Journal of Structural Geology, Vol. 3, No. 3, pp. 309 to 315, 1981.
Printed in Great Britain

0191-8141/81/030309-07 $02.00/0

The West European stress field: new data and interpretation

T. J. SCHMITT

14005 Cove Lane Apt. 204, Rockville, Maryland 20851, U.S.A.

(*Received 27 November 1979; accepted in revised form 14 May 1981*)

Abstract—The mapped stress field of Western Europe reflects the tectonic process active there. A traverse of stress measurements from the Alps and through their northern foreland to the southern border of the Lower Rhine Embayment identifies three distinct stress sub-provinces; the Western Alps, the blocks on both flanks of the Rhinegraben, and the Rhenish Shield. The Alps have high magnitude stresses up to 35 MPa in the direction of maximum compression, here called σ_{1h}. The general direction of σ_{1h} is about 140°. The foreland has the same directional trend of σ_{1h} with a magnitude reduced to about 2.0 MPa. Local anomalies in magnitude and direction occur along the course of the Rhinegraben which is a site of active sinistral shear. The Rhenish Shield shows an internal zonation of the stress field. The magnitudes of the stresses are low (usually negative) along the axis connecting the northern end of the Rhinegraben with the rifting of the Lower Rhine Embayment. The direction there is about 150°. On the eastern and western flanks of the shield the stress directions are essentially the same as in the southern blocks. These zones are distant from the belt of active strain release, consequently stresses of up to 4.0 MPa have accumulated.

INTRODUCTION AND METHODOLOGY

In situ *stress measurements*

THE DETERMINATION of the regional tectonic stress field by *in situ* techniques has become an important aspect of crustal dynamic studies. The three most common techniques are overcoring, hydraulic fracturing, and flat jacks. They were originally developed for engineering purposes. These three methods operate by three different processes and, as rocks deviate from a simple elastic behavior, each method has its advantages and disadvantages (McGarr & Gay 1978).

Most of the measurements used in this study were made by the doorstopper technique (Leeman 1970). The technique is an overcoring method in which a 3 or 4 element electrical resistance strain gauge rosette is affixed to the flattened end of a borehole. The end of the borehole is then mechanically decoupled from the rock mass by an overcoring process and the resultant strains are determined. The stresses can be calculated with a knowledge of the modulus of elasticity and Poissons ratio. A correction must be made for the stress concentration at the end of the borehole (Bonnechere & Fairhurst 1968, Van Heerdon 1968, Crouch 1969). Such a measurement or series of measurements in a borehole will determine the stresses acting in a plane normal to the axis of the hole. The complete tensor can be determined by measurements in three nonparallel holes.

The use of the above method in geodynamic studies requires considerable caution. Secondary effects which may be of no consequence in engineering and mining applications become important when the state of tectonic stress and not the simple state of stress is sought. These additional effects are of two types, those due to external factors, and those due to the intrinsic non-ideal behaviour of the rocks. Some of the factors are also of some concern for engineering studies.

Extrinsic effects

Exaggerated topographies can produce stress concentration and gravitational stresses (Hooker *et al.* 1972, Sturgul *et al.* 1976). These factors can be evaluated using a two- (or three-) dimensional finite element simple elastic model of the area surrounding a proposed site, so that a correction can be made. However, if the effects are of major proportions, it may be better to seek another site.

With surficial measurements, seasonal and diurnal temperature changes induce thermal stresses in rocks (Hooker & Du Vall 1971). While the effect on the direction is probably minimal, the magnitude effect can be large. The magnitude of these temperature effects decreases rapidly with depth (Carslaw & Jaeger 1959) and measurements at several metres depth are reasonably free of such influences. Systematic trends of magnitudes with depth may be used to determine at what depth the effect can be neglected.

Possible mechanical decoupling of a block by joints and stress concentrations around solution features in carbonates can induce false values. Comparison of values from two or more bore-holes separated by tens to hundreds of metres offers an excellent control for this effect. It is unlikely that such possible stress concentrations and/or released areas would have the same effect at widely separated sites (Greiner 1978).

Surficial weathering can create non-systematic joints which could partially decouple a block from a rock mass. This could affect both the direction and magnitude of stress. If the weathering induces an expansion of the rock, then the stress at the surface could be higher than at depth. These effects should decrease considerably with depth, and thus measurements made in mines or deep quarries should be fairly free of weathering-induced or altered stress. Again a comparison of measurements from different depths and widely-spaced boreholes is a check for these problems as it is unlikely that weathering would have the same effect at widely separated places.

Intrinsic effects

Rocks can record previous stress fields even when mechanically decoupled from external forces. This phenomenon is called residual stress, an engineering term in common usage. In the strict sense it refers to stresses within a body in the absence of applied forces. Difficulty has arisen in applying the term geologically as a consequence of the scale factor. In engineering it generally refers to a small (macro- or microscopic) scale phenomenon. Geologically, it has also been used to describe forces locked in a body due to elastic strains on a structure that were unable to relax because of large-scale boundary conditions when the external forces become extinct. In such a situation the whole structure may be regarded as being in a residual stress state, although an individual small component would not be. In this article I use residual stress in the engineering connotation; as a macro- or microscopic phenomenon.

Residual stress can act upon the doorstopper in addition to the active tectonic stress and thus results in a false estimate. Residual stress in rocks is a poorly understood, and apparently complex process. They apparently reside along grain boundaries, or boundaries of clusters of grains (Holzhausen & Johnson 1979). They have been identified in granites by X-ray diffraction (Friedman 1972), via lattice parameter distortions of strained quartz. Detection is most often reported on the basis of double overcoring, a process through which a mechanically decoupled block is subjected to a blank over- (or under-) coring experiment. If any strains are observed, then there is a possibility of residual stress. The phenomenon is lithologically dependent. As far as the author is aware they have not been detected in limestones, although they sometimes occur in sandstones (Engelder & Sbar 1976) and apparently are often present in granite (Nichols 1975). It is also possible that there is some sort of microstructural control. In numerous double overcoring experiments in Germany no residual stress has been detected though it appears to be common in the Appalachians (Greiner 1978). In general, the rocks of Germany tend to be much more highly fractured than those of the Appalachians and thus it is possible that the numerous microfractures that cut the rocks of the Alpine foreland serve to release the residual stresses.

In some porous rocks, such as poorly-cemented sandstones, a change in the amount of pore water, which can be induced by drilling, can induce strains in the bore hole core. The effect is rather complex, as it can induce expansion or contraction, and is highly time dependent. Detailed laboratory experiments are necessary to determine the presence of the effect. Caution is necessary in that this effect can produce strains similar to residual strains in double overcoring. Preliminary investigations by the author show that this effect is largely hydrostatic, but an anisotropy of porosity and/or permeability could induce directional strains. If these pore-pressure strains are present, it can be extremely difficult to estimate the true stress magnitude.

Typical tectonic stress determination

A typical tectonic stress determination in this data-base consists of observations from about 20 doorstoppers set in 3 to 5 boreholes to a depth of up to 10 m. Usually at least one double overcoring experiment was made to test for residual stresses. A two-dimensional finite element model was made if there was doubt about topographic effects. Where practicable the stress determinations were made in deep quarries to get as far below the level of surface weathering as was possible. Complete details on the application of the doorstopper technique to this class of problems is given in Greiner (1978).

Conventions used

The following conventions are used in this study: $\sigma_{1h} > \sigma_{2h}$; this is identical to the common usage in geology but opposite to that often used in engineering. The majority of the stress determinations were from shallow (5–10 m), horizontal measurements. In the case of three-dimensional studies the horizontal tensor was used. In the case of deep measurements the magnitude of stress induced by the overburden was subtracted from the measured stress.

THE STRESS FIELD

Regional consistency

A map of the horizontal stress field of Western Europe derived from *in situ* techniques is shown in Fig. 1. Rapid inspection of the map (Fig. 1) shows a NNW–SSE direction of maximum compression acting throughout the area. Earthquake fault plane solutions (Ahorner 1978) also indicate this direction for the Alps and their northern foreland. The NNW–SSE trend can be considered the regional stress field, in the sense used by Sbar & Sykes (1973). A closer look, however, shows strong localization of the magnitude and to a lesser extent the direction of the field. The highest stresses are in the Alps and the lowest along the seismic axis connecting the northern end of the Rhinegraben with the rifting of the Lower Rhine Embayment. Intermediate magnitudes are dominant in the foreland. This zonation correlates directly with the tectonic processes active in those regions.

The Alps

Magnitudes of σ_{1h} up to 35 MPa are found in the Western Alps. The general direction of maximum compression is about 140°. This direction is nearly normal to the contours of rates of recent uplift which in places reaches 1.7 mm/yr (Gubler 1976). The uplift is probably of isostatic origin as folding and thrusting in the Neo-Alpine

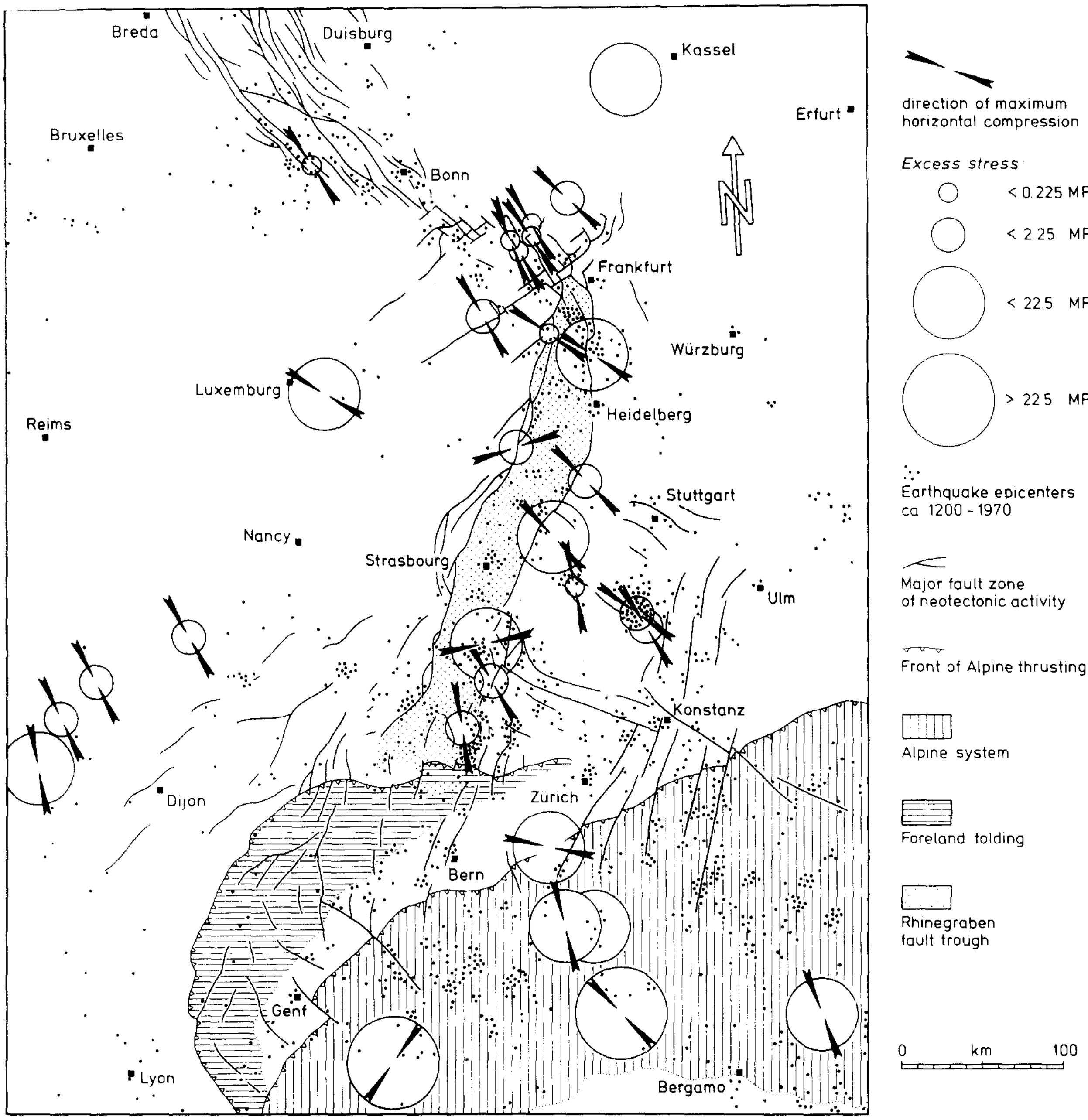

Fig. 1. The stress field of Central Europe as derived from *in situ* techniques. An overall NW direction of the maximum compression is dominant on a regional scale. Three distinct magnitude sub-provinces can be recognized, the Western Alps, the Rhinegraben and its flanking blocks, and the Rhenish Shield. Modified from Illies & Greiner (1978). (Further data sources are given in Fig. 2 and Table 1).

orogen had ended by the Early Pliocene (Trümpy 1973). Some later tectonic activity is known (Jäckli 1965), however, it is of distinctly different character and can be attributed to gravity effects induced by the uplift. Stresses will have been induced in the Alps from the topography and also from expansion and contraction related to phase changes brought about by the uplift (Illies & Greiner 1978). There is some variation in the measured direction of σ_{1h} in the Alps. This can be expected in an area with such an uneven topography (Hooker *et al.* 1972, Sturgul *et al.* 1976). The seismic fault plane solutions, which come from greater depths where these topographic effects are almost completely attenuated, are much more consistent and yield essentially the same mean direction (Ahorner 1978).

The Northern Alpine foreland and the Rhinegraben system

The direction of σ_{1h} in the foreland is basically the same as in the Alps, however, the magnitudes drop drastically to an average of around 2.0 MPa. This value is probably close to the background stress in Western Europe. Local stress anomalies occur in the Rhinegraben which is a zone of seismic strain release along its entire trace (Ahorner 1975). The Rhinegraben has been the dominant structure of the northern foreland since at least Mid-Eocene times. Illies (1977) and Illies & Greiner (1978) have reviewed its mechanical evolution. From Mid-Eocene to Early Miocene times it was an active extensional rift valley. It was inactive until the late Pliocene when (because of its NNE trend) it became a zone of sinistral shear in response to the

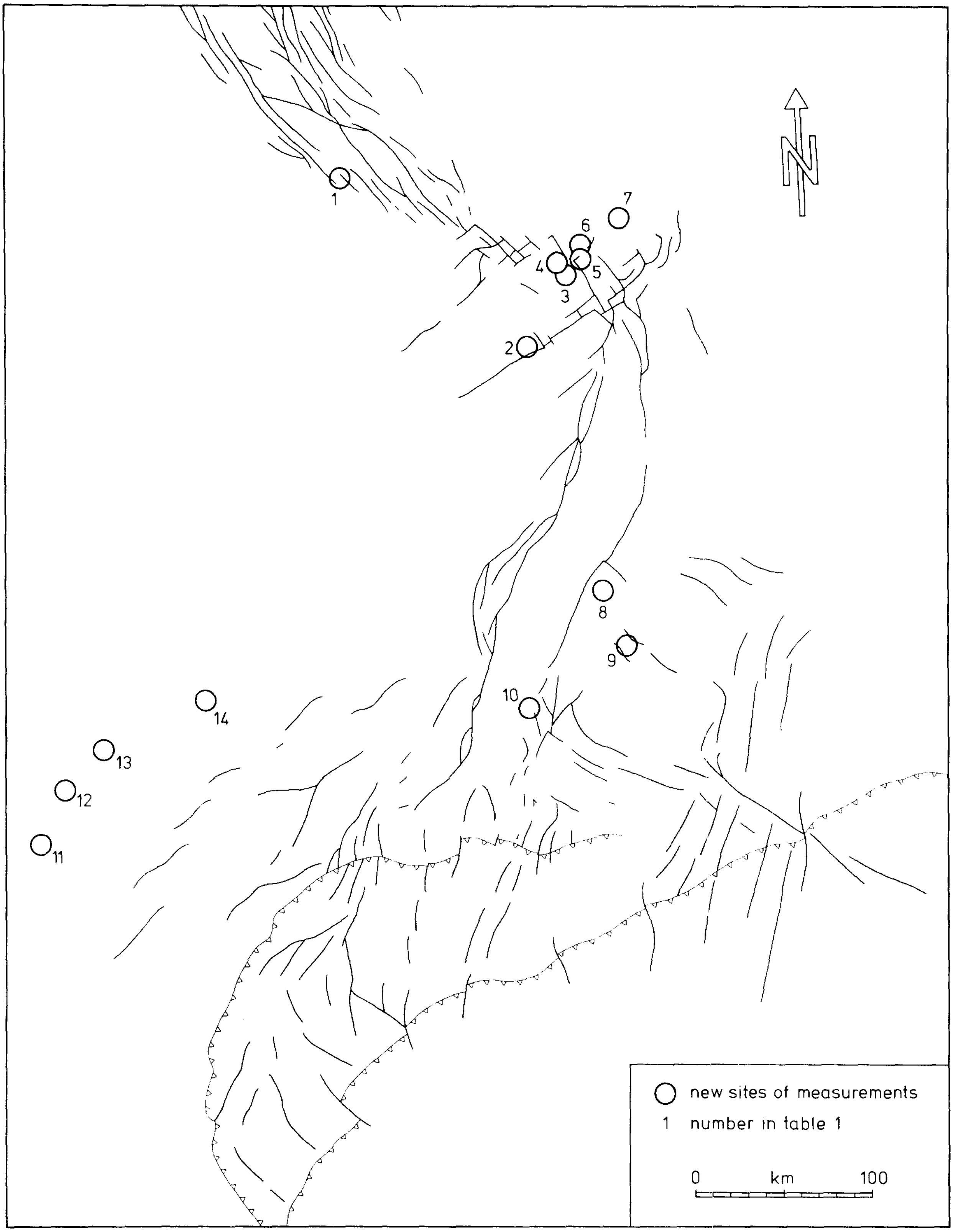

Fig. 2. Location of the new data points.

still active NNW stress regime.

The extensional Rhine rift was not a straight trace but contained several bends. Subsequently, when subjected to shear, the affected segments became sites of local tectonic anomalies. Fault plane solutions of earthquakes show components of compression or tension superimposed on the general shear (Ahorner 1975). Geologically, these areas are further defined by local anomalies in uplift and subsidence of the graben fill (Illies & Greiner 1978). These local anomalies are also reflected in the stress field. The Baden-Baden measurement (Schirmer 1979) (NE of Strasbourg on the map) yields a value of σ_{1h} of 5.4 MPa which is about twice the background value. This area is a site where the course of the graben takes on a more E–W

Table 1. Test sites and results

Location	Latitude	Longitude	Direction σ_1	Magnitude σ_1	Magnitude σ_2	Reference	Lithology
1 Gressenich	50°46′	6°18′	146°	−1.3	−4.5	this paper	
2 Stromberg	49°57′	7°46′	150°	0.2	−0.2	this paper	L. S. Mid. Devonian
3 Hanstätten	50°19′	8°04′	150°	0.02	−0.26	Greiner (pers. comm.)	L.S. Mid. Devonian
4 Fachingen	50°22′	7°54′	160°	−0.01	−0.3	Greiner (pers. comm.)	L.S. Mid. Devonian
5 Villmar	50°24′	8°11′	145°	−0.8	−1.7	Illies & Greiner (1979)	L.S. Mid. Devonian
6 Wirbelau	50°26′	8°13′	150°	−1.3	−2.3	Illies & Greiner (1979)	L.S. Mid. Devonian
7 Wetzlar	50°33′	8°38′	136°	1.3	0.9	Elmohandes (pers. comm.)	L.S. Mid. Devonian
8 Baden-Baden	48°42′	8°15′	138°	5.4	N.A.	Schirmer (1979)	Rhy. L. Permian
9 Freudenstadt	48°27′	8°29′	170°	×	×	this paper	S.S.U. Buntsandstein
10 Kaiserstuhl	48°09′	7°45′	78°	2.4	1.1	Leopoldt (1979)	L.S. Bajocian
11 Massangis*	47°38′	3°58′	167°	2.4	−0.2	Paquin *et al.* (1979)	L.S. Jurassic
12 Ravières*	47°43′	4°13′	152°	2.2	0.5	Paquin *et al.* (1979)	L.S. Jurassic
13 Etrochey*	47°53′	4°31′	149°	1.2	0.03	Paquin *et al.* (1979)	L.S. Jurassic
14 Choignes*	48°07′	5°10′	148°	1.1	−0.19	Paquin *et al.* (1979)	L.S. Jurassic

Numbers refer to points in Fig. 2. * indicates that measurements were made by flat jacks, all others were made by doorstoppers. Magnitudes of stresses are given in MPa. Magnitude data for point 9 not reliable due to pore pressure strains. Other symbols: L.S., limestone; Rhy, rhyolite and S.S.U., sandstone.

trend, thus becoming an area of compressional shear. Two points in the graben, the Kaiserstuhl (Leopoldt 1979), south of Strasbourg, and Albersweiter (Illies & Greiner 1979), south west of Heidleberg show significantly deviations from the general 140° trend. They are probably related to the complicated behaviour of the graben blocks when subjected to shear.

The Rhenish Shield

The Rhenish Shield (approximately the area on the map bordered by Luxemburg, Bonn, Kassel, & Frankfurt) shows an internal stress zonation. The seismicly active axis connecting the northern end of the Rhinegraben with the rifting of the Lower Rhine Embayment is a zone of very low, usually negative, stresses. The direction of σ_{1h} in this area trends about 150°. On the eastern and western flanks the magnitude of stress is considerably higher, about equal to one, that is slightly greater than the values from the blocks adjacent to the Rhinegraben. The direction of σ_{1h} is also the about 140°, that is it conforms to the 'regional trend'.

The Rhenish Shield is an area of Quaternary uplift and volcanism. Pliocene river terraces have been raised to 300 m above the level of the present-day Rhine and Mosel rivers. The seismic axis is often considered to be a zone of extension where true rifting is hindered because of plastic behaviour in the upper crust of this unit (Illies *et al.* 1979).

Hence, although the cause for the low stress is not precisely known, it is possibly related to crustal extension, which is not expressed as a rupture, but as a plastic, crustal attenuation. The tensile stresses are not likely to exist at great depths as the overburden pressure would counteract them. The meaning of the slight directional change is also not perfectly clear. However, the intense folding of the Hercynian orogen imparted a high degree of mechanical anisotropy upon the region and under the extensile regime the effect of the anisotropy may be amplified, and this alone could cause the slight directional deviation.

Interpretation of stress data, when tensile stresses are measured, requires a great deal of caution. In all techniques of estimating stress the conversion of the measured data into a calculated stress depends upon the assumption that the material is behaving as a continuum. Rock, at best, behaves in a non-ideal manner, but in judiciously selected sites under moderate compressive stresses the continuum assumption may be made. This is not necessarily true when the rock is subjected to tension. In this case the pre-existing fractures become an important factor during deformation.

An example of such a problem is given in Fig. 3. The values of the measurements from the Limburg basin (Illies & Greiner 1979, Greiner pers. comm.) (points 3, 4 5 and 6 in Fig. 2) are plotted as σ_{1h} vs σ_{2h}. The sites are separated by a maximum distance of about 30 km. The direction of σ_{1h} from the four sites is practically identical, yet the magnitude varies far beyond what could be considered reasonable for such a small, tectonicly quiet, area. I tentatively interpret this to be the result of some sort of yielding of the rocks under tension, and thus the value of the measured stress is a function of local material behaviour. Field observations support this model. The quarries of Hanstätten and Fachingen show a relatively high fracture density with open fissures. These sites show the stresses closest to zero. The quarries at Villmar and Wirbelau are almost fracture free and have both been worked for monumental stone. They show the most negative stresses. *In situ* determined moduli of elasticity (Greiner pers. comm.) show an analogous relationship. The relatively unfractured rocks show an average *E*-

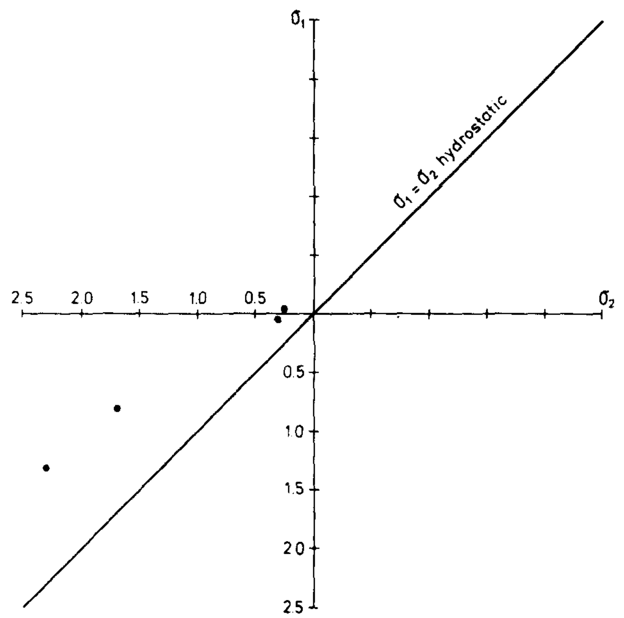

Fig. 3. Stress values from the Limburg basin northwest of Frankfurt points 3, 4, 5 and 6 on Fig. 2 are plotted as σ_{1h} vs. σ_{2h}.

modulus of 32 × 10^3 MPa, and the fractured works values of 10 × 10^3 MPa. For similar strains, the stresses would reflect the differences in E. This model of lithologically controlled stresses in the Limburg Basin illustrates a point that can also apply on a regional scale. Wherever inelastic processes occur, the measured stresses will be a function of complex geological processes.

CONCLUSIONS

By proper selection of sites, standard engineering *in situ* stress measurement techniques can be applied to problems concerning not only the state of stress but also the state of active regional stress. Attention must be paid to extrinsic and intrinsic variables which can result in a false estimate of the stress field.

Factors such as topographic loading, stress concentrations related to the shape of the landscape, and strain release due to lithology or microstructure and degree of weathering will result in an error when the *in situ* stress field is interpreted in terms of a regional stress field. Secondary stresses or apparent stresses due to temperature, residual, or water induced effects will result in a false estimate in the measured stress field.

The mapped stress field of Western Europe from the Alps through their northern foreland to the southern border of the Lower Rhine Embayment shows that the average direction of σ_{1h} is about northwest. Local variations in the magnitude and direction of the stress field reflect the active tectonic processes in the region. The agreement of the observed stress field and local geological processes warrants both a supportive and cautionary note. *In situ* stress measurements are indeed an extremely useful tool in geological investigations. However, given the very close correlations between the measured values and crustal processes of an apparent sub-regional nature, such as post-orogenic uplift or shear on pre-existent faults, the conversion of that data for discussions of large scale stress fields should be done with caution.

Acknowledgements—I gratefully acknowledge the support of the Deutsche Forschungsgemeinschaft which financed this and other research while I was at the Geologisches Institut, Universität Karlsruhe. Prof. J. H. Illies first introduced me to stress measurements and constantly guided and stimulated me through my two years in Karlsruhe. Frau Illies cheerfully hosted with unending hospitality countless nights of geological discussions. Gerhard Greiner did most of the developmental work on the application of the doorstopper method to tectonic stress measurements in Germany. He introduced me to the practice of the technique. S. E. Elmohandes has assisted on virtually every field measurement done by the Karlsruhe group. It is hard to believe that the measurements would have been possible without his scientific and moral support.

REFERENCES

Ahorner, L. 1978. Horizontal compressive crustal stresses in Central Europe. In: *Alps, Apennines, Hellenides* (edited by Closs, H., Roeder. D. & Schmidt, K.) Inter-Union Commission on Geodynamics Scientific Report No. 38, Schweizerbartsche Verlagsbuchhandlung, Stuttgart, 17–19.

Ahorner, L. 1975. Present-day stress field and seismotectonic block movements along major fault zones in Central Europe. *Tectonophysics* **29**, 233–249.

Bonnechere, F. & Fairhurst, S. 1968. Determination of the regional stress field from "doorstopper" measurements: *Jl. S. Afr. Instn. Min. Metall.* **68**, 520–544.

Carslaw, H. S. & Jaeger, J. C. 1959. *Conduction of Heat in Solids.* Clarendon Press, London.

Crouch, S. L. 1969. A note on the stress concentrations at the bottom of a flat-ended borehole: *Jl. S. Afr. Instn. Min Metall.* **70**, 100–102.

Engelder, J. T. & Sbar, M. L. 1976. Evidence for uniform strain orientation in the Potsdam Sandstone, Northern New York, from *in situ* measurements: *J. geophys. Res.* **81**, 3013–3017.

Friedman, L. 1972. Residual elastic strain in rocks. *Tectonophysics* **15**, 297–330.

Greiner, G. 1978. Spannungen in der Erdkruste—Bestimmung und Interpretation am Beispiel von in situ-Messungen im Suddeutschen Raum. Doctoral thesis, Universität Karlsruhe.

Gubler, E. 1976. Beitrag des Landesnivellements zur Bestimmung vertikaler Krustenbewegungen in der Gotthard-Region. *Schweiz. miner. petrogr. Mitt.* **56**, 675–678.

Holzhausen, G. & Johnson, A. 1979. On residual stress in rocks. *Tectonophysics* **58**, 237–267.

Hooker, V. E., Bickel, D. L. & Aggson, J. J. 1972. *In situ* determination of stresses in mountainous topography *U.S. Bur. Mines* **7654**, 1–12.

Hooker, V. E. & Du Vall, W. I. 1971. *In situ* rock temperature. Stress investigations in rock quarries. *U. S. Bur. Mines* **7589**, 1–12.

Illies, J. H. 1977. Ancient and Recent rifting in the Rhinegraben. *Geologie Mijnb.* **56**, 329–350.

Illies, J. H. & Greiner, G. 1979. Holocene movements and state of stress in the Rhinegraben rift system. *Tectonophysics* **52**, 349–359.

Illies, J. H. & Greiner, G. 1978. Rhinegraben and the Alpine system. *Bull geol. Soc. Am.* **89**, 770–782.

Illies, J. H., Prodehl, C., Schmincke, H.-U. & Semmel, A. 1979. The Quaternary uplift of the Rhenish shield in Germany. *Tectonophysics* **61**, 197–225.

Jäckli, H. L. A. 1965. Pleistocene glaciation of the Swiss Alps and signs of postglacial differential uplift. *Spec. Pap. geol. Soc. Am.* **84**, 153–157.

Leeman, E. R. 1970. Experience throughout the world with the C.S.I.R. "doorstopper" rock stress measuring equipment *Proc. 2nd Cong. Int. Soc. Rock Mech.* **3**, 419–425.

Leopoldt, W. 1979. *In situ* Spannungsmessungen am Michaelsberg bei Riegel, Nordöstlicher Kaiserstuhl, Südwestdeutschland. *Oberrhein. geol. Abh.* **28**, 17–28.

McGarr, A., & Gay, N. C. 1978. State of stress in the Earth's crust. *A. Rev. Earth Planet. Sci.* **6**, 405–436.

Nichols, T. C. 1975. Deformations associated with relaxation of residual stresses in a sample of Barre Granite from Vermont. *Prof. Pap. U.S.*

geol. Surv. **875,** 1–32.

Paquin, C., Froidevaux, C. & Souriau, M. 1978. Mesures directes des contraintes tectoniques en France septentrionale. *Bull. Soc. geol. Fr.* 7, *Ser.* **20,** 727–731.

Sbar, M. L. & Sykes, L. R. 1973. Contemporary compressive stress and seismicity in Eastern North America: An example of Intra-Plate Tectonics, *Bull. geol. Soc. Am.* **84,** 1861–1882.

Schirmer, P. 1979. *In situ* Spannungsmessungen in Baden-Baden. *Oberrhein. geol. Abh.* **28,** 7–15.

Sturgul, J. J., Scheidegger, A. E. & Grinsphan, 1976. Finite-element model of a mountain massif. *Geology* **4,** 439–442.

Trümpy, R. 1973. The timing of orogenic events in the Central Alps: In: *Gravity and Tectonics* (edoted by De Jong, K. A. & Schotten, R.) Wiley, New York, 229–225.

Van Heerden, W. L. 1968. The effect of end of borehole configuration and stress level on stress measurements using "doorstoppers". *Rep. Coun. scient. ind. Res. S. Afr. Meg.* **626.**