Crustal strain in the Southern Alps, France, 1948–1998

E. Calais a,*, L. Galisson a,b, J.-F. Stéphan a, J. Delteil a, J. Deverchère a, C. Larroque c, B. Mercier de Lépinay a, M. Popoff a, M. Sosson a

a UMR 6526, Centre National de la Recherche Scientifique, Université Pierre et Marie Curie, Université Nice-Sophia Antipolis, 250 Rue Albert Einstein, 06560 Valbonne, France
b Ecole Nationale Supérieure des Arts et Industries de Strasbourg, Strasbourg, France
c Université de Reims, Laboratoire de Sciences de la Terre, Reims, France

Received 9 June 1999; accepted for publication 14 January 2000

Abstract

Active tectonics in the Western Alps is revealed by a moderate level of seismic activity and geological evidences for Quaternary deformation. We present new geodetic determinations of the current strain rates in the southern part of the Western Alps, that complement existing results in Provence, the Jura, and the Subalpine and Belledone massifs around Grenoble. We combined first- and second-order triangulation data collected in 1948 with global positioning system (GPS) data collected in 1998. We estimate shear strain rates of 0.1–0.2 μrad yr⁻¹ over distances on the order of 30 km, significantly different from zero in most of the network. We obtain NW–SE to N–S compressive shear strain directions over most of the area, in agreement with the seismological and geological data. These geodetic estimates correspond to long-term overall N–S to NW–SE shortening of 2–4 mm yr⁻¹ over the study area, on the same order as the far-field shortening measured by continuous GPS on the Grasse–Torino baseline (−2.0 ± 0.5 mm yr⁻¹), that incorporates our geodetic network. These results provide new constraints on interseismic strain in the Western Alps. © 2000 Elsevier Science B.V. All rights reserved.

Keywords: active deformation; Alps; global positioning system; triangulation

1. Introduction

Active deformation in the Alpine mountain range is commonly interpreted as a consequence of the present-day convergence between the African and Eurasian plates (Fig. 1; e.g. Philip, 1987). This convergence currently occurs at a rate of 6.2 ± 0.5 mm yr⁻¹ in a N17°W ± 9° direction at the longitude of the Western Alps according to the Nuvel-1A plate motion model (DeMets et al., 1990) a value that has been confirmed by space geodetic measurements in Sicily (Ward, 1994; Zarraoa et al., 1994). Although this far-field kinematic framework seems well constrained, the kinematics of the present-day deformation within the Africa–Eurasia plate boundary in the Western Mediterranean remains poorly documented. This hampers the understanding of the physical mechanisms responsible for active deformation and earthquakes within the plate boundary zone. For instance, the amount of Africa–Eurasia convergence currently accommodated in the Alps is not known, although this information is crucial for estimating the force balance currently driving its deformation (e.g. Lyon-Caen and Molnar, 1989).
Fig. 1. Regional tectonic setting (after Armijo et al., 1986; digitized data courtesy G. Ferhat). Areas in light gray indicate stable blocks or plates, areas in dark gray indicate active deformation regions. Squares and circles indicate deep and shallow earthquakes, respectively; filled circles before 1950, open circles after 1950. The large white triangles show permanent GPS stations, in particular Grasse and Torino, mentioned in this paper. Arrows at the bottom of the figure indicate the velocity of the African plate relative to Eurasia according to the Nuvel-1A plate kinematic model (DeMets et al., 1994). The white square indicates the study area.

The present-day tectonic activity of the Western Alps (Fig. 1) is demonstrated by a moderate seismicity (Thouvenot, 1996; Maddedu et al., 1997) and sparse geological and geomorphological observations of recent deformation (e.g. Ménard, 1988; Labaume et al., 1989; Ritz, 1991; Dubar and Pérez, 1989; Combes, 1984; Grellet et al., 1993; Sue et al., 1999). The diffuse spatial distribution of the seismicity and the difficulty to characterize active faults in an environment where erosion and anthropic activity compete with very low strain rates makes the application of classical seismotectonic analysis techniques difficult (Sébrier et al., 1997). In addition, the available geodetic measurements of crustal deformation in the Western Alps are so far controversial. Jouanne et al. (1994) in the Jura and Martinod et al. (1996) around Grenoble report strain rates ranging between 0.5 and 1 strain yr$^{-1}$, as high as those measured around such well-known active faults as the San Andreas fault in southern California (Feigl et al., 1990; Shen et al., 1996) or the North Anatolian fault in Turkey (Straub and Kahle, 1995). In the Southern French Alps and Provence, Ferhat et al.
(1998) place a much lower upper bound of 0.1 μrad yr⁻¹ on active deformation. Reilly and Gubler (1990) report similar strain rates of 0.21 ± 0.04 μrad yr⁻¹ in the Helvetic Alps and 0.05 ± 0.02 μrad yr⁻¹ in the Jura mountains. Finally, the analysis of 2.5 years of continuous global positioning system (GPS) data at Torino (Italy), Grasse (France), and Zimmerwald (Switzerland) suggests that the present-day differential motion across the Western Alps does not exceed 3 mm yr⁻¹ (Calais, 1999).

There is clearly a need for direct geodetic determinations of crustal strain and block motion in the Alps, but the low expected rates, close to the accuracy of the geodetic techniques, make such measurements challenging. GPS campaign measurements of crustal deformation in the Western Alps started in 1993 (Chéry et al., 1995). However, unambiguous detection of a tectonic signal greater than the GPS noise will take a minimum of three observation epochs over at least 5 years. Continuous measurements at permanent stations have also started (Calais et al., 1998), but their spatial resolution is coarse (>50 km between stations). An alternate strategy involves the comparison of old triangulation data with recent triangulation or GPS measurements. This paper presents the results of a comparison between triangulation and GPS measurements made 50 years apart in the southernmost part of the Western Alps (Alpes Maritimes, France). It complements similar studies by Reilly and Gubler (1990) in the Helvetic Alps, Jouanne et al. (1994) in the southern Jura, Martinod et al. (1996) in the Subalpine and Belledone massifs around Grenoble, and Ferhat et al. (1998) in the southern Alps and Provence.

2. Seismotectonic setting

The southern part of the Western Alps (Alpes Maritimes) is seismically one of the most active part of the Alpine range (Fig. 1). The seismicity is mostly concentrated on the northeastern side of the Argentera massif along the Pennic front, with a variety of compressive, extensive, and strike-slip focal mechanisms and P-axes trending SW–NE to E–W (Fig. 2; Béthoux et al., 1998; Madeddu et al., 1997). The largest instrumental earthquakes in the area, however, occurred along the northern margin of the Ligurian basin (Fig. 2; 1963, magnitude 6.0; 1989, magnitude 4.5; 1995, magnitude 4.7; Ritz et al., 1990; Béthoux et al., 1992; Courboulex et al., 1998). Focal mechanisms generally show compression and strike–slip, with P-axes trending NW–SE to N–S (Béthoux et al., 1992). A number of historical earthquakes are also known in the Alpes Maritimes, some of them associated with severe casualties, such as the 1887 Ligurian earthquake, that reached an intensity of X MSK (estimated magnitude 6.4; Levret et al., 1994) and affected most of the French Riviera between Nice and Menton. The occurrence of such events represents a serious threat to this densely populated and highly touristic area. In addition to the geodynamic objectives cited already, this study also aims at quantifying present-day strain rates in the Alpes Maritimes in order to contribute to seismic hazard analyses.

From a geological standpoint, the Alpes Maritimes are located at the crossroads between the highly elevated crystalline massif of the Argentera, the sedimentary units of the southern Subalpine Massifs (the so-called Castellane and Nice arcs, or fold-and-thrust belts), and the Ligurian oceanic domain and its steep and narrow margin (Fig. 2). Our study area is centered on the Castellane and Nice arcs, a series of S to SSW-verging fold and thrusts involving the Mesozoic to Paleogene sediments, translated above a basal decollement zone in upper Triassic evaporites and shales [Fig. 2B; Goguel (1936); Géze (1960, 1963); Pérez (1975); Gidon and Pairis (1986); Pelline-Chabert (1996); Laurent (1998)]. This decollement is connected to the north, in the inner part of both arcs, with a blind north-dipping crustal ramp associated with basement uplifts of the Barrot Dome and the Argentera Massif [Fig. 2B; Siddans et al. (1984); Labaume et al. (1989); Ritz (1991); Laurent et al. (1999)].

Although most of the deformation in the Castellane and Nice arcs occurred during the Miocene, there is evidence for shortening and uplift continuing into the Pliocene and the
Quaternary, under a stress-field similar to the Miocene one (Ritz, 1991, 1992). Marine lower Pliocene sediments that discomformably overlie the western edge of the Nice arc along the Var river are locally folded and underthrust (Campredon et al., 1977; Irr, 1984; Clauzon et al., 1996; Schroetter, 1998). Northeast of the Var and Vésoublie rivers confluence (ca. 25 km SW of site MOUL, Fig. 2A), Pliocene conglomerates infilling the Messinian paleo-Var valley were uplifted to

Fig. 2. (A) Geodetic network superimposed on a structural map of recent/active faults and seismicity in the Southern French Alps. The open triangle is the GRAS permanent GPS site. BSMF, Breil–Sospel–Monaco fault; STF, Saorge–Taggia fault; VMBA, Vésoublie–Mont Férion–St Blaise–Aspermont fault. (B) Geological cross-section from the Ligurian Sea to the Argentera massif. Structural and geological contours for A and B are from Laurent (1998), Chaumillon et al. (1994), and Stéphan (unpublished data), seismological informations from Béthoux et al. (1998), Moho depth from Recq (1973, 1974), Calais et al. (1993) and Chamot-Rooke et al. (1999). The geometry of the crustal ramps in (B) is speculative.
Fig. 2. (continued)
1030 m above a south-vergent thrust (Irr, 1984; Clauzon et al., 1996; Schroetter, 1998). In the Castellane arc, Gilli and Delange (1999) found evidence for very recent southward thrusting from offsets in speleothems built in karstic galleries along a north-dipping thrust plane. Recent movements are also evidenced by deformed Plio-Pleistocene sediments along the left-lateral Breil–Sospel–Monaco strike–slip fault zone and the Saorge–Taggia right-lateral strike–slip fault [Fig. 2A; Rebai (1988); Ritz (1991); Cosani (1997)]. Quaternary tectonic activity is also evidenced by uplifted and faulted marine and river terraces along the coast from the Maures–Tanneron massif eastwards and on both flanks of the lower Var valley (Dubar, 1987; Dubar and Pérez, 1989). Uplift was greater on the east bank of the Var (the 1.5–1.8 My terrace is 50 m higher) with a transpressional regime to the east (NNE–SSW shortening) and a mostly extensional one to the west (reactivation of Oligocene N–S graben) in the Provençal foreland of the Castellane arc (Dubar and Pérez, 1989). These differential movements on both sides of the lower Var valley are probably related to the current activity of the so-called Vésubie–Mont Férion–St Blaise–Aspremont lineament (Fig. 2), an inherited N–S fault zone that shows radon anomalies (Borchiellini et al., 1991) and microseismic alignments (Tressols, 1996). Offshore along the Ligurian margin, seismic reflection data between the Argentera and the coast, incision of the hydrographic system, interpreted by Chaumillon et al. (1994) as an uplift of the margin at a rate of 0.3–0.5 mm yr$^{-1}$. North of the Castellane and Nice arcs, apatite fission track data show that the Argentera Massif is currently uplifting at a minimum rate of 0.1 mm yr$^{-1}$ (Bigot-Cormier et al., 1999). Between the Argentera and the coast, incision of the hydrographic system, together with local anomalies in longitudinal profiles of some rivers, also suggests recent tectonic activity (Julian, 1980; Guglielmi, 1993).

In spite of these numerous indications for active deformation in the southern part of the Western Alps, no direct determinations of the current strain rates are so far available. We present in the following the methodology that we used in order to determine strain rates in that area, discuss the results and their tectonic implications.

3. Data acquisition and analysis

Triangulation data come from first- and second-order surveys performed in 1948 by the French Institut Géographique National (IGN) at 35 sites (Fig. 3). The 1948 measurements consist of angular values only. They were acquired with Wild T3 theodolites, with an associated uncertainty of 0.5 arcsec, or about 2.4 microradians ($\mu$rad) (Lejeune, 1950). We performed an adjustment of the 1948 triangulation network in order to check the quality of the data and select a list of sites that had the appropriate planimetric accuracy (Fig. 3). We decided to retain only the points for which it was possible to ensure a millimeter per year level accuracy over 50 years, that is, points that were defined with an accuracy better than 5 cm for their east and north coordinates. We also performed a Helmert transformation of the adjusted triangulation network to the GPS coordinates (see below) in order to identify sites with abnormal coordinate differences. Fourteen sites did not pass these two tests because of insufficient or bad quality triangulation data. They were not used in the rest of this study. Moreover, we chose to measure only the exact benchmarks that had been occupied during the 1948 surveys in order to avoid using local ties to auxiliary benchmarks that would degrade the overall accuracy of our solution. Among the remaining points, two were unreachable because of late snow falls in spring 1998 (Valdieri, Punta Maladecia), three had been destroyed (Savi, Aiera, and Fascia), and for one site we did not obtain permission to observe it (Peille). We finally observed a total of 14 sites (Fig. 2). The two-dimensional rms of the final triangulation network adjustment is 3 cm, which makes this data set appropriate for a comparison with GPS data collected 50 years later in 1998.

GPS data were collected between 7 and 16 June 1998, using four dual-frequency Ashtech Z12 receivers sampling at a 30 s interval during 8–10 h sessions. We processed pseudorange and phase GPS data in single-day solutions using the ‘GAMIT’ software (version 9.6, King and Bock,
1997). We solved for regional station coordinates, satellite state vectors, and phase ambiguities using doubly-differenced ionosphere-free (‘LC’) GPS phase measurements. We used IGS final orbits, IERS Earth orientation parameters, and applied azimuth and elevation dependent antenna phase center models, following the tables recommended by the IGS. We included in the processing seven stations of the International GPS service for Geodynamics (IGS) (Grasse, Matera, Kootwijk, Zimmerwald, Graz, Torino, Villafranca) to serve as ties with the International Terrestrial Reference Frame 1994 (ITRF94, Boucher et al., 1996). The least squares adjustment vector and its corresponding variance–covariance matrix for station positions and orbital elements estimated for each independent daily solution were then passed to a Kalman filter (‘GLOBK’, Herring et al., 1990) in order to perform a global network adjustment and to estimate station positions and their formal errors. We imposed the reference frame by minimizing the position and velocity deviations of IGS stations with respect to the ITRF94 while estimating an orientation, translation and scale transformation.

Since none of the sites were occupied during more than one session, it was not possible to compute measurement rms scatter (or ‘short-term repeatability’), a value that is commonly used to estimate the accuracy of GPS measurements (Larson and Agnew, 1991). However, a number of tests on permanent GPS stations now operating
in the Western Alps show a short-term repeatability on the order of 5 mm in the planimetric components with 10–12 h sessions on 30–100 km long baselines using the processing strategy described above (which are the characteristics of the network analyzed in this study). Since all the sites were located in a low-multipath environment with a good sky visibility, we assume that this accuracy also holds for our measurements.

4. Strain rates estimation

We estimate the strain rates using the forward modeling network deformation analysis software FONDA (Dong, 1993) following the standard procedure used by Ferhat et al. (1998). The input data sets are the angles from the triangulation measurements and the baseline representation of the coordinate estimates from the GPS measurements, both with their associated variance. In a first step, FONDA analyses each data set and produces loosely constrained estimates of site positions at each observation reference epoch. We constrained the vertical position components to their GPS-derived value because these parameters cannot be estimated from the triangulation data. All other parameters were loosely constrained in order to allow the reference frame to be consistently defined at the combination stage. In a second step, FONDA combines individual loosely constrained solutions at each observation reference epoch into a single solution, allowing time-dependent variation of the parameters. The result is a horizontal position (relative to a reference epoch) and a velocity for each station, together with strain and rotation rates over the network. A full description of the FONDA procedure can be found in Dong (1993) and Dong et al. (1998).

The deformation can then be derived from a steady homogeneous horizontal velocity field given by:

\[ V_n = L_{11} x_n + L_{12} x_e + t_n \]
\[ V_e = L_{21} x_n + L_{22} x_e + t_e \]

where \( V_n \) and \( V_e \) are the north and east components of velocity at a site located at \((x_n, x_e)\), \( L_{ij} \) are the components of the velocity gradient tensor \( L \) that describes how velocity varies spatially, and \( t_n \) and \( t_e \) represent the rigid body translation velocity common to all points. The velocity gradient tensor can be decomposed into a symmetric part, \( E \), the horizontal strain rate tensor, and an asymmetric part \( W \), the horizontal rotation rate matrix:

\[
L = E + W = \begin{pmatrix}
\dot{E}_{11} & \dot{E}_{12} \\
\dot{E}_{21} & \dot{E}_{22}
\end{pmatrix} + \begin{pmatrix}
0 & \omega \\
-\omega & 0
\end{pmatrix}
\]

The strain rate tensor \( E \) can be diagonalized and represented in a coordinate system defined by the principal strain rate axes using its eigenvalues \((\dot{e}_i)\) and unit eigenvector \((\hat{e}_i)\):

\[
\dot{E} = \sum_{i=1}^{2} \dot{e}_i \hat{e}_i^T
\]

By convention, extension is positive, \( \dot{e}_1 > \dot{e}_2 \), and the principal shortening axis is oriented at an azimuth \( \Theta \) measured clockwise from north. This direction is oriented at 45° angle from the maximum shear direction.

The 1948 surveys involved only angular observations, with no distance or absolute azimuth measurements. Consequently, the derived site coordinates and the velocity field obtained from a comparison with GPS measurements cannot yield rotation \((\hat{\omega})\) and dilatation rates \((\Delta = \dot{E}_{11} + \dot{E}_{22})\). The velocity gradient tensor and even the strain rate tensor are therefore ambiguous unless two sites of the measured network are fixed. Since the choice of the fixed sites can have a large impact on the strain rates and azimuths, and because we have no a priori way of knowing which sites experienced zero relative motion between the 1948 and 1998 surveys, we chose to use a parameterization of the velocity gradient tensor in terms of two angular shear strain rates \( \gamma_1 \) and \( \gamma_2 \):

\[
\gamma_1 = (\dot{E}_{11} - \dot{E}_{22})
\]
\[
\gamma_2 = (2\dot{E}_{12})
\]

The meaning of \( \gamma_1 \) and \( \gamma_2 \) and their relation with the velocity field gradient and the conventional strain rate tensor are explained for instance in Feigl et al. (1990), to which we refer the reader
for further details. This parameterization is independent on any assumption of fixed sites and is not affected by the unknown rotation and dilatation rate. The drawback is that the interpretation of $\gamma_1$ and $\gamma_2$ in geological terms is not straightforward. We therefore decided to follow the approach of Ferhat et al. (1998) and calculated maximum shear strain rates $\gamma$:

$$\gamma = \sqrt{\gamma_1^2 + \gamma_2^2}$$

together with their azimuth $\Psi$ given by:

$$\tan(2\Psi) = \gamma_1 / \gamma_2.$$  

We plotted in Figs. 4–6 the azimuth $\Theta = \Psi - 45^\circ$, which is the direction of the axis of the maximum compressional principal strain rate, corresponding to the 'P-axis' in seismological nomenclature. Such a representation has the advantage of being easily readable and directly comparable to seismological results, but the disadvantage of not being able to distinguish pure shear from simple shear.

5. Results analysis and robustness tests

In order to assess the quality of our solutions and the influence of the processing strategy on the strain rate estimates, we performed a series of robustness tests on the quality of the points used in the solutions, the a priori uncertainties used for the station positions, and the geometry of the subnetworks.

5.1. Quality of benchmarks

We observed on the field that all the geodetic benchmarks were not in the same state of conservation and consequently had to be weighted accordingly. We rank the sites into three categories on the basis of the following criterias: category A sites: no ambiguity on the benchmark, millimeter-level centering possible, monument stable and undisturbed (7 points); category B sites: no ambiguity on the benchmark, millimeter-level centering possible, monument stability questionable (4 points); category C sites: ambiguity on the benchmark, and/or millimeter-level centering not possible.

Fig. 4. Test of the influence of the site quality on shear strain rate estimates. The shear strain rate is shown in each subnetwork with the following symbols: thick black bars indicate the value of the shear strain rate $\gamma$, the superimposed thick gray bars give the one-sigma uncertainty. The two symmetrical thin black lines give the one sigma uncertainty on the azimuth of the 'geodetic P-axis' $\Theta$. (A) Solution using category A sites only; (B) solution using category A and B sites; (C) solution using category A, B, and C sites.
It appears clearly that including all category A, B, and C sites (Fig. 4C) results in an unreasonable solution, with maximum shear strain rates $>1 \mu \text{rad yr}^{-1}$, on the same order as the values found for instance in southern California (Feigl et al., 1990; Dong, 1993), not compatible with the seismotectonic information in the Alpes Maritimes (e.g. Maddedu et al., 1997; Béthoux et al., 1998). The same conclusion holds when including category C sites one by one in the solution. We also observe that the solution with category A and B sites is not significantly different from the solution with category A sites alone for the subnetworks common to the two solutions. This proves that category B sites do not affect the quality of the solution. We will therefore base our interpretation on solutions that include both category A and B sites.

5.2. A priori uncertainties

We performed a series of solutions by multiplying the a priori uncertainties on the vertical and horizontal station position by a factor up to 5. We found that these perturbations do not modify the values of $\dot{c}_1$ and $\dot{c}_2$ by $>10\%$ of their uncertainty.

5.3. Geometry of subnetworks

In a region where the strain directions may vary significantly over short distances, as observed on Fig. 4, it is important to test the influence of the subnetwork geometry on the strain rate field. A correct interpretation of the results would require the strain rate to be homogeneous within each individual subnetwork, but this condition of spatial homogeneity is unfortunately not known a priori. Therefore, we try to reach the best compromise between the spatial resolution and the accuracy of the strain rate estimates. We show in Fig. 5A the shear strain rate field with the highest spatial resolution (smallest possible subnetworks). A comparison with the larger subnetworks of Fig. 5B and C shows that the shear strain rates and directions remain consistent at different scales in the western part of the network, where the deformation seems therefore to be homogeneous. The comparison between Figs. 5A and 4B, that use the
same sites but a different subnetwork geometry in the western part of the network, also shows that the strain rates and directions are largely independent of the subnetwork geometry in that area. The strain field over the CHEI-TOUR-NICE-CAST-MOUL-CLAN area (western part of the network) is therefore homogeneous, with N–S compression and a maximum shear strain rate to the order of 0.10–0.20 μrad yr\(^{-1}\). The eastern part of the network does not show that homogeneity in the shear strain rates and directions. At the highest spatial resolution (Fig. 5A), it shows triangles with NW–SE compression and triangles with NE–SW compression. As shown in Fig. 5B, the CAST–ALTO–FRON–SANR subnetwork shows consistent NE–SW compression at a rate of 0.15 μrad yr\(^{-1}\), whereas the ORME–TEND–MOUL–CAST–ALTO–FRON subnetworks shows consistent NW–SE compression at a rate of 0.16 μrad yr\(^{-1}\). The CLAN–MOUL–TEND triangle shows insignificant strain.

6. Interpretation and discussion

On the basis of the above analysis, we chose to represent our results as shown in Fig. 6 (numerical values are given in Table 1). We obtain maximum
Table 1

<table>
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<tr>
<th>Subnetwork</th>
<th>$\dot{\gamma}_1$</th>
<th>$\sigma_{\dot{\gamma}_1}$</th>
<th>$\dot{\gamma}_2$</th>
<th>$\sigma_{\dot{\gamma}_2}$</th>
<th>$\dot{\gamma}$</th>
<th>$\sigma_{\dot{\gamma}}$</th>
<th>$\Theta$</th>
<th>$\sigma_{\Theta}$</th>
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<td>CHEI–TOUR–NICE–CLAN</td>
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<td>0.07</td>
<td>0.09</td>
<td>0.05</td>
<td>0.20</td>
<td>0.07</td>
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<td>0.07</td>
<td>0.09</td>
<td>74.2</td>
<td>41.7</td>
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<td>0.12</td>
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<tr>
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<td>0.45</td>
<td>0.26</td>
<td>22.2</td>
<td>16.8</td>
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shear strain rates on the order of 0.1–0.2 $\mu$rad yr$^{-1}$ over the network. The $\gamma$ values are statistically significant at the one sigma level in all subnetworks but one, the CLAN–MOUL–TEND triangle. The analysis of the spatial distribution of the shear strain shows three domains with homogeneous behavior: N–S compression in the western half of the network, NW–SE compression along a stripe trending NE–SW across the network in its eastern half, NE–SW compression in the CAST–ALTO–FRON–SANR part of the network. The large $\gamma$ value obtained in triangle NICE–CAST–SANR (0.45±0.26 $\mu$rad yr$^{-1}$), three times larger than in the neighboring subnetworks, might be related to the very acute geometry of that triangle and may not be representative of the actual strain rate in that area. There is indeed no evidence in the geology or the seismic activity of larger strain rates in the area covered by this subnetwork.

6.1. Comparison with other local geodetic strain rates in the Alps

When reduced to two subnetworks of similar size as those used by Ferhat et al. (1998) just west of our study area, we obtain maximum compressive shear strain rates of 0.15±0.06 and 0.13±0.09 $\mu$rad yr$^{-1}$ (Fig. 5C). These values agree well with the $\gamma$ values of Ferhat et al. (1998) for the Embrunais–Ubaye region (typically 0.15±0.1 $\mu$rad yr$^{-1}$), an area of relatively high seismicity like the Alpes Maritimes region. In addition, the $\gamma$ value and direction in the CHEI-TOUR-NICE-CLAN subnetwork (Fig. 6) are very close to its exact neighbor in Ferhat et al.’s (1998) study (see their fig. 3, triangle L). We also note that our values are similar to the maximum compressive shear strain rate of 0.21±0.04 $\mu$rad yr$^{-1}$ obtained in the Helvetic Alps by Reilly and Gubler (1990). The comparison with the results of Jouanne et al. (1994) in the southern Jura and Martinod et al. (1996) in the Subalpine and Belledone chains around Grenoble is not directly possible because these authors chose to parameterize the strain field using conventional strain rates and velocities, but Ferhat et al. (1998) transformed their velocity field into shear strain rates. On the basis of Jouanne et al.’s results, he obtains $\gamma$ values ranging between 0.3 and 1.25 $\mu$rad yr$^{-1}$, but statistically significant (at the one sigma level) only in very few subnetworks over 22 (0.45±0.40 and 0.79±0.42 $\mu$rad yr$^{-1}$). On the basis of Martinod et al.’s results, he obtains $\gamma$ values ranging between 0.3 and 0.5 $\mu$rad yr$^{-1}$, but statistically significant (at the one sigma level) in one subnetwork over 25 (0.48±0.21 $\mu$rad yr$^{-1}$). In both cases shear strain rates are three to five times larger than the values found in our study area. However, the fact that the $\gamma$ values obtained by Jouanne et al. and Martinod et al. are significant (at the one sigma level) only in very few subnetworks renders these results difficult to interpret in terms of deformation. It could mean that the triangulation measurements that they used do not have the necessary accuracy and/or that their strain estimation strategy was not appropriate to detect the current level of deformation in these regions.
6.2. Coseismic versus interseismic deformation

Assuming 0.15 μrad yr⁻¹ of N–S shortening over the whole network, coseismic deformation as the only contribution to the geodetic strain rates measured here would require about 35 cm of displacement within the network in the 50 years between the surveys. However, no earthquake larger than  Mₛₐₜₚₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜ这儿的齿轮能旋转吗？
Argentera massif from a N–S direction in the western part of the network to a NW–SE direction in the eastern part and remains perpendicular to the proposed blind crustal thrust ramp beneath the Argentera massif. We also note in Fig. 4B a N–S gradient of the maximum compressive shear strain rate from higher values along the Ligurian coastline to lower ones toward the Argentera in the western part of the network. These observations, together with geological evidences for N–S Quaternary compression in the Castellane arc and along the Ligurian margin (cf. supra), could indicate that the geodetic compressional shear strain is rather related to active N–S shortening along E–W-trending folds and/or thrusts, at least in the western part of the network.

6.4. Relation to larger scale geodetic deformation

Fig. 7 shows the result of continuous GPS measurements of the 160 km long baseline between Grasse and Torino, that crosses our study area [details on the processing strategy can be found in Calais (1999)]. From October 1996 to December 1999 it shows a rate of baseline length change of $-2.0 \pm 0.5 \text{ mm yr}^{-1}$ (one sigma). The shear strain rates found in the southern part of the Western Alps imply an overall N–S to NW–SE shortening of $2–4 \text{ mm yr}^{-1}$ over the study area. Using a N350E average direction for the maximum compressive shear strain (from Table 1, except subnetwork NICE–CAST–SANR), this corresponds to $1.5–3.1 \text{ mm yr}^{-1}$ of shortening along the direction of the Grasse–Torino baseline (N30E azimuth). Since our network samples only the less seismically active part of the Grasse–Torino baseline, one could expect the shortening determined in the Alpes Maritimes to be significantly lower than the far-field shortening rate. The fact that shortening in the study area along the Grasse–Torino direction is comparable to the Grasse–Torino rate of baseline length change could indicate that most of the strain along the Grasse–Torino baseline is localized within our network. The lower level of instrumental seismicity in the external part of the southern French Alps (Castellane and Nice arcs), where our network is located, compared to the inner zones (northeastern flank of the Mercantour Massif), could either mean that this strain is released aseismically or during infrequent, but moderate or large, earthquakes. The existence of several earthquake of large to moderate magnitude along the Ligurian coast (magnitude 6.4 in 1887; magnitude 6.0 in 1963; magnitude 4.5, 1989; magnitude 4.7, 1995), suggests that the deformation is not released aseismically (at least not entirely) and tends to support the latter hypothesis.

However, such a comparison between locally determined strain rates and regional-scale displacements, which integrate strain over a large region, has several limitations. An initial limitation is the fact that deformation in the Western Alps is dis-
tributed on several active faults of different orientations and kinematics, as shown by the instrumental seismicity (Béthoux et al., 1998), with simultaneous active thrusting in the external zones and widespread radial extension in the internal zones (Sue et al., 1999). In addition, block rotations have been evidenced in the internal parts of the Western Alps and could also play an important role in the internal deformation of the Alpine range (Thomas et al., 1999).

Acknowledgements

The authors would like to thank Alain Harmel from IGN/SGN for his explanations on the old triangulation data and for giving them access to the triangulation site descriptions. They are particularly grateful to Gilbert Ferhat for providing them with the triangulation data set and for his help with the fonda software. They thank Georges Buffet, Chi Chuan Tang, Olivier Laurent, Olivia Lesne, Anthony Lomax, Philippe Motte, Vincent Dupuis, and Rafaela Montelli, from Géosciences Azur, as well as Yves Mehr from the Conseil Général des Alpes Maritimes for their efficient, and sometimes sportive, help during the GPS measurements. This work greatly benefited from discussions with Nicole Béthoux and Jean-François Ritz. Nicole Béthoux provided the earthquake focal mechanisms and locations. They thank INSU for providing the GPS equipment. The authors also thank A. Geiger, P.A. Ziegler and J.-P. Burg for their constructive reviews, that significantly contributed to improve this paper. This work was funded by the French national program ‘Géofrance 3D-Alpes’ (BRGM, MENR, INSU). Contribution Géofrance 3D No. 69 and Géosciences Azur No. 305.

7. Conclusions

Using a rigorous combination of GPS and triangulation data collected 50 years apart in the southern part of the Western Alps, we detect shear strain rates to the order of 0.1–0.2 μrad yr⁻¹, significantly different from zero in most of the network. We obtain NW–SE to N–S maximum compressional strain directions over most of the area, in agreement with the seismological and geological data, that both indicate a Quaternary N–S to NW–SE compressive strain and stress regime in the southern part of the Western Alps. The observed strain rate values are about twice as large as those obtained in Provence by Ferhat et al. (1998). This is consistent with a very low level of seismicity and few geological evidence for active deformation in Provence as compared to the Alpes Maritimes. Interpreted as interseismic elastic strain accumulation on locked faults, the shear strain rates measured here imply long-term fault slip rates on the order of 2–4 mm yr⁻¹ and an overall N–S to NW–SE shortening of 2–4 mm yr⁻¹ over the study area. However, an interpretation of the strain rates measured here in term of interseismic deformation is difficult due to a poor morphological expression of active faults in the study area and the lack of information on their current activity and geometry at depth. A better identification of potentially active faults and paleoearthquakes (recurrence time of large events) is therefore important in order to integrate geodetic, geological and seismological information into a deformation model and gain a better understanding of seismic hazard in the Alpes Maritimes.

References


