TECTONIC OVERPRESSURE IN
FRANCISCAN TERRAINS

BY

RAZEL ANN WITTELS

SUBMITTED IN PARTIAL FULFILLMENT
OF THE REQUIREMENTS FOR THE
DEGREE OF MASTER OF
SCIENCE

at the

MASSACHUSETTS INSTITUTE OF
TECHNOLOGY

January, 1969

Signature of Author........................... ...
                    Department of Geology and Geophysics,
                                    January 20, 1969

Certified by.....................   .
                                      Thesis Supervisor

Accepted by......\   ....
                    Chairman, Departmental Committee
                            on Graduate Students

TABLE OF CONTENTS

TECTONIC OVERPRESSURE IN FRANCISCAN TERRAINS

Razel Ann Wittels

The possibility has been tested experimentally that
tectonic overpressure could have raised the mean pressure
during metamorphism of the Franciscan terrain to values
commensurate with the development of the glaucophane-
schist facies. Since tectonic overpressure directly de-
pends on the strength of the rocks, the strength of sam-
ples of the graywacke, possibly the hardest rock of the
terrain, was measured in triaxial tests under conditions
which would maximize the strength (fast strain rate, low
temperature) and at a variety of confining pressures and
pore pressures appropriate to a depth of 15 km. From the
results of these experiments, an upper limit on tectonic
overpressure was set.

The upper limit and probable reductions of it due to
the increased temperatures and decreased strain rates
found in the earth were compared with the amount of over-
pressure needed to raise the mean pressure to the minimum
value necessary for formation of glaucophane-schist, about
8 kb. This pressure is reached at moderate pore pressure
(λ = .69) under the cold, fast conditions used in the
laboratory, or at hydrostatic (λ = .37) or lower pore
pressure under the warmer, slower conditions of the crust,
and with the restriction $\sigma_2 = \sigma_1 = \sigma_D + \sigma_v$. However, it
would be unlikely that tectonic overpressure would pro-
vide the necessary additional pressure if the pore pres-
sure was high (λ > .69), if σ_2 was significantly less than
σ_1, or if the strength of the terrain is less than the
strength of the rock used in these measurements.

The Franciscan terrain of California consists of Jurassic and Cretaceous eugeosynclinal sediments of a varying nature. About 80% of the terrain is a graywacke of variable composition, accompanied by siltstone and shale, conglomerate, limestone, mafic volcanic rocks, chert and shale, and glaucophane and accompanying schist (Bailey, 1960). The glaucophane schist of the Franciscan is apparently the result of high pressure, low temperature metamorphism of the graywacke (Ernst, 1965). Basically, it occurs in three ways: (1) as tens of square miles of regionally metamorphosed rock, (2) as patches grading laterally into unmetamorphosed rock, and (3) as generally rounded tectonic inclusions (Bailey, 1961). The schist contains glaucophane, jadeitic pyroxene, and lawsonite, all characteristic of the glaucophane-schist facies of Eskola*.

In spite of the fact that the glaucophane-schist facies is considered to be formed as a result of burial metamorphism, or metamorphism resulting from the increased pressure and temperature arising from burial in the crust, the pressure and temperature conditions at which the facies forms do not lie on a normal geothermal gradient. A mechanism for attaining simultaneously the

---

* The glaucophane-schist facies is also known as the blue-schist facies. Winkler (1967) uses the term "lawsonite-glaucophane" facies, since glaucophane may be found in facies other than the glaucophane-schist facies.

high pressures and low temperatures required for its formation is necessary to explain the facies' occurrence. One possibility is that high pressures occurred abnormally in a normally low pressure, low temperature environment. Tectonic forces could conceivably create enough additional pressure, or _tectonic overpressure_, to raise the mean pressure on the rocks to the required values.

In this paper we will evaluate tectonic overpressure as a possible mechanism for the Franciscan metamorphism by establishing in the laboratory an upper limit for tectonic overpressure in a subgraywacke from the Franciscan.

TECTONIC OVERPRESSURE AND PRESSURES WITHIN THE EARTH.

Rocks in the crust exist in a state of stress which can be considered as having hydrostatic and non-hydrostatic components. For a situation in which there are no tectonic forces, the approximation may be made that the rock is in a state of hydrostatic stress, $\sigma_v = \bar{\rho}gh$, where h is depth of burial, g is acceleration due to gravity, and $\bar{\rho}$ is an average density of the overlying material. We will consider compressive stresses as positive. For thermodynamic purposes, a mean stress $\sigma_M = \frac{1}{3}(\sigma_1 + \sigma_2 + \sigma_3)$ may be defined, where σ_1, σ_2, σ_3 are principal stresses such that $\sigma_1 \geqslant \sigma_2 \geqslant \sigma_3$. For the hydrostatic case, $\sigma_1 = \sigma_2 = \sigma_3$ and $\sigma_M = \bar{\rho}gh$. If there are tectonic forces, however, $\sigma_1 \neq \sigma_2 \neq \sigma_3$. Allowing $\sigma_3 = \sigma_v$, we have as a maximum

limit $\sigma_1 = \sigma_v + \sigma_D$, where σ_D is the strength of the rock, $\sigma_D = \sigma_1 - \sigma_3$. Any greater value of σ_1 would cause failure of the rock. It is also possible that $\sigma_1 = \sigma_v - \sigma_D$, but since these values could possibly be less than zero, or tensile stresses, and because we are considering overpressure and not "underpressure", we will consider only $\sigma_1 = \sigma_v + \sigma_D$.

σ_2 may have a value which ranges from σ_1 to σ_3. If $\sigma_2 = \sigma_1 = \sigma_v + \sigma_D$, we find that $\sigma_M = \sigma_v + \frac{2}{3}\sigma_D$ (Birch, 1955). Compared with the hydrostatic case, there is an additional average pressure of $\frac{2}{3}\sigma_D$ on the rock. This is the maximum tectonic overpressure, the maximum additional pressure which can be experted on the rock by tectonic forces.

Because the maximum overpressure depends on the strength of the rock, we can place a maximum limit on the tectonic overpressure by establishing the strength of the rock under conditions which would maximize the strength.

AN UPPER BOUND ON STRENGTH

The strength of a rock is dependent on several variables, including confining pressure, pore pressure, strain rate, temperature, and the properties of the rock itself. These variables must be considered in establishing conditions which would maximize the strength of the Franciscan.

The material chosen was a subgraywacke collected by W. G. Ernst fron San Luis Obispo County, California. It

is typical of the unmetamorphosed Franciscan graywackes,
as shown in Table 1. Specimens of the graywacke were
chosen because the graywacke is the most common of the
unmetamorphosed rocks of the formation, and is also per-
haps among the strongest. Studies on the strengths of
sedimentary rocks (Handin and Hager, 1957; Handin, et
al., 1963) indicate that at 2kb confining pressure, sand-
stone is often stronger than shale or siltstone and about
as strong as limestone. Some sandstones are strong
enough to compare in strength with strengths reported
for some igneous rocks (Griggs, et al., 1960). Certainly
a sequence of rocks cannot be stronger than its strongest
member; possibly it will not be stronger than its weakest
member. In either case, by choosing what appears to be
the strongest rock type for our experiments, we can be
assured that the strength of the terrain will not be under-
estimated.

The strength of a rock generally will decrease as the
strain rate decreases (Heard, 1968). Geologic strain
rates are usually estimated as slower than 10^{-14} per sec.
By choosing a strain rate much faster than this, the strength
of the rock as determined in the laboratory would be greater
than its actual strength in the crust.

The effect of increased temperature would be to de-
crease the strength of the rock. It is desirable, there-
fore, to determine the strength at lower temperatures than

Table 1   Description of Rocks

Blocks I and II		Bailey (1960)	
Quartz	27%	Quartz	10 - 50%
Albite	35%	Feldspar	33 - 66%
Rock Fragments	3%	Mafic Volcanic Rocks	0 - 33%
Muscovite	15%	Micas	
Biotite	5%		
Chlorite	15%	Chlorite	5-20%
Chert	<1%	Kaolinite	

(Muscovite + Biotite = 20%)

Blocks I and II: Angular to subangular grains. Poorly sorted.

Bailey (1960): Highly angular to subangular grains. Poorly sorted.

those necessary for the development of the facies. The glaucophane-schist facies can develop within a temperature range of 200° to 450°C (Winkler, 1967). By measuring the strength of the rock at room temperature we obtain an upper limit to strength at the temperatures of the facies.

Pressure effects pose more of a problem. First, the hydrostatic component was assumed to be $\sigma_v = \bar{\rho}gh$. The density of our rock was determined as 2.7 g/cc. For depth of burial, we shall use the thickness of the Franciscan as reported by Bailey (1961), about 15 km. This depth gives $\sigma_v = 4.05$ kb as a maximum. However, strength is dependent not on the confining pressure but on the effective confining pressure, $\bar{\sigma} = \sigma_v - p = \sigma_v(1 - \lambda)$, where p is the pore pressure and λ is the ratio of pore pressure to hydrostatic pressure, p/σ_v. Pore pressure is generally not less than its hydrostatic value at the appropriate depth, which for our rock gives a value of $\lambda = .37.$ Elsewhere, values of λ as great as 1.0 have been reported (Hubbert and Rubey, 1959). Almost all metamorphic petrologic studies are made under the assumption that $p = \sigma_v$, or $\lambda = 1$, resulting in $\bar{\sigma} = 0$, if we assume that the effective stress law held under these conditions.

Unfortunately, knowledge of pressure conditions for rocks undergoing metamorphism is far from complete. It will be necessary to consider values of λ varying from zero to one, and confining pressure to a maximum of 4.05 kb.

EXPERIMENTAL PROCEDURE AND RESULTS

Strength was measured in triaxial experiments similar to those used by Handin et al. (1963) and Brace and Martin (1968). Cylinders of the rock, between 1.00 and 1.44 in. long and .625±.05 in. in diameter, were subjected to confining pressure through a 3 mm polyurethane jacket. The pore pressure fluid was admitted to the sample through a hollow piston. The piston was driven at a constant displacement rate during the experiments.

Measurements of strength were made on specimens cut from two blocks of the graywacke. Specimens from Block I were used for the majority of the determinations of strength; comparison with specimens from Block II showed the variation possible in rock from one locality.

The tests were conducted at varying confining pressures and pore pressures to simulate possible conditions within the earth. In addition, some experiments were made without pore pressure. For these experiments the samples were dried in a vacuum overnight before the tests were made. The samples used in tests using pore pressure were saturated prior to testing. The early saturated tests were allowed to remain for at least four hours under the confining pressure and pore pressure of the test conditions. This amount of time was increased to about twelve hours after it was decided that pressure equilibrium was not being achieved throughout the sample.

The majority of the tests were run at a strain rate of 10^{-6} per sec, and the remaining at 10^{-5} or 10^{-7} per sec. Tests run at 10^{-6} per sec lasted no longer than a day, making this a convenient speed.

All experiments were made at room temperature, about 21°C.

The results from the experiments are listed in Table 2 and graphed in Figure 1 and Figure 2. Figure 1 shows the results of all experiments performed on Block I. The results of dry experiments performed on samples whose ratio of length to diameter was less than 2.00 were plotted after being corrected according to criteria established by Mogi (1966). The curve is fitted to the results of the dry experiments. Figure 2 shows the results of experiments performed on Block II, compared with the curve of Figure 1.

The experiments with pore pressure were performed on Block I only. All the results from these experiments do not fit the curve of Figure 1 well. Of those experiments run at a strain rate of 10^{-6} per sec, the results which are far from the curve are those with pore pressures of at least 1.75 kb. Some of these did not remain at their test conditions for at least twelve hours. The most notable of these was Experiment 6, which was held at its pore pressure of 1.88 kb for only four hours. If the pore pressure in the rock was not as high as was indi-

cated on the gage, the true effective stress will be greater than what was calculated, and the rock should be stronger than would be expected at the calculated effective stress.

It is probable that a discrepancy between actual and measured pore pressures is responsible for the variation of most of the anomalous pore pressure tests. Brace and Martin (1968) report an unexpected increase in strength with increased strain rate which they believe is due to dilatancy. This strengthening first appeared at strain rates faster than some critical value, between 10^{-7} and 10^{-8} per sec for most rocks tested with water as a pore fluid. Below the critical strain rate there was no anomalous strengthening, and this effect is therefore not expected to occur in most geologic situations, where strain rates are on the order of 10^{-14} per sec or slower.

Experiments 6 and 9 were repeated at different strain rates to try to ascertain whether dilatancy strengthening was the cause of the anomalous strengths. Because dilatancy strengthening does not appear at strain rates slower than about 10^{-7} per sec, tests made under the same pressure conditions but at strain rates of 10^{-7} per sec or slower should group around the curve for $\lambda = 0$. Tests made at strain rates faster than the critical strain rate should show a departure from that curve which increases as the strain rate increases.

Table 2   Experimental Conditions and Results

Experiment no.	Block	L/d	Strain Rate (per sec)	σ_v (kb)	p (kb)	λ	$\bar{\sigma}$ (kb)	σ_D (kb)	corrected* σ_D (kb)
1	I	1.95	10^{-5}	1.80	0	0	1.80	6.99	6.75
2	I	1.98	10^{-6}	1.51	0	0	1.51	6.20	6.02
3	I	1.96	10^{-6}	1.42	0.50	.36	0.92	3.04	--
4	I	1.98	10^{-6}	1.44	1.26	.88	0.18	4.41	--
5	I	1.98	10^{-6}	2.93	1.00	.34	1.93	6.68	--
6	I	1.99	10^{-6}	2.87	1.88	.66	0.99	7.50	--
7	I	1.99	10^{-6}	4.31	1.49	.35	2.82	10.11	--
8	I	1.99	10^{-6}	2.82	0	0	2.82	8.37	8.20
9	I	1.76	10^{-6}	4.23	0	0	4.23	11.17	10.74
10	I	1.97	10^{-6}	4.25	1.75	.41	2.50	10.88	--
11	I	1.64	10^{-6}	2.85	0	0	2.85	9.34	8.89
12	I	2.09	10^{-6}	0.54	0	0	0.54	4.34	--
13	I	2.23	10^{-6}	1.42	0.50	.35	0.92	5.47	--
14	I	2.29	10^{-6}	4.21	1.50	.36	2.71	9.29	--
15	I	2.24	10^{-5}	2.87	1.88	.66	0.99	7.87	--
16	I	2.19	10^{-7}	4.23	1.77	.36	2.46	8.96	--
19	II	2.21	10^{-6}	1.03	0	0	1.03	5.84	--
20	II	2.19	10^{-6}	2.57	0	0	2.57	9.09	--
21	II	2.18	10^{-6}	3.49	0	0	3.49	10.40	--
22	II	2.23	10^{-6}	0.42	0	0	0.42	4.57	--

* by criteria set forth by Mogi (1966).

Figure 1   Experiments on Block I

Circles indicate a strain rate of 10^{-6}/; per sec
triangles, 10^{-5}/; per sec squares, 10^{-7}/. per sec Open symbols
represent dry experiments ($\lambda = 0$)   For experi-
ments with pore pressure, the confining pres-
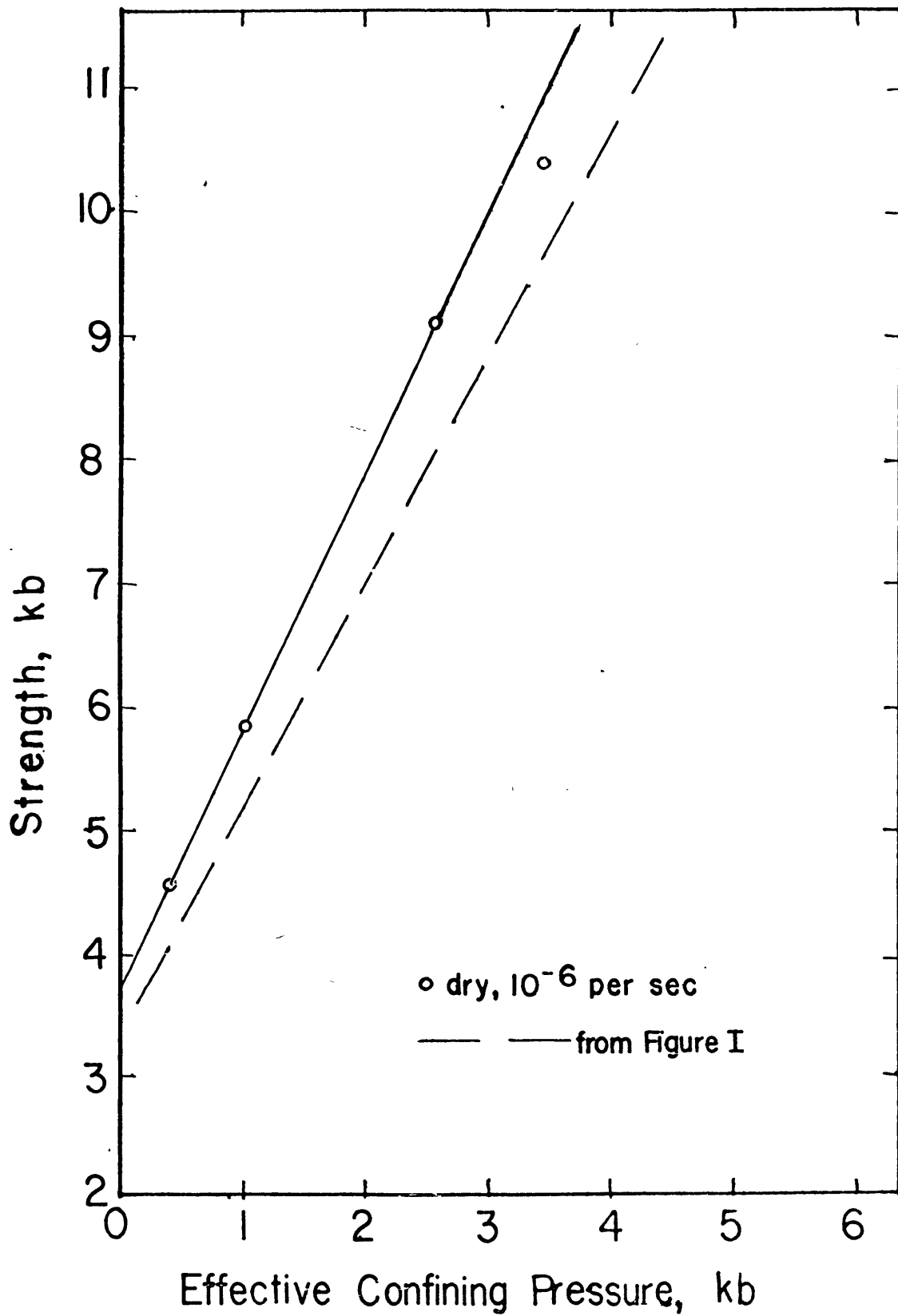sure and pore pressure are listed in that order.
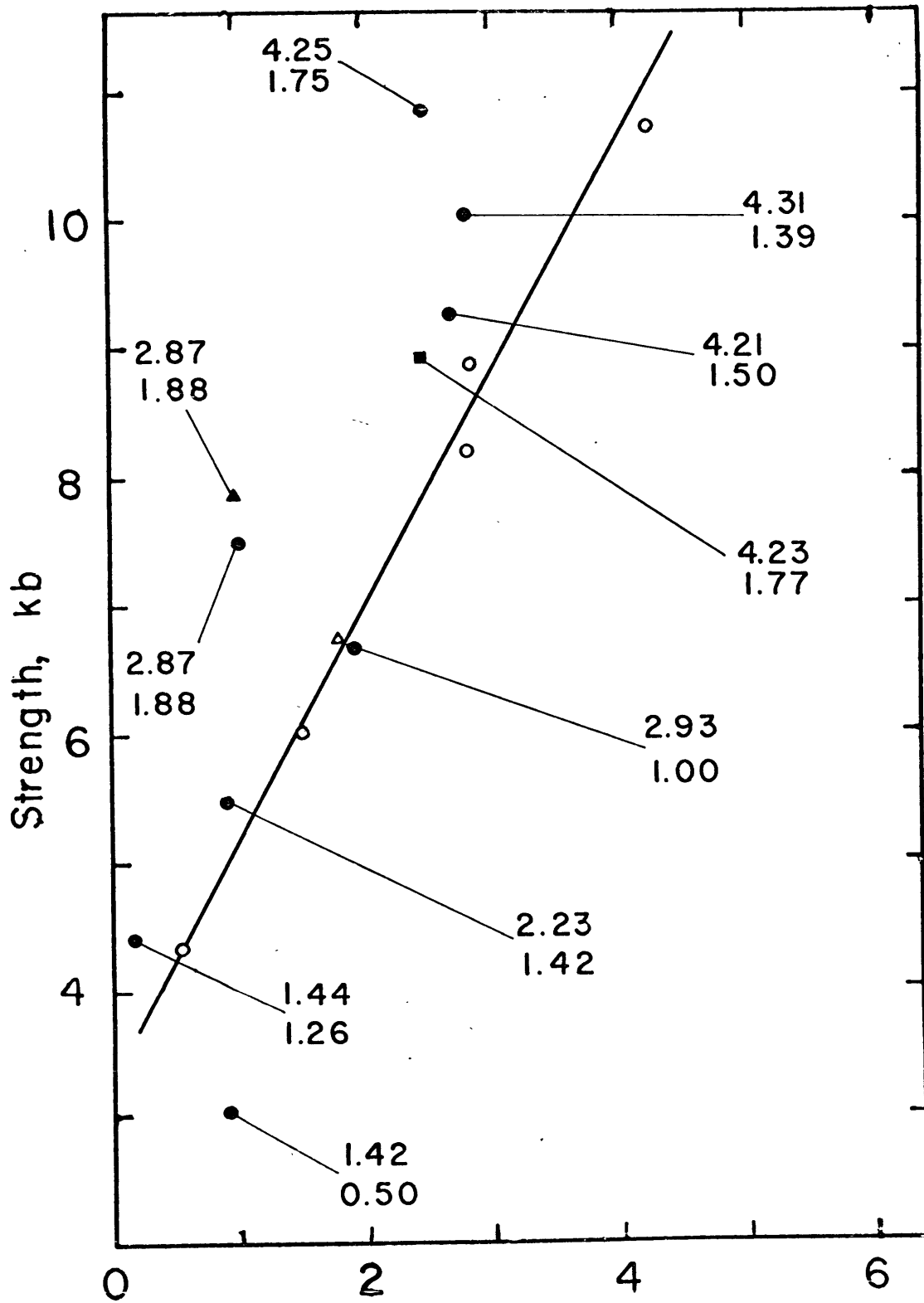
Figure 1   Experiments on Block I

Figure 2   Experiments on Block II

Experiment 16 was performed under the same conditions
as Experiment 9, except that a slower strain rate of 10^{-7}
per sec was used. The strength determined by Experiment
16 was less than the strength determined by Experiment 9.
Also, Experiment 15 was performed under the same condi-
tions as Experiment 6, except that a faster strain rate
of 10^{-5} was used. The strength determined by Experiment
15 was _greater_ than that determined by Experiment 6. These
are just the expected results if dilatancy strengthening
were involved. We emphasize that at geologic strain rates
this effect would not appear, and therefore base all fur-
ther calculations on the zero pore pressure curves of
Figure 2.

DISCUSSION

We shall now consider the amount of overpressure
available from our rock as compared with the amount nec-
cessary for the development of the glaucophane-schist
facies.

The lower bound on pressure for the glaucophane-
schist facies has not been exactly determined. It is
governed by the minimum pressure needed for the formation
of jadeite. Birch et al. (1960) provide the equilibrium
curve, albite = jadeite + quartz, at temperatures from
800° to 1000°C. However the error margins on the slope
of the curve are too broad to provide for a reliable

extrapolation of these results to lower temperatures. In
addition, the exact location of this curve will doubt-
lessly vary with varying jadeite composition of the pyrox-
ene. A more helpful aid is the phase relations of $CaCO_3$.
When the glaucophane-schist facies contains $CaCO_3$, it is
usually in the form of aragonite. The lawsonite-albite
facies, a lower pressure assembledge, may contain either
aragonite or calcite. Thus, it is reasonable to expect
the reaction, albite = jadeite + quartz, to be at higher
pressures than the transition of calcite to aragonite.
Winkler (1967) concludes that pressures on the order of
8 to 8.5 kb at temperatures of 250° to 350°C must have
been exceeded for the formation of the glaucophane-schist
facies. More recently, Boettcher and Wyllie (1968) have
located the transition, albite = jadeite + quartz, be-
tween 15 and 20 kb at 600°C. If we consider the lower
pressure of 15 kb, Birch's curve extrapolates to about
8 kb at 300°C.

Since we are considering a hydrostatic load pressure
of not greater than 4.05 kb, a mean pressure of 8 kb de-
mands a tectonic overpressure of at least 3.95 kb, or a
strength of 5.93 kb. From Figure 2 we can see that this
strength is reached at an effective confining pressure
of about 1.25 kb, or λ = .69. If the pore pressure were
at its hydrostatic value, the effective confining pres-
sure would be 2.55 kb, at which the strength is about

8.50 kb. Thus, for the conditions we have used to establish a maximum strength, namely fast strain rates, low temperatures, and $\sigma_2 = \sigma_1$, and provided that λ is less than .69, our rock is strong enough to support the necessary overpressure. As our rock moves from these conditions toward conditions which were likely during the metamorphism, its strength will decrease.

When rocks are strained at slow rates, they are weaker than when they are strained at fast rates. Heard (1968) and Brace and Martin (1968) show for granite and diabase a decrease in strength of .5 to 1 kb for a decrease of strain rate of 10^{-4}. Because geologic strain rates are 10^{-8} times slower than the strain rates used in our experiments, a decrease of strain rate to geologic rates would decrease strength by at least 1 kb.

The strength of our rock should also be less at the temperatures of the facies than at the relatively cold laboratory temperatures at which the experiments were performed. For silicate rocks whose strengths at 5 kb confining pressure are comparable to the strength of our rock, this reduction is about 4 kb when the temperature is increased from 25° to 300°C (Griggs et al., 1960; Baidyuk, 1967; Heard, 1968). Tests on some sedimentary rocks (Baidyuk, 1967) indicate that the amount of decrease, increases with increasing confining pressure.

Let us consider a decrease in strength of 5 kb at a

confining pressure of 5 kb due to the combined effects of decreased strain rate and increased temperature. This will result in the strength of 5.93 kb being reached at an effective confining pressure of greater than 2.55 kb. Since 2.55 kb is the effective confining pressure for which the pore pressure is at its hydrostatic value, the pore pressure must be less than its hydrostatic value if the rock is to be at the minimum necessary strength.

Another consideration is that we have taken an upper limit in the overpressure by assuming the $\sigma_2 = \sigma_1$. In the laboratory experiments $\sigma_2 = \sigma_3$, and the tectonic overpressure produced was equal to $\frac{1}{3}$ of the strength instead of $\frac{2}{3}$ of the strength. Using $\frac{1}{3}$ instead of $\frac{2}{3}$, halves the overpressure available from a given strength; a strength of 11.85 kb is necessary to produce the necessary 3.5 kb of overpressure. As can be seen from Figure 2, this is stronger than our rock is at an effective confining pressure of 4.05 kb. It is possible that $\sigma_2 = \sigma_1$, but it should be noted that σ_2 cannot equal σ_3 if the desired effect is to be achieved.

So far we have seen that the strength of this rock type is adequate only under very specified conditions. It is adequate if $\sigma_2 = \sigma_1$, and at moderate pore pressures if measured under the cold, fast conditions used in the experiments. If, however, we consider the slower, warmer conditions which were more likely to occur during meta-

morphism, λ must equal .37, its hydrostatic value, or be less than .37. The strength is inadequate if $\sigma_2 = \sigma_3 = \sigma_v$ or if λ is much greater than .37.

The problem of pore pressures is critical. It must be remembered that we are dealing with sediments which were saturated when they were deposited. The pressure on the pore fluid would tend to increase due to reduction of pore space resulting from compaction of the sediment. It has been suggested that this increase continues until the pore pressure exceeds the load pressure; at higher pore pressure the pore fluid would tend to escape (Yoder, 1955). Furthermore, low pressures would lower reaction rates and greatly increase the time necessary for the metamorphism. Turner and Verhoogan (1960) suggest that high fluid pressures are necessary for metamorphism to the glaucophane-schist facies. If that were the case, the effective confining pressure would be low, and the strength of our rock would be insufficient to provide the necessary overpressure.

CONCLUSIONS

When we consider the metamorphism in the Franciscan terrain, the possibility of tectonic overpressure increasing the mean pressure to the minimum values necessary for the development of the glaucophane-schist facies seems remote. It would be necessary to have geologically unrealistic strain rates, to have the restriction $\sigma_2 = \sigma_1$,

and to have temperatures too low for the facies to develop. There will be some increase in mean pressure due to tectonic forces, but this increase will probably not be more than 1 kb, calculated from our data considering appropriate pore pressures, strain rates, and temperatures. This is much less overpressure than is necessary.

It is possible that the necessary pressure and temperature conditions were reached by the converse of our proposition, and that low temperatures were found in a normally high temperature, high pressure environment. Ernst (1965) advocates this theory, and suggests that it would be possible for rapid sinking of the crust combined with thermal disequilibrium to result in the required temperatures and pressures.

ACKNOWLEDGEMENTS

BIBLIOGRAPHY

Baidyuk, B.V., 1967, Mechanical Properties of Rocks at
    High Temperatures and Pressures: New York,
    Consultants Bureau, 75 p.

Bailey, E.H., 1960, Metamorphic facies of the Franciscan
    formation of California and their geologic signifi-
    cance: Geol. Soc. America Spec. Paper 68, p. 4-5.

Bailey, E.H., Franciscan formation of California as an
    example of eugeosynclinal deposition: Geol. Soc.
    America Bull., v. 71, p. 2046-2047.

Birch, F., 1955, Physics of the crust: Geol. Soc. America
    Spec. Paper 62, p. 101-117.

Birch, F., and P. Le Compte, 1960, Temperature-pressure
    plane for albite composition: Am. Jour. Sci., v. 258,
    p. 209-217.

Boettcher, A.L., and P.J. Wyllie, 1968, Melting of granite
    with excess water to 30 kilobars pressure: Jour.
    Geology, v. 76, p. 235-244.

Brace, W.F., and R.J. Martin,III, 1968, A test of the law
    of effective stress for crystalline rocks of low
    porosity: Int. J. Rock Mech. Min. Sci., v. 5, p. 415-
    426.

Ernst, W.G., 1965, Mineral paragenesis in Franciscan meta-
    morphic rocks, Panoche Pass, California: Geol. Soc.
    America Bull., v. 76, p. 879-914.

Griggs, D.T., F.J. Turner, and H.C. Heard, 1960, Deforma-
    tion of rocks at 500° to 800°C: Geol. Soc. America
    Memoir 79, p. 39-104.

Handin, J., and R.V.Hager,Jr., 1957, Experimental deforma-
    tion of sedimentary rocks under confining pressure:
    tests at room temperature on dry samples: Bull. Amer.
    Assoc. Petrol. Geol., v. 41, p. 1-50.

Handin, J., R.V.Hager,Jr., M. Friedman, and J.N. Feather,
    1963, Experimental deformation of sedimentary rocks
    under confining pressure: pore pressure tests:
    Bull. Amer. Assoc. Petrol. Geol., v. 47, p. 717-754.

Heard, H.C., 1968, Experimental deformation of rocks and the problem of extrapolation to nature: NSF Advanced Science Seminar in Rock Mechanics, v. 2: Bedford, Mass., Terrestrial Sciences Laboratory, Air Force Cambridge Research Laboratories, p. 439-507.

Hubbert, M.H., and W.W. Rubey, 1959, Role of fluid pressure in mechanics of overthrust faulting, II: Geol. Soc. America Bull., v. 70, p. 167-206.

Mogi, K., 1966, Some precise measurements of fracture strength of rocks under uniform compressive stress: Rock Mech. and Eng. Geology, v. IV/1, p. 41-55.

Turner, F.J., and J. Verhoogen, 1960, Igneous and Metamorphic Petrology: New York, McGraw-Hill Book Company, Inc., 694 p.

Winkler, H.G.F., 1967, Petrogenesis of Metamorphic Rocks: New York, Springer-Verlag New York, Inc., 237 p.

Yoder, H.S., 1955, Role of water in metamorphism: Geol. Soc. America Spec. Paper 62, p. 505-524.