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Structural and rheological implications of lower-crustal earthquakes below northern Switzerland

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ABSTRACT

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A detailed study of the recent seismicity in the northern Alpine foreland of Switzerland reveals that earthquakes with magnitudes between 0.9 and 4.2 occur not only in the upper part of the crust, but, contrary to observations in most other intracontinental settings, seismicity extends down to depths of about 30 km in the lower crust. Observed P-wave velocities in the focal region of even the deepest earthquakes are less than 6.3 km s^{-1}. The average value of Poisson's ratio for the entire focal-depth range (6–30 km) lies between 0.23 and 0.24. Focal mechanisms are mostly a combination of strike-slip and normal faults, with consistent orientations of P- and T-axes, but without any systematic dependence on focal depth. From geological observations, regional strain rates are of the order of 10^{-13} s^{-1}. Two-dimensional temperature modelling, indicates that temperatures in the lower-crustal seismogenic zone are well above 450°C.

Under the assumption of a dry environment, the observed P-wave velocities and Poisson's ratios imply either a granitic or quartz-rich gneissic composition or possibly a gabbroic (quartz-free) granulite. The deformation would be governed either by the rheology of quartz or of plagioclase. The onset of ductility observed in laboratory studies lies around 300°C for quartz and around 450°C for plagioclase. Thus, conditions in the lower crust beneath northern Switzerland appear to be incompatible with the occurrence of either brittle failure or plastic instabilities, as is implied by the existence of earthquakes.

Alternatively, lower-crustal earthquakes could be a manifestation of the presence of fluids in the lower crust. Fluids, under near-lithostatic pressure, either evolved in situ from metamorphic reactions or from infiltration from the mantle, would tend to decrease the effective stress on pre-existing faults and thus allow brittle friction failure to occur at depths where, under dry conditions or low fluid pressures, only ductile deformation would be expected.

1. Introduction

The occurrence or absence of earthquakes in a particular depth range of the Earth's crust constitutes an indication of the rheological behaviour and thus of the material properties of the focal region. Whether a rock will undergo brittle or ductile deformation when subjected to shear stress depends on petrological composition, temperature, strain rate, fluid content and tectonic regime. Kobayashi (1977) appears to have been one of the first to have recognized the possible link between the range of earthquake focal depths in the crust and surface heat flow (Ito, 1990). In most intracontinental areas with moderate to high heat flow, seismicity is restricted to the upper 10 or 20 km of the crust. The few areas where crustal earthquakes are observed to occur at greater depths are associated with relatively low heat flow (e.g. Shudofsky et al., 1987; Fadaie and Ranalli, 1990; Wong and Chapman, 1990). These observations have led to the postulation of a simple two-layered model consisting of a seismogenic and brittle upper crust over an aseismic and ductile lower crust with a single brittle-ductile transition, whose depth is temperature dependent. (Brace and Kohlstedt, 1980; Meissner and Strehlau, 1982; Sibson, 1982; Chen and Molnar, 1983). In two

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earlier papers (Deichmann, 1987a; Deichmann and Rybach, 1989) it is shown that the northern Alpine foreland of Switzerland, with significant lower-crustal seismicity and relatively high heat flow, represents a notable exception to this simple model of crustal rheology. Recently, several more sophisticated rheological models, which include multiple layering due to compositional variations, have been proposed, and a much larger number of laboratory results have become available (Smith and Bruhn, 1984; Carter and Tsenn, 1987; Meissner and Kusznir, 1987; Ranalli and Murphy, 1987). In a comprehensive review, Ord and Hobbs (1989) synthesize most of the available data into a series of crustal rheology models and discuss the consequences of plastic instabilities and of a pos-

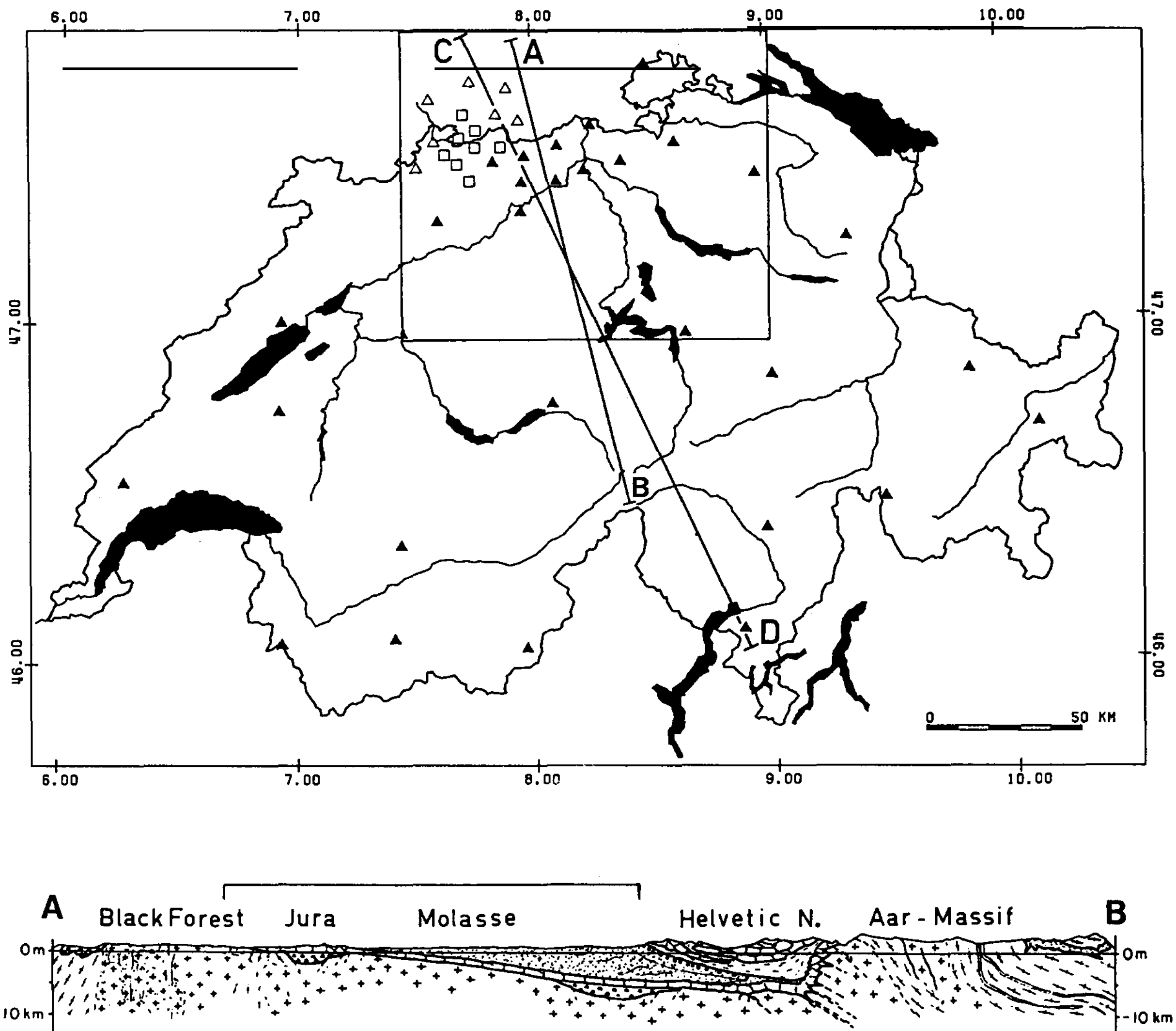

Fig. 1. Map of Switzerland with geological cross-section (after Diebold and Müller, 1984). ▲, seismometer stations (FM-telemetry with digital recording) operated by the Swiss Seismological Service; △, digital stations operated by the University of Karlsruhe; □, an 8-station temporary array near Basel, in operation from November 1986 to December 1988. The bracket above the cross-section delimits the focal-depth profile in Fig. 3. Line C–D corresponds to the profile in Fig. 4.

sible breakdown of Byerlee's law of brittle failure at higher temperatures and pressures. In the present paper we want to review the evidence regarding seismicity, crustal structure and temperatures beneath northern Switzerland in the light of some additional data and show that this evidence contradicts even the more recent models of crustal rheology, and then discuss the role high-pressure fluids might play in triggering earthquakes at depths where ductile deformation would be expected.

2. Tectonic setting, strain rates and temperatures

The region of interest, comprising part of the northern Alpine foreland, consists essentially of the Jura Mountains and the Molasse Basin, bounded in the north by the Rhinegraben and Black Forest and in the south by the Helvetic Nappes of the Alps (Fig. 1).

From geological considerations, regional strain rates must lie in the range between 10^{-15} s^{-1} and 10^{-13} s^{-1} (Pfiffner and Ramsey, 1982). For the Jura overthrust Müller and Briegel (1980) calculated a strain rate of 10^{-13} s^{-1}.

The surface heat flow determined from temperature measurements in numerous bore-holes corresponds to a pronounced positive thermal anomaly, with values ranging between 75 and 145 mW m^{-2} (Rybach et al., 1987). Part of this anomaly is the result of the extensive hydrothermal activity and can be accounted for by convective heat transfer due to extensive groundwater flow throughout the upper 8 km of the crust. Taking this effect into account, the average regional heat flow amounts to about 80 mW m^{-2}. Temperatures inferred from thermo-hydraulic model calculations reach 250°C at a depth of 8 km (Griesser and Rybach, 1989). Two-dimensional steady state model calculations, based on the assumption of purely conductive heat flux but taking into account the temperature dependence of thermal conductivity and the depth dependence of radioactive heat production, indicate that temperatures between 20 and 30 km in the lower crust range from 450 to 650°C (Deichmann and Rybach, 1989).

3. Crustal thickness and seismic velocities

The results of seismic refraction experiments and gravity surveys show that the crust–mantle boundary below northern Switzerland dips gently in a NNW–SSE direction from a depth of 25–26 km below the Black Forest to about 33 km below the front of the Helvetic nappes. Further south, the Moho dips more steeply, reaching depths of 50–55 km below the culmination of the Alps (Mueller et al., 1980; Maurer, 1989). Upper crustal P-wave velocities derived from refraction measurements range between 5.5 km s^{-1} near the top of the basement to about 6.1 km s^{-1} at mid-crustal depths between 10 and 15 km (Sierro, 1989). The asymptotic behaviour of the Pg phase of earthquakes (the direct wave) indicates that even for the deepest events, situated at a depth of 30 km just above the Moho, lower-crustal P-wave velocities do not exceed 6.3 km s^{-1} (Deichmann, 1987a; Deichmann and Rybach, 1989). Ratios of P- to S-wave velocities (V_p/V_s) derived from Wadati diagrams vary between 1.69 and 1.71, which implies Poisson's ratios between 0.23 and 0.24 for the entire focal depth range from 6 to 30 km (Deichmann and Rybach, 1989). Modelling P- and S-wave travel-time differences between direct waves and reflections at the Moho for one of the earthquake swarms in northern Switzerland resulted in a Poisson's ratio of 0.25 for the lower crust (Pfister, 1990).

Thus seismic velocities derived from refraction measurements and earthquake observations do not give any indication of pronounced velocity discontinuities between the top of the basement and the crust–mantle transition. Both wide-angle and normal incidence reflection data, on the other hand, show evidence of enhanced reflectivity in the lower crust below a depth of about 20 km (Mueller et al., 1987).

4. Seismicity

The distribution of seismograph stations in Switzerland and southern Germany is shown on the map in Fig. 1. Details of the data acquisition and location procedures, as well as a discussion of the accuracy of the hypocenter determinations can be found in the earlier papers (Deichmann, 1987a; Deichmann and Rybach, 1989). The epicenters of the earthquakes recorded between January 1984 (the onset of digital data acquisition) and December 1990 are plotted in Fig. 2, together with a representative selection of fault-plane solutions.

The magnitude of the earthquakes recorded from 1984 to 1990 range between 0.5 and 4.2. Except for a concentration of activity at the southern end of the Rhinegraben, the epicenters are not associated with any visible fault systems. Instead, earthquakes in northern Switzerland tend to occur in individual clusters. In fact, roughly half of the total number of events occurred as doublets or as multiplets of up to about 50 events with almost identical signal character. By means of high-precision relative locations of the events within each cluster, based on a cross-correlation method of the digital signals, it is possible to show that the hypocenters of such similar events lie on a plane which corresponds exactly to one of the nodal planes of the fault-plane solution (Deichmann, 1987b; Deichmann, 1990). Such repeated slip on pre-existing faults are seen both at upper-crustal depths of less than 10 km and in the lower crust at depths greater than 20 km.

Strike-slip focal mechanisms predominate, although several events have a strong normal faulting component. However, there is no systematic variation with depth of either type or orientation of the focal mechanisms. The orientation of the P- and T-axes are consistent with the well-established regional stress field characterized by a NNW–SSE oriented compression and an ENE–WSW oriented extension (Pavoni, 1980, 1987).

In order to visualize the focal depth distribution in relation to crustal temperatures, hypocenters and calculated isotherms are projected onto a vertical cross-section running roughly parallel to

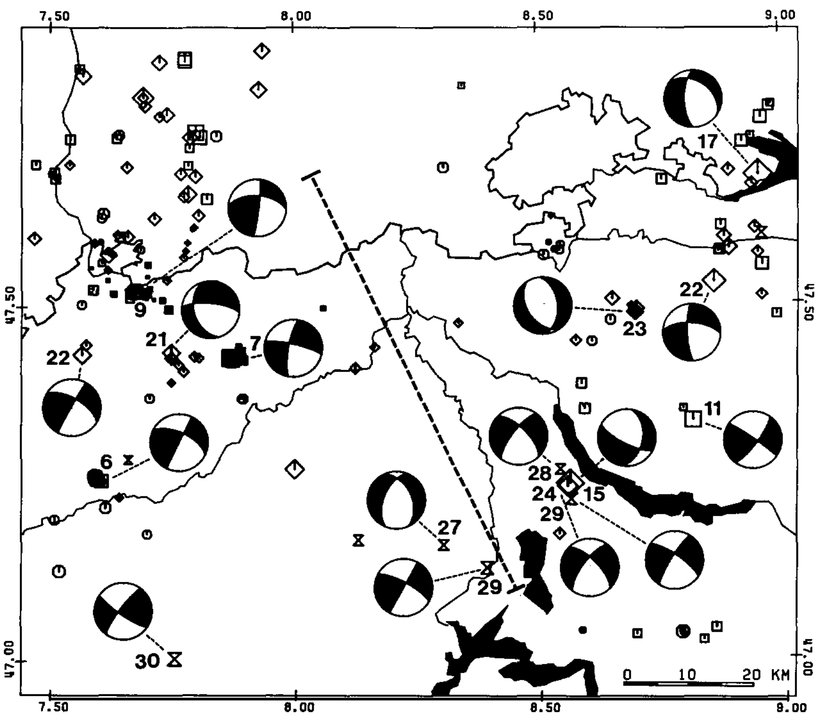

Fig. 2. Epicenter map with fault-plane solutions (lower hemisphere, equal area projections) for the period from January 1984 to December 1990. The numbers next to the events with focal mechanisms denote focal depths in kilometer. The different symbols correspond to different focal-depth ranges (see Fig. 3) and their size is proportional to the magnitude (between 0.5 and 4.2). A complete list of focal-mechanism parameters and detailed fault-plane solutions can be found in Deichmann (1987b, 1990).

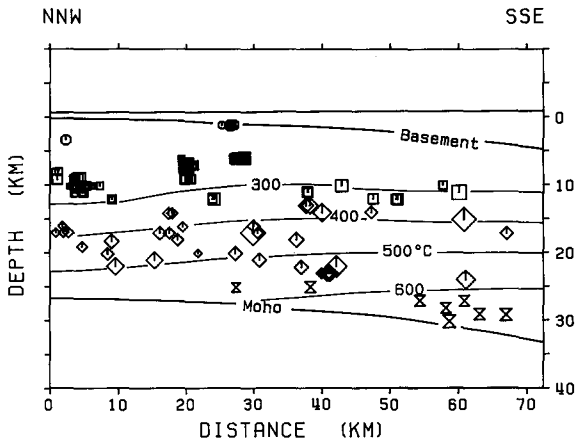

Fig. 3. Hypocenters below northern Switzerland for the period from January 1984 to December 1990, projected on to a cross-section roughly parallel to the dip of the Moho, together with the calculated isotherms (Moho depths after Maurer (1989) and temperatures from Deichmann and Rybach (1989)).

the dip of the Moho in a NNW–SSE direction (Fig. 3). Only those events whose focal depths could be calculated with an estimated confidence of better than ± 5 km are included in Fig. 3. Since, in general, focal depths for deeper earthquakes can be determined with more confidence than for shallower events, most focal depths greater than 20 km are accurate within ± 2 km. Of the excluded events (17 in all) a large part have calculated depths of less than 5 km. Consequently, the relative lack of hypocenters in the uppermost part of the crust in Fig. 3 may be the result of a bias in the dataset caused by the exclusion of poorly located events. Based on the interpretation of new refraction data (Maurer, 1989), the Moho at the SSE margin of the profile is about 2 km shallower than on previously published cross-sections. Thus, focal depths beneath the northern Alpine foreland of Switzerland extend fairly uniformly from close to the surface all the way down to the Moho. Moreover, the hypocenters in the lower crust are situated in a depth range where calculated temperatures are well above 450°C.

Lower-crustal earthquakes are not restricted solely to the area discussed in this paper: hypocenters at depths of 22 to 24 km are found below the southern Black Forest with a crustal thickness of 25 to 26 km (Bonjer et al., 1984, 1989), and earthquakes 20 to 30 km deep have been located both to the SW beneath the Franco–Swiss Jura and to the NE in the Lake of Constance region, where the Moho lies at depths of 30–35 km.

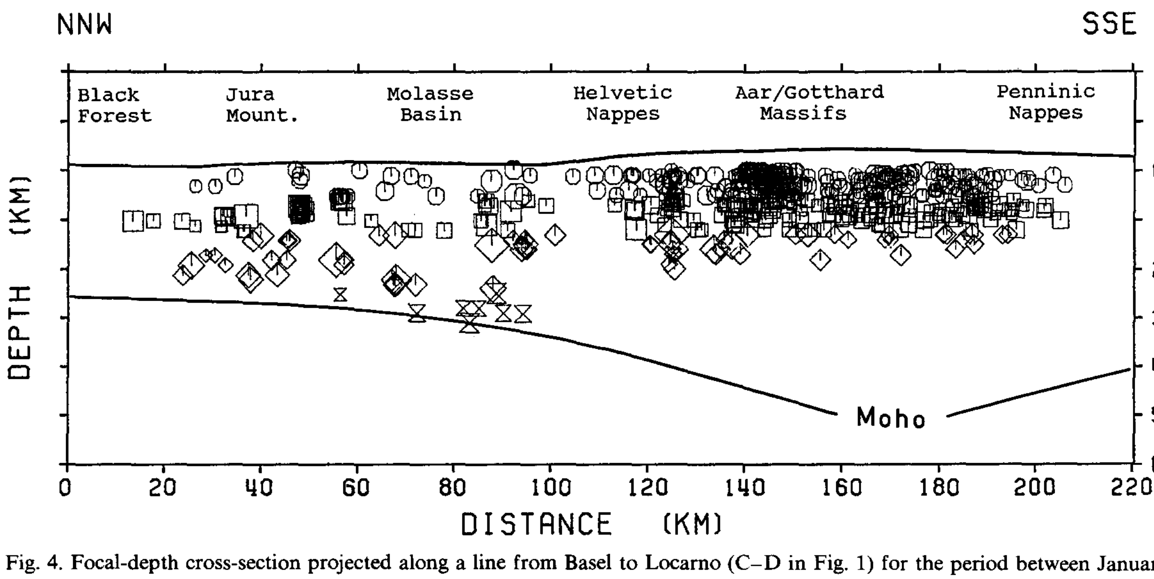

Fig. 4. Focal-depth cross-section projected along a line from Basel to Locarno (C–D in Fig. 1) for the period between January 1975 and September 1990 (after Deichmann and Baer, 1990). Only the more reliably located events are included here, selected according to the following criteria: GAP ⩽180°, number of observations ⩾11, minimum epicentral distance ⩽ 30 km, r.m.s. ⩽0.4 s. Moho depth after Mueller et al. (1980) and Maurer (1989).

However, further south at the front of the Helvetic Nappes and in the Alpine Massifs, where the Moho dips steeply to a depth of more than 50 km, the depth cut-off of seismic activity appears to lie between 15 and 20 km (Fig. 4). During a detailed seismicity survey with nine additional portable seismographs in the Helvetic and Penninic domains of the eastern Swiss Alps, out of a total of about 300 recorded events none of the hypocenters was located deeper than 13 ± 1 km (Roth et al., 1991). Moreover, there is no evidence either below the Alpine foreland or below the Alps for any seismicity in the upper mantle.

5. Discussion

5.1. The discrepancy between focal depths and models of crustal rheology

Both the low P-wave velocities and low Poisson's ratios found throughout the crust of northern Switzerland are generally associated with rocks of granitic or quartz-rich gneissic composition (Birch, 1960; Simmons, 1964; Kern, 1982). Some laboratory measurements of seismic velocities in granulite facies rocks of gabbroic (quartz-free) composition, when corrected for temperature effects, also closely match the values found here in the lower crust (e.g. Chroston and Evans, 1983). All the more mafic rocks are characterized by significantly higher P-wave velocities and Poisson's ratios, and are thus incompatible with the observed values. Therefore, the deformation of the crust beneath northern Switzerland is likely to be governed by the rheology of a combination of quartz and plagioclase.

Ord and Hobbs (1989) discuss in detail the mechanical behaviour of a three-layered crustal model for various thermal gradients. The rheology of their model is governed by quartzite in the upper crust, anorthosite in the middle crust, diopsidite in the lower crust and dunite in the upper mantle. Their calculations are based on the rheology of dry rock, except for the upper crust and mantle, which are assumed to be fluid saturated. Since, in our case, the mafic material seems to be absent in the lower crust or confined to at most a

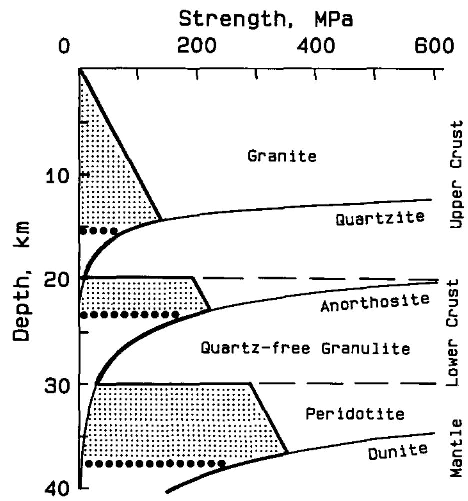

Fig. 5. Strength of the crust and upper mantle (adapted from Ord and Hobbs, 1989). The upper and lower crust and the upper mantle are governed by a quartz, plagioclase and olivine rheology respectively. The heavy dots indicate the lower bound of plastic instabilities in each layer. Earthquakes are expected to occur only in the stippled depth intervals.

1 or 2 km thick crust–mantle transition, and since the measured seismic velocities preclude a quartz-free crust above about 20 km, their model has been modified accordingly and is reproduced here in Fig. 5.

The heavy black line in Fig. 5 denotes the yield strength for the respective rocks as a function of depth. The straight portions of the line are derived from Byerlee's (1968, 1978) relation for brittle friction failure (the Coulomb–Navier criterion for shear failure in a rock-mechanical context), while the curved portions correspond to the power law for plastic flow for each rock type (Brace and Kohlstedt, 1980). The three intersections of the linear brittle behaviour with the flow-law curves mark three transitions from brittle to ductile behaviour. Assuming that plastic instabilities as well as brittle failure can generate earthquakes (Hobbs et al., 1986; Hobbs and Ord, 1988), the lower bounds of seismicity extend slightly into the ductile domains, as shown by the horizontal rows of heavy dots. For the calculations in the brittle domain, strike-slip faulting was assumed, in accordance with the observed tectonic regime. According to this model, but contrary to observation, the crust should be entirely ductile and thus aseismic between 16 and 20 km and below about 23 km, while earthquakes are expected to occur again in the upper mantle. Both the assumed temperature gradient of 20°C km^{-1} and the strain rate of 10^{-12} s^{-1} are on the conservative side. Whereas somewhat higher temperatures and lower strain rates would tend to eliminate the seismicity below the Moho, they would also even further reduce the depth intervals at which earthquakes can occur within the crust. If in addition one allows for a breakdown of Byerlee's law owing to a change from velocity weakening to velocity strengthening behaviour above some values of confining pressure and temperature (e.g. Tse and Rice, 1986), earthquakes would be restricted to the uppermost part of the crust (see fig. 9 of Ord and Hobbs, 1989).

One way to shift the brittle–ductile transition in any rheological model to greater depths, and thus reconcile the model with observations, is by increasing the resistance to ductile flow. As already mentioned, laboratory measurements of P-wave velocities on dry samples of the stronger mafic rocks are significantly higher than those observed in the crust of northern Switzerland. On the other hand, Hyndman and Klemperer (1989) suggest that a porosity of a few percent and the presence of fluids can lower seismic velocities by at least 5% (see also the review article by Christensen, 1989). Thus some fluid-saturated mafic rocks could be compatible with the observed P-wave velocities. However, the presence of pore fluids has a significant weakening effect on all rocks in the ductile regime, thus cancelling the added strength gained from a more mafic composition (e.g. Meissner and Strehlau, 1982; Ord and Hobbs, 1989). Plastic instabilities as a cause for earthquakes are possible only below a certain critical temperature which is only slightly higher than the transition temperature to ductile deformation. Thus such plastic instabilities can not contribute significantly to the extension of the depth range of expected seismic activity. Lower temperatures and higher strain rates, which would result in greater ductile strength, are difficult to reconcile with the temperatures inferred from heat flow measurements and with the regional strain rates estimated from geological evidence. One could postulate that deformation is not taken up uniformly, but that strain is concentrated in localized shear zones, where strain rates would be much higher than in the surrounding rock. However, if the estimates for the regional strain rates are correct, this would imply that strain rates in the rock around such shear zones and consequently also the overall shear stress would have to be very low, so that differential stresses would be insufficient to generate any earthquakes. Therefore, based on the foregoing arguments, greater ductile strengths than those considered in the model adapted from Ord and Hobbs (1989) are very unlikely.

5.2. The role of fluids in brittle failure

An alternative way of lowering the brittle–ductile transition consists of reducing the frictional resistance to brittle deformation. In general, earthquakes in the brittle zone of the crust are attributed to stick-slip frictional failure on pre-existing faults. In dry rock, the shear stress necessary to induce such failure is proportional to the normal stress acting on the fault. In the presence of fluids, on the other hand, the required shear stress is proportional to the effective stress, which is equal to the normal stress minus the fluid pressure (Hubbert and Rubey, 1959). As the fluid pressure approaches the least compressive stress in the crust, the resistance to frictional failure will approach zero, thus dropping well below the corresponding resistance to plastic flow. Consequently, even in a tectonic setting where the crust is subjected to relatively low differential stress, fluids under sufficiently high pressure can in principle shift the lower bound of the seismogenic layer to arbitrarily large depths.

The importance of fluids in the context of seismic faulting has been recognized for a long time. The exact role that fluids play is, however, still a subject of intense debate. Fluid migration within the crust can either be caused by the action of earthquakes (seismic pumping) or the earthquake process may act as a valve, which is triggered episodically by increases in fluid pressure (seismic valving) (Sibson, 1981). Either mechanism will entail a decrease of the effective stress and the fluids involved may be both liquids (e.g. water, hydrocarbons, magma) and gases (e.g. carbon dioxide, methane). Evidence for the involvement of fluids in faulting processes can be found in direct observations of the effects of earthquakes that rupture the surface, in the seismic activity associated with magmatic intrusions, from the examination of hydrothermal vein systems in relation to ancient faults and by actually triggering earthquakes in hydrofracturing experiments (e.g. Hill, 1977; Sibson, 1981).

It is clear that as long as fluids occupy a system of interconnected cracks which reach the Earth's surface, the fluid pressure will be roughly equal to the hydrostatic pressure throughout. As depth increases, the difference between hydrostatic pressure and the least compressive stress will necessarily become larger than the rigidity of the rock and the cracks will close. The rock rigidity, in this context, corresponds to its crushing strength, which effectively is the propping strength of the irregularities or asperities on the fault surfaces. In the absence of fluid pressures greater than hydrostatic, the tendency of cracks to close will be enhanced by the decreasing rigidity of rock as a consequence of increasing temperatures with depth. Below the depth where the cracks close, the crust will be impermeable to the downward migration of fluids. The exact level at which fractures become sealed, being dependent on the inherent strength of the rock, on the nature and orientation of the fractures and on the tectonic regime, will be subject to large lateral variations and is difficult to quantify (for a discussion of the depth extent and stability of tensional fractures in the crust (see e.g. Secor, 1965; Secor and Pollard, 1975). However, it is highly unlikely that interconnected fluid-filled cracks can extend from the surface all the way down into the lower crust.

Nevertheless, there is ample evidence from petrological considerations (e.g. Fyfe et al., 1978) and from seismic reflection and magnetotelluric results (e.g. Hyndman and Shearer, 1989) that fluids are present also in the lower crust. The source of these fluids may be water entrained to greater depths by a subsiding crust below a sedimentary basin or by a subducting slab, metamorphic reactions that release fluids in situ, or ascent of fluids from degassing of the mantle. In any case, deep crustal fluids, if present in sufficient amounts, will occupy discrete fluid domains, which are systems of interconnected cracks and pores, isolated from the upper-crustal hydrostatic fluid column and from other fluid domains by a pressure discontinuity (Gold and Soter, 1985). Such a pressure discontinuity will effectively act as a cap rock, in analogy to the impermeable layers found in sedimentary sequences, without however, necessarily corresponding to lithological changes. According to the model of Gold and Soter (1985), within each domain the least compressive stress, which controls the closing of fractures, will be balanced by the sum of the fluid pressure and the propping strength of the asperities on the interconnected fractures. Owing to the effect of the buoyant force of the fluid in the denser rock, fluid domains will have the tendency to migrate upwards, opening new cracks at the top as cracks at the bottom are squeezed shut. Either by the addition of more fluids or through the upward migration of whole domains, fluid pressure will aproach the least compressive stress and the effective stress will drop to zero. Under such circumstances, even the low shear stresses that can accumulate at depths normally associated with plastic flow will be released in form of an earthquake.

If the Earth's crust is assumed to be fluid saturated down to the Moho, it must be viewed as a stack of several fluid domains, with pore pressure increasing stepwise across layers of impermeable rock. Such a system is basically unstable: as a fluid domain migrates upward, it will penetrate the impermeable layer and ultimately join with the domain above it. As a consequence, the fluid pressure in the lower domain will drop, and, if no additional fluids are evolved in situ or enter the domain from below, the cracks will gradually close and the whole domain will vanish. If some shear stress acts on the fracture connecting two fluid domains, an earthquake invariably will be triggered as fluids pass through. By the combined effect of a drop in pressure in the immediate neighbourhood below the fault, of the deposition on the fault of minerals carried upward by the fluids and of the development of a fault gouge from frictional sliding, the fault is likely to heal in a much shorter time than it would take for all the fluid to pass from one domain to the other. From an analysis of microearthquake swarms involving events with similar signal forms, repeated slip on the same fault has been observed to occur with inbetween time intervals ranging from less than a minute to more than a year (Deichmann, 1987b, 1990). As the fractures at the bottom of the lower domain gradually close or as new fluids are added to the domain, fluid pressure will gradually rise again, until the least compressive stress acting at the bottom of the next higher domain is reached and the faulting process can be repeated. In this way fluids in the crust can migrate upwards from one domain to the next by a sum of episodic bursts until they reach the uppermost domain, which extends to the Earth's surface.

6. Conclusions

This model of fluid migration in the lower crust, as proposed by Gold and Soter (1985), not only extends Sibson's (1981) idea of seismic valving to greater depths, thus offering a plausible explanation for the occurrence of lower-crustal earthquakes, but also provides a simple mechanism for the occurrence of earthquake clusters caused by repeated slip on the same fault. Whether fluids are a significant constituent of the lower crust in northern Switzerland is, however, still an open question. Several possibilities exist to investigate this problem. Electromagnetic measurements are very sensitive to the permeability and fluid content of rocks and, if areas can be found which are free of man-made disturbances, can provide very informative results even down to lower crustal depths. Modelling impedance contrasts of seismic reflection data is another possibility to estimate whether the observed lower-crustal reflectivity could be caused by fluid inclusions. A further parameter sensitive to the presence of fluids is Poisson's coefficient. The effect of pore fluids on the ratio of P- to S-wave velocities has been investigated by Nur and Simmons (1969) and by Spencer and Nur (1976) for a limited range of temperatures and pressures. Since shear-wave velocity seems to depend critically on pore pressure it is difficult to extrapolate their results to lower-crustal conditions. Further laboratory measurements of seismic velocities not only at high confining pressures but also at high temperatures and in the presence of pore pressure near or equal to the confining pressure are necessary to clarify this important point (see Christensen, 1989). Such measurements would in fact add a new dimension to the interpretation of S-wave experiments in many areas of the world.

One of the most puzzling problems concerns the variations of maximum focal depths between the Alpine foreland and the culmination of the Alps (Fig. 4). Considering the lower heat flow in the Alps (e.g. Bodmer, 1982), the observations are exactly the reverse of what commonly accepted models of crustal rheology would predict. At any rate, the seismological evidence clearly demonstrates an important lateral heterogeneity in rheological behaviour between the crust below the Alps and below the northern foreland. Considering the tectonic history of the two regions, this heterogeneity is not surprising. However, there is no simple explanation for the observed variations in maximum earthquake focal depths in terms of either temperature distribution or seismically visible structural differences. As mentioned before, the temperature extrapolation down to the base of the crust have been calculated on the assumption of a steady-state situation. A conclusive explanation should involve a thermal model which includes possible transient effects associated with the dynamics of the Alpine orogeny. Moreover, identification of petrological composition should be reevaluated in the light of future laboratory results from seismic velocity measurements on rock samples at high temperatures and pressures and in the presence of high-pressure pore fluids. Nevertheless, the model of the lower-crustal earthquakes below the northern foreland involving high-pressure fluids remains at present the most plausible explanation. The different rheological behaviour of the crust below the Alps and the northern foreland might then be the result of the different tectonic evolution of the two regions. The uplift of the Alps and subsidence of the Molasse basin could have caused significant lateral fluid migration, either in the crust or upper mantle (G.

Ranalli, personal communication, 1990), or result in different metamorphic conditions and thus different rates of fluid release in the lower crust.

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References

Birch, F., 1960. The velocity of compressional waves in rocks to 10 kilobars, part 1. J. Geophys. Res., 65: 1083–1102.

Bodmer, P., 1982. Beitraege zur Geothermie der Schweiz. Ph.D. Thesis, Zürich, 201 pp.

Bonjer, K.-P., Gelbke, C., Gilg, B., Rouland, D., Mayer-Rosa, D. and Massinon, B., 1984. Seismicity and dynamics of the upper Rhinegraben. J. Geophys., 55: 1–12.

Bonjer, K.-P., Faber, S. and Apopei, I., 1989. Seismizität als Zugang zu räumlichen und zeitlichen Anomalien der Spannung der Lithosphäre. In: Spannung und Spannungsumwandlung in der Lithosphäre, Sonderforschungsbereich 108, Berichtsband 1987–1989, Universität Karlsruhe, pp. 65–131.

Brace, W.F. and Kohlstedt, D.L., 1980. Limits on lithospheric stress imposed by laboratory experiment. J. Geophys. Res., 85: 6248–6252.

Byerlee, J.D., 1968, Brittle–ductile transition in rocks. J. Geophys. Res., 73(14): 4741–4750.

Byerlee, J.D., 1978. Friction of rocks. Pure Appl. Geophys., 116: 615–626.

Carter, N.L. and Tsenn, M.C., 1987. Flow properties of continental lithosphere. Tectonophysics, 136: 27–63.

Chen, W.-P. and Molnar, P., 1983. Focal depths of itracontinental and intraplate earthquakes and their implications for the thermal and mechanical properties of the lithosphere. J. Geophys. Res., 88: 4183–4214.

Christensen, N.I., 1989. Pore pressure, seismic velocities and crustal structure. In: L.C., Pakiser, and W.D. Mooney, (Editors), Geophysical Framework of the Continental United States. Geol. Soc. Am., Mem., 172: 783–798.

Chroston, P.N. and Evans, C.J., 1983. Seismic velocities of granulites from the Seiland petrographic province, N. Norway: implications for Scandinavian lower continental crust. J. Geophys., 52: 14–21.

Deichmann, N., 1987a. Focal depths of earthquakes in northern Switzerland. Ann. Geophysicae, 5B: 394–402.

Deichmann, N., 1987b. Seismizität der Nordschweiz, 1983–1986. Nagra Technischer Bericht, NTB 87–05. NAGRA, Baden, 40 pp.

Deichmann, N., 1990. Seismizität der Nordschweiz, 1987–1989, und Auswertung der Erdbebenserien von Günsberg, Läufelfingen und Zeglingen. Nagra Technischer Bericht, NTB 90–46. NAGRA, Baden, 52 pp.

Deichmann, N. and Rybach, L., 1989. Earthquakes and temperatures in the lower crust below the northern Alpine foreland of Switzerland. In: R.F. Mereu, St. Mueller and D.M. Fountain (Editors), Properties and Processes of the Earth's Lower Crust. Am. Geophys. Union, Geophys. Monograph 51, IUGG 6: 197–213.

Deichmann, N. and Baer, M., 1990. Earthquake focal depths below the Alps and northern Alpine foreland of Switzerland. In: R. Freeman, P. Giese, and St. Mueller (Editors), The European Geotraverse: Integrative Studies. European Science Foundation, Strasbourg, pp. 277–288.

Diebold, P. and Müller, W.H., 1984. Szenarien der geologischen Langzeitsicherheit: Risikoanalyse für ein Endlager für hochaktive Abfälle in der Nordschweiz. Technischer Bericht, NTB 84–26. NAGRA, Baden, 110 pp.

Fadaie, K. and Ranalli, G., 1990. Rheology of the lithosphere in the East African Rift System. Geophys. J. Int., 102: 445–453.

Fyfe, W.S., Price, N.J. and Thompson, A.R., 1978. Fluids in the Earth's Crust. Elsevier, Amsterdam, 383 pp.

Gold, T. and Soter, S., 1985. Fluid ascent through the solid lithosphere and its relation to earthquakes. Pure Appl. Geophys., 122: 492–530.

Griesser, J.-C. and Rybach, L., 1989. Numerical thermohydraulic modelling of deep groundwater circulation in crystalline basement: an example of calibration. In: A.E. Beck, L. Stegena and G. Garven (Editors), Hydrogeological Regimes and their Subsurface Thermal Effects. Am. Geophys. Union, Geophys. Monograph 47, IUGG 2, pp. 65–74.

Hill, D.P., 1977. A model for earthquake swarms. J. Geophys. Res., 82: 1347–1352.

Hobbs, B.E., Ord, A. and Teysier, C., 1986. Earthquakes in the ductile regime? Pure Appl. Geophys., 124: 309–336.

Hobbs, B.E. and Ord, A., 1988. Plastic instabilities — implications for the origin of intermediate and deep focus earthquakes. J. Geophys. Res., 93: 10521–10540.

Hubbert, M.K. and Rubey, W.W., 1959. Role of fluid pressure in the mechanics of overthrust faulting. Bull. Geol. Soc. Am., 70: 115–166.

Hyndman, R.D. and Klemperer, S.L., 1989. Lower-crustal porosity from electrical measurements and inferences about

composition from seismic velocities. Geophys. Res. Let., 16: 255–258.
Hyndman, R.D. and Shearer, P.M., 1989. Water in the lower continental crust: modelling magnetotelluric and seismic reflection results. Geophys. J. Int., 98: 343–365.
Ito, K., 1990. Regional variations of the cutoff depth of seismicity in the crust and their relation to heat flow and large inland-earthquakes. J. Phys. Earth, 38: 223–250.
Kern, H., 1982. Elasic-wave velocity in crustal and mantle rocks at high pressure and temperature: the role of the high-low quartz transition and of dehydration reactions. Phys. Earth and Planet Int., 29: 12–23.
Kobayashi, Y., 1977. A relationship between the distribution of focal depth of micro-earthquakes and surface heat flow in southwestern Japan and central Japan. Proc. Symp. Earthquake Prediction Res. 1976, National Committee of Geophysics and Seismological Society of Japan, pp.184–193 (in Japanese with English abstract).
Maurer, H.-R., 1989. Die Struktur der Erdkruste unter dem Schweizerischen Alpennordrand. Diploma Thesis, Institute of Geophysics, ETH-Zürich, 108 pp.
Meissner, R. and Strehlau, J., 1982. Limits of stresses in the continental crusts and their relation to the depth-frequency distribution of shallow earthquakes. Tectonics, 1: 73–89.
Meissner, R. and Kusznir, N.J., 1987. Crustal viscosity and the reflectivity of the lower crust. Ann. Geophysicae, 5B: 365–374.
Mueller, St., Ansorge, J., Egloff, R. and Kissling, E., 1980. A crustal cross section along the Swiss Geotraverse from the Rhinegraben to the Po Plain. Eclogae Geol. Helv., 73: 463–483.
Mueller, St., Ansorge, J., Sierro, N., Finckh, P. and Emter, D., 1987. Synoptic interpretation of seismic reflection and refraction data. Geophys. J.R. Astron. Soc., 89: 345–352.
Müller, W.H. and Briegel, U., 1980. Mechanical aspects of the Jura overthrust. Eclogae Geol. Helv., 73: 239–250.
Nur, A. and Simmons, G., 1969. The effect of saturation on velocity in low porosity rocks. Earth and Planet. Sci. Let., 7: 183–193.
Ord, A. and Hobbs, B.E., 1989. The strength of the continental crust, detachment zones and the development of plastic instabilities. Tectonophysics, 158: 269–289.
Pavoni, N., 1980. Crustal stresses inferred from fault-plane solutions of earthquakes and neotectonic deformation in Switzerland. Rock Mechanics, Suppl. 9: 63–68.
Pavoni, N., 1987. Zur Seismotektonik der Nordschweiz. Eclogae Geol. Helv., 80: 461–472.
Pfiffner, O.A. and Ramsey, J.G., 1982. Constraints on geological strain rates: arguments from finite strain states of naturally deformed rocks. J. Geophys. Res., 87: 311–321.
Pfister, M., 1990. Gemeinsame Auswertung von Erdbeben-, Refraktions- und Reflexionsdaten in der Nordschweiz. Diploma Thesis, Institute of Geophysics, ETH-Zürich, 104 pp.
Ranalli, G. and Murphy, D.C., 1987: Rheological stratification of the lithosphere. Tectonophysics, 132: 281–295.
Roth, P., Pavoni, N., Deichmann, N., 1991. Seismotectonics of the eastern Swiss Alps and evidence for precipitation induced variations of seismic activity. Tectonophysics (European Geotraverse, Special Issue), in press.
Rybach, L., Eugster, W. and Griesser, J.-C., 1987. Die geothermischen Verhälnisse in der Nordschweiz. Eclogae Geol. Helv., 80: 521–534.
Secor, D.T., 1965. Role of fluid pressure in jointing. Am. J. Science, 263: 633–646.
Secor, D.T. and Pollard, D.D., 1975. On the stability of open hydraulic fractures in the earth's crust. Geophys. Res. Lett., (2)11: 510–513.
Shudofsky, G.N., Cloeting, S., Stein, S., Wortel, R., 1987. Unusually deep earthquakes in East Africa: constraints on the thermo-mechanical structure of a continental rift system. Geophys. Res. Lett., 14, 7: 741–744.
Sibson, R.H., 1981. Fluid flow accompanying faulting: field evidence and models. In: D.W. Simpson and P.G. Richards (Editors), Earthquake Prediction: an International Review. Am. Geophys. Union, Maurice Ewing Series, 4. pp.593–603.
Sibson, R.H., 1982. Fault zone models, heat flow and depth distribution of earthquakes in the continental crust of the United States. Bull. Seismol. Soc. Am., 72: 151–163.
Sierro, N., 1989. Regionale Struktur der Erdkruste in der Nordschweiz. Ph.D. Thesis, Institute of Geophysics, ETH-Zürich, 203 pp.
Simmons, G., 1964. Velocity of shear-waves in rocks to 10 kilobars, 1. J. Geophys. Res., 69: 1123–1130.
Smith, R.B. and Bruhn, R.L., 1984. Intraplate extensional tectonics of the eastern Basin and Range: Inferences on structural style from seismic reflection data, regional tectonics, and thermal-mechanical models of brittle-ductile deformation. J. Geophys. Res., 89: 5733–5762.
Spencer, J.W. and Nur, A.M., 1976. The effects of pressure, temperature, and pore water on velocities in Westerly granite. J. Geophys. Res., 81: 899–904.
Tse, S.T. and Rice, J.R., 1986. Crustal earthquake instability in relation to the depth variation of frictional slip properties. J. Geophys. Res., 91(B9): 9452–9472.
Wong, I.J. and Chapman, D.S., 1990. Deep intraplate earthquakes in the western United States and their relationship to lithospheric temperatures. Bull. Seismol. Soc. Am., 80, 3: 589–599.