Strait (SNED & SECEG 1990) stresses that there is no conclusive evidence of any tectonic structure with associated seismicity crossing the area of the Strait; and concludes that on both shores of the Strait, systems of conjugate faults that indicate an E–W compression that has persisted up to the Pliocene are to be found. We might infer from our results that, far from ceasing in the Pliocene, E–W compression (and N–S extension) may be continuing at the present day, detectable even on time-scales as short as 5 yr. It is likely, however, that the interplate motion in the region of the Strait of Gibraltar is being accommodated in a zone of diffuse deformation, perhaps over 100 km in width; and the present measurements across the Strait may have sampled only a minor part of this zone.

#### Strain-rate determinations elsewhere in the western Mediterranean area

There appear to be few geodetic strain determinations in the western Mediterranean; the nearest is a study by Ruegg *et al.*

(1982) of the area of the El Asnam (Algeria) earthquake of 1980 October 10. These authors have reported the results of remeasurement (after an interval of 5 yr) of a geodetic network of about  $30 \times 30 \text{ km}$  in extent around the epicentre, and have presented ‘mean strain tensors’ deduced for the constituent triangles of the network. It is possible to derive a single mean strain-rate tensor, comparable with the results for Gibraltar presented above, by an analysis of the displacement vectors for 14 network stations given in their Table 3 (Ruegg *et al.* 1982, p. 2232). This tensor yields:

$$\begin{aligned} \text{Dilatation rate} & \\ & -(0.9 \pm 1.4) 10^{-6} \text{ yr}^{-1}; \\ \text{Shear strain rate} & \\ & (2.1 \pm 1.4) 10^{-6} \text{ rad yr}^{-1}; \text{ and} \\ \text{Axis of maximum relative shortening} & \\ & 158^\circ/338^\circ \pm 19^\circ, \end{aligned}$$

where the quoted accuracies (to 1 SD) are derived from the distribution of residuals of the published displacement vectors, rather than from the original data. These values are an order of magnitude greater than those derived for the Strait of Gibraltar, no doubt largely because the geodetic observations at El Asnam closely encompass the earthquake of magnitude  $M_s = 7.3$ . Ruegg *et al.* (1982) give the strike azimuth of the surface fault break as 230° (with normal 140°/320°); the strain deduced above suggest a mean rate of extensional strain normal to the fault is  $-(2.6 \pm 1.9) 10^{-6} \text{ yr}^{-1}$ , in accordance with the inferred thrust faulting mechanism, and there is a possible residual component of left-lateral shear of  $(1.2 \pm 1.4) 10^{-6} \text{ rad yr}^{-1}$ . Ruegg *et al.* (1982) also recommended discarding five of their 14 stations for various reasons; in this case the axis of maximum relative shortening becomes  $175^\circ/355^\circ \pm 18^\circ$ . Their shortening axis thus lies in the SE–S octant rather than the E–W direction of our result at the Strait of Gibraltar. At El Asnam, the NUVEL-1 value for the azimuth of the relative motion of the Eurasian with respect to the African plate is

147°, rotated 11° clockwise from that at the Strait of Gibraltar (136°).

Yielding (1985) reinterpreted the El Asnam main shock as a multiple event, and has given focal mechanism solutions for the main shock ( $M_s = 7.3$ ), and for the largest aftershock ( $M_b = 6.1$ ) that occurred 3 hr later: the fault planes are inferred to have had a dip of 54° (northwest), and a strike azimuth of 40°/220° for the main shock, but of 80°/260° for the aftershock. In each case, the quoted rake angle of 90° implies that the slip was normal to the strike axis, and that the shortening (compressive) directions will therefore be 130°/310° for the main shock, and 170°/350° for the aftershock; the latter is closer to the revised maximum relative shortening axis at 175°/355° ± 18° deduced from the data of Ruegg *et al.* (1982).

Apart from the work of Ruegg *et al.* (1982) at El Asnam, there appear to be few geodetic strain determinations in the western Mediterranean. Elsewhere, in the Alpine region, also considered to be a part of the Eurasian–African collision zone, Reilly & Gubler (1990) have reported a shear strain rate in central Switzerland of magnitude similar to that at Gibraltar,  $(0.10 \pm 0.02) 10^{-6} \text{ rad yr}^{-1}$ , with an axis of relative shortening approximately NW–SE, derived primarily from a comparison of triangulation measurements extending over 100 yr.

From the area of the Friuli earthquake of 1976, Reilly & Arca (1987) determined a shear strain rate of  $(0.38 \pm 0.10) 10^{-6} \text{ rad yr}^{-1}$  with an axis of maximum relative shortening of 176°/356° ± 7°, determined from a repetition of triangulations over an interval of 28 yr and an area of 40 × 60 km. The area surveyed is similar to that at El Asnam, the time interval is greater by a factor of 6, and the total shear strain in the measurement interval is (coincidentally?) similar:  $(10 \pm 7) 10^{-6} \text{ rad}$  at El Asnam, and  $(10 \pm 3) 10^{-6} \text{ rad}$  in Friuli.

## CONCLUSIONS

The results obtained demonstrate the capabilities of both EDM and GPS measurements to determine the rate of strain in the Earth's crust over intervals as short as 5 yr. The problem now, however, is how to assess the significance of such results in the geodynamic setting of the Strait of Gibraltar. The contemporary shear strain appears to be consistent in direction with long term (ca.  $10^7$  yr) trend evidenced in the faulting patterns. It would, however, be much easier to assess the rate of crustal strain at the Strait of Gibraltar in a regional context if analogous results were available from a wider encompassing area. Further evidence over short (ca. 5 yr) time intervals can be expected from the results of the WEGENER-MEDLAS Project (e.g. Baldi & Zerbini 1988). For a longer time interval (ca. 50–100 yr), repeated triangulation surveys would be ideal. The national first-order triangulation surveys of the Iberian Peninsula would appear to be of an age and an accuracy to offer valuable results from even partial repetition.

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## REFERENCES

- Anderson, H. & Jackson, J., 1987. Active tectonics of the Adriatic Region, *Geophys. J. R. astr. Soc.*, **91**, 937–983.
- Argus, D. F., Gordan, R. G., DeMets, C. & Stein, S., 1989. Closure of the Africa–Eurasia–North America plate motion circuit and tectonics of the Gloria Fault, *J. geophys. Res.*, **94**, 5585–5602.
- Baldi, P. & Zerbini, S. (eds), 1988. *Proceedings of the Third International Conference on the WEGENER-MEDLAS Project*, Esculapio, Bologna, 427 pp.
- Bibby, H. M., 1973. The reduction of geodetic survey data for the detection of earth deformation, *Geophysics Division Report 84*, Wellington.
- Bibby, H. M., 1982. Unbiased estimate of strain from triangulation data using the method of simultaneous reduction, *Tectonophysics*, **82**, 161–174.
- Caturla, J. L., Mourón, A. D., Barrado, J., Padilla, R. & Almazán, J. L., 1990. Campaña de observación GPS de la Red Geodésica para Observaciones Geodinámicas del Estrecho de Gibraltar, *Unpublished Report*, Instituto Geográfico Nacional de España, Madrid.
- DeMets, C., Gordon, R. G., Argus, D. F. & Stein, S., 1990. Current plate motions, *Geophys. J. Int.*, **101**, 425–478.
- Fredrich, G. & Hein, G. W., 1991. *Eurasian–African plate motion: results of a four-dimensional integrated adjustment of the Strait-of-Gibraltar network*, Institute of Astronomical and Physical Geodesy, University FAF Munich, D-8014 Neubiberg.
- Grimson, N. L. & Chen, W.-P., 1988. Source mechanisms of four recent earthquakes along the Azores–Gibraltar plate boundary, *Geophys. J.*, **92**, 391–401.
- IGNE & SECEG, 1987. *Trabajos Geodésicos en el Estrecho de Gibraltar Años 1985–1986*, Madrid, Instituto Geográfico Nacional de España, Sociedad Española de Estudios para la Comunicación fija a través del Estrecho de Gibraltar.
- Landau, H., 1991. *TOPAS User's Guide*, Institute of Astronomical and Physical Geodesy, University FAF Munich, D-8014 Neubiberg.
- Landau, H., Hehl, K., Eissfeller, B. & Hein, G. W., 1988. OPERA 2.4—User's Guide, in *Operational Geodesy Software Packages*, Schriftenreihe des Studiengangs Vermessungswesen der Universität der Bundeswehr München 34, pp. 5–250.
- Purdy, G. M., 1975. The eastern end of the Azores–Gibraltar plate boundary, *Geophys. J. R. astr. Soc.*, **43**, 973–1000.
- Reilly, W. I., 1982. Three-dimensional kinematics of earth deformation from geodetic observations, *Veröffentlichungen Deutsche Geodätische Kommission*, **B. 258/VII**, 207–221.
- Reilly, W. I., 1987. Continuum models in earth deformation analysis, in *Determination of Heights and Height Changes*, pp. 557–569, eds Pelzer, H. & Niemeier, W., Bonn, Dümmler.
- Reilly, W. I., 1988. A User's Guide to GEONET, in *Operational Geodesy Software Packages*, Schriftenreihe des Studiengangs Vermessungswesen der Universität der Bundeswehr München 34, pp. 253–323.
- Reilly, W. I., 1990a. Geodetic analysis of motion at a convergent plate boundary, in *Developments in Four-Dimensional Geodesy*, Lecture Notes in Earth Sciences 29, pp. 139–159, eds Brunner, F. K. & Rizos, C., Springer, Berlin.
- Reilly, W. I., 1990b. A User's Guide to GEONET (2nd. Revised Ed.), *Geophysics Division Technical Report 110*, Department of Scientific and Industrial Research, Wellington, New Zealand, 36 pp.
- Reilly, W. I. & Arca, S., 1987. Crustal strain in Friuli, 1949–77, *Boll. Geod. e Sc. Aff.*, **46**, 247–254.

- Reilly, W. I. & Gubler, E., 1990. Crustal strain in Switzerland 1870–1970, *Geophys J. Int.*, **103**, 251–256.
- Ruegg, J. C., Kasser, M., Tarantola, A., Lepine, J. C. & Chouikrat, B., 1982. Deformations associated with the El Asnam earthquake of 10 October 1980: geodetic determination of vertical and horizontal movements, *Bull. seism. Soc. Am.*, **72**, 2227–2244.
- SNED & SECEG, 1990. Rapport informatif: A. Milieu Physique, *Troisième Colloque International sur la Liaison Fixe à travers le Détroit de Gibraltar*:
- Yielding, G., 1985. Control of rupture by fault geometry during the 1980 El Asnam (Algeria) earthquake, *Geophys. J. R. astr. Soc.*, **81**, 641–670.