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Seismicity and dynamics of the Upper Rhinegraben

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Abstract. In this paper we present the results of a 10-year period (1971–1980) of research on the seismicity of the Upper Rhinegraben. Our investigations are exclusively based on instrumentally recorded earthquakes with local magnitudes between $0.5 < M_L < 5$. The increase in the number of high-gain seismic stations during the past 2 decades improved the quality of the observations considerably, thus allowing detailed recognition of the spatial distribution of the earthquake loci in focal areas deduced from the analysis of historical events. No region, regarded up to now as aseismic, revealed itself as seismic, not even at the level of microearthquakes. Excluding the focal area of the Swabian Jura, the northernmost and southernmost parts of the Upper Rhinegraben show the highest degree of seismic activity. The middle part of the Rhinegraben, between Strasbourg and Karlsruhe, reveals only modest activity, somewhat in contrast to the historical record. The number of earthquakes increases towards the east of the river Rhine relative to the west. An even more pronounced asymmetry is shown in the southern graben by different maximum focal depths perpendicular to the strike of the Rhinegraben. In the Vosges mountains and in the graben proper, depths of 13 and 16 km, respectively, are not exceeded. Maximum depths down to about 20 km are found in the Black Forest. No earthquake was detected in the lower gabbroic crust or in the mantle. The maximum focal depth seems to be governed by variations in the temperature-depth distribution.

Fault plane solutions of more than 30 earthquakes demonstrate that the seismic dislocations take place predominantly as strike slip mechanisms in the southern graben area whereas normal faulting prevails in the north. In the northern graben, most of the seismic dislocations occur on fault segments striking N30° W whereas in the south a strike of N20° E or N60° W (the conjugate direction) is dominant. Furthermore, the fault plane solutions indicate a clockwise rotation of the principle stress directions from north to south by about 40°.

Key words: Seismicity – Continental rift – Stress in the crust – Seismogenic zone – Upper Rhinegraben

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Introduction

The Upper Rhinegraben was initially formed as a tensional feature in mid-Eocene to early Miocene times, apparently following pre-existing zones of weakness in the basement (Illies, 1965, 1977). Starting in the middle Pliocene, the tensional graben was remodelled into a broad shear zone with dominant strike-slip character (Illies, 1965; Illies and Greiner, 1978). It has been shown by Ahorner (1970) and by Ahorner and Schneider (1974), by various fault plane solutions, that strike-slip dislocations govern the dynamics in most parts of the graben at the present time.

The seismicity of the Upper Rhinegraben is low in comparison to other continental graben systems (Ahorner, 1975). Although the historical seismic record covers about the past 1,000 years (e.g. Sieberg, 1940a, b; Sponheuer, 1952, 1969), no great earthquake is documented. The stron-

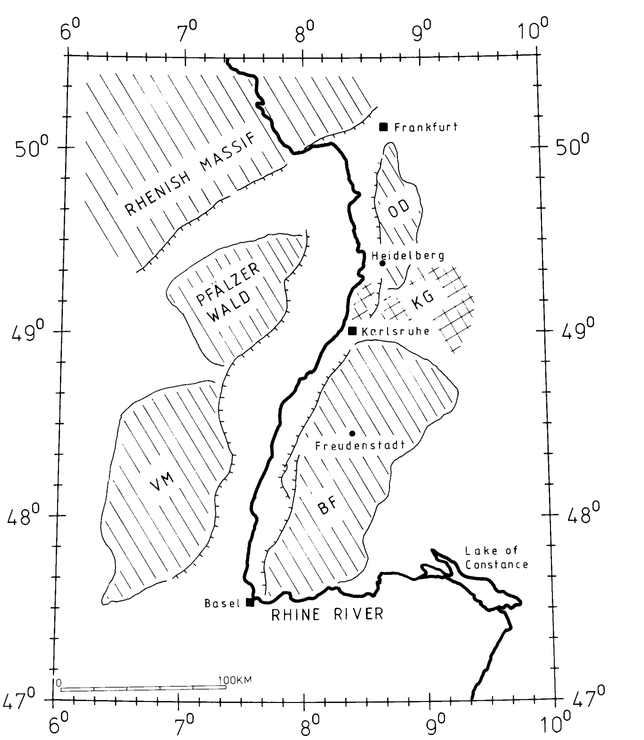

Fig. 1. General map of the Upper Rhinegraben area (OD = Odenwald, BF = Black Forest, VM = Vosges mountains, KG = Kraichgau)

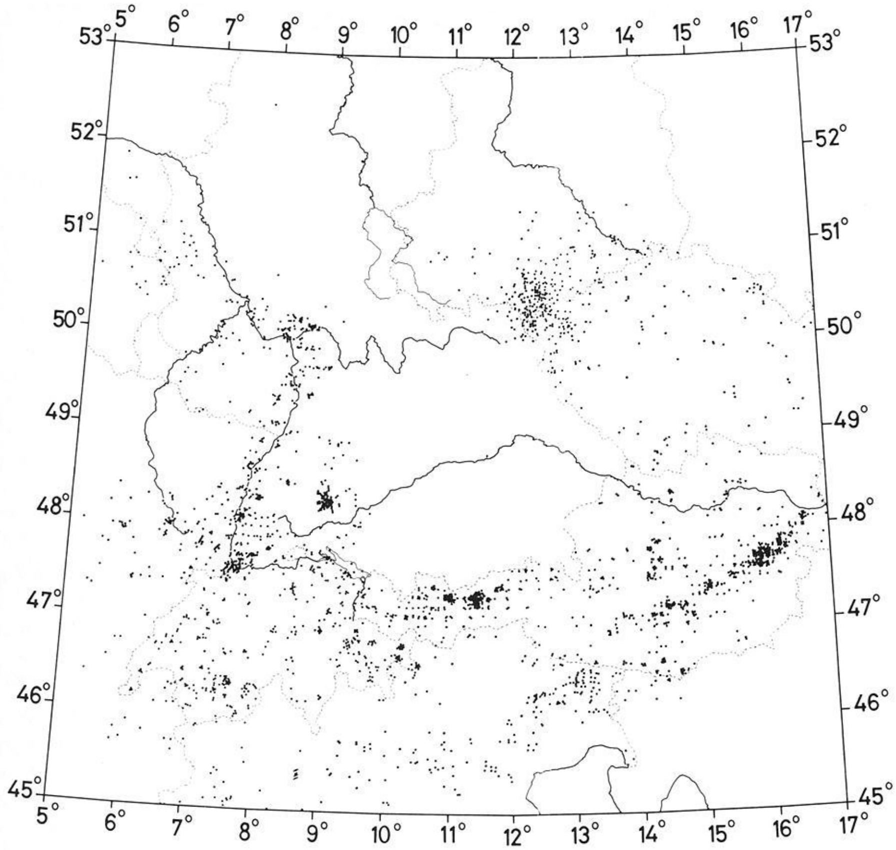

Fig. 2. Seismicity of central Europe (1000–1970). (References: Sieberg, 1940a, b; Sponheuer, 1952; Hiller et al., 1967; Ahorner, 1970; Sponheuer et al., 1968; Caputo and Postpischl, 1972; Prochazkova, 1974; Gutdeutsch and Aric, 1976; Pavoni, 1977; and others

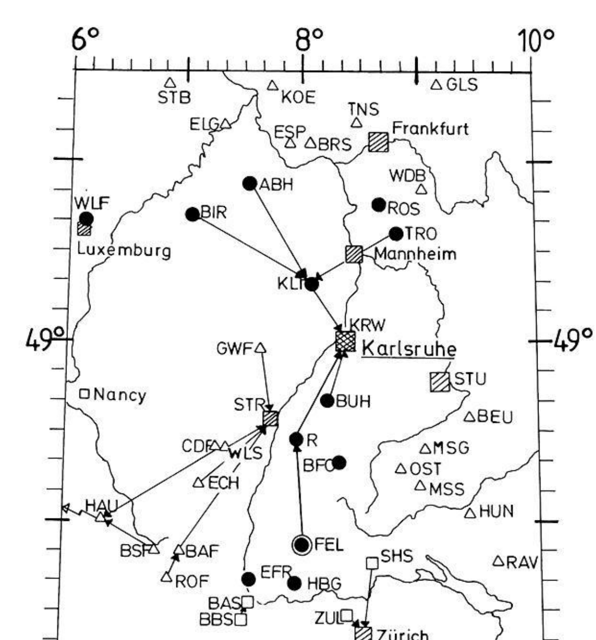

Fig. 3. The seismic stations in the Rhinegraben area. Three characters: seismic stations. *Black circles*: stations of the seismic network operated by the Geophysical Institute, University of Karlsruhe. *Triangles* and *squares:* stations operated by other Institutes. *Lines with solid arrows:* telemetry lines

gest event with a maximum intensity (MSK) of at least $I_o = \mathrm{IX}$ (Mayer-Rosa and Cadiot, 1979) occurred in 1356 A.D. near Basel, in the southern end of the Rhinegraben (Fig. 1). Knowledge of the seismicity up to 1960 is based mainly on macroseismic information. Installation of seismic stations started around the turn of the century. Due to poorly known crustal and upper mantle structure, low magnification and great station separations, the intrumentally determined focal coordinates were of scanty precision and often less accurate than the macroseismic determinations (e.g. Sponheuer, 1960). However, these data have already allowed the Upper Rhinegraben to be recognized as the seismic tie between the Alps in the south and the Rhenish Massif in the north. The Rhinegraben also separates the South German block in the east from the Lorrainese block in the west (Fig. 2), which both show little to negligible seismic and orogenic activity. On the other hand, the precision of the "historical" data in general is insufficient to study the seismic dislocations in detail. The installation of a dense network of high-gain seismic stations in the Rhinegraben area (Fig. 3) and the progress of crustal structure studies in the past 20 years (e.g. Prodehl et al., 1976) provided a sufficient basis for a thorough investigation of the seismicity in this region for the first time.

Data

Magnitude-frequency relationships

Figure 3 reproduces the short-period seismic network in the Rhinegraben area. The majority of these 37 stations have

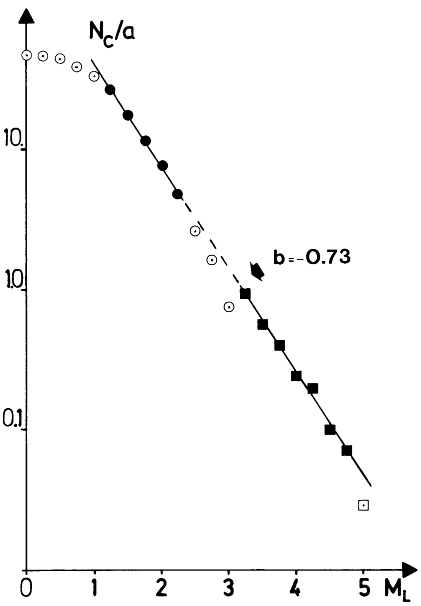

Fig. 4. Cumulative magnitude-frequency distribution for Upper Rhinegraben earthquakes (circles: instrumental data 1971–1979; *squares:* macroseismic and instrumental data 1900–1970)

high magnifications of about 100,000 (1 Hz) and were set up by French, German and Swiss institutes between 1965 and 1980. A considerable lowering of the mean detection threshold, to at least a magnitude of $M_L = 1.2$, was achieved by this network. Lippert (1979) demonstrated that this threshold holds for the entire Upper Rhinegraben area, using the events recorded between 1971 and 1978. The local magnitudes are determined from coda length scaling laws deduced by Lippert (1979) for the Rhinegraben area. As it has been shown recently by digital data of the station FEL (Gruber, 1983), the coda-magnitudes are in surprisingly good agreement with simulated Wood-Anderson magnitudes when applying Richter's (1935) procedure. Nevertheless, because of the different geological settings in California and the Upper Rhinegraben, a bias with respect to the original local California magnitudes cannot be excluded (Bonjer et al., 1982). For the graben proper, Fig. 4 (Bonjer, 1979b) demonstrates that the magnitude-frequency distribution of the historical events, labelled as macroseismic data (e.g. Ahorner, 1975; Leydecker and Harjes, 1978), can be extended by short-term but high-sensitive observations, labelled as instrumental data. For the whole graben area, the historical, mainly macroseismic data, and the new instrumental data yield the same b-value of -0.73. By extrapolating both data sets towards one another, they show compatible annual occurrence rates. However, as shown by Lippert (1979), the $b = -0.73$ value is only an average value which can vary significantly (i.e. $-0.5 > b > -1.0$) if the graben area is divided into smaller focal regions. However, in performing such a subdivision, care has to be taken to define the appropriate location and size of every focal region so that it covers an area with a uniform seismotectonic regime. The strong variation of b-values and the annual occurrence rate of earthquakes (with respect to a fixed magnitude) found by Lippert (1979) might either reflect true seismotectonic differences or an artifact of an inappropriate choice of the focal regions and/or insufficient data. The latter possibility might be true, given the rather short period of 9 years of instrumental observations. The proper discrimination of individual focal regions is not only important for the assessment of seismic hazard but may also serve as a tool in understanding the present-day dynamics of the Upper Rhinegraben.

The average P-wave velocity model

In parallel with the increase in the number of high-gain seismic stations, the structure of the crust and upper mantle has been studied in detail during the past 2 decades (e.g. Prodehl et al., 1976; Prodehl, 1981). On the basis of the crustal models deduced by Edel et al. (1975) from refraction profiles of different regions in the Rhinegraben area, an average model for the southern graben was constructed by Gelbke (1978). From the results of P_g-time term analysis, Gelbke decreased the basement velocity of the model by Edel et al. down to 6.0 km/s and the velocity gradient, in the depth range between 15 and 20 km, was slightly lowered to explain crustal phases of deeper events recorded at distances of up to about 140 km. This average model was again refined by Gilg (1980), who replaced the remaining velocity gradient zones between 15 and 20 km, as well as between 20 and 26 km, by layers with constant average velocities in order to match crustal phases which often dominate the records up to distances of about 500 km. In this way, the middle crust, regarded as characteristic for a continental graben by Mueller (1977), disappeared as a separate crustal unit.

In the construction of an average crustal model for the northern part of the Upper Rhinegraben, Gilg's model was modified slightly by increasing the depth of the crust-mantle boundary by 1.5 km to a depth of 27.5 km. The uppermost mantle velocities were taken from Ansorge et al. (1979). A v_p/v_s-ratio of $\sqrt{3}$ was used throughout this study. In Fig. 5, the velocity model, corresponding travel times and P-arrival times of the Sierentz mainshock of July 15, 1980 are shown to demonstrate the agreement between observed and computed travel times. For distances greater than 260 km (i.e. outside the Rhinegraben area) the upper mantle velocity model is no longer valid, but the average crustal model may still be used.

The focal coordinates are calculated with the computer program HYPGRAD (Gelbke, 1978), a modified version of HYPO-71 (Lee and Lahr, 1972). The program HYPGRAD utilizes a travel-time subroutine which allows the use of arbitrary velocity-depth distributions (e.g. gradients, velocity inversions). In addition to the first P- and S-arrivals, later P- and S-phases, including overcritically reflected waves, can be used to augment the number of observations and to considerably increase the reliability of the focal parameter determinations (Gelbke, 1978). For most of the events their accuracy is better than ± 3.5 km. This has been verified by relocating calibration shots at various locations within the Rhinegraben area (Gelbke, 1978).

The epicentre distribution

Figure 6 summarizes the epicentres of the period 1971–1980. During this time the southern and northern end of the Upper Rhinegraben were considerably more active than the middle part between Strasbourg and Karlsruhe. On the other hand, the historical epicentre distribution (Hiller et al., 1967) shows a seismic activity for the

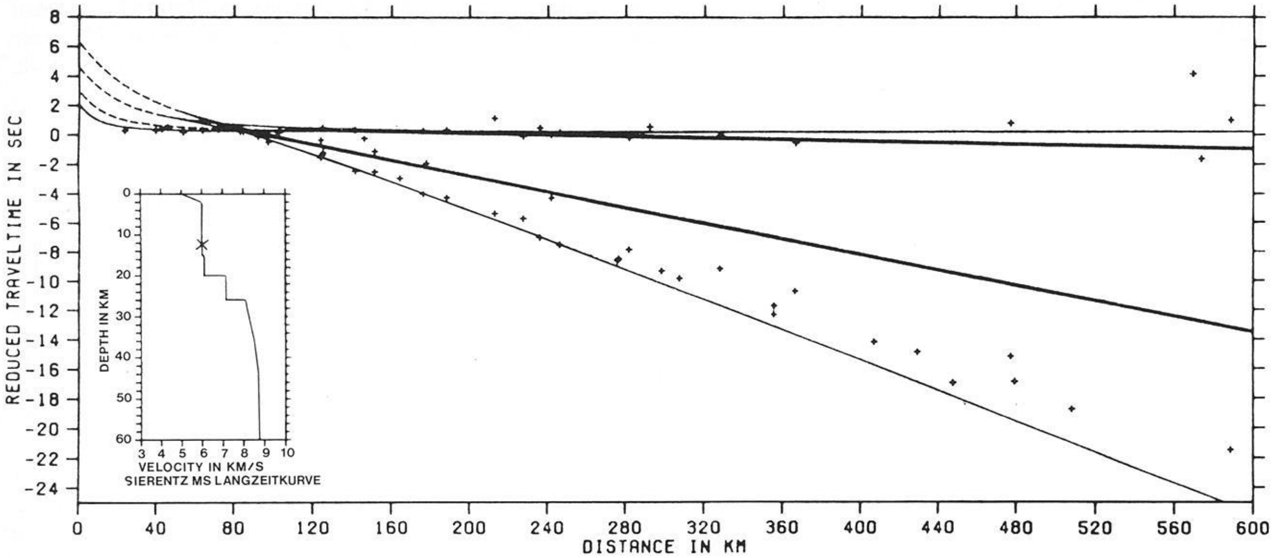

Fig. 5. Average velocity-depth model and travel times used for hypocentre determinations (+ = observed travel times of the Sierentz main shock July 15, 1980 – after Gruber, 1983). Reduced travel times ($v_r = 6.0$ km/s) as a function of epicentre distance

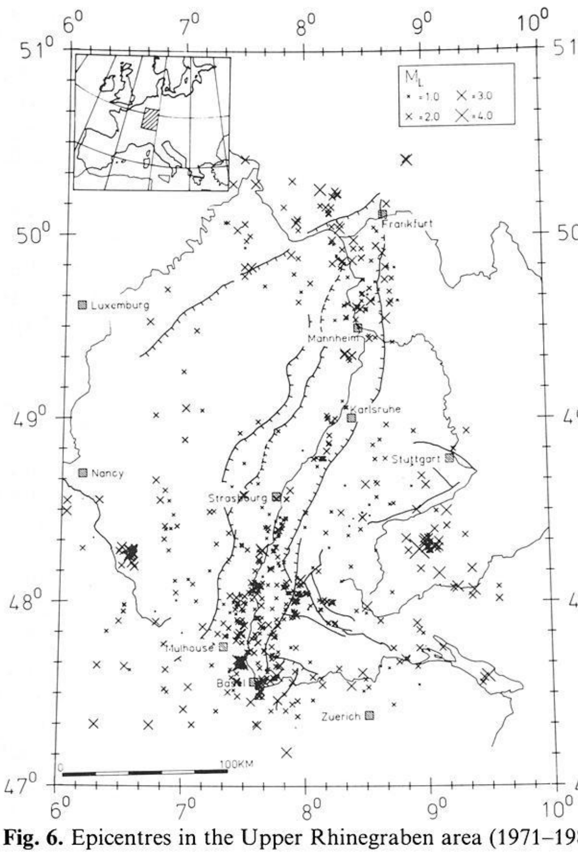

Fig. 6. Epicentres in the Upper Rhinegraben area (1971–1980)

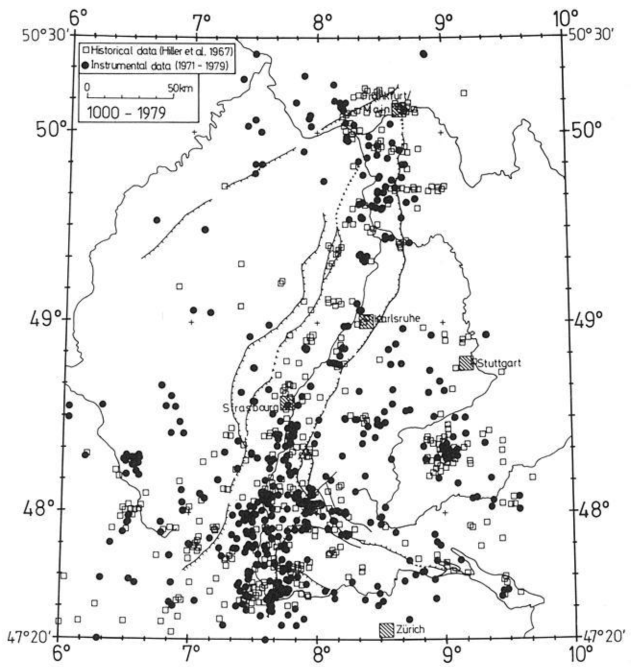

Fig. 7. Comparison between historical and present day seismic activity. *squares:* historical (mainly macroseismic) epicentres (1000–1970); *circles:* instrumentally determined epicentres (1971–1980)

middle part which is of about the same order as at the graben ends (Fig. 7). This may be seen more convincingly when the seismic energy flux is considered (Fig. 1 of Hägele and Wohlenberg, 1970). Thus the short observation time may account for the low activity reported here.

The areas with the highest seismicity are found in the Swabian Jura and near Saulgau, south of Stuttgart between the Danube and Neckar rivers. The data are taken from *Earthquakes in the Federal Republic of Germany* (Seismological Central Observatory Graefenberg, 1971–1980). It is partly due to these two areas that the number of earthquakes east of the Rhine river is greater than west of it. Such an asymmetry can be recognized in the historical data, too (Fig. 7). Neither the geological setting nor the topography (Illies, 1974) give any hint in furnishing an explanation for this asymmetric distribution of foci. Although no hypocentre has been detected in the lower crust near the Moho (see next section), it is worthwhile noticing that the border between the more and the less active areas closely follows the crest line of the updomed crust-mantle boundary (Fig. 8) with a strike direction of about N30°E (Gelbke, 1978; Bonjer, 1979a).

One of the remarkable features of the seismicity pattern

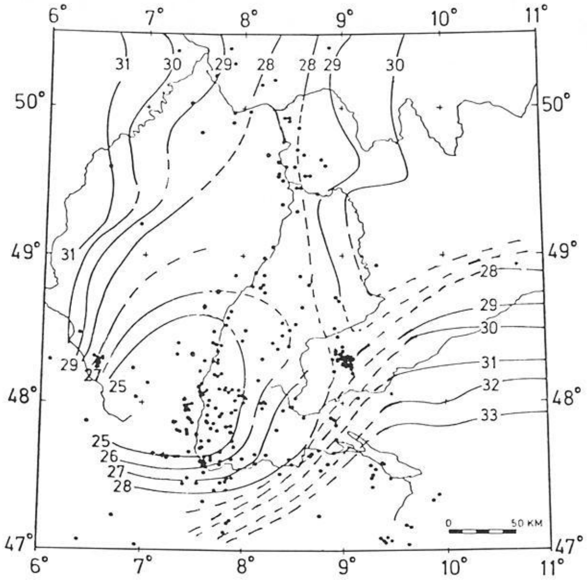

Fig. 8. Epicentres (1971–1980) and depth contour-lines of the crust-mantle boundary (Edel et al., 1975)

is connected with the western and eastern main border fault systems of the graben. Both fault systems create only negligible seismic activity, if at all. The low activity of the bordering faults has already been stated by Hiller et al. (1967), on the basis of epicentres from 1021 to 1965 (Fig. 7, open squares). As regards the instrumental data presented in this paper, already this lack of activity seems to be obvious at the western border faults on rough inspection of the epicentre map (Fig. 6). Apparently, such a deficit of seismicity does not exist at the eastern border faults, if only the epicentre distribution is considered. However, it emerges, when discussing the focal depths (Figs, 9, 10), that even the eastern border faults do not show earthquakes, at least down to a depth of about 10 km. Furthermore, the most prominent faults at the graben flanks do not show enhanced seismic activity.

It remains an open question whether or not the foci apparently following the strike of the Bonndorf graben zone between Freiburg (48°00′N/7°50′E) and Lake Constance (Fig. 1) may be attributed to this fault system, as favoured by Gelbke (1978) and Bonjer and Fuchs (1979), or to a number of parallel arranged conjugate faults. Fault-plane solutions (Fig. 11) show that both directions are possible. At the southernmost end of the Rhinegraben (Basel-Dinkelberg area, 47°35′N/7°40′E) a short lineation of epicentres with an abrupt change of strike from N to NE is found (Fig. 6). As a sufficient number of reliable fault-plane solutions is not yet available, further discussion is not justified at this time.

The distribution of epicentres seems to be even more complicated in the graben proper (Fig. 6), especially in the northern and southern parts. Here the foci are spread out, approximately over the entire width of the graben, and no clear alignment of epicentres exceeding a length of 10 km is noticeable. If there were fault segments active over a length of 10 km or more, they should have been detected due to the sufficient accuracy of the localizations.

The most pronounced seismic activity is located in the western Swabian Jura (48°17′N/9°01′E) and the Epinal (48°18′N/6°30′E) – Remiremont (47°59′N/6°31′E) regions, roughly situated at the same distance from the Rhinegraben "hinge-lines" (Mueller, 1970) of the uplifted graben shoulders. Comparing historical and instrumental data (Fig. 7), a shift of activity from south (Remiremont) to north (Epinal) can be seen. Quite a similar migration of the strongest events towards the north was found in the western Swabian Jura, starting around 1900 (Brenner, 1966; Schneider, 1971). Another point of comparison between

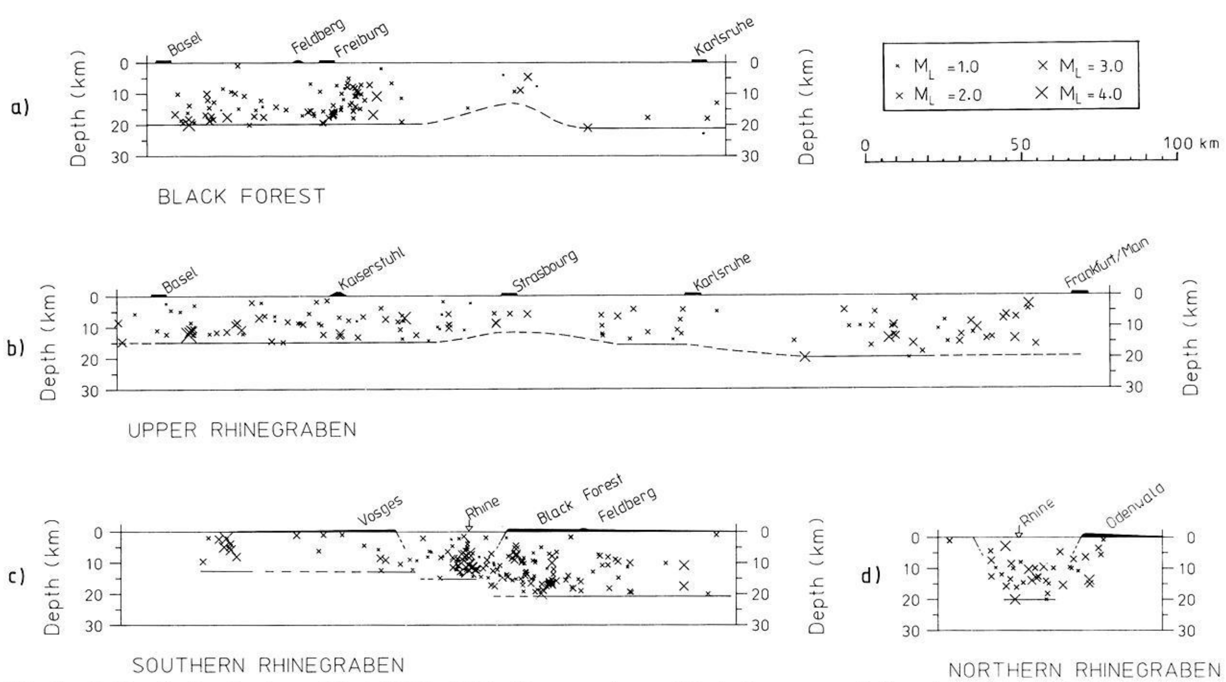

Fig. 9a–d. Depth distribution of foci (1971–1980). Cross section **a**: Black Forest, parallel to the strike of the graben; **b**: graben proper, average strike direction N20°E; **c**: perpendicular to the strike of the southern graben. Hypocentres between Strasbourg and Basel were projected; **d**: northern graben, strike direction approximately EW. The boundaries of greatest focal depths are marked by *solid lines*

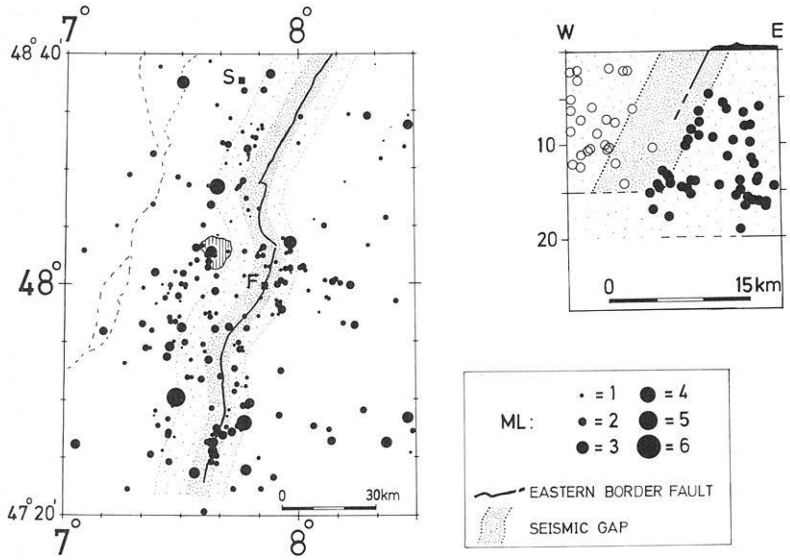

Fig. 10. The zone of seismic quiescence at the eastern main border fault system. *Left:* epicentre map with eastern main border fault system (*solid line*). *Right:* Hypocentres in the *stippled areas* are projected onto a vertical master section

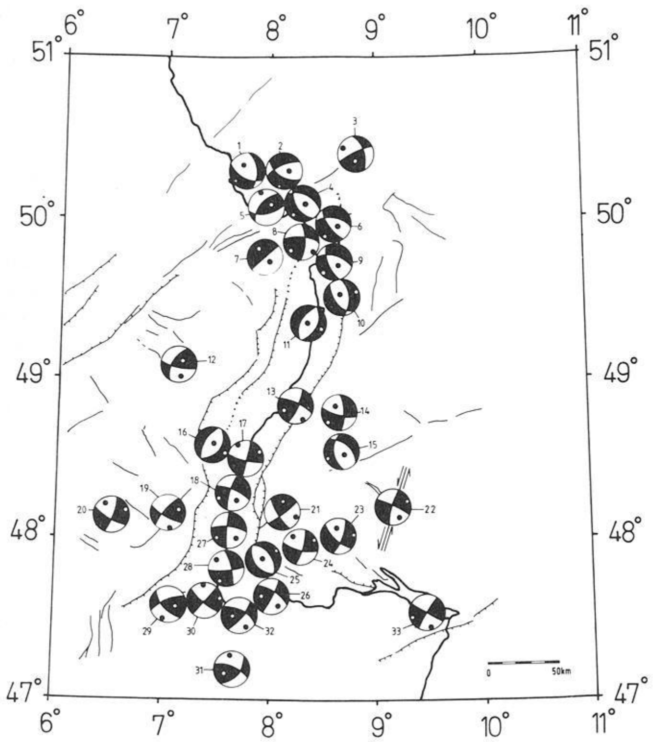

Fig. 11. Fault-plane solutions (Wulff projection) in the Upper Rhinegraben area. (References: [1, 5, 6, 8, 9, 21, 23, 24, 27] = Ahorner et al., 1983; [3] = Neugebauer and Tobias, 1977; [4] = Baier and Wernig, 1983; [22] = Turnovsky, 1981; [31, 33] = Mayer-Rosa and Pavoni, 1977; [2, 7, 10–20, 25, 26, 28–30, 32] = this paper). *Black quadrants:* compression. *White quadrants:* dilatation. *Black dots:* *P*-axes. *White dots:* *T*-axes

these two areas is the dominance of strike-slip mechanisms (Fig. 11). Even though the seismotectonics of the western Swabian Jura have been studied in detail during the past decades (e.g. Schneider, 1968; Schick, 1968, 1970; Haessler et al. 1980; Turnovsky, 1981; Turnovsky and Schneider, 1982; Illies, 1982; Illies and Baumann, 1982), no convincing correlation of the strike directions of seismic shear zones at depth with those of surface geology was found. As far as the corresponding shear zone between Remiremont and Epinal is concerned, no statement can be made yet because of the scarcity and uncertainty of the seismological observations.

The comparison of historical and instrumental data (Fig. 7) reveals three facts. Firstly, local source areas exhibit a significant geographical consistency for the past millennium. Secondly, the short-time, high-sensitivity observations show that most of these focal areas are marked by persistent microseismicity which reflects the general pattern of the Rhinegraben seismicity. This pattern has already been discussed by Bonjer and Fuchs (1974). Statistical analyses of $N(M_L)$ corroborate this result when historical data are matched with instrumental observations (Fig. 4). Thirdly, the Kraichgau area, in the middle part of the Rhinegraben (Fig. 1) north of Karlsruhe, which was already considered to be aseismic by Hiller et al. (1967) and Schneider (1968), is still seen to be a zone of seismic quiescence. No earthquake with a local magnitude $M_L > 0.5$ was detected there.

Focal depths

The accuracy of focal-depth determinations has been increased due to the enlargement of the seismic network, to much more detailed knowledge of the crust and uppermost mantle structure, and to the additional use of later body-wave phases, which give strong constraints when no near-by observations are available (Gelbke, 1978). Focal depths based on instrumental data for certain time spans and graben regions have already been reported by Gelbke (1978), Bonjer (1979a), Gilg (1980) and Gruber (1983). In Fig. 9 a presentation of the focal depths for the years 1971–1980 in five different cross-sections is given. The most conspicuous details are presented in Fig. 9c, which shows a cross-section perpendicular to the strike of the southern graben, roughly half-way between Strasbourg and Basel. All hypocentres of that region are projected onto this cross-section. The greatest focal depths of about 20 km are found in the Black Forest, whereas in the graben proper and in the Vosges mountains focal depths of 16 and 13 km, respectively, are not exceeded. This maximum depth of about 20 km is not only found in the southern, but also in the northernmost part of the Black Forest at the latitude of Karlsruhe (Fig. 9a). Shallower depths seem to prevail in the middle

Table 1. Fault-plane solutions of earthquakes 1971–1980

No.	Region	Date	Lat. N	Long. E	h (km)	M_L	Plane 1		Plane 2		*P*-Axis		*T*-Axis		*B*-Axis	
							Strike	Dip	Strike	Dip	AZ	Plunge	AZ	Plunge	AZ	Plunge
1	St. Goar	Dec. 31, 1980	50.18	7.70	6	2.8	113	64SW	170	42NE	336	57NW	227	13SW	130	30SE
2	Idstein	Mar. 7, 1977	50.25	08.14	13.3	3.6	62	42NW	118	60SW	80	44NE	185	8SW	280	30NW
3	Echzell	Nov. 4, 1975	50.41	8.87	11	3.6	63	80NE	162	50NE	300	20NW	195	34SW	51	57NE
4	Wiesbaden	Nov. 4, 1979	50.04	8.30	5	3.2	140	35NE	153	56SW	90	78E	237	12SW	338	6NW
5	Geroldstein	May 8, 1977	50.07	7.94	11	3.1	45	45NW	45	45SE	137	0SE	0	90	45	0NE
6	Darmstadt	Jun. 7, 1979	49.91	8.66	6	2.3	100	42NE	145	58SW	106	65SE	215	10SW	308	24NW
7	Undenheim	Feb. 11, 1981	49.85	08.28	7.4	1.9	44	88NW	60	10SE	130	55SE	318	35NW	224	3SW
8	Oppenheim	Jul. 31, 1979	49.87	8.40	10	2.9	2	85NE	94	75NE	135	8SE	228	14SW	25	75NE
9	Lorsch	Apr. 9, 1977	49.65	8.56	15	2.6	105	50NE	153	49SW	130	63SE	41	0NE	310	25NW
10	Großsachsen	Jul. 12, 1971	49.55	08.67	13.7	3.3	181	52NE	142	45SW	330	70NW	73	5NE	165	20SE
11	Speyer	Feb. 28, 1972	49.36	08.34	20	3.2	201	65NE	195	25NW	300	78NW	108	21SE	200	2SW
12	Hambach	Dec. 6, 1976	49.06	07.01	11.7	2.3	28	29NW	103	80SW	168	28SE	38	49NE	277	28NW
13	Rastatt	Dec. 18, 1971	48.81	08.21	9.6	2.1	25	80SE	290	70NE	156	7SE	249	21SW	48	68NE
14	Enzklösterle	Oct. 12, 1980	48.65	08.50	20	3.3	100	60SW	177	70NE	325	45NW	234	3SW	150	50SE
15	Dornstetten	Nov. 23, 1979	48.46	08.48	4.6	2.3	160	50SW	305	45NE	135	72SE	233	3NW	325	18NW
16	Molsheim	Dec. 16, 1977	48.59	07.52	8.9	2.9	30	54SE	198	35NW	326	79NW	115	10SE	206	6SW
17	Rhinau	Oct. 27, 1979	48.29	7.65	7	3.9	12	90	102	80SW	326	8NW	57	8NE	192	80SW
18	Wyhl	Jul. 26, 1976	48.17	07.59	6.7	1.7	190	70NW	296	54NE	147	41SE	246	10SW	345	46NW
19	Schlucht	Aug. 18, 1980	48.13	07.02	8.1	1.6	37	65NW	117	70SW	168	3SE	77	34NE	262	58SW
20	Epinal	Nov. 12, 1974	48.28	06.58	2.2	3.9	20	80SE	119	70SW	338	23NW	69	7NE	180	68SE
21	Waldkirch	Jan. 27, 1979	48.12	7.97	17	3.1	60	82SE	163	60SW	110	15SE	12	26NE	225	58SW
22	Albstadt	Setp. 3, 1978	48.29	9.03	7	5.7	20	75NW	113	77NE	157	19SE	66	2NE	333	71NW
23	Wolterdingen	Febr. 29, 1976	47.98	8.51	11	3.1	12	60SE	133	50SW	340	55NW	77	6NE	168	33SE
24	Waldau	Jan. 15, 1976	48.00	8.22	10	2.1	13	80SE	106	60SW	328	23NW	62	5NE	170	68SE
25	Heitersheim	Mar. 5, 1975	47.87	07.64	17.5	2.1	145	40SW	330	50NE	285	83SW	59	5NE	150	4SE
26	Schopfheim	Aug. 15, 1977	47.65	07.79	16.8	1.8	22	80SE	120	40NE	150	30SE	255	30SW	30	40NE
27	Kaiserstuhl	Apr. 30, 1978	48.10	7.63	11	2.8	4	80NW	105	45NE	133	39SE	242	22SW	355	45NW
28	Bantzenheim	Nov. 7, 1975	47.79	07.44	11.5	2.1	90	70S	170	60NE	120	15SE	218	8SW	120	54SE
29	Delle	Feb. 11, 1978	47.55	7.06	8	2.4	128	60SW	62	54NW	186	4SW	92	50SE	279	41NW
30	Sierentz	Jul. 15, 1980	47.68	7.48	12	4.7	40	90	130	80SW	354	8N	86	8E	220	80SW
31	Langenthal	Apr. 26, 1974	47.20	07.90	20	3.0	110	65NE	45	35SE	354	18NW	240	51SW	90	45E
32	Dinkelberg	May 21, 1974	47.60	07.77	19.9	3.7	42	72SE	115	41NE	162	18SE	275	44W	55	39NE
33	Bodensee	Mar. 2, 1976	47.6	9.4	20	3.7	30	0	120	0	346	0NW	76	0NE	–	90

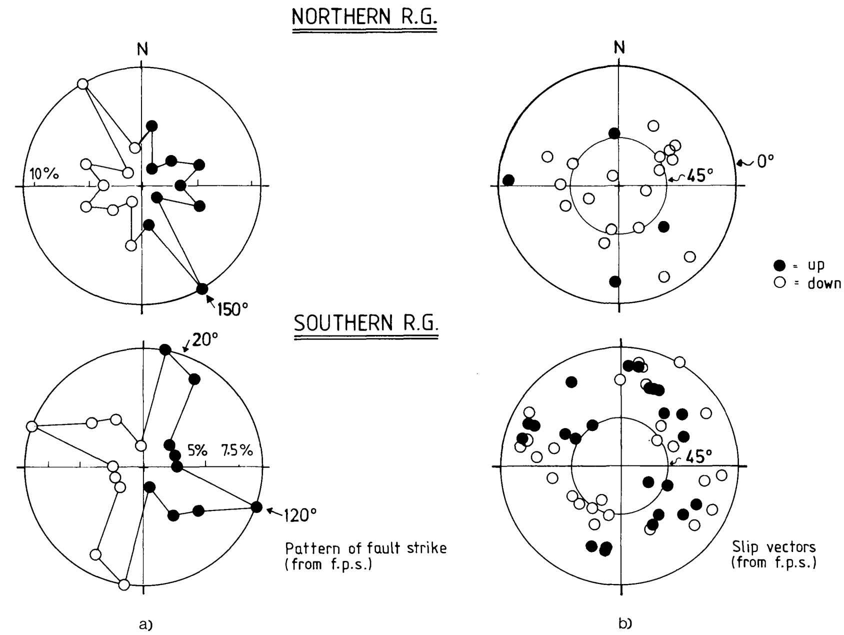

Fig. 12a + b. a Rose diagram of fault and auxiliary planes; **b** distribution of slip vectors deduced from fault-plane solutions (Wulff net)

part around Freudenstadt (Fig. 9a). In contrast to the southern graben with greatest depths of 16 km, the northern part shows focal depths down to 20 km (Fig. 9b, d). Only few data are so far available from the central part of the graben between Karlsruhe and Strasbourg but they indicate a restriction of the foci to the uppermost 10–15 km of the crust (Fig. 9b).

Since the focal depths of about 20 km practically coincide with the sharp velocity increase at the boundary between upper and lower crust (Fig. 5), we checked the error in focal depth due to inaccuracies in the depth of this discontinuity. It was varied in both directions with appropriate velocity modifications in order to correct for travel times. In this way we could confirm these great focal depths, i.e. no earthquake occurred below 20 km. Hence, the lower crust in the Rhinegraben area is free of seismic activity. (On the basis of prevailing macroseismic investigations, this fact has already been mentioned by Ahorner et al. (1972).) This important result will be discussed in detail later. As documented in Fig. 9, the thickness of the seismogenic part of the crust varies laterally, but no fine structure of the depth distribution of the hypocentres is recognizable with certainty. Whether or not the relatively small number of events in the uppermost 5 km is only an artifact (earthquakes with small magnitudes are likely to be recorded by an insufficient number of stations resulting in unreliable depth determinations), or is caused by the seismotectonic regime, can only be resolved by a greater concentration of seismic stations in areas with supposedly shallow foci.

In discussing the epicentre map (Fig. 6), it was mentioned that the main western and eastern border faults are lacking seismic activity. This is hard to accept with regard to the eastern fault, but it emerges from Fig. 9c that this might indeed be the fact. Gelbke (1978) was able to prove convincingly the lack of seismicity by correcting the crooked course of the eastern masterfault and projecting the hypocentres onto cross-sections perpendicular to the local strike of the fault and assembling these sections into a master-section. From Fig. 9 it is quite obvious that the eastern master fault is free of earthquakes down to a depth of approximately 10 km. The lateral extent of the earthquake-free zone is about 8 km. Due to an insufficient number of events a similar experiment cannot be done on the western border fault of the graben.

Focal mechanisms

It has often been questioned whether fault-plane solutions derived from microearthquakes can serve as a clue in understanding the seismotectonic regime of a broader region, e.g. the Upper Rhinegraben, or whether they reflect solely the seismic dislocation at a particular site. Whatever the case, we are forced to use microearthquakes when investigating the seismic dislocations and the regional stress field acting in the Rhinegraben area. Otherwise, we would not have sufficient data in an acceptable short period of time (see Fig. 4) for a reliable interpretation.

Several studies of focal mechanisms in the Rhinegraben area were performed in the past (e.g. Schneider et al., 1966; Schneider, 1968; Ahorner and Schneider, 1974; Mayer-Rosa and Pavoni, 1977; Bonjer, 1979b; Bonjer and Fuchs, 1979; Haessler et al., 1980; Rouland et al., 1980; Turnovsky, 1981). In this paper, older data were omitted, mainly because of their poor quality, and as far as possible replaced by more recent solutions. However, the former solutions did display the general trend of the regional stress field (Schneider et al., 1966; Ahorner, 1970; Ahorner and Schneider, 1974).

Figure 11 summarizes the fault-plane solutions deduced from earthquakes which occurred during the period 1971–1980. Details are given in Table 1. The type of seismic dislocation changes from north to south at a latitude of about 49° N. In the north, normal faulting is predominant, whereas in the south, strike-slip mechanisms partly prevail, associated with small dip components (Bonjer, 1979b). In the middle section of the graben, the fault-plane solutions are indicative of a greater amount of dip. From south to north, i.e. with increasing distance from the Alps, the strike of the P-axes rotates counter-clockwise by about 40°. In the southern graben area the strike of the P-axes closely approaches the north-south direction. A further detail should be stressed. At a latitude of about 48° N, the strike of the P-axes remains constant from east to west, i.e. we find the same seismotectonic regime in the western Swabian Jura, the Black Forest, the graben proper and the Vosges mountains. Despite the ambiguity of associating the actual rupture plane (the same holds for the slip vector) with one of the two nodal planes, it is worthwhile constructing a rose diagram with the strike of the nodal planes in order to look for prevailing directions. In Fig. 12a such rose diagrams are shown for the northern and the southern graben. Note that there is a striking difference between the north and the south. In the north, only one direction, of N150° E, is dominant. This direction is nearly perpendicular to the strike of the SE border fault of the Rhenish Massif and can be followed by fault-plane solutions, alignment of epicentres (Ahorner et al., 1983) and tectonic features through the Massif up to the Lower Rhine Embayment. In the southern graben area, two prominent conjugate directions of N20° E (Rhenish) and N120° E (Hercynian) emerge.

All these directions well exceed the scatter of the data. The slip vector, another seismotectonic parameter, can also be extracted from the fault-plane solutions. Again, this was performed separately for the northern and the southern graben areas (Fig. 12b). In the north the majority of slip vectors dip more or less steeply with an average trend towards SW. In the southern graben area, the situation seems to be chaotic. Even a trend of the slip is not recognizable, and the number of directions downwards and upwards is about the same. It is impossible to calculate an average annual slip rate of the seismic dislocation unless one can define source areas with uniform slip. Seismic slip rates, calculated simply by adding up the amount of slip (without taking into account the azimuth, dip angle and sense) are grossly in error in a region like the southern Rhinegraben.

Discussion

In the magnitude-frequency distribution of earthquakes from 1971 to 1978 for the area of the graben proper (after Lippert, 1979),

$$\log(N_c/a) = 2.24 - 0.73\, M_L \qquad 1.0 < M_L < 2.5 \qquad (1)$$

(N_c/a = cumulative frequency per annum of earthquakes with magnitudes greater than M_L),

no fore- or aftershocks have been included, when they could be recognized. The magnitude-frequency relationship (1) is in excellent agreement with the one deduced by Ahorner (1975) for the period 1700–1969; thus even a short time of observation can reveal the general pattern of seismicity in the Rhinegraben. On the other hand, it should be kept in mind that both relationships are only evaluated for a range of $1\,^1/_2$ orders of magnitude.

On average, a total of about only 23 earthquakes with local magnitudes equal to or greater than $M_L = 1.2$ are recorded in the Rhinegraben proper per annum. Because of the above-mentioned restriction, in deducing the $\log N_c(M)$-distribution, this is indicative of up to 23 different focal sites being active once a year. Even when fore- and aftershocks are included, the average annual number of events with $M_L > 0.5$ is less than 150. This is not very impressive for an area of about 16×10^4 km^2. Furthermore, since a local magnitude greater than $M_L = 5$ is seldom reached, and omitting the exceptional Basel event of 1356 A.D., we can regard the Rhinegraben as an area of low seismicity compared with other continental rift systems (e.g. Ahorner, 1975; Fairhead and Stuart, 1982).

As was discussed in the preceding sections, no convincing evidence of seismic activity in the southern area was found to be associated with the main border faults and the master faults of the graben shoulders. Furthermore, the earthquakes do not follow the contact zone between granite and gneiss in the southern Black Forest (Gelbke, 1978). Delays of P_g-waves crossing the eastern border faults led Gelbke (1978) to the conclusion that the uppermost 5–10 km of this zone are strongly fractured and, therefore, unable to accumulate sufficient strain to produce earthquakes. If this zone can be loaded at all by the regional stress field, strain should be released by creep processes or "nano-earthquakes" which are below the detection level of the existing seismic network. A different explanation is given by Illies et al. (1981) for the absence of seismic activity. They argue that the frictional resistance of faults dipping at about 63° is high enough to withstand the shear stress produced by the actual regional stress field. However, this model does not explain the occurrence of earthquakes which are below a depth of approximately 10 km. At this depth, due to an increase of confining pressure, the frictional resistance might be even greater if temperature and pore pressure effects are neglected. On the other hand, we do not want to deny either the inaccuracy of the position of the foci by projecting them onto a vertical master section as shown in Fig. 10, or a possible mislocation due to lateral velocity changes, e.g. by a crushed zone (Gelbke, 1978; Gilg, 1980). Anyway, it is a challenge for future research to map the spatial extent and the rheology of this zone of seismic quiescence more precisely, in order to answer the implicit question of whether we have to regard this zone as being highly strained or just the opposite.

In the northern part of the graben, adjacent to the Odenwald south of Frankfurt (Fig. 1), the eastern border fault does not show a zone of seismic quiescence, at least not as pronounced as in the southern graben (Gilg, 1980). A corresponding zone at the western border fault system, as suggested by the epicentre distribution, cannot be traced due to the small number of events at adjacent sides of the fault.

Illies has frequently pointed out that active faulting in the graben proper is found mainly along faults striking about 170°–180° (N-S), oblique to the general trend of the graben (e.g. Illies and Greiner, 1978). They act as en-échelon faults, Riedel shears or secondary faults in terms of Chinnery's model (1966) with left lateral sense of movement trending around NS, and right lateral sense in the conjugate direction. A dense pattern of these second-order shears intersects the graben along its whole length (e.g. Fig. 8 of Illies and Greiner, 1978). Superposition of this figure and the epicentre distribution in Fig. 6 of this paper show that practically all epicentres coincide with one of the Riedel shears or even that several foci join the same fault segment. The problem now arises as to whether this is just a fortuitous coincidence or whether in fact the actual seismic activity mimics the pattern of the Riedel shears at deeper crustal levels as advocated by Illies (e.g. Illies et al., 1981). However, it has already been pointed out how much care has to be taken, if epicentre maps are interpreted without taking the focal depths into account. In the southern graben, at least, the depths are known for a considerable number of earthquakes. But what is known about the dip and the extension with depth of the particular Riedel

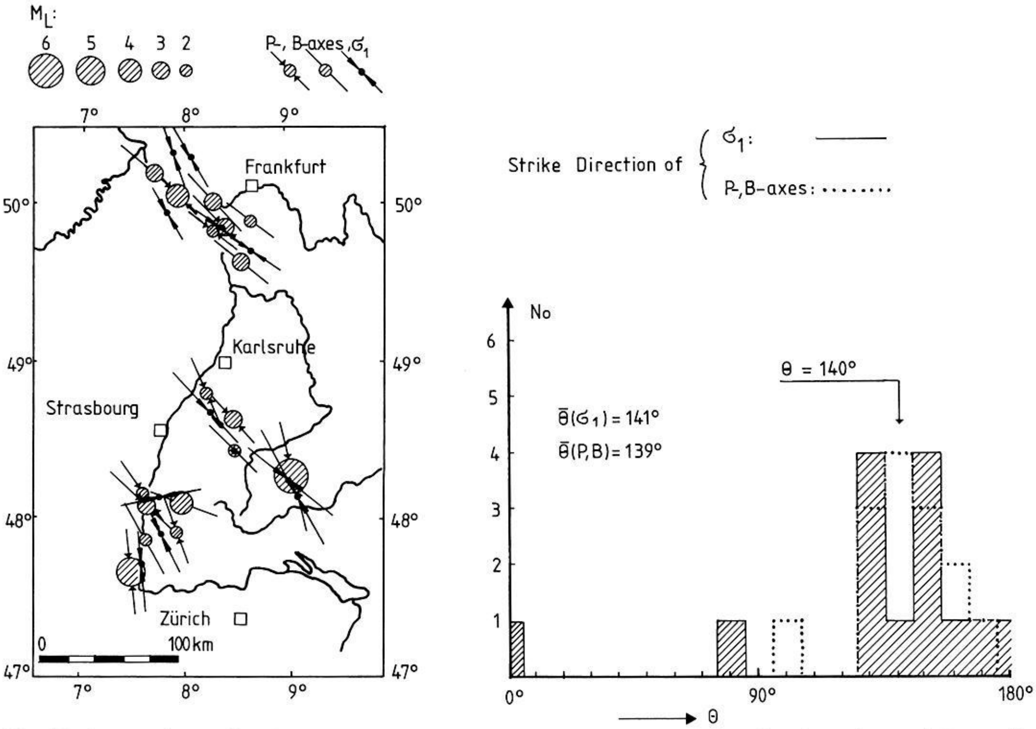

Fig. 13. Comparison of in situ stress measurements (σ_1-direction) and strike direction of *P*- and *B*-axes from fault-plane solutions

shears? At least, the strike directions are known. We will come back to this point, when we discuss the focal mechanisms.

According to Fig. 9 the thickness of the seismogenic part of the crust varies laterally in the Upper Rhinegraben area. The lower boundary was found by mapping the greatest focal depths. If they represent maximum depths down to which earthquakes are occurring in the Rhinegraben area, we can assume that such a boundary marks the transition from brittle, or frictional, to ductile behaviour of rocks. In the Black Forest, this transition zone coincides with the rapid increase of *P*-velocity from slightly more than 6 km/s to about 7 km/s, starting at a depth of approximately 20 km (Edel et al., 1975).

A *P*-wave velocity of about 7 km/s in the depth range of 20–25 km can be assigned to gabbroic rocks (Ringwood and Green, 1966). Above this transition zone, the average velocity of 6.0–6.1 km/s of the upper crust can be attributed to granitic rocks, and in the lowermost few kilometres velocities of up to 6.5 km/s in the original model of Edel et al. (1975) correspond to granodioritic rocks (Edel, 1975). Thus the die-off of seismic activity at a depth of about 20 km in the Black Forest may be linked to the change from granitic to gabbroic rocks (Bonjer, 1979b). In the graben proper, the same velocity-depth distribution is found, as in the Black Forest. However, here the drop-off in the number of earthquakes occurs in the granitic part of the crust at a depth of about 16 km. Upwarping of the isotherms from the Black Forest towards the Rhinegraben (Mueller and Rybach, 1974) results in a temperature of about 460° C at 16 km depth in the graben. The heat flow in the Vosges mountains is supposed to be even higher than in the graben (e.g. Gelbke, 1978), and accordingly the maximum focal depths are even shallower. It seems obvious that the maximum focal depths are intrinsically connected with the temperature at those particular depths (Bonjer, 1977; Gelbke, 1978).

Combining Byerlee's law (Byerlee, 1968) with flow laws for quartz or olivine to account for temperature effects and correction of pore pressure, Brace and Kohlstedt (1980) derived upper limits for the lithospheric stress as a function of depth. The introduction of the quartz flow law results in a very abrupt drop in frictional resistance, close to zero, at shallow depths between 10 and 20 km. This shallower depth offers an explanation for the rather sharp transition from seismic to aseismic behaviour, beneath the Black Forest, Upper Rhinegraben and Vosges mountains. The model of Brace and Kohlstedt (1980) has been successfully applied to several seismically active regions outside subduction zones (e.g. Meißner, 1980; Sibson, 1982; Meißner and Strehlau, 1982). Although Kirby (1980) has questioned the use of the quartz flow law for the middle and lower crust, it can at least give a lower bound to the crustal shear stresses (Brace and Kohlstedt, 1980).

As mentioned earlier, most events which have been investigated for the deduction of fault-plane mechanisms are microearthquakes. The mechanism of a single event is not necessarily representative for a broader region. On the other hand, Fig. 11 shows that the mechanisms of events in particular areas, e.g. the northern or southern graben, the eastern graben shoulders etc. reveal a remarkable consistency. This fact is supported by Fig. 12. The mean strike directions of the nodal planes (Fig. 12) follow, with surprisingly small scatter, the prevailing direction of the border-fault system as well as the conjugate direction in the southern graben area. A strike direction of N170°–180° E, claimed by Illies et al. (1981) for the present seismotectonic active direction of the Riedel shears, is not found in the rose diagram. We, therefore, favour the idea that the two expressions of active tectonics occur at different levels of the crust, i.e. second-order shear seems to be restricted to the uppermost part of the crust. In the northern graben area, an average strike direction of N150° E for the fault planes prevails (Fig. 12a). The same strike also dominates in the Rhenish Massif and the Lower Rhine Embayment (Ahorner et al., 1983). These areas, including the northern Upper Rhinegraben, are presently experiencing active rifting (Ahorner et al., 1983).

Figure 13 shows a comparison of the σ_1-directions from in situ stress measurements in the Upper Rhinegraben area, as summarized by Baumann (1981), and strike directions of P-axes derived from earthquakes, which occurred close (less than about 20 km) to the site of stress measurements. The maxima of the most frequently obtained strike directions from in situ measurements and P-axes differ only by a few degrees. Both parameters may reflect the same cause: (the strike of) the regional stress field in the Upper Rhinegraben. If this is the case, the regional stress field at the surface and at depth would have the same strike. On the other hand, in the southernmost graben, more than an order of magnitude difference exists for the amplitude of the maximum horizontal shear stress (<4 bars) deduced from in situ data (Illies and Baumann, 1982) and the average static stress drops ($30<\Delta\sigma(\text{bars})<300$) of strike-slip earthquakes (Durst, 1981; Gruber, 1983).

Although considerable progress has been made during the past decade in understanding the seismotectonics of the Upper Rhinegraben, essential points concerning the dynamical processes remain unsolved, especially the mode of strain accumulation.

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