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ΓΕΩΔΑΙΣΙΑ.— **Tectonic strain in Greece from geodetic measurements,** by *G. Veis, H. Billiris, B. Nakos* and *D. Paradissis**, διὰ τοῦ Ἀκαδημαϊκοῦ κ. Περικλέους Θεοχάρη.

ABSTRACT

A geodetic determination of tectonic strain in Central Greece is presented, using data from two triangulation nets measured in 1895 and 1975. The two nets had 80 common points and after being expressed on a common free reference system, the displacement vectors were computed at those points. A unique strain tensor was calculated for the whole area, which showed a general extension of 0.031 μstr/y in an azimuth of 33^0 and a shear strain $\gamma=0.048$ μstr/y. Assuming a deformation without discontinuities over the same area a displacement field was numerically derived by interpolation showing clearly the distortion pattern. Finally the area was divided into 9 blocks assuming discontinuities on their boundaries and the complete strain tensor including block displacement and rotation was evaluated for each block. The results appear to be in general agreement with what is known for the tectonics of this region and with results obtained by satellite techniques using only 15 points.

Introduction

During the last fifteen years it has been widely recognized, that although the theory of plate tectonics can be used successfully to describe a variety of large scale geological phenomena, it can not be applied to parts of the continents. Deformation is distributed over wide areas on the continents rather than being concentrated in narrow boundaries as in the oceans.

* Γ. ΒΕΗ, Χ. ΜΠΙΛΛΗΡΗ, Β. ΝΑΚΟΥ, Δ. ΠΑΡΑΔΕΙΣΗ, **Γεωδαιτικὸς προσδιορισμὸς τεκτονικῶν παραμορφώσεων στὸν Ἑλληνικὸ Χῶρο.**

The determination of such deformations is essential for the study of the earth's crust especially in relation to earthquake activity, because of the strong correlation between the two.

Most of the present knowledge on crustal deformation comes from geological and paleomagnetic data, spread over thousands and millions of years, or from seismic activity analyses, which cover the last century. However, the determinations of crustal deformations are needed independently from seismicity, in order to relay them to earthquake activity.

An independent estimate for the rates of deformation can come from geodesy. If the positions of benchmarks have been determined precisely at suitable time intervals using geodetic networks, changes in their relative positions can be derived and hence crustal deformations can be computed.

In order to determine reliable displacements, directly from repeated geodetic measurements, if the time interval is short, we need to have measurements of high accuracy. If we have measurements of medium accuracy, we need an extended period between the two measurements.

Considering that the strain rates are of up to $0.1 \cdot 10^{-6}$ per year and that classical ground geodetic methods provide accuracies of about $2\text{-}10 \cdot 10^{-6}$ we need a time interval of 20 - 100 years to detect the expected strain rates. However, new satellite geodesy techniques, can reach accuracies of better than 10^{-6} and up to 10^{-7}, which means that an interval of a few years should be enough.

These «geodetic» methods although appear to be very simple, should be applied with extreme care, in order to secure that results are expressed always to the same geodetic reference system. It should of course be realized that geodetic methods provide displacements, representing the upper surface of the crust.

Greece is a region with intensive tectonic activity with displacements of some centimeters per year and earthquakes producing local displacements of decimeters and even meters. There is also a rather good geodetic data record, going a century back, based on field measurements made by the Greek Army Geographic Service. Some of the old data must be used with extra caution and only after a careful investigation, but recent geodetic data are considered to be one of the highest quality.

It appears thus that Greece offers an excellent natural laboratory to experiment with the geodetic determination of tectonic deformations. This paper deals with such a determination by using geodetic data from two different epochs, one in the 1890's and one in the 1970's. A similar analysis using only 15 points of the 1890 triangulation and comparing their positions with those obtained by satellite methods, has already been presented elsewhere (*Billiris et al, 1991*).

The tectonic setting

Greece, unlike some other well known seismic regions (e.g. California) does not lie on a boundary between tectonic plates but in a region of intraplate deformation (*Jackson and McKenzie, 1988 / England and Jackson, 1989*) within what is usually considered the Eurasian plate but close to the border with the African plate.

It is one of the most tectonically active regions in the world with hundrends of active faults (many of which lie under water) and an average of eight earthquakes of magnitude five or greater every year (*Ambraseys and Jackson, 1990*).

The whole area appears to be stretching in a roughly north-south direction at approximately 5 cm/year, compared to the slower convergence (about 1 cm/year) between the African and the Eurasian plate that bound the region (*Jackson and McKenzie, 1988*) and contrary to the primarily strike-slip motion of the North Anatolian fault that is extending into the Northern Aegean (*Dewey and Sengor, 1979*).

All this extensive tectonic activity has raised up questions such as: if the extension is linear of radial, (*Le Pichon and Angelier, 1979*), what is the actual delineation of the different crustal blocks that appear to be separated by long normal faults, how are they moving with respect to each other, is the deformation primarily seismic or aseismic etc.; and the mechanics behind it is not understood as yet.

For the above reasons it is natural that great interest has been expressed by geoscientists to work in this area and study its tectonics.

Long research work has already been done covering Greece or parts of it and the Mediterranean in general (e.g. *Papazachos and Comninakis, 1971 / McKenzie, 1972/Makris, 1978/Lyon-Caen et al, 1988/Ekstrom and England, 1989*) and many models (in some cases even conflicting) for deformation and strain distribution have been proposed, all of them based on geological, paleomagnetic, tectonic and seismological data, rather than direct measurements.

Now that new satellite geodesy methods promise an independent and accurate technique for the direct determination of crustal deformations there is a strong interest in applying these new methods. A few national and international research programs are already in progress, (e.g. *Veis et al, 1989 / Kahle et al, 1990*) but there is still a lot of work that needs to be done.

The geodetic setting

Geodetic measurements started in Greece before 1890 with the establishment of a first order triangulation network. After a decade, 93 triangulation points had been observed covering the central and southwest part of Greece. The work of extending this net continued and by the 1920's the net covered the whole country. Several adjustments and computations of the network had been performed but these computations, were of a lower standard (although the observations were of high quality) because of lack of computing facilities. In 1950 the U.S. Army Map Service using electronic computers for the first time performed a large triangulation adjustment and computation for the whole of Europe, using all available geodetic data. The result known as «European Datum 1950» (or ED 50) had been used in Greece primarily by the military, but also served some limited civilian applications.

A new series of geodetic measurements of high standards started around 1970 by the Army Geographic Service and completed around 1980. Under the auspicies of the National Committee for Geodesy and Geophysics a work started for the definition and establishment of a new optimum Geodetic Reference System using the above measurements. The work was the cooperative effort between the Higher Geodesy Laboratory of the National Technical University of Athens, the Army Geographic Service and the Hellenic Mapping and Cadastral Organization, under the direction of one of the authors. The result known as «Greek Geodetic Reference System 1987» (GGRS'87 or ΕΓΣΑ'87), (*Veis and Paradissis, 1990*), is a set of coordinates for about 25000 geodetic points, covering the whole of the country, given to an accuracy of a few centimeters $+10^{-6}$, while the scale and orientation are to within a few parts 10^{-7} to that of the internationally accepted reference system ITRF 89 (*Boucher and Altamini, 1989*).

Displacement fields. deformations, strain tensors and their rates

We are considering a two dimensional deformation of the earth's crust in time, assuming that the crust is a thin deformable calotte on a spherical earth. The analysis could be simplified if we use a conformal mapping of the sphere on a plane, and take into consideration the distortions arrising from this mapping. However, for a small area (radius of say less than 5°) the mapping distortions (except for the orientation parameters) will not exceed 10^{-3} in scale and thus can be ignored without any practical loss of rigour. Since the area covered in the present work is of only a few degrees, the deformations will be treated as if they were on a plane.

If (x_1, y_1) are the plane coordinates of a point on the crust, converted from its geodetic coordinates (φ_1, λ_1) expressed on a well defined reference system at time t_1, and (x_2, y_2) the coordinates of the same point on the same reference system at time t_2, $\delta x = x_2 - x_1$ and $\delta y = y_2 - y_1$ represent the displacement vector for the same point, expressed in the same system and $\delta x/\delta t$ and $\delta y/\delta t$ are the displacement rates.

By plotting on a map the displacement vectors for all the points that their position could be determined at epochs t_1 and t_2 one could have a first indication of the deformation, giving a picture of the displacement field.

The displacement field can be better visualised by plotting on a regular grid the displacement vector, through interpolation from the directly observed values. One could further, apply the displacement vectors to the grid itself, providing thus the image of the deformed grid. These presentations require a proper interpolation algorithm.

Geodetic methods, both ground and/or satellite, can provide coordinates to a well defined reference system for a number of points properly selected and monumented on the earth's surface. Coordinates are derived from measured distances, angles (both horizontal and vertical), directions, or their differences, between the monumented points and to satellites if satellite geodesy methods are used.

In determining positions and displacements on a plane at least one fixed point and a fixed direction is needed. If these do not exist one could take arbitrarily any point and any direction as fixed, without loss of generality, realizing that all coordinates and all displacements are relative. This is the general rule in similar studies, although sometimes more points are considered fixed, based on external information for their stability (e.g. *Billiris et al, 1991*).

However, instead of arbitrarily selecting a fixed point and direction, it is more appropriate to adjust the set of coordinates at time t_2 to the set of coordinates at time t_1 by solving for a shift and a rotation that will minimize the discrepancies between the two, i.e. minimize the displacements. Mathematically the two methods are identical and one could go from the one to the other. In the present work, the second method is used as we prefer not to make any a priori assumptions on stability.

Crustal deformations have discontinuities both in space (e.g. faults) and in time (e.g. earthquakes). However, in most cases depending on the space and time scale we treat them at first as if we had a continuous field and treat the discontinuities separately. To investigate discontinuities in detail we need a very dense in space and time network of control points.

It follows from the above, that the displacements within a region will be known if the coordinates of all points within the region at epoch t_2 (x_2, y_2) could be given as a function of the coordinates of the same points at epoch t_1 (x_1, y_1), or if we know the relations:

$$\begin{aligned} x_2 &= f(x_1, y_1) \\ y_2 &= g(x_1, y_1) \end{aligned} \tag{1}$$

the simplest relation being a linear one, or:

$$\begin{aligned} x_2 &= a_0 + a_1 x_1 + a_2 y_1 \\ y_2 &= b_0 + b_1 x_1 + b_2 y_1 \end{aligned} \tag{2}$$

The displacements then will be given by:

$$\begin{aligned} \delta x &= f(x_1, y_1) - x_1 \\ \delta y &= g(x_1, y_1) - y_1 \end{aligned} \tag{3}$$

Although in reality f (x, y) and g (x, y) could be any function, since we assume no discontinuities we can expand them in a Taylor series keeping only the linear terms and so use always the simple relation (2). Linearity could hold to the required degree of accuracy for an appropriate small region.

Using the geodetically derived displacements vectors, we can study the deformation of the crust, following the approach used for studying the distortions in the theory of map projections (e.g. *Tissot, 1880* / *Driencourt et Laborde, 1932*), or the theory of elasticity (e.g. *Love, 1944*) and continuum mechanics (e.g. *Theocharis, 1981*). Since the topic is closely related to geodesy we will follow more or less the cartographic approach. Following also the geodetic practice, angles will be measured clockwise from North (or the Υ-axis) corresponding to azimuths.

Assuming that the relations between the two sets of coordinates at t_1 and t_2 are given by (2) we could trace the axes parallel to X, and Y of a point (x, y) at epoch t_1, as they will be at epoch t_2. For each axis we have a shift d, a rotation ε, and an extension, expressed by a new scale along the axis K=1+e. It is easy to derive:

$$d_x = a_0$$

$$d_y = b_0$$

$$\varepsilon_x = \arctan \frac{-b_1}{a_1}$$

$$\varepsilon_y = \arctan \frac{a_2}{b_2}$$

$$K_x = \sqrt{a_1^2 + b_1^2} = 1 + e_x \text{ or } e_x = \sqrt{a_1^2 + b_1^2} - 1$$

$$K_y = \sqrt{a_2^2 + b_2^2} = 1 + e_y \text{ or } e_y = \sqrt{a_2^2 + b_2^2} - 1$$

In crustal deformations we have annual displacements of the order of up to a few cm and relative changes in lengths of up to a few parts of 10^{-7}. Accordingly for time intervals of up to some centuries d will be up to a few meters, ε and e up to 10^{-4}. So ignoring square terms of the small quantities we can write the expression (2) as:

$$\begin{aligned} \delta x &= x_2 - x_1 = d_x + e_x x + \varepsilon_y y \\ \delta y &= y_2 - y_1 = d_y - \varepsilon_x x + e_y y \end{aligned} \tag{4}$$

or

$$\begin{pmatrix} \delta x \\ \delta y \end{pmatrix} = \begin{pmatrix} d_x \\ d_y \end{pmatrix} + \begin{pmatrix} e_x & \varepsilon_y \\ -\varepsilon_x & e_y \end{pmatrix} \begin{pmatrix} x \\ y \end{pmatrix} \tag{5}$$

or

$$\delta_i = d_i + E\, x_i \tag{6}$$

where $d_x=a_0$, $d_y=b_0$, and $e_x=a_1-1$, $e_y=b_2-1$, $\varepsilon_x=-b_1$, $\varepsilon_y=a_2$ are all small quantities.

ε_x and ε_y represent the (small) orientation deformation (i.e. rotation) of the original X and Y axes, while e_x and e_y the (small) linear deformation, expressed as per cent extension of the original length along the X and Y axes. The percent extension $e = \Delta l / l$ is called strain, and the rotational orientation deformation is called orientation strain.

The (asymmetric) two dimensional matrix E is called strain tensor and, as will be demonstrated, characterizes the local surface deformation at point (x, y). The elements e_x, e_y, ε_x, and ε_y are called strain parameters.

The asymmetric tensor E could be analysed as the sum of a symmetric and an asymmetric one as $E = E_0 + \Omega$ where Ω represents a solid body rotation.

If we set

$$\omega = \frac{\varepsilon_y + \varepsilon_x}{2} = \bar{\varepsilon} \quad \text{and} \tag{7}$$

$$\varepsilon_0 = \frac{\varepsilon_y - \varepsilon_x}{2},$$

we have

$$\varepsilon_y = \omega + \varepsilon_0, \qquad \varepsilon_x = \omega - \varepsilon_0$$

and then

$$E = \begin{pmatrix} e_x & \varepsilon_y \\ -\varepsilon_x & e_y \end{pmatrix} = \begin{pmatrix} e_x & \varepsilon_0 \\ \varepsilon_0 & e_y \end{pmatrix} + \begin{pmatrix} 0 & \omega \\ -\omega & 0 \end{pmatrix} \tag{8}$$

Here $\omega = \bar{\varepsilon}$ is the mean rotation of the X and Y axes (total solid body rotation) and ε_o is the equal and opposite rotation of the two axes in addition to the mean rotation. The right angle of the X, Y axes will be reduced by $2\varepsilon_o$.

Now eq. (5) is written as:

$$\begin{pmatrix} \delta x \\ \delta y \end{pmatrix} = \begin{pmatrix} d_x \\ d_y \end{pmatrix} + \begin{pmatrix} e_x & \varepsilon_o \\ \varepsilon_o & e_y \end{pmatrix} \begin{pmatrix} x \\ y \end{pmatrix} + \begin{pmatrix} 0 & \omega \\ -\omega & 0 \end{pmatrix} \begin{pmatrix} x \\ y \end{pmatrix} \tag{9}$$

or

$$\delta_i = d_i + E_o x_i + \Omega x_i \tag{10}$$

The above expressions can be used only if the deformed area can be connected (by measurements) to a fixed in position and orientation reference system. If not, which is often the case, it is obvious that vector d_i and matrix Ω are undetermined and the relations (9) and (10) have to be written:

$$\begin{pmatrix} \delta x \\ \delta y \end{pmatrix} = \begin{pmatrix} e_x & \varepsilon_o \\ \varepsilon_o & e_y \end{pmatrix} \begin{pmatrix} x \\ y \end{pmatrix} \tag{11}$$

or

$$\delta_i = E_o x_i \tag{12}$$

and thus only the three strain parameters e_x, e_y, ε_o could be determined.

Having the strain tensor E (or E_o) one could derive the local deformations at point (x, y). If we assume a unit vector at (x, y) in the direction of an azimuth α (measured clockwise from the Y-axis), the vector will have original components at t_1 ($\sin\alpha$, $\cos\alpha$) and at t_2, after the deformation (expressed by E) has been applied, will become:

$$E \begin{pmatrix} x + \sin\alpha \\ y + \cos\alpha \end{pmatrix}$$

and the distortion of the unit vector:

$$E \begin{pmatrix} \sin\alpha \\ \cos\alpha \end{pmatrix}$$

We can analyze this distortion in two components. One, ε, in the direction across the original unit vector, which expresses the (clockwise) rotation or orientation strain of the unit vector and one, e, along the vector which expresses the (linear) strain in that direction.

We have

$$\begin{pmatrix} \varepsilon \\ e \end{pmatrix} = \begin{pmatrix} \cos\alpha & -\sin\alpha \\ \sin\alpha & \cos\alpha \end{pmatrix} E \begin{pmatrix} \sin\alpha \\ \cos\alpha \end{pmatrix} \tag{13}$$

or

$$e = e_x \sin^2\alpha + e_y \cos^2\alpha + (\varepsilon_y - \varepsilon_x) \sin\alpha \cos\alpha \tag{14}$$

and

$$\varepsilon = \varepsilon_x \sin^2\alpha + \varepsilon_y \cos^2\alpha + (e_y - e_x) \sin\alpha \cos\alpha \tag{15}$$

Expression (14) gives the (linear) strain at a point (x, y), for which the strain tensor E is given, as a function of the azimuth.

By setting $\frac{de}{d\alpha} = 0$ we find the max and min values of e as well as the orientation (azimuths) of the max (α_1) and min (α_2) values.

We obtain :

$$\begin{aligned} e_{max} = e_1 = \tfrac{1}{2}\left[(e_x + e_y) + \sqrt{(e_x - e_y)^2 + (\varepsilon_y - \varepsilon_x)^2} \right] \\ e_{min} = e_2 = \tfrac{1}{2}\left[(e_x + e_y) - \sqrt{(e_x - e_y)^2 + (\varepsilon_y - \varepsilon_x)^2} \right] \end{aligned} \tag{16}$$

and

$$\begin{aligned} \tan 2\alpha_1 = \frac{\varepsilon_y - \varepsilon_x}{e_y - e_x} \quad \text{for the maximun} \\ \tan 2\alpha_2 = \frac{-(\varepsilon_y - \varepsilon_x)}{-(e_y - e_x)} \quad \text{for the minmum} \end{aligned} \tag{17}$$

which proves that the max and min values of e correspond to perpendicular directions. They are called principal directions.

If we use $\bar{e} = [(e_x + e_y)]/2$ for the mean value of the strain along the X and Y axes, and call $\gamma_e = e_y - e_x$, $\gamma_\varepsilon = \varepsilon_y - \varepsilon_x$, and $\gamma = \sqrt{\gamma_e^2 + \gamma_\varepsilon^2}$ we obtain the simplified expressions:

$$
\begin{aligned}
e_1 &= \bar{e} + \tfrac{1}{2}\gamma \\
e_2 &= \bar{e} - \tfrac{1}{2}\gamma \\
\alpha_1 &= \tfrac{1}{2}\arctan\frac{\gamma_\varepsilon}{\gamma_e} \\
\alpha_2 &= \tfrac{1}{2}\arctan\frac{-\gamma_\varepsilon}{-\gamma_e}
\end{aligned}
\tag{18}
$$

From where it follows that:

$$
\begin{aligned}
\bar{\varepsilon} &= \frac{e_y + e_x}{2} \\
\gamma &= e_1 - e_2 = e_{max} - e_{min}
\end{aligned}
\tag{19}
$$

If we add 1 to the expression (14) we get the equation of an ellipse with semi-major axis $a = 1 + e_1$, and semi-minor axis $b = 1 + e_2$, oriented along the principal directions with azimuths α_1 and α_2. This proves that any small unit circle will be distorted into an ellipse. Such an ellipse can be used as a strain indicatrix, since it reflects the strain tensor E, the same way as the Tissot indicatrix is used in map projections (*Tissot, 1880*). This ellipse is called strain ellipse.

The strain parameters e and ε of (14) and (15) can also be expressed as a function of the parameters defining the strain ellipse. Instead of the azimuth α we will use the angle β measured clockwise from the minor axis (fig. 1), or:

$$\beta = \alpha - \alpha_2$$

Then the expressions giving e and ε will be:

$$
\begin{aligned}
e &= e_1 \sin^2\beta + e_2 \cos^2\beta \\
&= \bar{e} - \tfrac{1}{2}\gamma \cos 2\beta
\end{aligned}
\tag{20}
$$

and

$$
\begin{aligned}
\varepsilon &= \tfrac{1}{2}(e_1 - e_2)\sin 2\beta \\
&= \tfrac{1}{2}\gamma \sin 2\beta
\end{aligned}
\tag{21}
$$

If we take the perpendicular to β direction, we find that its orientation strain ε_{90} will be

$$\varepsilon_{90} = \frac{1}{2}\gamma \sin 2(90 + \beta) = -\varepsilon$$

This means that the right angle will be distorted by $\psi = 2\varepsilon$, which represents the shear on lines parallel to the direction of β. The angle ψ is called shear angle and is given by

$$\psi = \gamma \sin 2\beta \tag{22}$$

The $\tan\psi$ is called shear strain but since ψ is a small angle, shear angle and shear strain can be used indiscriminately.

From eq. 22 we find that the maximum value of angular shear will be $\psi_{max} = \gamma$ and will correspond to the right angles defined by the bisectors of the principal axes.

Geodetic determination of the strain parameters

Geodetic methods can provide by direct measurements both the linear e and orientation ε, strain parameters.

If the horizontal distance between two monumented markers A and B in the direction of azimuth α is measured and found to be l_1 at time t_1 and l_2 and time t_2, we have:

$$e = \frac{l_2 - l_1}{l_1}$$

the strain in azimuth α.

Every such measurement provides an observation equation of the form:

$$e = e_x \sin^2\alpha + e_y \cos^2\alpha + (\varepsilon_y - \varepsilon_x) \sin\alpha \cos\alpha \tag{22}$$

Here we can see that ε_y and ε_x cannot be separated in the solution and the unknown parameters to be determined will be e_x, e_y and $(\varepsilon_y - \varepsilon_x) = \gamma_\varepsilon = 2\varepsilon_o$.

Any three distance measurements will suffice to solve for the three unknowns, assuming that the measured baselines are not parallel. Of course no parallel shift d_x, d_y and no total rotation ω can be derived from such measurements since there is no connection to any outside fixed points.

Similarily if a horizontal angle from a point P is measured between points A and B and found to be θ_1 at time t_1 and θ_2 at time t_2, we have the angular strain $\varepsilon_\theta = \theta_2 - \theta_1$ which corresponds between the directions α_A and α_B, the azimuths of points A and B.

The value of ε_θ will be found as:

$$\varepsilon_\theta = \varepsilon_B - \varepsilon_A$$

The value of ε is given by (15), but as in the case of no outside control, the total rotation ω cannot be determined which means that $\varepsilon_x = -\varepsilon_y = \varepsilon_0 = = \frac{1}{2}\gamma_\varepsilon$ and equation (15) will be written:

$$\varepsilon = \tfrac{1}{2}\gamma_\varepsilon \cos 2\alpha - \tfrac{1}{2}\gamma_e \sin 2\alpha \tag{23}$$

Finally we have:

$$\varepsilon_\theta = \tfrac{1}{2}\gamma_\varepsilon \left(\cos 2\alpha_B - \cos 2\alpha_A\right) - \tfrac{1}{2}\gamma_e \left(\sin 2\alpha_B - \sin 2\alpha_A\right) \tag{24}$$

For every measured angular change $\varepsilon_\theta = \theta_2 - \theta_1$ we have an observation equation of the form of (24). Any two such measurements will provide the two unknowns γ_ε and γ_e (and of course $\gamma = \sqrt{\gamma_\varepsilon^2 + \gamma_e^2}$). It is apparent that from angular measurements alone one could not obtain a solution for linear strain parameters, but only shear strain, since no distances have been measured.

More observations of distances and/or angles could be used as odservation equations and solve for the best values of the strain parameters by least squares. However, in case of mixing distances and angles, the observation equation for angles (24) has to be written as:

$$\begin{aligned} \varepsilon_\theta = \tfrac{1}{2} e_x \left(\sin 2\alpha_B - \sin 2\alpha_A\right) - \tfrac{1}{2} e_y \left(\sin 2\alpha_B - \sin 2\alpha_A\right) \\ + \tfrac{1}{2}\left(\varepsilon_y - \varepsilon_x\right) \cos 2\alpha_B - \cos 2\alpha_A) \end{aligned} \tag{25}$$

to be in the same form as the equation (22) for distances.

With this method, from the differences of measured distances and angles at two different epochs, we can estimate the strain parameters e_x, e_y and ε_o directly but not any solid body (total) displacement and rotation.

There is another method that can be used to estimate the strain parameters from geodetic data. We first compute from the geodetic measurements the coordinates, expressed in the same reference system, of the control points at the two epochs. We then find the displacement (δx, δy) for each point. These displacements are related to the strain parameters by the two equations (4). If there are three points we have a set of 6 equations (2 for each point) with 6 unknowns (d_x, d_y, e_x, e_y, ε_x, ε_y) and the unknowns can be determined. If there are more than three points we use the least squares method for the estimation of the unknowns.

This presumes that the two sets of coordinates refer to the same geodetic reference system, which requires that the geodetic networks used for the computation of the coordinates at the two epochs include a number of points fixed between the two epochs.

If there are no fixed points, we can always take arbitrarily one point as fixed (by setting the same coordinates at both epochs) and one direction (by setting the one coordinate the same for a second point) and so give arbitrary values to satisfy the three degrees of freedom (two translations and one rotation).

A better solution in case there are no fixed points in the networks (such networks are called free-nets), is to determine and then apply a translation (d_x, d_y) and a rotation (ω) between them that will minimize, in a least squares sence, all the displacements, and arrive at a zero mean displacement. Such a solution is statistically justified, especially if the area covered is very large.

Sometimes (especially in the case of old triangulations) the geodetic networks are based on high quality angular measurements but very poor distances. Then the scale of the network will be rather poorly defined and will be necessary to include, in addition to the translation (d_x, d_y) and rotation (ω), an additional unknown, a scale factor $K=1+\varkappa$, to relate the two sets of coordinates.

Let (x_1', y_1') be the given set of coordinates at epoch t_1. In order to best fit them to the coordinates (x_2, y_2) at epoch t_2 by applying a translation d_x, d_y, a rotation ω and a scale correction $\varkappa$, we have to solve by least squares the following observation equations:

$$\begin{pmatrix} x_1' - x_2 \\ y_1' - y_2 \end{pmatrix} = \begin{pmatrix} d_x \\ d_y \end{pmatrix} + \begin{pmatrix} \varkappa & \omega \\ -\omega & \varkappa \end{pmatrix} \begin{pmatrix} x_1' \\ y_1' \end{pmatrix} \tag{26}$$

If $\hat{d}_x$, $\hat{d}_y$, $\hat{\omega}$ and $\hat{\varkappa}$ are the best estimates obtained by least squares, then the new «adjusted» set of coordinates for epoch 1 will be:

$$\begin{pmatrix} x_1 \\ y_1 \end{pmatrix} = \begin{pmatrix} x_1' \\ y_1' \end{pmatrix} - \begin{pmatrix} \hat{d}_x \\ \hat{d}_y \end{pmatrix} + \begin{pmatrix} \hat{\varkappa} & \hat{\omega} \\ -\hat{\omega} & \hat{\varkappa} \end{pmatrix} \begin{pmatrix} x_1' \\ y_1' \end{pmatrix} \tag{27}$$

Although we can not determine any translation and rotation for the total area under consideration, it is possible to determine the complete strain tensor including translation and rotation, for smaller regions within the given area. This is possible since a common fixed reference system has been indirectly defined by minimizing the displacements over the total area. The strain parameters to be derived will not be absolute, but the relative strain between the regions will be correct.

The geodetic data used and the determination of the displacement vectors

The starting geodetic material used in this study were the horizontal angles measured during the first Greek Triangulation performed in the period 1890-1900 and referring to a mean epoch of 1895, (*Hartl, 1901*). These angles were used, after a new adjustment of the triangulation net, to calculate the coordinates of the 93 triangulation points that were included in the 1895 triangulation.

Since the baseline and the astronomic orientation of that triangulation were of poor quality, in contrast to the high quality of the angles, measured with an average standard deviation of about $0''.3$ or 1.5 ppm, the net was scaled and oriented to a first approximation by fixing two points to the coordinates of the new Greek Geodetic Reference System (GGRS' 87) the one used for the second epoch. This will not affect the final analysis since, as the net was a free net, it would be reoriented and rescaled, at a second stage. Using the GGRS' 87 for the computations also minimizes the errors arrising from deflections of the vertical and other observation biases. For the adjustment of the net a software developed by P. Cross (*Cross, 1984*) was used.

For the second epoch the coordinates of GGRS' 87 were used directly. These coordinates were derived by adjusting in two steps all geodetic measurements performed between the years 1968-1982 with a mean epoch of 1975. The coordinates refer to the GRS' 80 reference ellipsoid optimally fitted to the greek geoid. In addition to the ground measurements, satellite data were also introduced to provide the overall scale and orientation (*Veis, 1987*).

Out of the 93 triangulation points of the 1895 triangulation, 80 were found to exist today and be included in the GGRS' 87 network. Triangulation pillars are often distroyed and lost. Most of the times they are reestablished on the same point using witness markers. However, one has to be carefull since the new marker may be displaced by several centimeters from its original position. The 80 common points found, were checked that they corresponded to the same original position.

While the 1975 coordinates refer to a well defined reference system, that of GGRS' 87, the 1895 coordinates are defined as a free-net, without position, orientation and scale. In order to combine the results from the two epochs t_1=1895 and t_2=1975 and compute the strain and strain rate, we used the method described earlier in this paper and produced a new set of coordinates for epoch t_1 from the original ones using equations (26) and (27). A scale correction of $\varkappa=-0.39$ ppm and a rotation of $\omega=+0''.08$ had to be applied in addition to a shift of 7 cm to the south and 54 cm to the west in order to elliminate the net displacement between the two epochs.

The coordinates (x_2, y_2) for epoch t_2 and the corrected for scale, rotation and shift coordinates (x_1, y_1) for epoch t_1, were used to derive the displacement vectors $(\delta x, \delta y)$ that were addopted for this study. These displacement vectors correspond to a time interval of $t_2-t_1=80$ years. Assuming a constant rate, they can be converted into annual displacement rates simply by dividing by 80. By the same way we can convert into annual rates the parameters of the strain tensor and obtain the strain rate tensor.

Strains are expressed in «microstrains» corresponding to a relative extension of 10^{-6}, or 1 μstr$=10^{-6}=1$ ppm. For orientation, angular or shear strain, 1 μstr$=0''.206=206$ marcsec. Strain rates will be expressed in μstr/y.

The accuracy of the 1895 coordinates was estimated by the adjustment procedure to be on the average ±25 cm (s.d.) except for the edges where it could be larger. The 1975 coordinates are more accurate and they were estimated to be ±10 cm (s.d.) over the region covered by this study. This means that the derived displacement vectors will have a standard deviation of ±27 cm, although in some areas towards the edges it could be larger. This means that the derived displacement rates will have an uncertainty of about ±3 mm/y.

Finally it should be remembered that since the 1895 triangulation was a free net without accurate scale, it is not possible to distinguish any total shift and rotation for the whole region, and that a systematic additional strain may be present. However, relative strains will be correct and it is always possible to determine the complete strain tensor for a subregion within the area covered by the triangulations under the assumptions already made.

The derived displacements appear on a map in fig. 2. In order to have a better visualisation of the displacements, the vectors are drawn to the map scale as they will correspond to a time interval of 2 My, although the rates derived over 80 years cannot be extrapolated to such a long period. A scale for the displacement rates in cm/year is also given.

Of the 80 points common to the two triangulations 5 were not used as they showed large displacements. Although those displacements may be real, it is suspected that their pillars may have been wrongly replaced and a special investigation is needed. The triangulation points not used are Velopoula, Ydra, Antinitsa, Kratsovo and Peristeri.

Presentation of results

If we consider the total area as been tectonically uniform, we can calculate a mean tensor of deformation for the whole area. We find for this tensor $e_1 = +0.031$ μstr/year in the direction of 33° and a minimum of $e_2 = -0.017$ μstr/year in the direction of 123°. This corresponds to a shear strain of $\gamma = 0.048$ μstr/year (or $0''.01$/year). Total displacement, total rotation and total expansion can not be calculated since there are no fixed points. Fig. 3. shows this tensor expressed as the deformation of a circle to an ellipse, over a time period of 5 million years.

However, it is not realistic to assume that one strain tensor can express the tectonics of the whole area. This became apparent also from the fact that the standard deviation of the residuals of the vectors after the least squares fitting for a single tensor, exceeded by 80% the a priori expected standard deviation of the observations. Considering the area covered by this work as an area with continuous deformation, we calculated, numerically, a displacement field by interpolation from the 75 geodetic points. The result is shown graphically in fig. 4. The length of the arrows, shows the displacement of a point on an imaginary 20 km grid and gives at the map scale the displacement over a million years. Fig. 5, gives another visual description of the same field as a distortion of the grid itself.

The variation of the deformations over the area under study, can be found if we calculate the strain tensor at different points using a region around them. By plotting the isopleths for each strain parameter a graphical presentation of the variation over the whole area will be given. Depending on the number and the magnitude of the regions used, in relation to the density of the original data, it will provide a more or less smoothed presentation of the variations. In this study, different number and magnitudes of regions were used with or without an overlap. Indicatively, fig. 6, 7 and 8, give respectively, the displacement in the east direction, the shear strain and the directions of the principal axes calculated by scanning in a half degree step in latitude and longitude, with a half degree overlap.

From the original displacement vectors given in fig. 2 one can identify blocks of more or less uniform deformations while their boundaries seem to have a lack of continuation. Accordingly the whole area was divided into 9 blocks, which appeared to justify the criteria of uniform deformation. This division shown in fig. 9, was based only on the criterion of uniformity, without taking into account any other prior knowledge related to the geology and tectonics of the area. Of course the boundaries of these blocks are uncertain since the distances between the points with original displacements are up to a few tens of kilometers. For each block the complete strain tensor was estimated with the methods already described, since it is now possible to calculate also a block shift and rotation with respect to the indirectly defined fixed reference system. Table I, gives the obtained results. Fig. 10, gives graphically the complete tensor for each block expressed as a block displacement, rotation and strain ellipse, corresponding at the map scale to changes over a period of 5 My and fig. 11, gives the total deformation of a normal grid over a period of 2 My.

Conclusions

The above analysis proves that geodetic methods can be successfully used for determining in a direct way the strain field of the earth's crust. Even older geodetic data, if properly used, can provide very valuable information, expanding considerably the time interval used for the determination of the strain rate.

The results obtained in this study are in rather good agreement with what is known for the tectonics of Central Greece. One can clearly delineate from fig. 9, the main tectonic features, although details cannot be obtained without a much denser network.

The only similar work that can be used for a direct comparison is the one reported in Nature (*Billiris et al, 1991*) in which the present authors have also participated. The original first epoch 1895 triangulation data was basicaly the same in both, while the second epoch data were quite different. They come from a completely different technique and they refer to different epochs. Also in the Nature paper only 15 triangulation points were used compared to the 75 used in the present work.

It should be realized that the derived displacement vectors in the two papers are expressed with different assumptions concerning the reference system. In the Nature paper the assumption was made, that some points were stable during the 1900-1988 period, while in the present work no fixed points were assummed but instead the fixed reference system was derived as a «statistical mean». So the results are not directly comparable.

If we convert, however, from the one system to the other, we find that the displacement vectors are in more or less the same direction, while their magnitudes are a little larger in the Nature paper. This is to be expected since it covers a period of 88 years compared to the 80 years of this work. Taking this into consideration we can conclude that the two results are in a quite good agreement.

It will be desirable to combine the two results in one solution. For this, it is planned to make new measurements by GPS on additional triangulation points in the summer of 1992 and proceed with a unified solution using data from all three epochs.

Up to very recently geodesy was used primarily for the establishment of a reference frame for mapping and engineering. In that context the design of geodetic networks and the required accuracy were limited to the specific application they were asked to serve. Now geodesy is challenged to provide faster and more accurate positional information to be used in geosciences. This challenge can today be met with the scientific and technological progress in satellite geodesy which, combined with fast computers and advanced adjustment techniques, can provide quickly distances of hundreds of kilometers to centimeter accuracies.

Properly designed satellite geodesy networks could give directly a detailed picture of the crust's strain rate in a few years. The geodetically derived strain rates could then be compared with those obtained by other geophysical methods and especially from the analysis of seismicity and earthquake mechanisms, in order to estimate the remaining accumulated stresses that have not yet been released and estimate the probability of future earthquake activity. Much work remains to be done in that direction.

The Higher Geodesy Laboratory and the Dionysos Satellite Observatory of the National Technical University of Athens, in cooperation with other national and international bodies, expect to continue their work in that direction.

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ΠΕΡΙΛΗΨΗ

Γεωδαιτικὸς προσδιορισμὸς τεκτονικῶν παραμορφώσεων στὸν Ἑλληνικὸ Χῶρο

Δεδομένης τῆς ἔντονης τεκτονικῆς δράσης στὸν Ἑλληνικὸ Χῶρο, ὑπάρχει ἰδιαίτερο ἐνδιαφέρον γιὰ τὸν προσδιορισμὸ τῶν πεδίων μετατοπίσεων καὶ τῶν τανυστῶν παραμορφώσεων τοῦ φλοιοῦ τῆς γῆς στὴν περιοχὴ αὐτή.

Κατὰ κανόνα γιὰ τοὺς προσδιορισμοὺς αὐτοὺς χρησιμοποιοῦνται γεωλογικά, παλαιομαγνητικά, τεκτονικὰ καὶ σεισμολογικὰ δεδομένα ἀπὸ τὰ ὁποῖα προκύπτουν οἱ μετατοπίσεις. Ἡ ἐφαρμογὴ ὅμως γεωδαιτικῶν μεθόδων, ποὺ ἀναπτύσσονται τὰ τελευταῖα χρόνια, ἐπιτρέπει τὸν ἄμεσο προσδιορισμὸ τῶν μετατοπίσεων, μὲ ἀποτέλεσμα τὴν δραστηριοποίηση τῆς ἐπιστημονικῆς αὐτῆς περιοχῆς, ἰδιαίτερα μὲ τὴ χρήση τῆς δορυφορικῆς γεωδαισίας.

Τὸ Κέντρο Δορυφόρων Διονύσου καὶ τὸ Ἐργαστήριο Ἀνώτερης Γεωδαισίας τοῦ Ε. Μ. Πολυτεχνείου, ἔχουν ἀναπτύξει σημαντικὴ δραστηριότητα, σὲ συνεργασία καὶ μὲ ἄλλα ἑλληνικὰ καὶ ξένα Πανεπιστήμια, καὶ Ὀργανισμοὺς γιὰ τὴν καταγραφὴ καὶ μελέτη τῶν τεκτονικῶν μετακινήσεων στὸν Ἑλληνικὸ Χῶρο μὲ γεωδαιτικὲς μεθόδους. Πρῶτο ἀποτέλεσμα τέτοιας συνεργασίας ἀποτελεῖ ἡ σχετικὴ δημοσίευση στὸ Nature (*Billiris et al, 1991*).

Στὴν παρούσα ἐργασία μελετῶνται τὰ ἀποτελέσματα ποὺ προκύπτουν ἀπὸ τὴν σύγκριση τῶν γεωδαιτικῶν μετρήσεων τοῦ πρώτου Ἑλληνικοῦ Τριγωνισμοῦ τοῦ 1890-1900 (1895) μὲ ἐκεῖνα τοῦ τελευταίου τῆς περιόδου 1968-1982 (1975), ἐκτιμᾶται ἡ κινηματικὴ τοῦ φλοιοῦ στὴν περιοχὴ τῆς Κεντρικῆς Ἑλλάδος μὲ γεωδαιτικὲς μεθόδους, ὑπολογίζονται οἱ τανυστὲς παραμορφώσεων καὶ διατυπώνονται πιθανὰ μοντέλα γεωτεκτονικῆς δομῆς του.

Χρησιμοποιήθηκαν 75 κοινὰ σημεῖα τῶν τριγωνισμῶν τῶν περιόδων αὐτῶν. Ὁ τριγωνισμὸς τοῦ 1975 χρησιμοποιήθηκε ὅπως ὑπολογίστηκε στὸ νέο Ἑλληνικὸ Γεωδαιτικὸ Σύστημα Ἀναφορᾶς (ΕΓΣΑ '87). Ὁ τριγωνισμὸς τοῦ 1895 χρησιμοποιήθηκε μετὰ ἀπὸ νέο ὑπολογισμὸ στὸ ἴδιο ἀκριβῶς σύστημα ἀναφορᾶς (ΕΓΣΑ '87) χρησιμοποιώντας μόνο τὶς γωνιομετρήσεις τῆς ἐποχῆς ἐκείνης, ὡς ἐλεύθερο δίκτυο, τοῦ ὁποίου τὰ στοιχεῖα θέσεως, προσανατολισμοῦ καὶ κλίμακας προσδιορίστηκαν μὲ τὴ μέθοδο τῆς «γενικῆς» προσαρμογῆς τῶν δύο δικτύων ὥστε νὰ μὴ ὑπάρξει a priori παραδοχὴ γιὰ τὴν σταθερότητα ὁρισμένων σημείων.

Τὸ σχ. 2 δίνει τὶς μετατοπίσεις ποὺ προκύπτουν ἀπὸ τὴν σύγκριση τῶν συντεταγμένων τῶν δύο δικτύων στὰ κοινά τους σημεῖα. Ἐπειδὴ εἶναι ἀδύνατον, ὅπως ἀποδεικνύεται, νὰ ἐκφραστοῦν ἱκανοποιητικὰ οἱ παραμορφώσεις ὅλης τῆς περιοχῆς μὲ ἕνα μόνο τανυστή, ὑπολογίσθηκε ἀριθμητικὰ ἕνα συνεχὲς πεδίο παραμορφώσεων. Τὰ ἀποτελέσματα ἐμφανίζονται γραφικὰ στὰ σχήματα 4-8.

Ἀπὸ τὰ πρωτογενῆ στοιχεῖα τῶν μετατοπίσεων διαπιστώνεται ἡ ὕπαρξη περιοχῶν μὲ ὁμοιόμορφη παραμόρφωση καὶ ἔντονη μεταβολὴ στὰ σύνορά τους ποὺ ὑποδηλώνουν ἀσυνέχεια. Γιὰ τὸν λόγο αὐτὸ ἡ ὅλη ἔκταση διαιρέθηκε σὲ 9 περιοχὲς ποὺ ἱκανοποιοῦν τὰ κριτήρια αὐτά, χωρὶς νὰ ληφθοῦν ὑπόψη ἄλλα ἐξωγενῆ στοιχεῖα (σχ. 9). Γιὰ κάθε περιοχὴ ὑπολογίστηκε ὁ πλήρης τανυστὴς παραμορφώσεως μὲ τὴν

Μέθοδο Ἐλαχίστων Τετραγώνων (Πίνακας I). Τὸ σχ. 10, δίνει γραφικὰ τοὺς τανυστὲς παραμόρφωσης γιὰ κάθε περιοχὴ καὶ τὸ σχ. 11, δίνει τὴν παραμόρφωση ὅλης τῆς περιοχῆς ὅπως μορφοποιεῖται μὲ τὴν παραμόρφωση ἑνὸς κανονικοῦ κανάβου.

Ἡ ἐργασία αὐτή, τὰ ἀποτελέσματα τῆς ὁποίας εἶναι συμβιβαστὰ μὲ ἐκείνης τοῦ Nature, ἐλπίζεται νὰ ἐπεκταθεῖ σὲ περισσότερα γεωδαιτικὰ σημεῖα τῆς Χώρας μὲ τὴν χρησιμοποίηση παλαιότερων μετρήσεων ἀλλὰ κυρίως μὲ τὸν ἐπαναπροσδιορισμὸ νέων θέσεων μὲ μεθόδους τῆς Δορυφορικῆς Γεωδαισίας. Ἐπίσης ἀποσκοπεῖται ἡ σύγκριση τῶν γεωδαιτικῶν μετατοπίσεων μὲ ἐκεῖνες ποὺ προκύπτουν ἀπὸ τὴν ἀνάλυση τῆς μέχρι σήμερα σεισμικῆς δράσης, ὥστε νὰ γίνει μιὰ ἐκτίμηση τῶν τάσεων ποὺ δὲν ἔχουν ἐκτονωθεῖ ὥστε νὰ ἐκτιμηθεῖ μὲ ἄλλον ἕνα τρόπο ἡ πιθανότητα ἀναμενόμενης σεισμικῆς δράσης.

Ὁ Ἀκαδημαϊκὸς κ. **Περικλῆς Θεοχάρης** εἰς τὴν ἀνωτέρω ἐργασίαν εἶπε τὰ ἑξῆς:

Εἰς τὸν Ἑλληνικὸν Χῶρον, ὅπου συγκρούονται ἡ Ἀφρικανικὴ μὲ τὴν Εὐρασιατικὴ πλάκα δημιουργεῖται ἔντονον τεκτονικὸν πεδίον μὲ ἀποτέλεσμα τὴν σημαντικὴν σεισμικὴν δράσιν, τὴν ἔντονον γεωλογικὴν ρηγμάτωσιν καὶ τὴν δημιουργίαν μικρῶν τοπικῶν πλακῶν μὲ ἰδιαιτέραν κινηματικήν, καθιστώντας τοιουτοτρόπως τὸν χῶρον αὐτὸν πρόσφορον Φυσικὸν Ἐργαστήριον διὰ τὴν δοκιμὴν νέων τεχνολογιῶν καὶ μεθόδων μετρήσεων, μελετῶν καὶ θεωριῶν, σχετικῶν μὲ τὴν κινηματικὴν καὶ γεωτεκτονικὴν δομὴν τῆς περιοχῆς.

Οἱ μετατοπίσεις αὐτές, ποὺ ἡ τάξις τους εἶναι τοῦ ἑκατοστοῦ τοῦ μέτρου ἀνὰ ἔτος, ἔχουν μὲν διαπιστωθῆ, χωρὶς ὅμως νὰ εἶναι εὔκολος ὁ ἀκριβὴς ἀπ' εὐθείας προσδιορισμός τους, λόγω τῆς ἀπαιτουμένης ὑψηλῆς ἀκριβείας τῶν μετρήσεων. Τοιουτοτρόπως διὰ τὴν μετακίνησιν τοῦ στεροῦ φλοιοῦ τῆς γῆς εἰς μίαν περιοχὴν θὰ ἔπρεπε νὰ περιμένη κανεὶς ἀρκετὰ χρόνια διὰ νὰ διαπιστώση μετακινήσεις.

Σήμερον ὅμως ἡ Γεωδαισία, μὲ τὰς Δορυφορικὰς Μεθόδους ποὺ δίδουν ἀκρίβειαν καλυτέραν ἀπὸ 10^{-6}, διὰ ἀποστάσεις μέχρι καὶ ἑκατοντάδες χιλιομέτρων, ἐπιτρέπει τὴν ἀνίχνευσιν μετακινήσεων μέσα σὲ λίγα χρόνια. Βεβαίως, αἱ παλαιαὶ γεωδαιτικαὶ μετρήσεις εἶναι πάντα πολύτιμοι, ἐφόσον διαπιστωθεῖ ὅτι εἶναι ἀκριβεῖς καὶ συμβιβαστὲς μὲ τὶς νεώτερες, ὁπότε εἶναι δυνατὸν νὰ συγκριθοῦν ἀποτελέσματα παλαιῶν (γηΐνων), καὶ νέων (δορυφορικῶν ἢ/καὶ γηΐνων) γεωδαιτικῶν μετρήσεων.

Τὸ Κέντρον Δορυφόρων Διονύσου καὶ τὸ Ἐργαστήριον Ἀνωτέρας Γεωδαισίας τοῦ Ε. Μ. Πολυτεχνείου, ἔχει ἀναπτύξει ἰδιαιτέραν δραστηριότητα στὴν ἐπιστημονικὴν αὐτὴν περιοχὴν μὲ τὴν συμβολὴν τῆς Γεωδαιτικῆς καὶ Γεωφυσικῆς Ἐπιτροπῆς τοῦ Κράτους καὶ ξένων Πανεπιστημίων καὶ Ἐρευνητικῶν Κέντρων, διὰ τὴν καταγραφὴν καὶ μελέτην τῶν τεκτονικῶν μετακινήσεων στὸν Ἑλληνικὸν Χῶρον. Περίπου 200 σημεῖα τοῦ Ἑλληνικοῦ Χώρου ἔχουν μετρηθεῖ μὲ δορυφορικὲς μεθόδους, μετρήσεις ποὺ θὰ ἐπαναληφθοῦν στὰ ἑπόμενα ἔτη διὰ τὸν προσδιορισμὸν τῶν μετατοπίσεων.

Ὡς πρῶτον ἀποτέλεσμα τῆς συνεργασίας αὐτῆς μὲ Ἀγγλικὰ Πανεπιστήμια ἀναφέρεται ἡ δημοσίευσις στὸ ἔγκριτον περιοδικὸν Nature (14-3-1991), μὲ τίτλον «Geodetic determination of tectonic deformation in Central Greece from 1900 to 1988», ὅπου διαπιστώνονται μετακινήσεις τῆς τάξεως τοῦ μέτρου διὰ χρονικὸν διάστημα περίπου 90 χρόνων, συγκρίνοντας τὰ ἀποτελέσματα μετρήσεων τοῦ πρώτου τριγωνισμοῦ τῆς χώρας, ποὺ ἔγινε εἰς τὴν περίοδον 1890-1900, μὲ δορυφορικὲς μετρήσεις μὲ τὸ σύστημα *Global Positioning System (GPS)*, ποὺ ἔγιναν τὸ 1988.

Εἰς τὴν παροῦσαν ἐργασίαν μελετῶνται τὰ ἀποτελέσματα ποὺ προκύπτουν ἀπὸ τὴν σύγκρισιν τῶν γηΐνων μετρήσεων τοῦ πρώτου Ἑλληνικοῦ Τριγωνισμοῦ 1890-1900, μὲ ἐκεῖνα τοῦ τελευταίου ἐπίγειου τριγωνισμοῦ τῆς περιόδου 1968-1982, ἐρευνᾶται ἡ κινηματικὴ τοῦ φλοιοῦ τῆς Κεντρικῆς Ἑλλάδος ἀπὸ τὴν σκοπιὰ τῆς Γεωδαισίας καὶ διατυπώνονται πιθανὰ μοντέλα γεωτεκτονικῆς δομῆς του.

Διὰ τὸν ὑπολογισμὸν τῶν μετατοπίσεων ἐχρησιμοποιήθησαν 80 κοινὰ σημεῖα, ποὺ βεβαιωμένα παρέμειναν τὰ ἴδια, τῶν τριγωνισμῶν τῆς περιόδου 1890-1900 (μέση ἐποχὴ περίπου 1895) καὶ 1968-1982 (μέση ἐποχὴ περίπου 1975). Ὁ τριγωνισμὸς 1975 ἐχρησιμοποιήθη ὅπως ὑπελογίσθη εἰς τὸ νέον Ἑλληνικὸν Γεωδαιτικὸν Σύστημα Ἀναφορᾶς, γνωστὸν ὡς ΕΓΣΑ '87. Τὸ σύστημα αὐτὸ στηρίζεται σὲ ἐπίγειες μετρήσεις, μὲ συνδυασμὸ δορυφορικῶν παρατηρήσεων καὶ ὑπελογίσθη μὲ τὴν συνεργασίαν τῆς Γεωγραφικῆς Ὑπηρεσίας Στρατοῦ (ΓΥΣ), τοῦ Ὀργανισμοῦ Κτηματολογίου καὶ Χαρτογραφήσεων Ἑλλάδος (ΟΚΧΕ), καὶ τοῦ Ε. Μ. Πολυτεχνείου (ΕΜΠ). Ἡ ἀκρίβειά του σὲ προσανατολισμὸ καὶ κλίμακα εἶναι καλύτερη ἀπὸ 10^{-6}, ἐνῶ ἡ τοπικὴ ἀβεβαιότης στὶς θέσεις εἶναι περίπου 2-3 cm.

Ὁ τριγωνισμὸς τοῦ 1895 χρησιμοποιήθηκε μετὰ ἀπὸ νέον ὑπολογισμὸν στὸ ἴδιο ἀκριβῶς σύστημα ἀναφορᾶς, ὅπως ἐκεῖνο τοῦ 1975 (ΕΓΣΑ 87), χρησιμοποιώντας μόνον τὶς γωνιομετρήσεις τῆς ἐποχῆς ἐκείνης, δεδομένου ὅτι οἱ πλευρὲς δὲν θεωροῦνται ἱκανοποιητικῆς ἀκρίβειας.

Ἐπειδὴ αἱ παλαιαὶ μετρήσεις ἦταν μόνον γωνίες, ὁ τριγωνισμὸς 1895 ἀποτελεῖ ἐλεύθερον δίκτυον καὶ ἑπομένως ἦτο ἐλευθέρας θέσεως, προσανατολισμοῦ καὶ κλίμακος. Τὰ στοιχεῖα αὐτὰ προσδιορίστηκαν οὕτως ὥστε νὰ ὑπάρξει ἡ καλυτέρα δυνατὴ προσαρμογὴ στὰ κοινὰ σημεῖα τοῦ τριγωνισμοῦ τοῦ 1975, μὲ τὴν ἔννοιαν τῶν ἐλαχίστων τετραγώνων.

Πρέπει νὰ σημειωθεῖ ὅτι ἡ ἐπιλογὴ τῆς προσαρμογῆς τοῦ τριγωνομετρικοῦ δικτύου τοῦ 1895 εἰς τὸ δίκτυον τοῦ 1975, εἶναι ἀπὸ γεωμετρικῆς ἀπόψεως ἰσοδύναμος μὲ τὴν παραδοχὴν τῆς σταθερότητος δύο σημείων εἰς τὸ χρονικὸν διάστημα αὐτό, παραδοχὴ ποὺ συνήθως γίνεται σὲ τέτοιες ἐργασίες. Ἐπελέγη ὅμως ἡ μέθοδος τῆς «γενικῆς» προσαρμογῆς τῶν δύο δικτύων, ὥστε νὰ μὴν ὑπάρξει a priori παραδοχὴ διὰ τὴν σταθερότητα ὡρισμένων σημείων. Ἡ μετατροπὴ πάντως τοῦ πεδίου μετατοπίσεων ποὺ προκύπτει, σὲ ἐκεῖνο ποὺ θὰ προέκυπτε ἂν ὁποιαδήποτε δύο σημεῖα ἐθεωροῦντο σταθερά, εἶναι δυνατή.

Συνολικῶς εὑρέθησαν 80 γεωδαιτικὰ σημεῖα κοινὰ στοὺς δύο τριγωνισμούς, ἀλλὰ τὰ 5 ἀπὸ αὐτὰ δὲν ἐχρησιμοποιήθησαν, ἐπειδὴ ἔδειξαν ἰδιαιτέρως μεγάλες μετατοπίσεις καὶ θὰ πρέπει νὰ ἐξεταστοῦν ἰδιαιτέρως διὰ ἐνδεχόμενα συστηματικὰ σφάλματα. Οἱ μετατοπίσεις, ὅπως προκύπτουν ἀπὸ τὴν σύγκρισιν τῶν δύο τριγωνομετρικῶν δικτύων τῶν ἐτῶν 1895 καὶ 1975 στὰ κοινά τους σημεῖα ἐμφανίζονται στὸ Σχῆμα 1. Στὸ σχῆμα αὐτὸ σημειώνονται μὲ βέλη οἱ ταχύτητες τῶν μετακινήσεων σὲ cm/y (ἢ m/cen), ἐνῶ τὸ μῆκος τοῦ βέλους δεικνύει εἰς τὴν κλίμακα τοῦ χάρτου τὴν μετατόπισιν τοῦ κάθε σημείου, ποὺ ἀντιστοιχεῖ σὲ 2 ἑκατομμύρια ἔτη.

Ἂν θεωρηθεῖ ἡ περιοχὴ ποὺ καλύπτεται ἀπὸ τὰ σημεῖα ποὺ ἐχρησιμοποιήθησαν, ὡς ἑνιαία ἐπιφάνεια, εἶναι δυνατὸς ὁ ὑπολογισμὸς μέσου τανυστοῦ παραμορφώσεων. Ὁ ὑπολογισμὸς αὐτὸς δίδει ὡς μεγίστην ἀνηγμένην παραμόρφωσιν ἴσην πρὸς +0.031 μstr/y (1 μstr $=10^{-6}$ ἢ 1 mm/km) ὡς πρὸς διεύθυνσιν μὲ ἀζιμούθιον 33° καὶ ἐλαχίστην -0.017 μstr/y εἰς ἀζιμούθιον 123°. Τοῦτο ἀντιστοιχεῖ σὲ μεγίστη γωνιακὴ παραμόρφωση 0.048 μstr/y. Ὁλικὴ μετακίνησις, ὁλικὴ στροφὴ καὶ ὁλικὴ ἀπόλυτος παραμόρφωσις δὲν εἶναι δυνατὸν νὰ ὑπολογισθεῖ, ἀφοῦ δὲν ὑπάρχουν σταθερὰ σημεῖα. Τὸ Σχῆμα 2 δεικνύει τὸν τανυστὴν αὐτόν, ὅπως σχηματοποιεῖται ἀπὸ τὴν παραμόρφωσιν κύκλου πρὸς ἔλλειψιν, διὰ χρονικὸν διάστημα 5 ἑκατομμυρίων ἐτῶν.

Δὲν εἶναι δυνατὸν ὅμως νὰ θεωρηθεῖ ὅτι ὅλη ἡ ἔκτασις ἐκφράζεται μὲ ἑνιαῖον τανυστὴν παραμορφώσεων. Ἡ τυπικὴ ἀπόκλισις τῶν ὑπολοίπων τῶν διανυσμάτων ὑπερβαίνει τὴν a priori ἀναμενόμενην τυπικὴν ἀπόκλισιν τῶν παρατηρήσεων (κατὰ 80%) καὶ ἄλλωστε, μὲ ἁπλὴν θεώρησιν τοῦ σχήματος 1 διαφαίνεται σαφῶς ἡ ἀδυναμία τοιαύτης ἑνιαίας λύσεως.

Θεωρώντας τὴν ἔκτασιν ποὺ καλύπτει ἡ μελέτη ὡς μίαν ἐπιφάνειαν μὲ συνεχεῖς παραμορφώσεις, εἶναι δυνατὸν νὰ ὑπολογισθεῖ ἀριθμητικῶς πεδίον μετατοπίσεων ποὺ προκύπτει μὲ παρεμβολὴν ἀπὸ τὰ 80 γεωδαιτικὰ σημεῖα, τῶν ὁποίων ἡ μετατόπισις προσδιορίστηκε ἀπ' εὐθείας. Τὸ ἀποτέλεσμα ἐμφανίζεται στὸ Σχῆμα 3, ὅπου τὸ μῆκος τοῦ βέλους, ποὺ δεικνύει τὴν μετατόπισιν στὰ κορυφὰς φανταστικοῦ κανονικοῦ κανάβου 20 χιλιομέτρων, δίδει στὴν κλίμακα τοῦ χάρτου τὴν μετατόπισιν δι' ἓν ἑκατομμύριον ἔτη. Μιὰ ἄλλη εἰκόνα τοῦ πεδίου αὐτοῦ, δίδει τὸ Σχῆμα 4, ὅπου φαίνεται ἡ παραμόρφωσις τοῦ κανάβου διὰ τὸ αὐτὸ χρονικὸν διάστημα.

Αἱ μεταβολαὶ τῶν παραμορφώσεων εἰς τὴν ἔκτασιν ποὺ μελετᾶται, μπορεῖ νὰ βρεθοῦν, ἂν ὑπολογισθοῦν οἱ τανυσταὶ παραμορφώσεων σὲ διάφορα σημεῖα, χρησιμοποιώντας ὡρισμένην περιοχὴν περὶ ἕκαστον σημεῖον. Ἀναλόγως μὲ τὴν πυκνότητα τῶν σημείων καὶ τὴν ἔκτασιν τῆς περιοχῆς ποὺ χρησιμοποιεῖται, ἐν σχέσει μὲ τὴν πυκνότητα τῶν δεδομένων, δημιουργεῖται μιὰ περισσότερον ἢ ὀλιγότερον ὁμαλοποιημένη ἀπεικόνησις τῆς μεταβολῆς ποὺ θὰ γίνεται μὲ τὴν χάραξιν ἰσαριθμικῶν καμπυλῶν διὰ τὰς διαφόρους παραμέτρους ποὺ ἐκφράζουν ἕκαστον τανυστήν.

Εἰς τὴν μελέτην αὐτὴν ἐχρησιμοποιήθησαν διάφοροι πυκνότητες σημείων καὶ ἐκτάσεις περιοχῶν διὰ τὴν ἐκτίμησιν τῆς μεταβολῆς τῶν παραμορφώσεων. Ἐνδεικτικῶς τὰ Σχήματα 5, 6 καὶ 7 δίνουν τὴν μετατόπισιν κατὰ τὴν διεύθυνσιν τῆς ἀνατολῆς, τὴν μεγίστην γωνιακὴν παραμόρφωσιν καὶ τὶς διευθύνσεις τῶν κυρίων ἀξόνων παραμορφώσεως ἀντιστοίχως, ὅπως προέκυψαν μὲ τὴν χρῆσιν κινητῆς μέσης τιμῆς καὶ μὲ σάρωσιν διαστήματος ἡμισείας μοίρας κατὰ πλάτος καὶ μῆκος.

Ἀπὸ τὴν εἰκόνα ποὺ δίνουν τὰ πρωτογενῆ στοιχεῖα τῶν μετατοπίσεων (σχ. 1) δύναται νὰ διαπιστωθεῖ ὅτι ὑπάρχουν περιοχαὶ ποὺ ἔχουν σαφῶς κατὰ τὸ μᾶλλον ἢ ἧττον ὁμοιόμορφον παραμόρφωσιν, ἐνῶ εἰς τὸ σύνορόν των φαίνεται νὰ ὑπάρχει ἔντονος μεταβολή, ποὺ μπορεῖ νὰ ὑποδηλώνει κάποια ἀσυνέχεια. Γιὰ τὸν λόγον αὐτὸν ἡ ὅλη ἔκτασις διηρέθη εἰς 9 περιοχὰς ποὺ ἔδειχναν νὰ ἱκανοποιοῦν τὰ κριτήρια ὁμοιομόρφου παραμορφώσεως. Ἡ διαίρεσις αὐτή, ποὺ ἐμφανίζεται στὸ Σχῆμα 8, ἔγινε μὲ μοναδικὸν κριτήριον τὸ πεδίον τῶν πρωτογενῶν μετατοπίσεων χωρὶς νὰ ληφθεῖ ὑπόψιν καμμία ἄλλη ἐξωτερικὴ πληροφορία, σχετικὴ μὲ τὴν γεωλογία ἢ τὴν τεκτονικὴ τῆς περιοχῆς. Ἂς σημειωθεῖ πάντως ὅτι τὰ ὅρια τῶν περιοχῶν ἔχουν σχετικὴν ἀσάφειαν, δεδομένου ὅτι οἱ ἀποστάσεις μεταξὺ τῶν σημείων ποὺ ὑπάρχουν οἱ πρωτογενεῖς μετατοπίσεις φθάνουν τὶς μερικὲς δεκάδες χιλιόμετρα.

Διὰ κάθε μία ἀπὸ τὶς 9 περιοχὲς ὑπελογίσθη ὁ πλήρης τανυστὴς παραμορφώσεως, δεδομένου ὅτι τώρα εἶναι δυνατὸν νὰ ὑπολογισθεῖ ἡ ὁλικὴ μετατόπισις καὶ στροφὴ ὡς πρὸς τὸν μέσον ὅρον τοῦ πεδίου τῶν μετατοπίσεων. Τὰ ἀποτελέσματα τῶν ὑπολογισμῶν ποὺ ἔγιναν μὲ τὴν Μέθοδο τῶν Ἐλαχίστων Τετραγώνων δίνονται σὲ σχετικὸν Πίνακα.

Τὸ Σχῆμα 9 δίδει γραφικῶς τὸν πλήρη τανυστὴν γιὰ τὴν κάθε περιοχὴν ἐκφρασμένον ὡς ὁλικὴ μετατόπισιν, στροφὴν καὶ ἔλλειψιν παραμορφώσεων. Εἰς τὴν κλίμακα τοῦ χάρτου τὰ μεγέθη, ὅπως ἐμφανίζονται, ἀντιστοιχοῦν σὲ μεταβολὲς 5 ἑκατομμυρίων ἐτῶν. Τέλος τὸ Σχῆμα 10 δίδει τὴν παραμόρφωσιν ποὺ θὰ ἔχει ὁ κανονικὸς κάναβος σὲ περίοδο 2 ἑκατομμυρίων ἐτῶν, συμπεριλαμβανομένης τῆς ὁλικῆς μετακινήσεως καὶ στροφῆς.

Τὰ ἀποτελέσματα τῆς ἐργασίας αὐτῆς εἶναι συμβιβαστὰ μὲ τὰ ἀποτελέσματα ποὺ παρουσιάσθησαν εἰς τὸ περιοδικὸν Nature, ἀλλὰ σ' αὐτὴν τὴν περίπτωσιν, ἔχοντας πενταπλάσια σημεῖα, δίδουν πολὺ καλυτέραν εἰκόνα διὰ τὸ πεδίον μετατοπίσεων. Οὕτω κατέστη δυνατὸν νὰ ὁριοθετηθοῦν περιοχαὶ μὲ ὁμοιόμορφον παραμόρφωσιν, ποὺ εὔκολα διαπιστώνεται ὅτι ἀντιστοιχοῦν εἰς τεκτονικὰς ἑνότητας.

Ἡ ἐργασία αὐτὴ ἐλπίζεται νὰ ἐπεκταθεῖ μὲ τὴν ἐπαύξησιν τῶν γεωδαιτικῶν σημείων, μὲ τὴν χρησιμοποίησιν παλαιοτέρων μετρήσεων, ἀλλὰ κυρίως μὲ τὸν ἐπαναπροσδιορισμὸν νέων θέσεων μὲ μεθόδους τῆς Δορυφορικῆς Γεωδαισίας. Ἐπίσης ἀποσκοπεῖται ἡ σύγκρισις τῶν γεωδαιτικῶν μετατοπίσεων, μὲ ἐκεῖνες ποὺ προκύπτουν ἀπὸ τὴν ἀνάλυσιν τῆς μέχρι σήμερον σεισμικῆς δράσεως, ὥστε νὰ εἶναι δυνατὴ ἡ ἐκτίμησις τῶν τάσεων ποὺ δὲν ἔχουν ἐκτονωθεῖ, καὶ νὰ ἐκτιμηθεῖ τοιουτοτρόπως μὲ ἄλλον ἕνα τρόπον ἡ πιθανότης ἀναμενόμενης σεισμικῆς δράσεως.

TABLE I

DISPLACEMENTS ROTATIONS & STRAIN IN CENTRAL GREECE

Block (n)	σ_o mm/y	$\dot{d}_x$ mm/y	$\dot{d}_y$ mm/y	$\dot{\bar{\varepsilon}}=\dot{\omega}$ marcsec/y	$\dot{e}_{max}=\dot{e}_1$ μstr/y	a_1 deg	$\dot{e}_{min}=\dot{e}_2$ μstr/y	a_2 deg	$\dot{\gamma}$ μstr/y	$\dot{\gamma}$ marcsec/y
A (11)	1.30	-1.43	-1.13	-0.94	+0.031	27.6	-0.041	117.6	0.072	14.75
B (16)	1.55	-1.65	-6.83	-0.90	+0.019	100.0	-0.015	10.0	0.034	7.0
C (5)	0.40	-8.21	-6.68	8.34	+0.028	77.7	-0.032	167.7	0.059	12.2
D (4)	3.41	-0.12	+9.65	33.55	+0.135	23.8	+0.019	113.8	0.116	23.9
E (8)	2.61	+6.60	+6.85	13.44	-0.016	20.8	-0.137	110.8	0.120	24.7
F (12)	1.69	+2.78	-0.36	2.34	+0.026	38.7	-0.039	128.7	0.064	13.2
G (4)	1.19	-7.03	+3.17	7.57	+0.087	108.7	-0.131	18.7	0.218	44.8
H (3)	--	-2.45	+3.34	-9.58	+0.120	7.6	-0.030	97.6	0.150	31.0
K (8)	4.00	+2.21	+8.22	-17.20	+0.057	179.1	-0.058	89.1	0.115	23.7

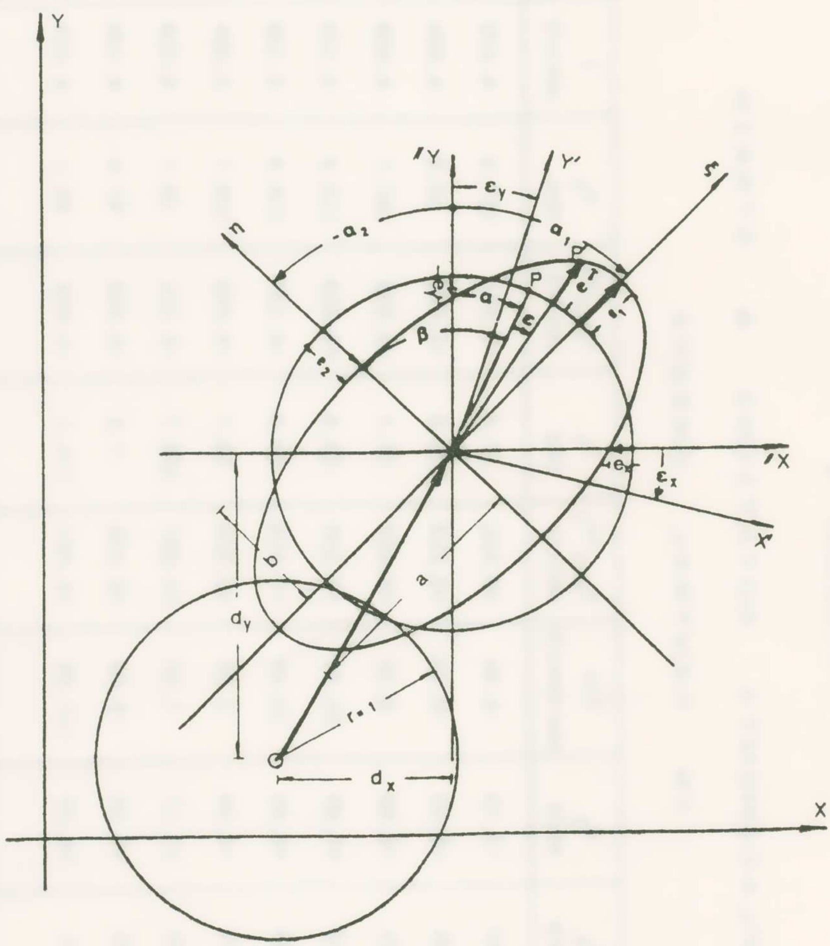

Fig. 1. The strain ellipse.

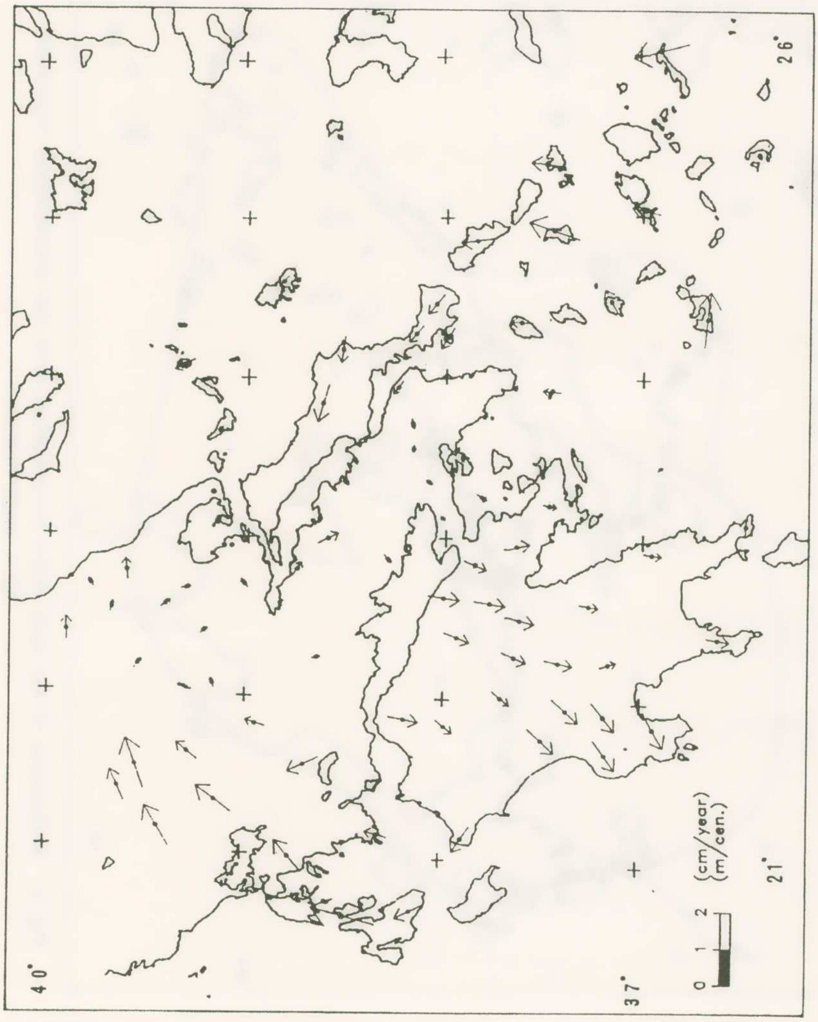

Fig. 2. The directly observed displacement vectors. The length of the vectors correspond to the displacement at the map scale over a period of 2 My.

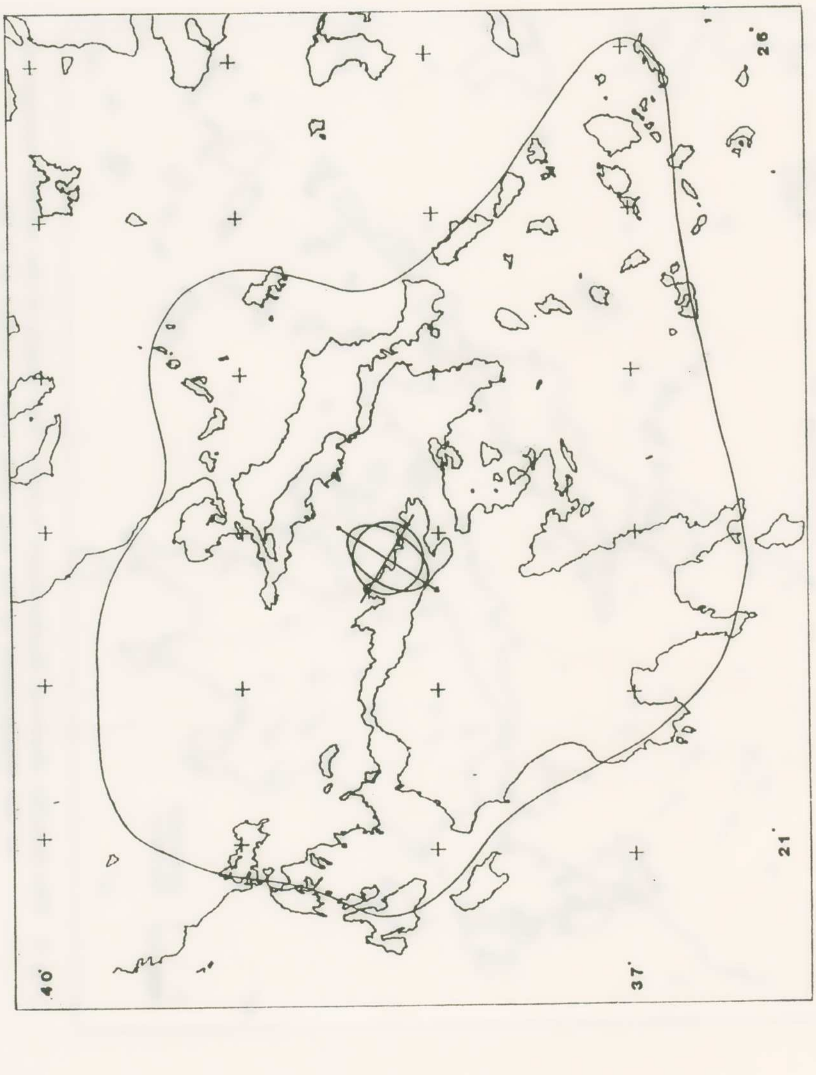

Fig. 3. Delineation of the area under investigation with the deformations expressed by a single tensor.

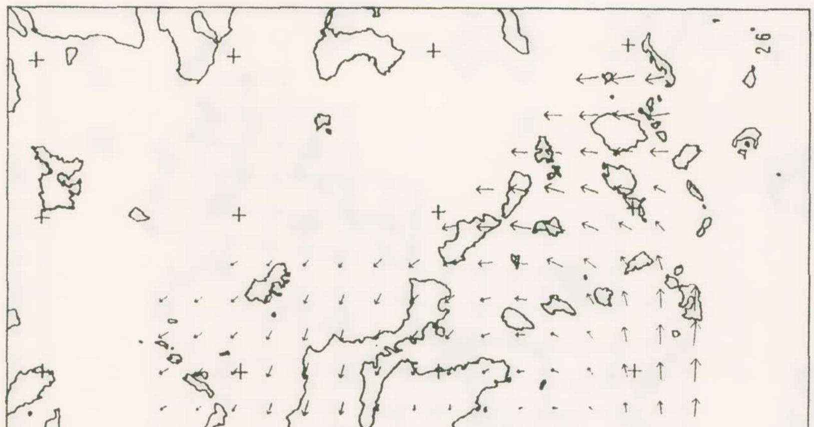

Fig. 4. The displacement field derived by interpolation from the observed displacements. The vectors correspond to the displacements at the map scale over a period of 1 My.

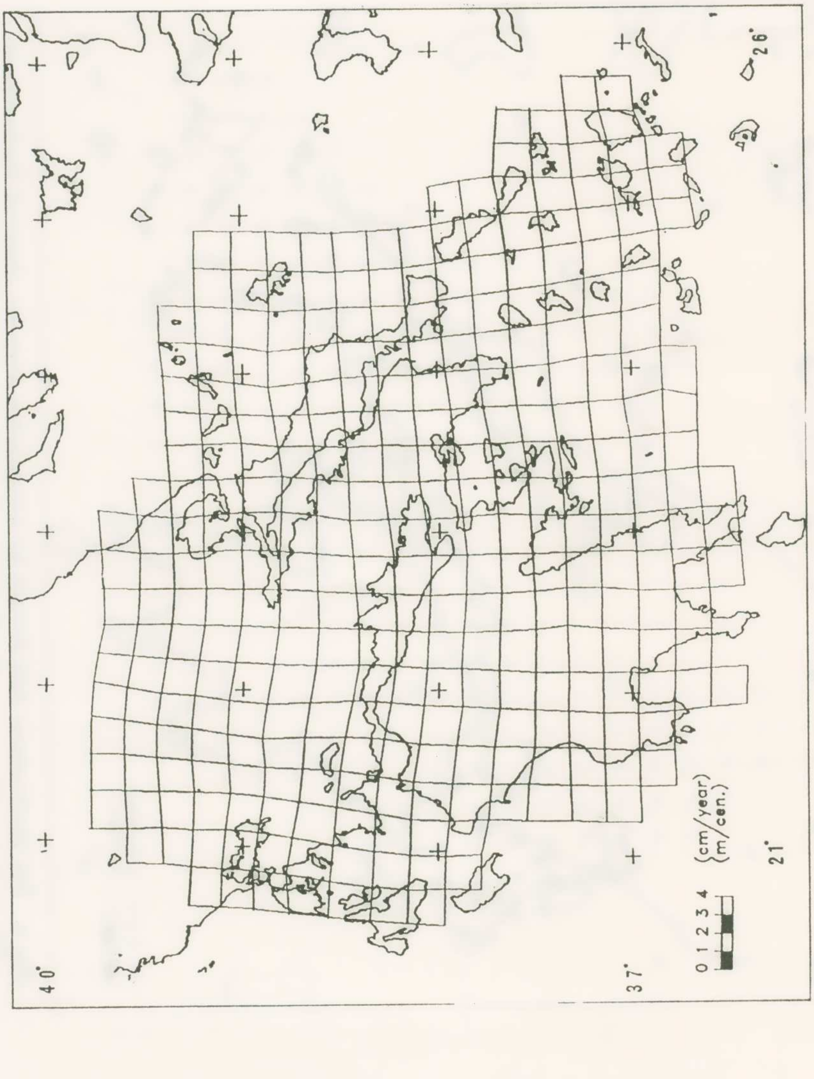

Fig. 5. The distorted grid 20 km × 20 km over a period of 1 My.

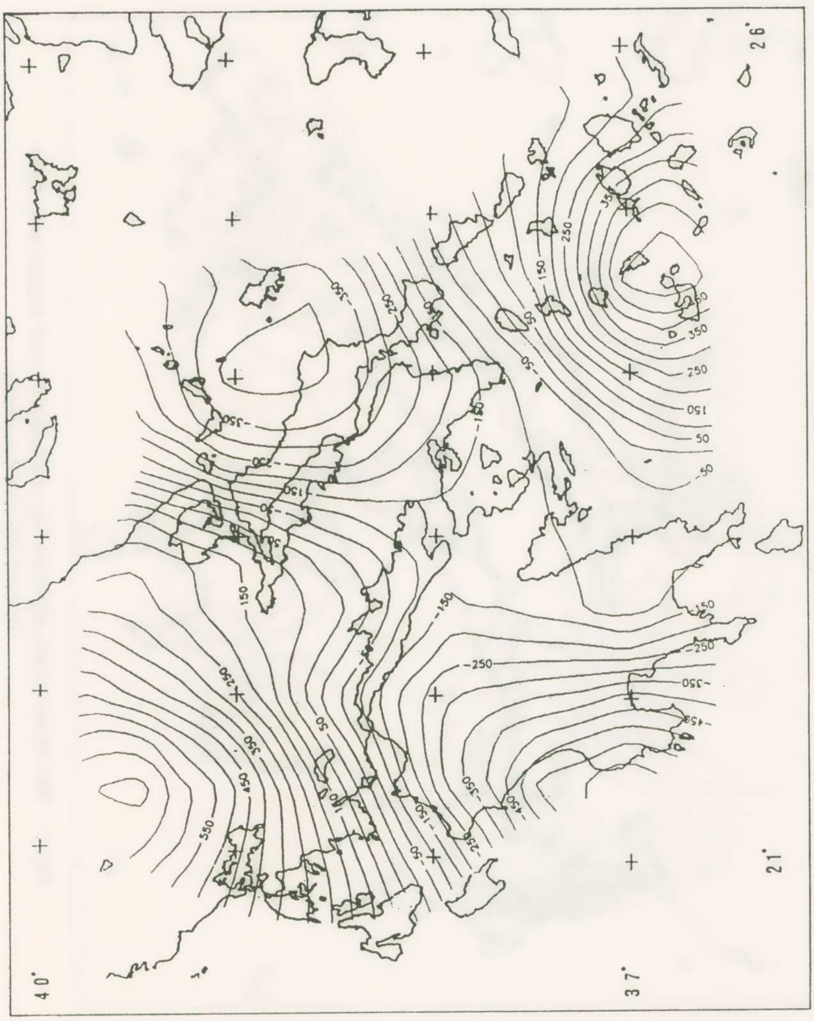

Fig. 6. The displacement in the East direction obtained by scanning in half degree step both in latitude and longitude.

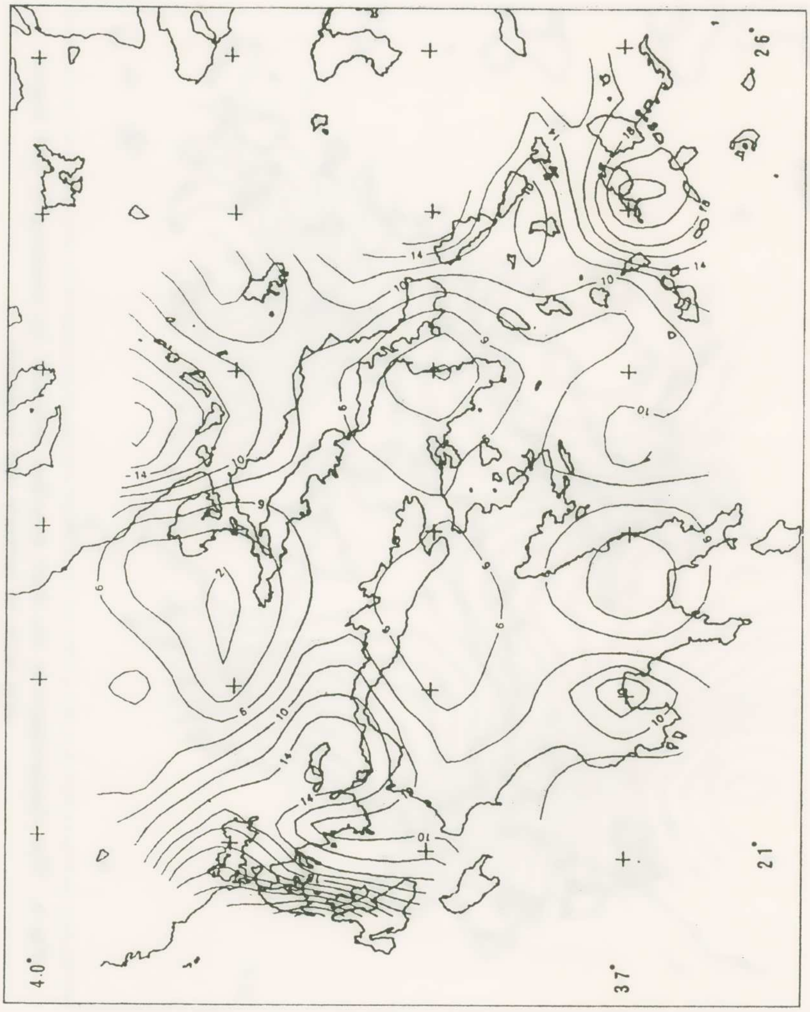

Fig. 7. The shear strain obtained by scanning in half degree step both in latitude and longitude.

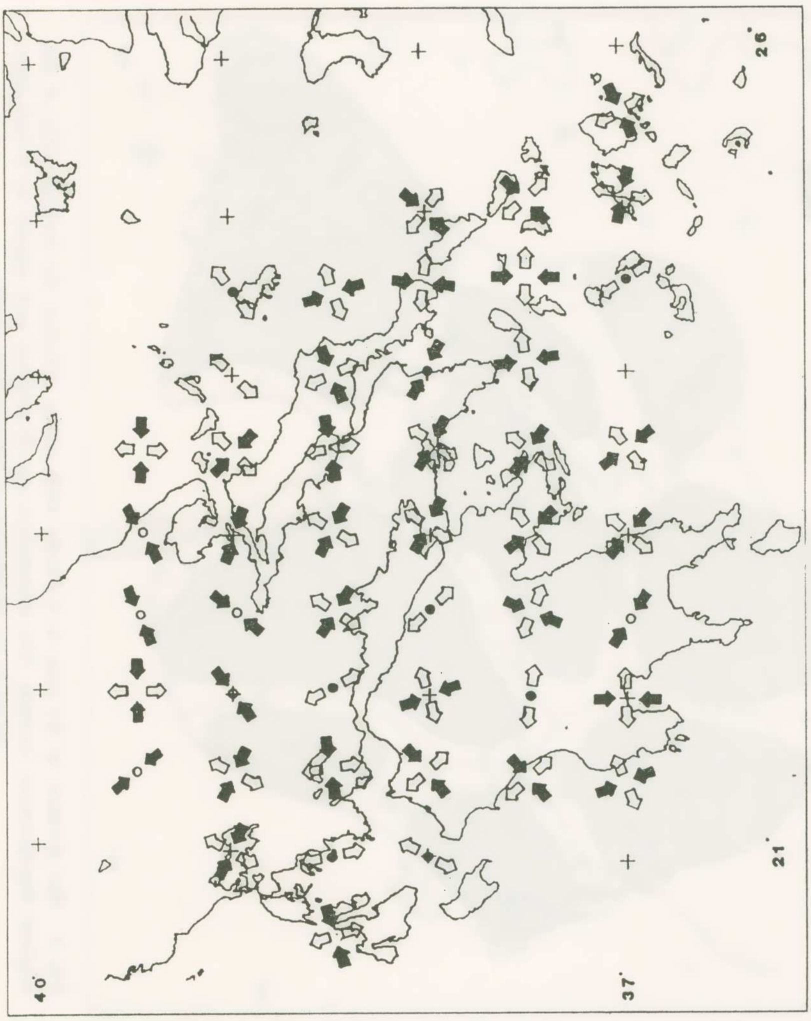

Fig. 8. The principal axes of strain obtained by scanning in half degree step both in latitude and longitude.

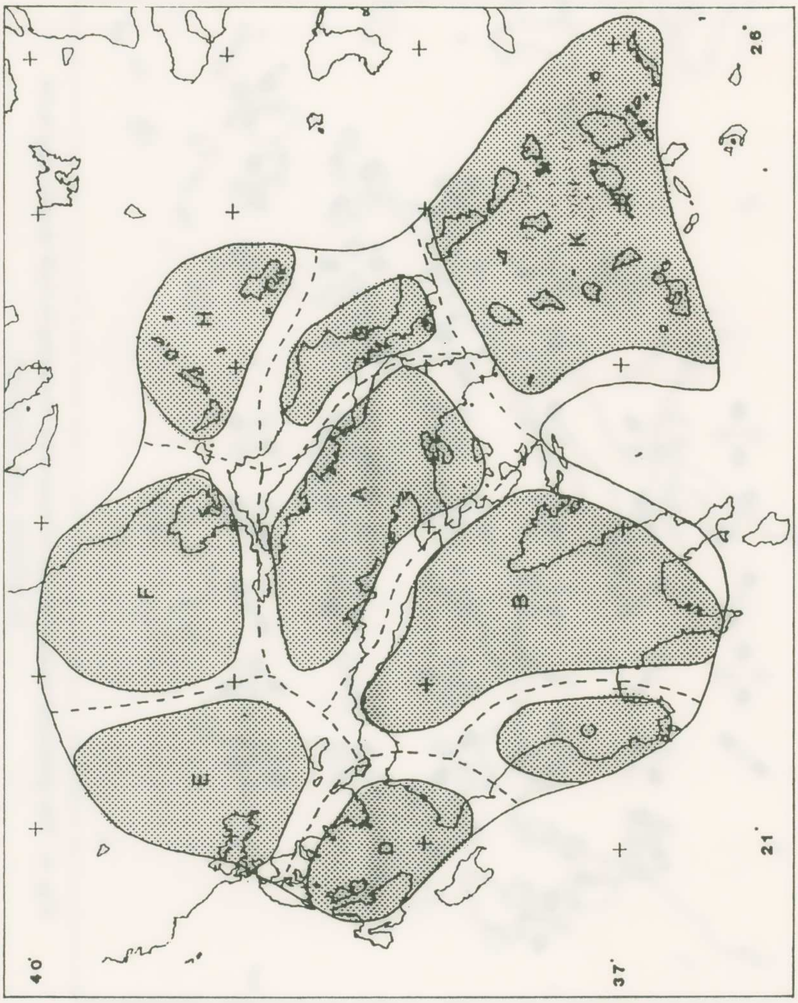

Fig. 9. The division of the area in 9 blocks with solely criterion the homogeneity of the original displacement vectors. The boundaries are ill defined and appear as a belt since there is no displacement information inside.

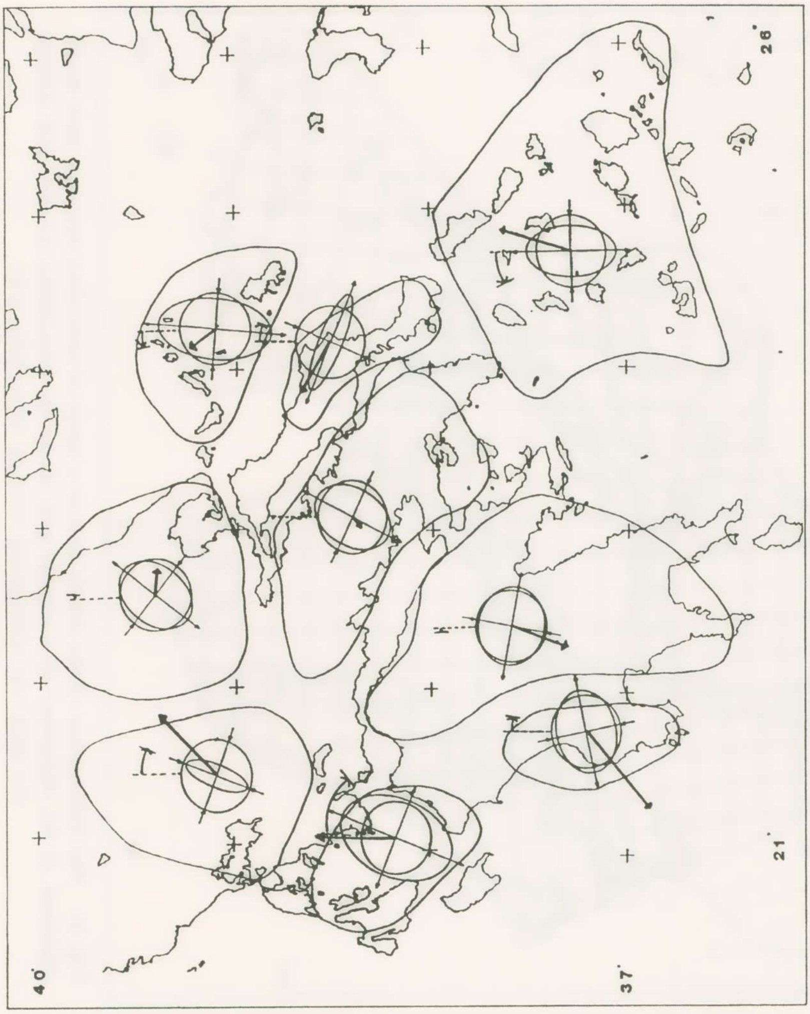

Fig. 10. The complete local strain tensor for each block presented as a block displacement, rotation and a strain ellipse. They are drawn to the map scale as they will correspond in 5 My.

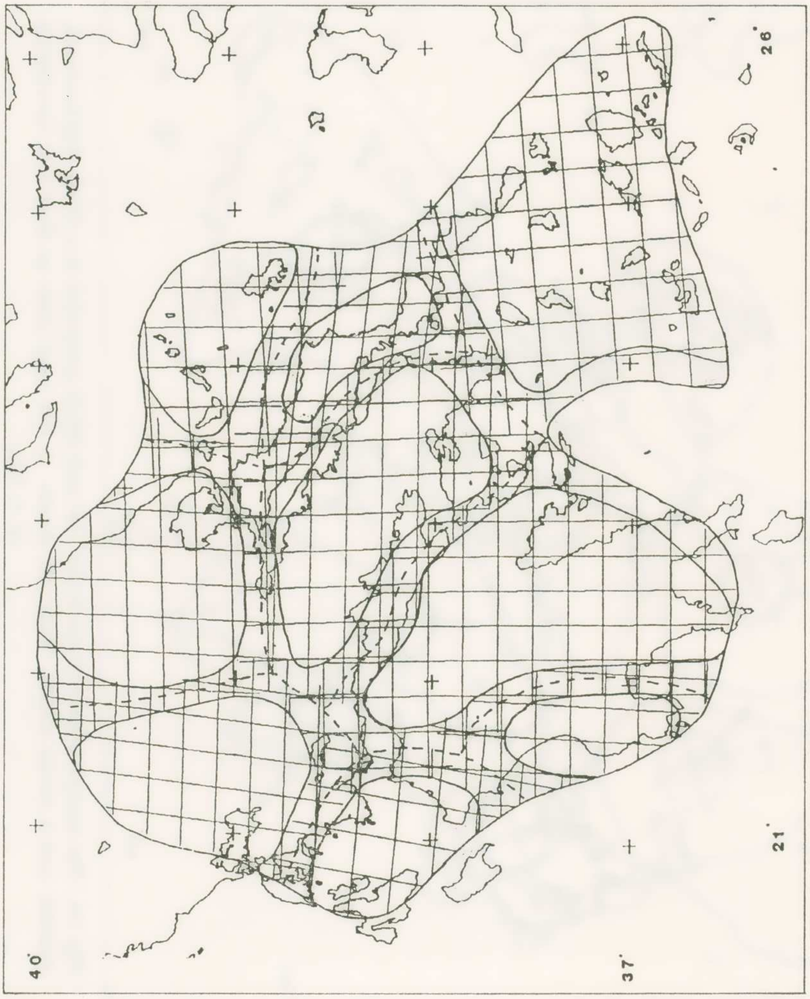

Fig. 11. The deformation of the whole area as a result of individual strain of each block, visualised as the distortion of a regular grid. Since the boundaries are not well defined the grids are overlapping over an uncertailty belt.