# The Stress Field in the Upper Crust as Determined from In Situ Measurements

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With 3 figures and 2 tables


## Zusammenfassung

Dieser Beitrag behandelt die Spannungsverhältnisse in der oberen Erdrinde, so wie sie sich bei Messungen von lokalen Spannungsfeldern herausstellen.

Die Streuung der Daten ist zwar beträchtlich, aber es zeichnen sich zwei Hauptlinien ab in den Beziehungen zwischen durchschnittlicher Horizontalspannung und Tiefe. In alten Grundgebirgen und in den deformierten Gesteinen der Faltungsgebiete findet man im allgemeinen Horizontalspannungen, die größer sind als die berechnete Belastung. In Sedimentärgesteinen und zerbrochenen massiven Gesteinen sind die Horizontalspannungen meistens geringer als die Belastung. In den meisten Fällen findet man eine unverkennbare Spannungsanisotropie in der Horizontalebene. In den Gebieten, für die wir über genügende Messungen verfügen (d. h. für Nordamerika und Fennoskandia) stellt es sich heraus, daß die Orientierung der maximalen horizontalen Hauptspannung ziemlich eindeutig und einheitlich ist, obgleich diese Orientierung sich in den meisten Fällen nicht aus einfachen, nach bestimmten Regeln zu berechnenden Spannungsverteilungen voraussagen läßt.

Die Deutung von örtlich bestimmten Spannungsverhältnissen wird von vielen Faktoren beeinträchtigt. Die Beziehung zwischen der Verteilung der Hauptspannungen und der Position der heutigen Oberfläche ist äußerst kompliziert. Die topographische Beschaffenheit und Abtragungsvorgänge dürften eine Anhäufung von Horizontalspannungen zur Folge haben. Sehr alte Restspannungen überlagern manchmal die Spannungen der rezenten Tektonik, während anderseits die heutige Spannungslage nicht mehr mit den Spannungsfeldern übereinstimmt, die zu den heute zu beobachtenden Bruch- und Faltungssystemen geführt haben. Messungen von Restspannungen in alten Gesteinen haben bewiesen, daß die Gesteine der oberen Erdrinde, sogar über geologische Zeitspannen heraus, eine gewisse Deformationsresistenz haben.

Vom Gesichtspunkt der Globaltektonik ist es zu empfehlen, alle *in-situ*-Spannungsbestimmungen nur mit größter Vorsicht zu verwerten, und auch dann nur unter Berücksichtigung der Bruchflächenlösungen der seismischen Daten. Das Spannungsfeld in der Erdrinde ist zwar im allgemeinen kompressiv, aber die Feststellung, daß undeformierte Sedimentdecken meistens keine horizontale Überspannung aufweisen, macht deutlich, daß die Annahme einer weltumfassenden horizontalen Kompression unnötig ist. Wo es in Gegenden, die einst heftigen Gebirgsbildungsvorgängen unterworfen waren, Horizontalspannungen gibt, die größer sind als der berechnete lithostatische Druck, so darf dies wohl zurückgeführt werden auf Restspannungen und auf den Einfluß der örtlichen Topographie und Struktur. Es scheint daher voreilig, auf lokalen Spannungsbestimmungen eine universelle tektogenetische Hypothese zu gründen. Um wirkliche Fortschritte in dieser Hinsicht zu erzielen, müssen weitere Messungen und eine quantitative Bewertung der jene Messungen beeinflussenden Faktoren durchgeführt werden.

## Abstract

This paper provides a synopsis of the state of stress in the upper parts of the earth's crust based upon *in situ* rock stress determinations.


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Despite the large scatter of the data, two dominant trends can be detected in the variations of average horizontal stress with depth in various geological environments. Basement rocks in ancient shields and deformed rocks in fold belts usually show horizontal stresses larger than the theoretical overburden pressure. Sedimentary cover rocks and fissured massive rocks show horizontal stresses smaller than the overburden pressure. The ratio of the maximum to the minimum horizontal stress exhibits a clear stress anisotropy in most cases. Directions of maximum horizontal compression are fairly consistent in areas where sufficient measurements are available (North America and Fennoscandia), although in many instances they do not conform to any simple predicted stress pattern.

Many factors complicate the interpretation of *in situ* stress determinations. There is no simple relationship between the stress trajectories and the free surface. Topographic features and erosional processes may cause horizontal stress concentrations. Remanent stresses of great age can be superimposed on current tectonic stresses, while sometimes current stresses no longer coincide with the stress systems that caused observable faulting and folding. Observation of remanent stresses in ancient rocks shows that rocks in the upper crust have finite strength even under geological time intervals.

From the viewpoint of global tectonics, *in situ* stress determinations ought to be used with great caution, and in conjunction with focal mechanism solutions of earthquakes. Although the state of stress is everywhere compressive, the fact that undeformed sedimentary cover rocks often show no excess horizontal stress would seem to indicate that no active global horizontal compression is required. Horizontal stresses larger than the overburden pressure in regions of intense palaeodeformation may be due to remanent stress effects and to the influence of the local structure. It is premature to advance any general statement on tectogenesis on the basis of *in situ* stress determinations. More measurements, and a quantitative evaluation of the factors affecting them, are required before further progress can be made.

## Résumé

Cette étude est une description de l'état des contraintes das les couches supérieures de l'écorce terrestre et est basée sur des déterminations *in situ* de contraintes dans les roches.

En dépit de la dispersion de données, on remarque deux tendances principales dans les variations des contraintes horizontales moyennes par rapport à la profondeur dans des milieux géologiques divers. Les roches cristallines des massifs anciens et les roches déformées des chaînes plissées montrent habituellement des contraintes horizontales supérieures à la pression causée par la pesanteur des roches superposées. Les roches de couches sédimentaires et les masses de roches fissurées montrent des contraintes horizontales inférieures à cette pression. Dans la plupart des cas, le rapport entre le maximum et le minimum de contraintes horizontales met en relief une anisotropie des contraintes. Les directions des compressions horizontales maximales sont assez régulières dans les regions où des mesures adéquates sont possibles (Amerique du Nord et Fennoscandie) bien que très souvent, elles ne sont pas conformes à aucune configuration théorique.

De nombreux facteurs viennent compliquer l'interprétation des déterminations de contraintes *in situ*. In n'y a aucune relation simple entre les trajectoires des contraintes principales et la surface libre. Les caractéristiques topographiques et les processus d'érosion peuvent engendrer des concentrations de contraintes horizontales. Des contraintes résiduelles très anciennes peuvent se superimposer à des contraintes tectoniques actuelles, alors que parfois les contraintes actuelles ne coincident plus avec les systèmes de pression qui sont à l'origine des phénomènes observables de plis et de failles. L'observation des contraintes résiduelles dans les roches anciennes montre que les

roches de la couche supérieure de l'écorce terrestre ont une résistance à la deformation même pendant des longues périodes géologiques.

D'un point de vue tectonique global, les déterminations de contraintes *in situ* ne doivent être utilisées qu'avec une grande prudence et seulement en conjonction avec les solutions des mécanismes focaux des secousses séismiques. Bien que l état de contraintes soit partout compressif, le fait qu'une couverture de roche sédimentaires non-déformée ne montre souvent aucun excès de pression horizontale semblerait indiquer qu'aucune compression horizontale globale n'est nécessaire. L'excès de contraintes horizontales dans les régions de déformation intense pourrait être causé par les effets des pressions résiduelles et l'influence de la structure locale. Il serait prématuré d'avancer une théorie générale de la tectogénèse sur la base de déterminations de contraintes *in situ*. Des mesures additionelles et une évaluation quantitative des facteurs qui les affectent seront nécessaires avant qu'un quelconque progrès puisse être réalisé.

**Краткое содержание**

В данном опубликовании приведены соотношения напряжений верхней Земной коры, полученные при измерении местных полей напряжения.

Рассеяние данных велико, но все же вырисовываются две основных линии взаимоотношений между средним напряжением по горизонтали и в глубину. В древних коренных горах и в деформированных породах складчатых гор установлено в общем напряжение по горизонтали, превосходящее подсчитанную нагрузку. В осадочных породах и породах разломанных массивов напряжение по горизонтали обычно ниже нагрузки. В большенстве случаев на горизонтальных равнинах наблюдают явную изотропию напряжения. В областях, где было проведено достаточное количество измерений (напр.: в Северной Америки и Фенноскандии) установили, что ориентация результирующего максимального напряжения по горизонтали достаточно явна и едина, хотя ее в большенстве случаев нельзя предвидеть на основании рассчетов распределения напряжений.

Интерпретация соотношения определенных местных напряжений зависит от многих факторов. Связь между распределением результирующего напряжения и расположением современной поверхности чрезвычайно сложна. Топографические строение и процессы сноса могут привести к накоплению напряжений по горизонтали. Древние остаточные напряжения перекрыты порой напряжениями современной тектоники, в то время, как положение напряжения сегодня не совпадает с полями напряжения, приведшими к наблюдаемым сегодня системам разломов и складок. Измерения остаточных напряжений в древних породах доказали, что породы верхней земной коры проявляют известное сопротивление деформации даже в течение геологических эпох.

С точки зрения мировой тектоники желательно все полученные на месте данные напряжения интерпртировать с большой осторожностью, учитывая прохождние сейсмических волн по границам отдельных сред. Напряжение в земной коре в общем проходит на сдавление, однако установленный факт того, что недеформированные осадочные покровы в большенстве случаев не проявляют горизонтального перенапряжения, явно говорит об отсутствии необходимости принимать теорию горизонтального сдавления, охватывающего весь земной шар.

В областях, где когда-то проходили сильные горообразовательные процессы и установлены горизонтальные напряжения, превышающие подсчитанное изостатическое давление, это повышение можно отнести за счет остаточного напряжения и влияния местной топографии и местных структур. Поэтому кажется преждевременным по данным определения местного напряжения создавать универсальную тектогенетическую гипотезу. Чтобы иметь успех в этом смысле, необходимо проводить дальнейшие измерения с количественной оценкой факторов, влияющих на каждое отдельное измерение.

## Introduction

The methods that are most frequently used in the determination of the present-day stress field in the earth's crust can be grouped as follows: (i) study of recent crustal movements, (ii) focal mechanism solutions of earthquakes, and (iii) *in situ* rock stress measurements. Recent crustal movements can be studied by means of geologic, geodetic, and geomorphologic observations. It is often impossible, however, to correlate uniquely an inferred displacement with the causative stress field; moreover, there is no theoretical basis to support extrapolation in time of the deduced stress field. Focal mechanism solutions of earthquakes, on the other hand, permit the determination of the current stress field in the seismic areas, subject only to the assuptions governing the laws of seismic wave propagation. But vast regions of the earth show small or no seismic activity, and consequently the crustal stress field in these areas cannot be determined solely from seismic studies.

The *in situ* measurement of rock stress has been used in the solution of problems in civil engineering, mining, and geotectonics. Several techniques are available (*cf.* JAEGER & COOK, 1969, for a review). Measurements are usually carried out in boreholes from the surface or from a mine wall, some at depths of more than 2000 m. In some cases the complete stress tensor is determined, but most frequently only the direction and magnitude of the principal horizontal stresses, or their average, and the vertical stress. Measurements are affected by many factors (distance from a free surface, assumed moduli of elasticity of the rock, instrumental limitations), and errors could be as high as 20—40% for principal stress magnitudes, and 5—20% for directions (BULIN, 1971). This should be kept in mind when interpreting the results.

Some confusion exists on terminology. In this paper, we adopt the following terms, a combination of those proposed by VOIGHT (1966 b), JAEGER & COOK (1969), and EISBACHER & BIELENSTEIN (1971). Natural stresses are the stresses present in the rock before excavation and drilling. They are composed of gravitational stresses, due to the weight of the overburden, and tectonic stresses (when present), usually acting in a horizontal or sub-horizontal direction. The latter are in turn composed of current tectonic stresses, related to present-day straining of the earth's crust, and remanent tectonic stresses, due to elastic strains that were locked in the rock at some stage of its history and have not been completely relieved during subsequent geologic events. Induced stresses are present in the rock as a result of artificial excavation and drilling. Their effects must be eliminated when determining the *in situ* state of stress for geological purposes, and this is usually accomplished by taking the measurement sufficiently removed from the excavation, or by applying correction factors based upon elasticity theory. Finally, residual stresses are self-equilibrating stresses that remain in the rock when all external loads are removed. Thus defined, they cannot be measured by standard *in situ* methods and require special techniques such as X-ray diffraction (FRIEDMAN, 1972), and are part of, but not the same as, the remanent tectonic stress.

The state of stress in a rock mass results from the superposition of stresses of various kinds as specified above. Boundary conditions also greatly affect the state of stress at any given point. Surface irregularities (valleys, hills, steep slopes), structural features (mechanical and thermal rock inhomogeneities, faults, folds), and erosional and depositional processes can alter the regional state of stress by causing stress concentrations and changing the orientations of the principal axes.

Knowledge of the natural stress field in the earth's crust is of great importance in the solutions of problems in geodynamics. As Voight et al. (1969) have pointed out, near-surface stress magnitudes and trajectories can help in establishing the type of large-scale stress fields acting on lithospheric plates, and therefore can be used, together with earthquake focal mechanisms and recent crustal movements to investigate the driving forces of global tectonics. The purpose of this paper is to analyse reliable *in situ* stress determinations and to discuss their relevance to the problem of tectogenesis. In order to achieve this, we have examined all *in situ* determinations available to date and konwn to us, retaining only those which appeared not to be greatly influenced by induced stresses or by limitations of measurement techniques. Attention is focused on horizontal stresses, and SI units are used throughout [1]).

Similar studies, based on the data available at the time, have been carried out, among others, by Hast (1969), Voight et al. (1969), Bulin (1971), Kropotkin (1972), and Turchaninov et al. (1972). Besides updating their results, we have done a retrospective search that has led to what we think is a fairly complete collation of data. Building upon their analysis, we study the variation of average horizontal stress with depth in various geologic environments, calculate the ratio of maximum to minimum normal horizontal stress, and discuss the directions of maximum horizontal stress.

## Variations of average horizontal stress with depth

In this paper, $\sigma_x$ and $\sigma_y$ are the maximum and minimum horizontal stresses, respectively (compression is taken positive). In over 60% of the measurements, one of the principal axes forms an angle of less than 30° with the vertical. In these cases, we have taken the two subhorizontal principal stresses to be equal to $\sigma_x$ and $\sigma_y$. For the purpose of comparison we have calculated the theoretical overburden (lithostatic) pressure

$$p = \varrho g z$$

for g (acceleration of gravity) $= 9.81\ \mathrm{m\ s^{-2}}$ and $\varrho$ (density) $= 2700\ \mathrm{kg\ m^{-3}}$. It should be noted that $\sigma_z$ (the observed vertical stress) in many cases does not coincide with the overburden pressure p. We shall examine the variation of $\sigma_z$ with depth and geologic environment in another paper. The difference between p and $\sigma_z$ has been studied theoretically by Howard (1966).

[1]) The adoption of Système International (SI) units has been recommended by several international scientific bodies and is finding increasing acceptance in the earth sciences. For the convenience of the reader, we give some conversion factors between SI units and units frequently used in rock stress determinations:

SI unit of stress: newton/metre² ($\mathrm{N\ m^{-2}}$)

1 bar $= 10^6$ dyn $\mathrm{cm^{-2}} = 14.5$ psi $= 10^5\ \mathrm{N\ m^{-2}}$

1 kgf $\mathrm{cm^{-2}} = 0.98$ bar $= 0.98 \cdot 10^5\ \mathrm{N\ m^{-2}}$

We consistently use the multiple meganewton/metre² ($\mathrm{MN\ m^{-2}}$) which is equal to $10^6\ \mathrm{N\ m^{-2}}$, that is, 10 bar or 10.2 kgf $\mathrm{cm^{-2}}$.

The determinations of average horizontal stress $(\sigma_x + \sigma_y)/2$ collated by us are indicated in table 1. The table gives, from left to right, measurement number (for reference to fig. 1), locality, average horizontal stress, depth, reference number (cf. appendix A), rock type in which the measurement was taken (when available), and tectonic environment.

The data are shown graphically in fig. 1, in which average horizontal stress is plotted versus the depth z. The measurements are represented by symbols expressing different tectonic zones. A five-fold subdivision has been adopted, namely, shields, Palaeozoic fold belts, Mesozoic and Cenozoic fold belts, sedimentary covers on platforms, and rift zones (Iceland). There are more than 150 rock stress determinations from North America, Fennoscandia, Eurasia, Africa, and Australia. The line OC represents the theoretical overburden pressure for $\varrho = 2700$ kg $m^{-3}$. Measurements to the right of this line exhibit "excess" horizontal stress. Measurements falling approximately along this line indicate a lithostatic state of stress. The line AB is the empiral relationship derived by HAST (1969) to describe results obtained in Fennoscandia and other parts of the world

$$(\sigma_x + \sigma_y)/2 = 9.31 + 0.05\,z\ (\mathrm{MN\,m^{-2}}).$$

Although the scatter is very large, the measurements clustering along this line are in their majority from shield areas and Palaeozoic folded belts with some from post-Palaeozoic folding. Only three measurements are from sedimentary cover rocks.

The line DE, differing only slightly from that proposed originally by BULIN (1971) for the sedimentary cover of platforms, can be represented by the equation

$$(\sigma_x + \sigma_y)/2 = 2.50 + 0.013\,z\ (\mathrm{MN\,m^{-2}}).$$

The measurements clustering along it represent localities where the average horizontal stress, except at very shallow depths, is less than the lithostatic pressure. The majority of these measurements are from the sedimentary mantle in the platforms and from fissured massive rocks of Palaeozoic fold belts, with some from shield and Palaeozoic areas of the U.S.S.R. and Fennoscandia, Mesozoic folding, and rift zones (Iceland). At depths approximately less than 500 m the results are widely scattered and many are close to the hydrostatic case. At larger depths, however, the deficiency in horizontal stress becomes apparent.

In general terms, shields and Palaeozoic fold belts are the main areas of excess horizontal stresses. However, many measurements from Palaeozoic fold belts, along with Mesozoic and Cenozoic fold belts, are widely scattered. The Fennoscandian and Canadian shields appear to be the regions in which bedrock stresses follow HAST's relationship most closely. Sedimentary overburden areas, undeformed or only slightly deformed, show typically less than lithostatic horizontal stress. However, the maximum and minimum horizontal stresses are appreciably different in most cases, showing that the stress field in these areas also has a current or remanent tectonic component. A tectonic stress field appears to exist in pratically all areas sampled so far by *in situ* rock stress measurements, even if these areas are not at present tectonically "active" in the geological and seismological sense.

Table 1: Average horizontal stress *vs.* depth.
(Information on rock type given only when available; for full references, *cf.* appendix A.)

| Measurement number | Locality | $(\sigma_x+\sigma_y)/2$ (MN m$^{-2}$) | z (m) | Reference number | Rock type | Tectonic environment |
|---|---|---|---|---|---|---|
| 1 | Stalldalen, Sweden | 49.0 | 900 | 18 | — | Baltic shield |
| 1a | Stalldalen, Sweden | 48.1 | 690 | 18 | — | Baltic shield |
| 2 | Hofors, Sweden | 39.7 | 650 | 18 | — | Baltic shield |
| 2a | Hofors, Sweden | 34.8 | 470 | 18 | — | Baltic shield |
| 3 | Vingesback, Sweden | 53.9 | 400 | 18 | Granite, amphibolite | Baltic shield |
| 3a | Vingesback, Sweden | 78.4 | 400 | 18 | Granite | Baltic shield |
| 3b | Vingesback, Sweden | 27.9 | 410 | 23 | Granite | Baltic shield |
| 4 | Laisvall, Sweden | 18.14 | 120 | 18 | Granite | Baltic shield |
| 4a | Laisvall, Sweden | 22.1 | 220 | 18 | Granite | Baltic shield |
| 4b | Laisvall, Sweden | 38.75 | 180 | 18 | Granite | Baltic shield |
| 5 | Malmberget, Sweden | 25.0 | 280 | 18 | Granite | Baltic shield |
| 5a | Malmberget, Sweden | 23.05 | 460 | 23 | Granite | Baltic shield |
| 6 | Halmstad, Sweden | 26.0 | 20 | 18 | — | Baltic shield |
| 7 | Hargshamn, Sweden | 21.1 | 250 | 18 | — | Baltic shield |
| 8 | Ekenas, Finland | 16.6 | 140 | 18 | — | Baltic shield |
| 9 | Kiruna, Sweden | 12.26 | 90 | 18 | — | Baltic shield |
| 9a | Kiruna, Sweden | 11.77 | 125 | 18 | — | Baltic shield |
| 9b | Kiruna, Sweden | 9.81 | 290 | 42 | — | Baltic shield |
| 10 | Jokkmokk, Sweden | 15.9 | 100 | 18 | Granite | Baltic shield |
| 11 | Nummi, Finland | 14.23 | 100 | 18 | — | Baltic shield |
| 12 | Borga, Finland | 12.75 | 50 | 18 | — | Baltic shield |
| 12a | Borga, Finland | 13.72 | 100 | 18 | — | Baltic shield |
| 13 | Kristianstad, Sweden | 12.26 | 5 | 18 | — | Baltic shield |
| 14 | Farosund, Sweden | 11.27 | 50 | 18 | — | Baltic shield |
| 15 | Kirkennes, Norway | 10.3 | 50 | 18 | — | Baltic shield |
| 16 | Idre, Sweden | 10.2 | 30 | 18 | — | Baltic shield |
| 17 | Lidingo, Sweden | 10.2 | 35 | 18 | — | Baltic shield |
| 18 | Karlshamn, Sweden | 10.3 | 10 | 18 | — | Baltic shield |
| 19 | Uddevalla, Sweden | 9.33 | 10 | 18 | — | Baltic shield |
| 20 | (Not specified) | 9.81 | 5 | 18 | — | Baltic shield |
| 21 | Tranas, Sweden | 8.4 | 8 | 18 | — | Baltic shield |
| 22 | (Not specified) | 6.87 | 20 | 18 | — | Baltic shield |
| 23 | Alma, New York | 18.5 | 512 | 34 | — | Appalachians |
| 24 | Elliot Lake, Ontario | 19.0 | 336 | 30 | Metamorphic succession, | Canadian shield |
| 24a | Elliot Lale, Ontario | 28.2 | 305 | 30 | quartzite, conglomer- | Canadian shield |
| 24b | Elliot Lake, Ontario | 29.35 | 700 | 30 | ates | Canadian shield |
| 25 | Churchill Falls Labrador | 12.36 | 305 | 21 | Gabbro, diorite | Canadian shield |
| 26 | Wawa, Ontario | 34.3 | 350 | 27 | Metamorphic | Canadian |
| 26a | Wawa, Ontario | 28.4 | 490 | 27 | tuff, meta- | shield |
| 26b | Wawa, Ontario | 29.4 | 550 | 27 | diorite, | |
| 26c | Wawa, Ontario | 32.9 | 550 | 27 | chert, and | |
| 26d | Wawa, Ontario | 41.2 | 550 | 27 | Precambian sediments | |
| 27 | North Bay, Ontario | 7.58 | 7.6 | 36 | — | Canadian shield |
| 28 | Ottawa, Ontario | 2.76 | 15.1 | 36 | — | Canadian shield |

Table 1 (continued)

| | | | | | | |
|---|---|---|---|---|---|---|
| 29 | South Africa | 2.94 | 1560 | 44, 40, 38 | — | Sedimentary cover |
| 29a | South Africa | 25.5 | 1750 | 44, 40, 38 | — | Sedimentary cover |
| 29b | South Africa | 48.0 | 1500 | 44, 40, 38 | — | Sedimentary cover |
| 30 | Liberia | 3.43 | 25 | 18, 42 | Gneiss, norite | African shield |
| 31 | South Africa | 38.2 | 2115 | 3 | — | Sedimentary cover |
| 32 | South Africa | 49.2 | 1770 | 9 | Lava | African shield |
| 33 | Carletonville, South Africa | 18.8 | 1320 | 29, 20 | Quartzite | Sedimentary cover |
| 33a | Carletonville, South Africa | 31.8 | 2320 | 29, 20 | Quartzite | Sedimentary cover |
| 34 | Boksburg, Transvaal | 27.0 | 2400 | 20 | Quartzite | Sedimentary cover |
| 35 | Harmoney Gold Mine, South Africa | 16.6 | 1500 | 20 | Quartzite | Sedimentary cover |
| 36 | Kafue Gorge, Zambia | 14.9 | 160 | 42, 18 | Granite | African shield |
| 36a | Kafue Gorge, Zambia | 22.55 | 400 | 42, 18 | Granite | African shield |
| 37 | Krivoy Rog Basin, USSR | 3.93 | 300 | 37, 38 | — | Shield USSR |
| 38 | Solikamsk, USSR | 4.42 | 260 | 38 | — | Shield USSR |
| 39 | Talnakh, USSR | 4.9 | 250 | 38 | — | Shield USSR |
| 40 | Gdov, Estonia | 1.96 | 200 | 28 | — | Shield USSR |
| 41 | Kola Peninsula, USSR | 39.25 | 105 | 37 | — | Shield USSR |
| 41a | Kola Peninsula, USSR | 45.2 | 600 | 37 | — | Shield USSR |
| 41b | Kola Peninsula, USSR | 22.5 | 100 | 37 | — | Shield USSR |
| 41c | Kola Peninsula, USSR | 25.0 | 300 | 37 | — | Shield USSR |
| 42 | Dahlen, Norway | 1.33 | 890 | 38, 41 | — | Scandinavian Caledonides |
| 43 | Lokken, Norway | 27.45 | 380 | 42 | — | Scandinavian Caledonides |
| 44 | Malm, Norway | 9.81 | 650 | 42 | — | Scandinavian Caledonides |
| 45 | Gol, Norway | 15.2 | 50 | 18, 42 | — | Scandinavian Caledonides |
| 46 | Sulitjelma, Norway | 9.71 | 850 | 19 | — | Scandinavian Caledonides |
| 46a | Sulitjelma, Norway | 5.0 | 900 | 19 | — | Scandinavian Caledonides |
| 47 | Lokken, Norway | 11.2 | 810 | 19 | Granite | Scandinavian Caledonides |
| 47a | Lokken, Norway | 13.05 | 870 | 19 | Granite | Scandinavian Caledonides |
| 48 | Rodsand, Norway | 2.94 | 320 | 19 | — | Scandinavian Caledonides |
| 48a | Rodsand, Norway | 3.73 | 620 | 19 | — | Scandinavian Caledonides |
| 48b | Rodsand, Norway | 5.88 | 700 | 19 | — | Scandinavian Caledonides |
| 49 | Mofjellet, Norway | 14.22 | 300 | 19 | — | Scandinavian Caledonides |
| 50 | Bidjovagge, Norway | 12.75 | 70 | 19 | — | Scandinavian Caledonides |
| 51 | Rock Chapel Mt., Georgia | 9.32 | 7.6 | 26 | Gneiss | Appalachians |

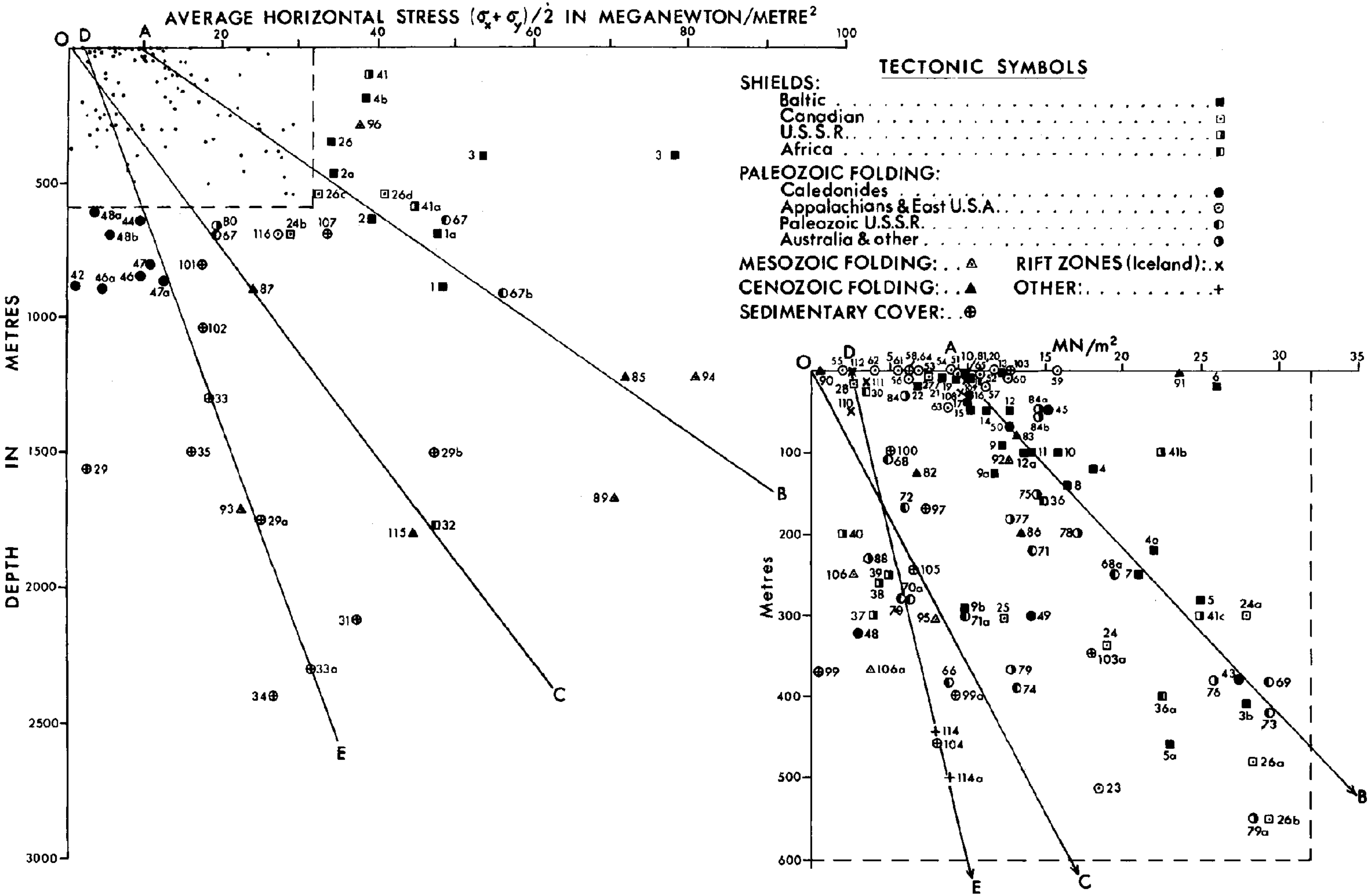


Fig. 1. Average horizontal stress *vs.* depth in various tectonic environments. Measurements numbers refer to table 1.


Fig. 2. Map of maximum horizontal stress directions. Measurement numbers refer to table 2.

Table 1 (continued)

| | | | | | | |
|---|---|---|---|---|---|---|
| 52 | Arabia Mt., Georgia | 11.7 | .2 | 26 | Gneiss | Appalachians |
| 53 | Pine Mt., Georgia | 6.7 | .2 | 26 | Gneiss | Appalachians |
| 54 | Stone Mt., Georgia | 9.03 | .6 | 26 | Granite | Appalachians |
| 55 | Douglasville, Georgia | 2.06 | .6 | 26 | Granite | Appalachians |
| 56 | Tewkesbury, Massachusetts | 6.37 | 11.6 | 25 | Paragneiss | Appalachians |
| 57 | Chelmsford, Massachusetts | 11.2 | 18.9 | 25 | Granite | Appalachians |
| 58 | Carthage, Missouri | 6.37 | 1.2 | 25 | Limestone | Sedimentary cover |
| 59 | Graniteville, Missouri | 15.8 | 1.4 | 25 | Granite | Paleozoic folding |
| 60 | Mt. Airy, N. Carolina | 12.55 | 10 | 25 | Granite | Appalachians |
| 61 | Troy, Oklahoma | 5.5 | 1.4 | 25 | Granite | Paleozoic folding |
| 62 | St. Peters, Pennsylvania | 4.02 | 1.4 | 25 | Norite | Appalachians |
| 63 | Barre, Vermont | 8.73 | 46 | 25 | Granite | Appalachians |
| 64 | Proctor, Vermont | 6.28 | .4 | 25 | Dolomite | Appalachians |
| 65 | Rapidan, Virginia | 10.78 | 2.6 | 25 | Diabase | Appalachians |
| 66 | Vsokogorsk, USSR | 8.82 | 380 | 38 | — | Paleozoic folding |
| 67 | Donetz Basin, USSR | 19.6 | 700 | 37, 38 | — | Paleozoic folding |
| 67a | Donetz, Basin USSR | 49.0 | 660 | 37, 38 | — | Paleozoic folding |
| 67b | Donetz Basin, USSR | 55.9 | 915 | 37, 38 | — | Paleozoic folding |
| 68 | Djezkazdan, USSR | 4.9 | 110 | 38 | — | Paleozoic folding |
| 68a | Djezkazdan, USSR | 19.6 | 250 | 38 | — | Paleozoic folding |
| 69 | Tashtagol, USSR | 29.4 | 380 | 38 | — | Paleozoic folding |
| 70 | Vsokogorsk, USSR | 5.88 | 280 | 38 | — | Paleozoic folding |
| 70a | Vsokogorsk, USSR | 6.37 | 280 | 38 | — | Paleozoic folding |
| 71 | Kazakhstan, USSR | 14.25 | 220 | 37, 38 | — | Paleozoic folding |
| 71a | Kazakhstan, USSR | 9.81 | 300 | 38 | — | Paleozoic folding |
| 72 | Sayan-Shushensk, USSR | 6.08 | 170 | 38 | Diabase | Paleozoic folding |
| 73 | Shoria Mts., Siberia | 29.4 | 420 | 37 | — | Paleozoic folding |
| 74 | Shoria Mts., Siberia | 13.2 | 390 | 37 | — | Paleozoic folding |
| 75 | Great Lake, Tasmania | 14.53 | 152 | 32 | Mudstones | Paleozoic folding |
| 76 | Spitzbergen | 25.8 | 380 | 42 | — | Paleozoic folding |
| 77 | New South Wales, Australia | 12.75 | 183 | 32 | Granitic gneiss | Paleozoic folding |
| 78 | Ireland | 17.15 | 200 | 42, 18 | — | Paleozoic folding |
| 79 | Cobar, N.S.W., Australia | 12.85 | 366 | 17 | — | Paleozoic folding |

Table 1 (continued)

| | | | | | | |
|---|---|---|---|---|---|---|
| 79a | Cobar, N.S.W., Australia | 28.5 | 550 | 17 | — | Paleozoic folding |
| 80 | Mt. Isa, Queensland, Australia | 19.6 | 665 | 8 | Shales, orebody | Paleozoic folding |
| 81 | Iran | 9.81 | 10 | 37 | — | Cenozoic folding |
| 82 | California | 6.77 | 125 | 38, 39 | — | Cenozoic folding |
| 83 | Alay Range, Uzbekistan, USSR | 13.25 | 80 | 37 | — | Cenozoic folding |
| 84 | Braga, Portugal | 6.08 | 30 | 18, 42 | Granite | Paleozoic folding |
| 84a | Portugal | 14.7 | 57 | 11 | Diorite | Paleozoic folding |
| 84b | Portugal | 14.7 | 52 | 2 | Diorite | Paleozoic folding |
| 85 | Mont Blanc, Alps | 72.6 | 1220 | 18 | Gneiss, granite | Alpine belt |
| 86 | Lago Maggiore, Italy | 13.55 | 200 | 1 | Paragneiss | Alpine belt |
| 87 | Ashio Mine, Japan | 24.5 | 900 | 4 | — | Cenozoic folding |
| 88 | St. Pierremont, France | 3.72 | 230 | 5 | Orebody | Peleozoic folding |
| 89 | Silver Summit Mine, Idaho | 71.2 | 1670 | 33 | Quartzite | Mesozoic folding |
| 90 | Rangely, Colorado | .49 | 1.8 | 31 | Sandstone | Cenozoic folding |
| 91 | Red Mt., Colorado | 23.51 | 9.2 | 24 | Granite | Cenozoic folding |
| 92 | Peace R., British Columbia | 12.75 | 110 | 18, 24 | — | Mesozoic folding |
| 93 | Burke, Idaho | 22.75 | 1715 | 16 | Ore, quartzite | Mesozoic folding |
| 94 | Wallace, Idaho | 81.6 | 1220 | 15 | Ore, quartzite | Mesozoic folding |
| 95 | Woh, Malaya | 7.95 | 305 | 14 | Granite | Mesozoic folding |
| 96 | Jor Power Station, Malaya | 38.25 | 290 | 13 | Biotite, granite | Mesozoic folding |
| 97 | Sigma Coal Mines, South Africa | 7.35 | 170 | 40 | — | Sedimentary cover |
| 99 | Luov-Volyn Coal Basin, USSR | .49 | 370 | 38 | — | Sedimentary cover |
| 99a | Luov-Volyn Coal Basin, USSR | 9.33 | 400 | 38 | — | Sedimentary cover |
| 100 | Illinois | 5.1 | 99 | 6 | — | Sedimentary cover |
| 101 | Ohio | 17.65 | 809 | 6 | — | Sedimentary cover |
| 102 | Allan, Saskatchewan | 17.92 | 1040 | 7 | Sylvanite potash | Sedimentary cover |
| 103 | Marble Falls, Texas | 12.75 | 1.4 | 25 | Granite | Sedimentary cover |
| 103a | Marble Falls, Texas | 17.95 | 346 | 12 | Granite | Sedimentary cover |
| 104 | Westvaco, Wyoming | 8.04 | 458 | 10 | Mineralized zone | Sedimentary cover |

Table 1 (continued)

| | | | | | | |
|---|---|---|---|---|---|---|
| 105 | Winnfield, Louisiana | 6.57 | 244 | 10 | Salt | Sedimentary cover |
| 106 | AEC Test site, Nevada | 2.75 | 250 | 10 | Granite | Mesozoic folding |
| 106a | AEC Test site, Nevada | 3.82 | 366 | 10 | Tuff | Mesozoic folding |
| 107 | Barberton, Ohio | 34.15 | 702 | 10 | Limestone | Sedimentary cover |
| 108 | Stokkenes, Iceland | 9.81 | 30 | 18, 42 | Igneous rock | Near rift zone |
| 109 | Hvalnes, Iceland | 9.81 | 10 | 18, 42 | Igneous rock | Near rift zone |
| 110 | Burfell, Iceland | 2.45 | 50 | 18 | Basalt | Near rift zone |
| 111 | Akureyri, Iceland | 3.43 | 10 | 18, 42 | Basalt | Near rift zone |
| 112 | Keflavik, Iceland | 2.45 | 2 | 18, 42 | Basalt | Near rift zone |
| 113 | (Not specified) | 4.42 | 15 | 18 | — | — |
| 114 | W. Germany | 7.84 | 440 | 38 | — | — |
| 114a | W. Germany | 8.83 | 500 | 38 | — | — |
| 115 | Rangely, Colorado | 45.03 | 1800 | 43 | — | Cenozoic folding |
| 116 | Morgantown, Pennsylvania | 27.5 | 700 | 35 | Diabase | Appalachians |

## Directions of maximum horizontal stress

The magnitudes of the horizontal stresses and the direction of the maximum horizontal compression included in this study are shown in table 2. The table lists, from left to right, measurement number (with reference to figs. 2 and 3), locality, $\sigma_x$ and $\sigma_y$ (where available), depth, direction of $\sigma_x$, reference number (cf. appendix B) and rock type in which the measurement was taken. The tectonic environment can be derived from the map in fig. 2.

Stress orientations available to date are considerably less numerous than stress magnitude determinations. With the possible exceptions of North America and Fennoscandia, many more measurements are necessary before meaningful conclusions can be drawn as to the principal horizontal stress directions over large areas. Fig. 2 gives an overall view of the situation. All the measurements listed in table 2 are plotted on it, although some directions on the map correspond to an average of two or more measurements at closely spaced stations, or to measurements at the same locality but at different depths. Incidentally, elastic analysis of lithospheric blocks subject to various boundary stresses (HAFNER, 1951; SANFORD, 1959) shows that stress trajectories are usually inclined with respect to the block's boundaries, and that they vary their orientation with depth. Therefore, as VOIGHT et al. (1969) have pointed out, an analysis of principal horizontal stress directions is strictly valid only if measurements have been carried out at the same depth below the surface. However, at a qualitative level a map such as that presented in fig. 2 gives an idea of near-surface maximum horizontal stress directions in the areas where they have been measured.

The possibility of error in the identification of maximum and minimum horizontal stresses must not be overlooked, especially when the ratio $\sigma_x/\sigma_y$ is close to unity. Fig. 3 shows this ratio (which we term the "horizontal stress anisotropy" or HSA ratio) for the measurements listed in table 2 (except a few deep ones). The HSA ratio varies widely, in some cases being close to unity and in others

Table 2: Principal horizontal stresses and maximum stress direction. (Information given only when available; for full references, *cf.* appendix B.)

| Measurement number | Locality | $\sigma_x$ ($MNm^{-2}$) | $\sigma_y$ ($MNm^{-2}$) | z (m) | Direction | Reference number | Rock Type |
|---|---|---|---|---|---|---|---|
| 1 | Silver Summit Mine, Idaho | 105.0 | 37.5 | 1670 | N 25 E | 1 | Quartzite |
| 2 | Great Lake, Tasmania | 16.6 | 12.4 | 152 | N 45 W | 2 | Mudstones |
| 3 | Rangely Anticline, Colorado | 1.34 | .36 | 1.8 | N 73 E | 3 | Sandstone |
| 4a | Elliot Lake, Ontario | 20.6 | 17.3 | 300–400 | N 90 E | 4 | Metamorphic succession, |
| 4b | Elliot Lake, Ontario | 36.4 | 20.0 | 305 | N 45 E | 4 | quartzite conglomerates |
| 4c | Elliot Lake, Ontario | 36.4 | 22.1 | 700 | N 90 E | 4 | |
| 5a | Carletonville, South Africa | 22.1 | 15.5 | 1320 | N 75 W | 5 | Quartzite |
| 5b | Carletonville, South Africa | 33.6 | 30.0 | 2320 | N 30 E | 5 | Quartzite |
| 6a | Kafue Gorge, Zambia | 16.7 | 13.3 | 160 | N 70 W | 6 | Granite |
| 6b | Kafue Gorge, Zambia | 26.5 | 18.7 | 400 | N 87 W | 6 | Granite |
| 7 | Liberia | 5.9 | 0.0 | 25 | N 40 W | 6 | Gneiss, norite, gabbro |
| 8 | Braga, Portugal | 6.86 | 6.86 | 30 | N 82 W | 6 | Granite |
| 9 | Spitzbergen | 25.8 (average) | | 380 | N 78 W | 6 | — |
| 10 | Wawa, Ontario | 32.9 | 27.2 | 365–570 | N 20 W | 7 | Metamorphic succession |
| 11 | Atlanta, Georgia: | | | | N 57 E (average) | | |
| 11a | Rock Chapel Mt. | 11.8 | 6.8 | 7.5 | N 56 E | 8 | Gneiss |
| 11b | Arabia Mt. | 15.2 | 6.0 | .2 | N 34 E | 8 | Gneiss |
| 11c | Pine Mt. | 7.65 | 5.9 | .2 | N 62 E | 8 | Gneiss |
| 11d | Stone Mt | 10.5 | 7.45 | .6 | N 10 E | 8 | Granite |
| 11e | Douglasville | 3.4 | 1.6 | .6 | N 55 W | 8 | Granite |
| 12 | Massachusetts: | | | | N 29 E (average) | | |
| 12a | Twekesbury | 8.3 | 4.55 | 11.6 | N 2 E | 9 | Paragneiss |
| 12b | W. Chelmsford | 14.7 | 7.65 | 18.9 | N 56 E | 9 | Granite |
| 13 | Carthage, Missouri | 7.3 | 5.35 | 1.2 | N 2 E | 9 | Limestone |
| 14 | Graniteville, Missouri | 22.0 | 9.6 | 1.4 | N 77 E | 9 | Granite |
| 15 | Mt. Airy, N. Carolina | 17.0 | 8.25 | 10 | N 87 E | 9 | Granite |
| 16 | Troy, Oklahoma | 7.4 | 3.6 | 1.4 | N 84 W | 9 | Granite |
| 17 | St. Peters, Pennsylvania | 5.6 | 2.3 | 1.4 | N 14 E | 9 | Norite |
| 18 | Vermont: | | | | N 5 E (average) | | |
| 18a | Barre | 11.95 | 5.4 | 46 | N 14 E | 9 | Granite |
| 18b | Proctor | 9.2 | 3.3 | .4 | N 4 W | 9 | Dolomite |
| 19 | Rapidan, Virginia | 11.6 | 9.55 | 2.6 | N 6 E | 9 | Diabase |
| 20 | Red Mountain, Colorado | 25.4 | 21.7 | 9.2 | N 12 W | 10 | Granite |
| 21 | St. Cloud, Minnesota | — | — | — | N 50 E | 10 | Granite |
| 22 | Gibisonville, Ohio | — | — | 1.4 | N 70 E | 11 | Sandstone |
| 23 | Stallberg, Sweden | 49 (average) | | 900 | N 60 W | 12 | — |

Table 2 (continued)

| | | | | | | | |
|---|---|---|---|---|---|---|---|
| 24 | Vingesback, Sweden | 44.1 | 11.0 | 410 | N 10 E | 12 | Granite, amphibolite |
| 25 | Malmberget, Sweden | 35.2 | 10.8 | 460 | N 80 W | 12 | Granite |
| 26 | Laisvall, Sweden | 24.5 | 12.8 | 120 | N 17 W | 12 | Granite |
| 27 | Churchill Falls, Labrador | 13.1 | 11.7 | 305 | N 50 W | 13 | Gabbro, diorite |
| 28 | Kirkenes, Norway | 10.3 (average) | | 50 | N 23 E | 6 | — |
| 29 | Kiruna, Sweden | 9.8 (average) | | 290 | N 17 W | 6 | — |
| 30 | Malm, Norway | 9.8 (average) | | 650 | N 18 W | 6 | — |
| 31 | Lokken, Norway | 27.45 (average) | | 380 | N 72 W | 6 | — |
| 32 | Lokken, Norway | — | — | — | N 89 W | 6 | — |
| 33 | Lokken, Norway | — | — | — | N 86 E | 6 | — |
| 34 | Lokken, Norway | — | — | — | N 66 W | 6 | — |
| 35 | Lokken, Norway | — | — | — | N 5 W | 6 | — |
| 36 | E. Iceland: | | | | | | |
| 36a | Stokkenes | 9.8 (average) | | 30 | N 72 W | 6 | — |
| 36b | Hvalnes | 9.8 (average) | | 10 | (average) | | |
| 37 | Akureyri, Iceland | 3.43 (average) | | 10 | N 2 W | 6 | — |
| 38 | Keflavik, Iceland | 2.45 (average) | | 2 | N 87 E | 6 | — |
| 39 | Mont Blanc, Alps | 72.6 (average) | | 1220 | N 40 E | 14 | Gneiss, granite |
| 40 | Ireland | 17.2 (average) | | 200 | N 16 W | 6 | — |
| 41 | Kirovsk, Kola Penin, USSR | 58.9 (average) | | 100–600 | N 60 W | 15 | Orebody |
| 42a | Cobar, N.S.W. Australia | 14.5 | 11.2 | 370 | N 86 E | 16 | — |
| 42b | Cobar, N.S.W. Australia | 32.6 | 24.5 | 550 | N 72 W | 16 | — |
| 43 | Wallace, Idaho | 91.2 | 72.0 | 1220 | N 45 W | 17 | Quartzite |
| 44 | Woh, Malaya | 8.4 | 6.95 | 305 | N 28 W | 18 | Granite |
| 45 | Marble Falls, Texas | 28.2 | 7.55 | 346 | N 23 E | 19 | Granite |
| 45a | Marble Falls, Texas | 15.4 | 10.3 | 1.4 | N 33 W | 9 | Granite |
| 46 | Barberton, Ohio | 44.8 | 23.4 | 700 | N 90 E | 20 | Limestone |
| 47 | Mount Isa, Australia | 26.9 | 12.4 | 665 | N 81 E | 21 | Orebody |
| 48 | Illinois | 7.8 | 2.4 | 99 | N 62 E | 27 | — |
| 49 | Moi-Rana, Norway | — | — | — | N 25 E | 22 | Mica- garnet schist |
| 50 | Lago Maggiore, Italy | 16.1 | 11.0 | 200 | N 27 W | 23 | Paragneiss |
| 51 | White Pine, Michigan | 8.8–17.0 | — | — | N 45 W | 24 | — |
| 52 | Nyack, New York | 1.2 | .5 | — | N 2 E | 28 | Diabase |
| 53 | Niagara Falls, New York | 6.8 | —.07 | — | N 55 E | 25 | Dolomite |
| 54 | Morgantown, Pennsylvania | 51.0 | 4.0 | 700 | N 27 E | 25 | Diabase |
| 55 | Urals, USSR | — | — | — | N 62 W | 26 | — |
| 56 | Djezkazgan, USSR | — | — | — | N 45 E | 26 | — |
| 57 | Temir Tau, USSR | — | — | 340 | N 70 W | 26 | — |
| 58 | Tashtagol, USSR | — | — | 410–550 | N 2 W | 26 | — |
| 59 | Burfell, Iceland | 2.45 (average) | | 50 | N 57 E | 14 | Basalt |

assuming relatively high values. Usually, however, the horizontal stress field is clearly anisotropic.

As to the principal horizontal stress directions, therefore, it can be concluded that although the data plotted in fig. 2 are fairly reliable, they are not sufficient

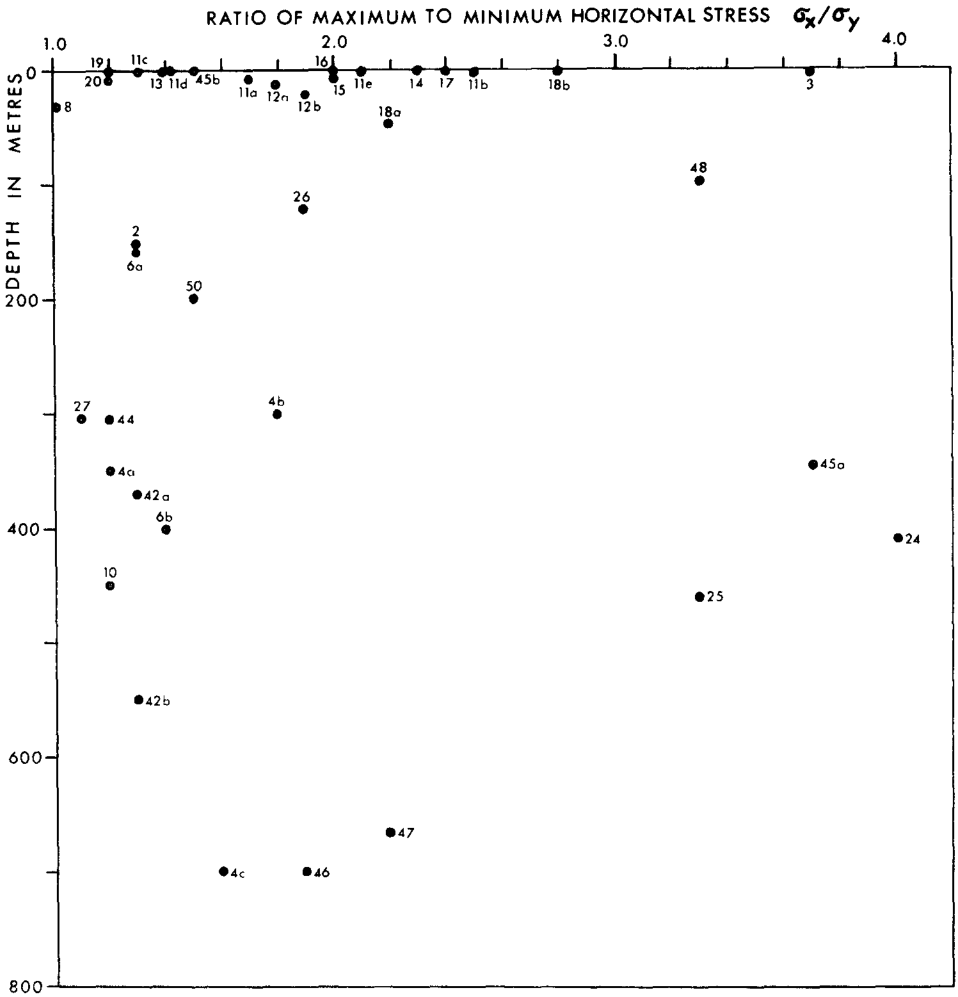


Fig. 3. Horizontal stress anisotropy (HSA) ratio *vs.* depth. Measurement numbers refer to table 2.

for a comprehensive evaluation of the orientation of the stress field in the uppermost layers of the earth. North America and Fennoscandia afford the most complete information to date. The following general observations can be made.

N o r t h A m e r i c a. A fairly consistent stress field, with the maximum horizontal compression oriented in an ENE direction, is present in the sedimentary cover of the mid-eastern and central areas of the United States. However, measurement to the west, east, and north of this region do not show the same stress pattern. In the New England states, the compression axis is often subparallel to the trend of the Appalachian fold belt. In the Precambrian shield rocks of the northeastern United States and Canada, some measurements show a trend in agreement to that to the south, but others do not. Perhaps these orientation reflect local irregularities due to complex topography, structure, and remanent stress.

Fennoscandia. Measurements can be divided into two groups, corresponding to two predominant stress fields. In the southern half of the Fennoscandian block the maximum horizontal compression usually assumes an approximately E—W direction, perpendicular or sub-perpendicular to the trend of the Caledonian fold belt. In the northern half, on the other hand, the trend is nearly meridional.

Other parts of the world. In Iceland, no consistent direction of maximum horizontal compression can be detected. The same can be said for the Alps, where more measurements are necessary before any conclusion can be reached. In the U.S.S.R., measurements in the Ukraine exhibit a WNW orientation not dissimilar from that observed in the southern half of the Fennoscandian block. In central Asia, the directions trend approximately N—S, an orientation possibly influenced by Palaeozoic and Cenozoic folding to the south. Finally, measurements in the South African and Australian blocks both show an approximately E—W direction.

## Factors affecting the observations

The evidence indicates that the magnitude and the orientation of the principal horizontal stresses are influenced by a variety of factors which must be examined before any conclusions relevant to geotectonics can be drawn from *in situ* rock stress measurements. We have already mentioned the variation of stress trajectories to be expected in any elastic block under given boundary conditions. Other important factors to be considered are: (i) topographic and structural features, (ii) erosional processes, and (iii) remanent stresses.

Topographic and structural features. Measurements taken close to the free surface in regions of rugged terrain show a wider scatter (in average horizontal stress, stress anisotropy, and directions) than measurements taken at deeper levels and in structurally simple areas. It is well known (cf. JAEGER & COOK, 1969) that a notch (valley) in an elastic plate running across the direction of maximum horizontal compression generates stress concentrations near its bottom that increase the HSA ratio. A protrusion (mountain) of sufficient height in a compressive stress field shows tension near the top. High stresses are generated at the foot of a wall. In many instances, therefore, large horizontal near-surface stresses could be induced by boundary conditions and not by tectonic processes. Since no precise correction factors for topographic effects are available, the best empirical rule for interpreting *in situ* stress measurements is to consider determinations sufficiently removed from free surfaces and in relatively undeformed rocks as more representative of the regional stress field.

Local variations in the stress patterns are also caused by geological inhomogeneities. These have been studied by TURCHANINOV et al. (1972) but even less is known about them than about the effect of topographic irregularities.

The influence of the shape of the free boundary on the near-surface stress field has an important effect in geomorphology. The common occurrence of high horizontal shear stress (often a few tens of MN $m^{-2}$) across vertical or sub-vertical planes near the surface may control the fracture pattern in the bedrock, as manifested in river courses and coastlines (HAST, 1969). Also, planes of fractures oriented along directions of maximum shear stress play a role in shaping the landscape in mountainous terrain (HAST, 1973). GERBER & SCHEIDEGGER (1969) have illustrated several instances, mainly from the Alps, in which landscape features are primarily determined by near-surface stress conditions rather than by erosion.

Erosional processes. Lage near-surface horizontal stresses, in excess of the overurden pressure, can be caused also by erosional factors. VOIGHT (1966 a) has shown that, if the depth of burial of an element of rock (assumed to be elastic) is reduced by denudation, the vertical normal stress is reduced more than the horizontal normal stress, depending on the value of Poisson's ratio for the rock. Consequently, if the initial conditions were, say, lithostatic, excess horizontal stresses are generated. Removal of 1 km of material can induce an excess horizontal stress of about 20 MN $m^{-2}$.

It appears therefore that large horizontal near-surface stresses (in excess of the overburden pressure) can be generated by a variety of nontectonic processes. Consequently, interpretations of measurements exhibiting such excess horizontal stresses in terms of a world-wide compressive tectonic stress field may not be valid. Actually, the stress field in sedimentary cover rocks on platforms would seem to indicate that such interpretations are unwarranted.

Remanent stresses. As stated in the introduction, the stress in a rock is the result of several components, one of them possibly being the remanent tectonic stress locked in the rock during a previous episode of deformation. These components cannot be separated from one another in *in situ* rock stress determinations. However, some measurements have been interpreted as being mainly due to remanent stress. EISBACHER & BIELENSTEIN (1971) have interpreted elastic strain recovery of quartzose sandstones in the Elliot Lake, Ontario, area of the Canadian shield as the consequence of remanent tectonic stress imprinted on the rocks during the Hudsonian orogeny (1700 my before present). Several measurements in shield rocks and fold belts could perhaps be interpreted in the same fashion, especially when principal horizontal stress directions do not conform to the trend of the latest episode of deformation in the area. On the other hand, it must be pointed out that also the opposite interpretation is theoretically possible, that is, stress orientations may change rather rapidly (geologically speaking) with time, and therefore the discrepancy could be caused by postorogenic stress reorientation (VOIGHT, 1966 b; TURCHANINOV et al., 1972).

The occurrence of "locked-in" stresses has been confirmed by X-ray diffraction studies, which allow the determination of residual stresses. FRIEDMAN (1972) has found residual differential stresses of 30—40 MN $m^{-2}$ in quartzites, sandstones, and granites. Some of these stresses relate to Mesozoic and possibly to Precambrian tectonic events [2]).

[2]) It is worth pointing out that X-ray diffraction techniques often indicate the occurrence of tensile residual stresses, whereas *in situ* measurements detect compressive stresses almost everywhere. If confirmed by more data, this could indicate that the occurrence of tensile components of stress is more common than previously thought.

The fact that rocks under dry upper crustal conditions can store potentially recoverable strain energy for time periods of at least $10^7$ and possibly $10^9$ years should not be overlooked in studies of their long-term rheological properties. It confirms experimental results by Price (1966), indicating that the long-term yield strength is considerably less than the instantaneous strength but does not vanish. This mechanical behaviour can be interpreted by means of a rheological model consisting of a firmoviscous (Kelvin) and a viscoplastic (Bingham) elements, i. e., a Schofield-Scott Blair model, the complete rheological equation of which are given *e. g.* by Ranalli (1972). Materials which do not possess a yield strength (that is, which in the long term flow under any differential stress no matter how small) show stress relaxation and eventually reach a hydrostatic state of stress (Jaeger & Cook, 1969). Therefore, the occurrence of residual stresses of great age shows that rocks in the upper crust have a non-vanishing yield strength and cannot be considered as simple viscous or viscoelastic bodies even under geological time intervals.

## Conclusions

From the viewpoint of global tectonics, it appears that *in situ* stress measurements, although theoretically capable of yielding relevant information, are affected by too many poorly known factors to be susceptible of unequivocal interpretation. Several authors (e. g., Hast, 1969; Voight et al., 1969; Kropotkin, 1972) have underlined the potentially important role that *in situ* rock stress determinations have in testing various geotectonic hypotheses. However, we think that excessive reliance on these determinations as a critical factor in accepting or rejecting a given hypothesis would be premature.

Attempts have also been made to determine the directions of maximum horizontal stress from a combined analysis of recent surface faulting and seismic focal mechanisms in the circum-Pacific belt (Lensen, 1960), and from a kinematic analysis of Late Cenozoic and Recent geologic structures (Pavoni, 1971). Unfortunately, the geographic distribution and the density of *in situ* stress measurement do no yet allow a meaningful comparison with these results.

A well-established method to study the current stress field in the crust is afforded by seismic focal mechanism solutions. As is well known, the earth's lithosphere is subdivided into a relatively small number of plates, with most tectonic and seismic activity taking place at, or near, the boundaries between them. A global analysis of earthquake focal mechanisms, therefore, yields information on stresses and relative displacements at these boundaries (Isacks et al., 1968). However, large areas within plates are aseismic or poorly seismic (approximately 90% of the world's seismic shocks occur near plate boundaries), and consequently information obtainable from seismology is insufficient over vast regions. Knowledge of the intra-plate state of stress must rely upon a combination of geologic, geodetic, seismic, and *in situ* data. It is in this context that the potentially great value of *in situ* stress measurements is evident, for the determination of the intraplate state of stress could help in the analysis of the driving mechanisms of global tectonics.

A study combining *in situ* stress measurements, seismic focal mechanisms, and analysis of postglacial geological features has been carried out by Sbar & Sykes

(1973) in central and eastern North America. They conclude that good agreement exists among the various techniques, indicating high horizontal compression trending nearly ENE. SYKES & SBAR (1973) have also examined the focal mechanisms of approximately 80 intra-plate earthquakes and have found that they are characterized by a predominance of thrust faulting, indicating the occurrence of high horizontal compressive stress in the interior of plates, as shown also by many *in situ* stress measurements.

The main points emerging from our analysis can be summarized as follows:

(a) Horizontal stresses are usually in excess of the overburden pressure in crystalline basement rocks and in many fold belts;
(b) Horizontal stresses in sedimentary cover rocks and in fissured massive rocks are less than the overburden pressure;
(c) The horizontal stress field is usually anisotropic;
(d) Determinations of maximum horizontal stress directions, although in some cases consistent over large areas, are not yet sufficient in number to allow reconstruction of the current tectonic stress field;
(e) Other factors besides tectonics such as topography and erosion can cause excess horizontal stresses, and they must be taken into consideration when interpreting the results from the geodynamic viewpoint;
(f) The occurrence of remanent stresses of great age shows that rocks in the upper crust possess a non-vanishing yield strength even under geological time intervals;
(g) The knowledge of the state of stress in the interior of lithospheric plates is potentially very useful in elucidating the driving forces of tectonic processes.

## Acknowledgements

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**Appendix B.** — References cited in table 2.

Note. — Entries appearing also in appendix A are identified only by author, year, and cross-reference.

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# GEOLOGISCHE VEREINIGUNG

## 65. Jahrestagung 1975

**27. Februar bis 1. März in Karlsruhe**

Thema: Tektonik kratonisierter Bereiche

Vorbereitung der Tagung: Prof. Dr. H. ILLIES, Karlsruhe

**Donnerstag, 27. Februar**

**9.00 bis 13.00 Uhr:**

Eröffnung der Tagung und Begrüßung der etwa 350 Teilnehmer durch den Vorsitzenden der Geologischen Vereinigung, Herrn Prof. Dr. H. MARTIN, durch den Rektor der Universität Karlsruhe, Magnifizenz Prof. Dr. Dr. h. c. H. DRAHEIM, und den Tagungsleiter, Herrn Prof. Dr. H. ILLIES.
Anschließend wissenschaftliche Sitzung.

Vorsitz: H. J. ZWART

A. E. M. NAIRN (Columbia, USA): Saxonische Tektonik unter dem Aspekt der Plattentektonik-Hypothese.

G. KELLER (Ibbenbüren): Saxonische Tektonik und Osning-Zone.

P. MEIBURG (Darmstadt): Schollenrotation und Abscherungstektonik in der Niederhessischen Senke.

B. SCHRÖDER (Bochum): Saxonische Tektonik im östlichen Süddeutschland.

V. J. DIETRICH (Zürich): Plattentektonik in den Ostalpen.