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Invited Review

Interaction between *in situ* stress states and tectonic faults: A comment

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Abstract: Understanding the *in situ* stress state is crucial in many engineering problems and earth science research. The present article presents new insights into the interaction mechanism between the stress state and faults. *In situ* stresses can be influenced by various factors, one of the most important being the existence of faults. A fault could significantly affect the value and direction of the stress components. Reorientation and magnitude changes in stresses exist adjacent to faults and stress jumps/discontinuities across the fault. By contrast, the change in the stress state may lead to the transformation of faulting type and potential fault reactivation. Qualitative fault reactivation assessment using characteristic parameters under the current stress environment provides a method to assess the slip tendency of faults. The correlation between *in situ* stresses and fault properties enhances the ability to predict the fault slip tendency via stress measurements, which can be used to further refine the assessment of the fault reactivation risk. In the future, stress measurements at greater depths and long-term continuous real-time stress monitoring near/on key parts of faults will be essential. In addition, much attention needs to be paid to distinguishing the genetic mechanisms of abnormal stress states and the type and scale of stress variations and exploring the mechanisms of pre-faulting anomaly and fault reactivation.

Keywords: *in situ* stress state; stress variation; fault reactivation; fault properties; interaction mechanism

1. Introduction

The *in situ* stress state is the virgin stress status in the earth's crust before engineering excavations or other perturbations [1–3]. In the past millions of years, the earth has experienced countless tectonic movements of different scales, and the stress fields produced by each tectonic movement have undergone countless superposition, traction, and transformation. Moreover, the stress field is affected by other factors, resulting in the notable complexity and variability of the stress field. Thus, to accurately determine the stress condition of a region of interest, reliable *in situ* stress measurements are necessary. In addition, as one of the key parameters describing the geological environment, a full knowledge of stresses is an important component of many underground engineering constructions and earth science research [4–7]. In particular, the analysis of *in situ* stresses is a necessary condition for understanding the dynamic process of the crust. The tectonism in shallow and deep parts and the related internal dynamic geological disasters (e.g., earthquakes, volcanoes, and structural fractures) are closely correlated with the crustal stress conditions [8–9].

In situ stresses can be influenced by many factors, one of the most important being the existence of faults [10–11]. Faults are typical structural traces preserved by crustal rock deformations and ruptures caused by tectonic stress. The crust contains widely distributed faults of different scales and directions due to the long tectonic process [12]. In fault areas, the fault effect can result in considerable local variability in the stresses [13]. Faults, regardless of their scales, have a relevant influence on the overall and local stress states involving value and direction, and this influence is relatively variable. In particular, in the vicinity of some large-scale active faults (faults that have been active in the present or modern geological period and may be active again in the future), the stress field is dominated by faults. The results of the *in situ* measurements demonstrate that the spatial distribution of stress in the fault area is uneven. That is, the *in situ* stress fields in fault areas are often inconsistent with regional stress settings, and reorientation and changes in the magnitudes of principal stresses occur near the faults; such jumps in the stress (or stress discontinuity) take place around the faults [14–15], such as abnormal increase or decrease in the principal stress value, sudden shifts of stress regimes (i.e., normal, reverse, and strike-slip faulting), and abnormal deviation in the maximum principal stress direction.

Conversely, the distribution of *in situ* stresses in terms of magnitude and orientation affects the geometry, shape, dimensioning, and orientation of faults [16]. The occurrence of faulting is closely related to changes in stress conditions, and

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crustal stress is the fundamental driving force that triggers fault reactivation [7]. Fault dislocation will occur under certain stress conditions [17], and different stress conditions induce fault activities with diverse characteristics. Under the dominance of stresses, when the stress acting on the fault plane accumulates to a level that exceeds the frictional strength of the fault, the fault is accordingly activated and slips, which may induce a series of geological disasters, such as earthquakes. Typically, the Wenchuan Ms 8.0 earthquake in 2008 is considered to be caused by the tectonic movement of the Longmenshan fault under the effect of crustal stress [18–19]. Moreover, the stress field controls the evolution process of faults [20], and different stress states may lead to different fault types in the formation. The stress field and fault geometry influence and interact with each other. Because the interaction between the *in situ* stress state and faults is extremely complex and controversial, the corresponding interaction mechanism is still not fully clear, which is, therefore, a topic of interest to researchers. In particular, the activity prediction of seismogenic faults is one of the ultimate goals of geoscience research, but we still lack a clear understanding of the seismogenic process and mechanism of seismogenic faults. Hence, revealing the regional crustal stress environment, particularly the *in situ* stress state near key tectonic positions or active faults, is of great scientific significance for the in-depth analysis and discussion of the mechanism of internal dynamic geological disasters, such as earthquakes, strength and kinematic characteristics of faults, evaluation of regional crustal stability, and dynamic background of plate movements [14,21–23].

Over the past decades, numerous *in situ* stress measurements for different purposes have been performed worldwide in the vicinity of faults and published, which are used for specific engineering projects and scientific research programs. These measurements provide an unparalleled opportunity to review and assess the stress state in fault areas and address the above-mentioned issues. However, due to different research emphases, different research schemes under various geological structures, and different fault geometries and properties, the current research results are relatively scattered. The primary objective of the present study is to gain a comprehensive understanding of the stress state (including magnitude and orientation), variability in the vicinity of the fault, the effect of *in situ* stresses on faulting, and the likelihood of the correlation between stress states and fault properties. This is important to reveal the interaction mechanism between the stress state and faults: How faults control the stress state and how *in situ* stress influences faulting and can provide some useful references for research in related fields.

2. Stress state near faults

2.1. Influence of faults on the stress magnitude

The stress field near a fault is a topic of interest for researchers. As early as 1923, Terzaghi [24] noticed that discontinuities, anisotropy, and heterogeneity caused a complex stress field in the crustal rock and believed that the open vertical structural plane in deep rocks might be evidence of low horizontal stress and small stress ratio. The magnitude of stress near faults, particularly active faults, has always been a controversial issue. Compared with the regional stress field, there are various opinions about whether the stress magnitude is high or low. In the past, it was generally accepted that earthquakes were caused by fault activities driven by *in situ* stress, and thus, it is inferred that there must be a stress concentration zone close to the active fault. However, in recent years, the results of *in situ* stress measurements near faults show that the change in stress magnitude close to faults is very complex and that the stress state at different segments of the same fault may be different.

The stress measurements using the hydraulic fracturing (HF) method in the Longyangxia dam site show that the stress value is higher near a fault but lower away from the fault [25]. To investigate the stress state near the famous San Andreas Fault in the United States, Zoback *et al.* [26] measured the stress of two sections near the fault using the HF method. Their measurement results show that the maximum and minimum horizontal principal stresses increase with the depth and distance from the fault. Furthermore, the magnitude of the maximum shear stress (i.e., half the difference in the determined horizontal stresses) has a functional relationship with the distance to the fault. Fig. 1(a) presents the variation of the maximum shear stress derived from different measurements at approximately 200 m depth with the distance from the fault. Clearly, the maximum shear stress is low close to the fault and increases with the increasing distance from the fault. The maximum shear stress value tends to rise from approximately 1.7 MPa at 2 km to approximately 5.4 MPa at approximately 20 km away from the fault, and then, it begins to descend and approaches approximately 4.8 MPa at approximately 33.8 km [26]. Overall, the maximum shear stress tends to be stable at a distance of approximately 20 km from the fault. Similar maximum shear stress distributions have been determined in the vicinity of the famous Tan–Lu fault in eastern China. From 1976 to 1979, Li *et al.* [27] conducted *in situ* stress measurements in seven places in the vicinity of the Tan–Lu fault, namely, Jinan, Anqiu, Qingdao, Xinyi, Huaibei, Dingyuan, and Lujiang locations, and detected the correlation between the maximum shear stress of each measuring point and the distance between the measuring point and fault, as illustrated in Fig. 1(b). The maximum shear stress values of several measuring points, such as Anqiu, Xinyi, and Lujiang locations, are between 0.15 and 0.55 MPa, while the maximum shear stress values of other measuring points away from the fault zone are approximately 1.0 MPa. In an area approximately 100 km away from the fault, the maximum shear stress tends to be stable. Feng *et al.* [18] investigated the change in the stress state in the Longmenshan fault after the Ms 8.0 Wenchuan earthquake in 2008 using the HF technique, and they found that the maximum shear stress at a 200 m depth is a function of

the vertical distance to from the Beichuan fault to the Yingxiu fault. The maximum shear stress value appears to sharply grow from 0.63 MPa at 4.5 km to approximately 2.85 MPa at 6.5 km, whereas it slowly drops and progressively trends toward a stable value with increasing distance, reaching a value of 1.54 MPa at 21.5 km (Fig. 1(c)). In addition, the correlation between the measured stress value in the vicinity of the Babaoshan fault and the distance from the fault (Fig. 1(d)) [28] indicates that the maximum and minimum horizontal stresses and maximum shear stress values near the fault zone are relatively low, and the stress values rise with the increase in the distance from the fault. The above examples show that the stress magnitude close to the fault is smaller than that in the area far away from the fault, which is obviously contrary to the inference that there must be a stress concentration zone adjacent to the active fault and provide evidence that there may also be low stress near the active fault. The stress, particularly the maximum shear stress, increases with the distance from a large-scale fault, which may be one of the signs of the present activity of the fault [29].

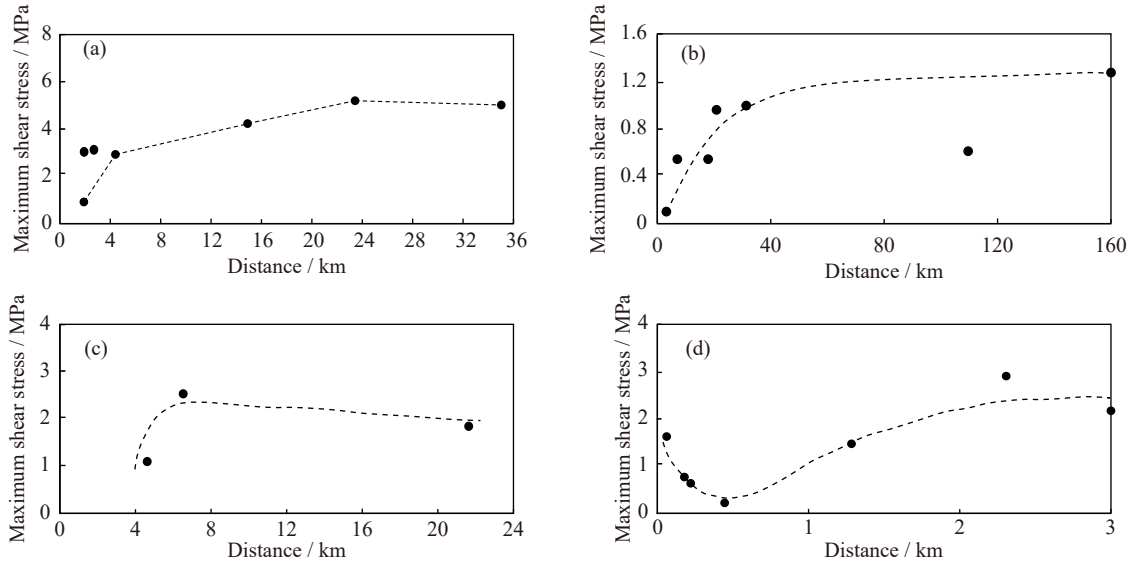

Fig. 1. Maximum shear stress as a function of distance from the San Andreas Fault (a) [26], Tan–Lu fault (b) [27], Longmenshan fault (c) [18], and Babaoshan fault (d) [28]. (a) M.D. Zoback, H. Tsukahara, and S. Hickman, *J. Geophys. Res. Solid Earth*, 85, 6157-6173 (1980) [26], Copyright Wiley-VCH Verlag GmbH & Co. KGaA. Reproduced with permission. (b) reprinted from *Chin. J. Rock Mech. Eng.*, 1, F.Q. Li, S.Z. Sun, and L.Q. Li, *In-situ* stress measurements in North China and Tancheng–Lujiang fault zone, 73-86, Copyright 1982, with permission from editorial board of *Chinese Journal of Rock Mechanics and Engineering*. (c) reprinted from *Int. J. Rock Mech. Min. Sci.*, 77, C. Feng, P. Zhang, X. Qin, W. Meng, C. Tan, and Q. Chen, Near-surface stress measurements in the Longmenshan fault belt after the 2008 Wenchuan Ms8.0 earthquake, 358-377, Copyright 2015, with permission from Elsevier. (d) reprinted from *Acta. Seismol. Sin.*, 6, J.M. Ding and G.P. Liang, On stress field in epicentral areas of 1976 Tangshan earthquake and 1679 Sanhe–Pinggu earthquake, 195-202, Copyright 1984, with permission from editorial board of *Acta Seismologica Sinica*.

The causes of low-stress anomalies adjacent to the fault can be attributed to the following possibilities: First, during an earthquake or fault activity, the stress value near the epicenter or fault is reduced due to the energy release. Second, some faults extend from the surface to the deep crust. If there is a high-pressure or high heat source in the deep crust, the pressure and temperature of the rock mass will rise, and thus, the formed high-pressure water will surge up along the fault plane and cause the surrounding rocks to produce relative elastic expansion, thereby reducing the stress level near the fault. Third, micro-fractures, fault gouges, and hydrochemical actions in the fault area and its vicinity can reduce the mechanical properties of rocks and lead to low-value stress anomalies. In addition, the stress value is supposed to be negatively correlated with the fault density at a regional scale, and the stress release caused by faulting may result in the reduction of the stress value in fault areas [8]. In particular, the Ms 8.0 Sanhe–Pinggu earthquake in 1679 and the Ms 7.8 Tangshan earthquake in 1976 occurred in low-value stress areas related to active faults [28]. From this understanding, we may predict the location of earthquakes by determining the low-stress area around the fault with a potential seismic risk through an *in situ* stress measurement.

The above can be regarded as the change in the stress value in the horizontal direction near the fault, and the change in stress value in the vertical direction adjacent to the fault is also analyzed below. According to the stress measurements [30], in the vicinity of the DF9 fault in the Sihe coal mine, China, the maximum (σ_H) and minimum (σ_h) horizontal principal stresses around the fault decrease with the increase in depth, which does not accord with the commonly accepted concept that stresses linearly increase with the overall depth. Stresses determined by the HF technique in a borehole near the neotectonic, postglacial Landsjärv fault indicate a noticeable stress magnitude anomaly compared to the average stress state in Fennoscandia [5]. That is, the σ_H and σ_h magnitudes decline to half the expected value near the fault at approximately 500 m depth, implying a stress relief due to

neotectonic faulting in the northern parts of the Fennoscandian Shield [5]. Zhou *et al.* [31] reported the stress measurement results in two boreholes (i.e., ZK1 and ZK2) in the vicinity of a fault, and the stress gradient in the results showed different distribution laws under the influence of the fault, as shown in Fig. 2. The stress value measured in a test hole decreases first in the measured depth of the hanging wall of the fault and increases in the footwall. However, the distribution of stress measured in another test hole is basically an "arc." That is, the stress value increases first within the tested depth of the hanging wall and suddenly decreases when the hole depth reaches the footwall. This discontinuous change in stress values in the hanging wall and footwall may be caused by the fault activity and the difference in geometric shapes of different fault segments.

In addition, the maximum horizontal stress jumped approximately 20 MPa across a major sub-horizontal fracture zone at a 320 m depth in a vertical borehole at Forsmark, central Sweden [32]. A similar phenomenon was observed at Lavia, central Finland, where a stress jump of approximately 20 MPa was deduced from the measurement results below a major fracture zone at 420 m depth in Proterozoic granodiorite rocks [32]. After the Ms 7.2 Kobe earthquake in 1995, Ito [33] measured the *in situ* stress using the HF technique at five locations along the seismogenic fault in the epicenter region. The measurement results demonstrate that the stress conditions of different parts of the same active fault are diverse, and the two ends of the fault have high differential stress, while the stress in the middle of the fault near the earthquake epicenter is uniform. A similar conclusion was also reached by Su and Stephansson [10] using a distinct element method. In addition to the measured results, several simulations [34–36] have also indicated that the change in fault occurrence will cause a variation in the local stress field near the fault. Moreover, Tian [37] suggested that the stress in the middle part of active faults appears to be released, and the stress concentration probably appears at both ends of faults. Due to the irregular geometry of a fault, the local stress condition of the fault is usually complex. That is, the stress state of different parts of the active fault is not necessarily the same. Notably, in the fault area, not only the stress release or increase may occur, but also the stress regime may change from the regional stress state.

In summary, a great deal of stress measurements imply that the stress value adjacent to the fault is not always defined by the regional boundary conditions of a block or plate, but it is also affected by the local geological environment, showing complex characteristics. This includes the stress-increasing area and stress-decreasing area, which is primarily related to the change in stresses adjacent to the fault with time. The increase and decrease in the stress magnitude highly depend on the geometry of the fault and the correlation between the fault and regional stress orientation, while the stress magnitude change is correlated with the scale of the fault. Generally, fault structures intersecting with the rock mass in a certain place or area can cause disturbance of the regional stress

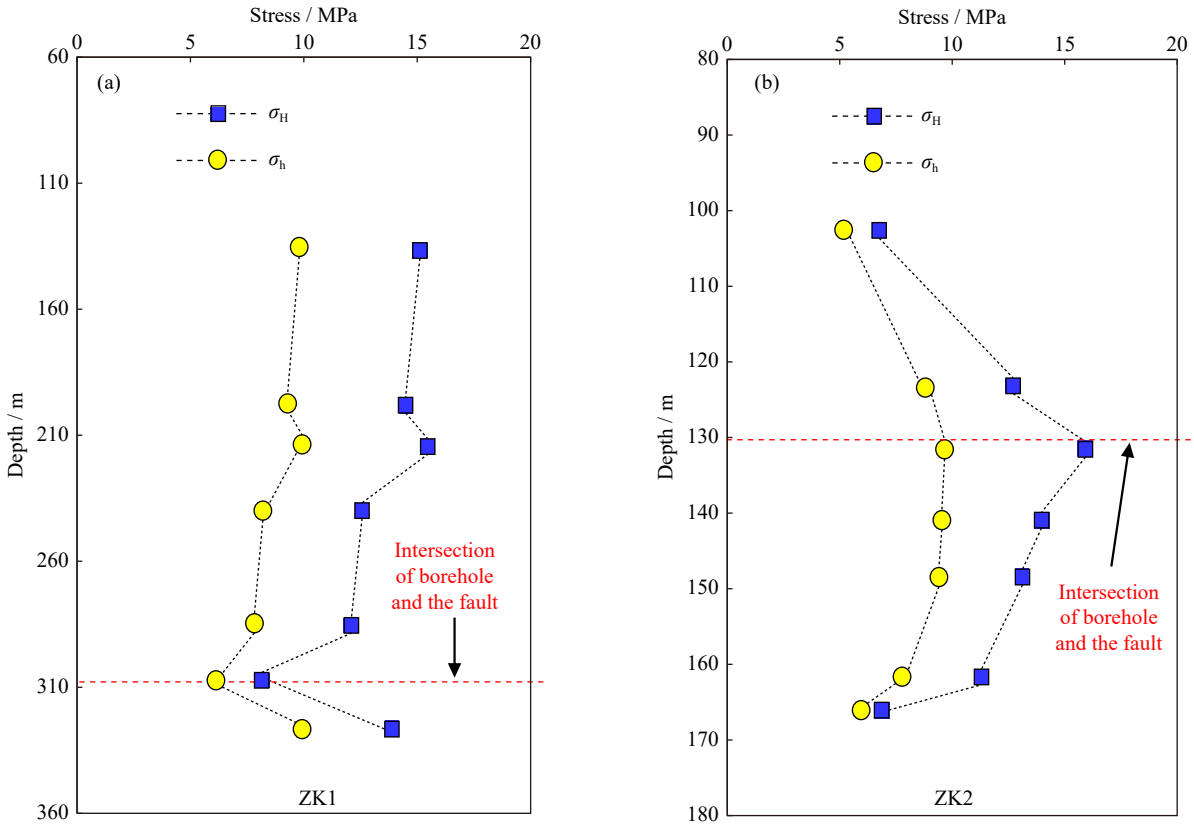

Fig. 2. Relations between the measured stress magnitudes and depth (modified from Zhou *et al.* [31]). Reprinted from *J. Yangtze River Sci. Res. Inst.*, 29, C.H. Zhou, J.M. Yin, J.Y. Luo, and G.Q. Xiao, Law of geo-stress distribution in the vicinity of fault zone, 57-61, Copyright 2012, with permission from editorial board of Journal of Yangtze River Scientific Research Institute.

state, and the spatial distribution of a fault structure exhibits a significant impact on the stress value near it. The amount of disturbance highly depends on the strength and deformability of the fault structure. The incidence of the fault on the stress magnitude is often proportional to the fault scale. The closer the distance from the fault is, the more obvious the control effect of the fault structure is. However, there are few examples in this regard, and more research and evidence are needed to examine the above conclusions. Moreover, the above examples suggest that stress discontinuities exist in the fault regions. In particular, the stress value along that depth profile changes discontinuously, showing a stress jump. That is, in fault areas, the horizontal principal stress magnitude does not always increase linearly with depth but solely as a function of the overburden. Thus, caution must be taken when using the relationship between generic stress and depth in fault areas.

2.2. Influence of faults on stress orientation

In many cases, the stress orientation near the fault area will also show significant variations. Previous studies have explained how faults perturb stress azimuths. For instance, Hudson and Cooling [38] determined three cases to reflect that a discontinuity (e.g., faults) exhibits a prominent local disturbance effect on stress condition, showing the perturbation of the principal stress trajectory. As plotted in Fig. 3, case 1 is an extreme case; namely, if the discontinuity is open (E_0 = 0), the maximum principal stress σ_1 orientation becomes parallel to the discontinuity, and the minimum principal stress σ_3 orientation converts to be perpendicular to the discontinuity, where σ_3 = 0. For case 2, when the filler in the discontinuity has basically the same properties as surrounding rocks ($E_0 = E$), the orientations of σ_1 and σ_3 are likely not to be disturbed. In addition, case 3 is another extreme case, namely, if the filler of the discontinuity is rigid ($E_0 \to \infty$), the σ_1 and σ_3 orientations will convert to be perpendicular and parallel to the discontinuity, respectively. In addition to these extreme cases, in other cases, the variation in the principal stress orientation is primarily governed by the physical and mechanical properties of the filling material in discontinuity [38].

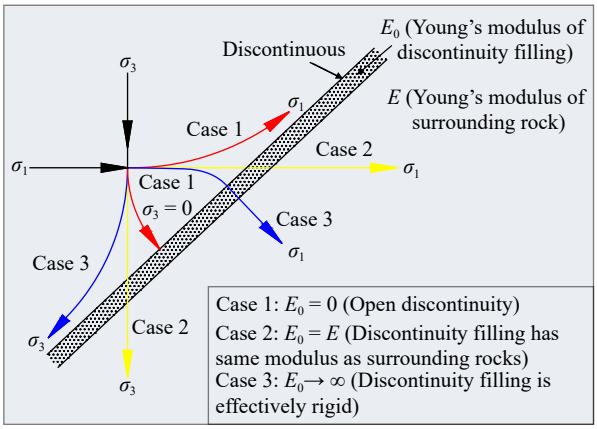

Fig. 3. Illustration of the influence of discontinuity on the stress orientation (modified from Hudson and Cooling [38]). Reprinted from *Int. J. Rock Mech. Min. Sci. Geomech. Abstr.*, 25, J.A. Hudson, and C.M. Cooling, *In situ* rock stresses and their measurement in the U.K.—Part I. The current state of knowledge, 363-370, Copyright 1988, with permission from Elsevier.

Similarly, Bell [39] schematically displayed and explained the role of geological structures (e.g., faults) in the perturbation of the maximum principal compressive stress azimuth by juxtaposing weak and stiff materials. If the stress trajectories encounter areas that are relatively stiffer than the surrounding rocks, they will be deflected so that the maximum principal compressive stress intersects the interface at right angles. Conversely, if the fault zone is relatively soft, with no shear and normal stresses parallel and perpendicular to the soft fault surfaces, the maximum principal compressive stress will be deflected and parallel to the interface. According to the findings reached by Hudson and Cooling [38] and Bell [39], the anomalous stress direction is the result of the structure and/or lateral changes in the elastic properties of rocks. The degree of the maximum principal compressive stress deflection is dependent on the comparison between the interface and geomechanical properties because fault sliding along faults and/or changing the geomechanical properties of nearby rocks can disturb the stress to varying degrees [40]. At present, it is unclear how far the stress deflection effect can extend from the property contrast interfaces. If there is enough stress orientation data to determine the regional stress orientation of a specific region and observe the abnormal direction, these data can be used to identify the open fracture and/or unsealed fault area. However, this type of application should only be tried in well-researched areas.

Similar phenomena naturally occur around open faults, as shown in Fig. 4. Stress state A refers to a state commonly existing in rocks. Nearer the fault, states B and C, the principal stress orientations rotate. For an open fault, no normal or shear stress can be sustained, respectively, perpendicular and parallel to the fault plane, so the fault plane becomes a principal stress plane with a zero principal stress magnitude, and

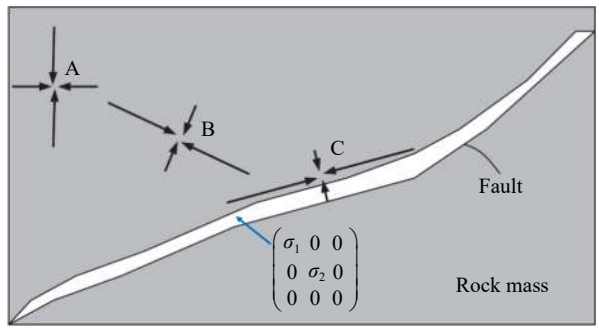

Fig. 4. Open fault perturbing the stress state and causing the principal stresses to be locally parallel and perpendicular to the fault (modified from Hudson *et al.* [41]), where σ_2 is the intermediate principal stress. Reprinted from *Int. J. Rock Mech. Min. Sci.*, 40, J.A. Hudson, F.H. Cornet, and R. Christiansson, ISRM suggested methods for rock stress estimation—Part 1: Strategy for rock stress estimation, 991-998, Copyright 2003, with permission from Elsevier.

the local stress direction significantly changes. When the fault is not fully opened or filled, the disturbance to the stress trajectories will be reduced [4,41]. This effect near many faults at all scales in rocks results in the expectation that local stress orientation may be variable, even highly variable [41]. In addition, the role of faults in the stress disturbances of different scales can be interpreted by the fault sliding of faults or by changing the geomechanical properties of nearby rocks [42–44].

In addition to the above-mentioned theoretical understanding of stress trajectory changes in the vicinity of faults, *in situ* stress measurement results around some faults reflect complex stress pattern changes. A famous example is the stress measurements (the measurement depth is approximately 2.79 m) close to the San Andreas Fault in Palmdale, California, using the overcoring method [45]. The measured results implied that the stress state adjacent to the active fault was quite complex; that is, the direction of the principal stress near the fault greatly deviated from the regional tectonic stress field, and only the stress direction of the measurement site far away from the fault was not affected by the fault (Fig. 5). The stress measurements along the Kobe earthquake-generating fault in Japan [33] indicated that the maximum principal stress orientation was nearly perpendicular to the fault at the southwest end of the fault, while it was nearly parallel to the fault or intersects at a small angle at the northeast end, which reflects the fact that the stress directions of different segments of active faults are not necessarily the same. Sun *et al.* [46] determined the stress state around the Meiling arc-shaped fault in the Babaoshan fault zone, Beijing, and found that the maximum principal compressive stress orientation changed regularly with different parts of the fault, and a local stress field existed near the fault, which was different from the regional stress setting. In the Cajon Pass scientific research borehole close to the San Andreas Fault zone in the United States, within a depth of approximately 1753–3264 m, the average maximum principal stress orientation deviated by 16°–38° compared with the regional stress field [47]. Furthermore, Barton and Zoback [48] provided additional examples of stress perturbations related to active faults. The principal stress direction at 1.25 km to the north of the Atotsugawa fault, Japan, was parallel to the fault plane but inconsistent with the dominant orientation of the regional stress field [49]. According to the measured stress information in the German Continental Deep Drilling Program, a pronounced change in the maximum horizontal principal stress orientation was observed close to a prominent cataclastic fault at approximately 7200 m, where it jumped by about 60° from N164° to N220° [50]. The *in situ* stress test results via HF in nearly 1000 wells in 86 fault-block oilfields in 13 major oil regions of China demonstrated that regardless of whether the faults in these oilfields are normal or reverse faults, the direction of fracturing fractures is basically perpendicular to the fault strike; that is, the minimum principal stress is horizontal and parallel to the fault trace [51]. In addition, in the San Andreas Fault Observatory at Depth pilot hole, Hickman and Zoback [52] reported that the local principal stress direction around the fault zones rotated violently, and the apparent azimuth angle of principal stress below a small fault progressively rotated by approximately 70° in an interval of only 5 m. This may not be surprising considering the deformation and seismic complexity of the San Andreas Fault Zone in Southern California [53].

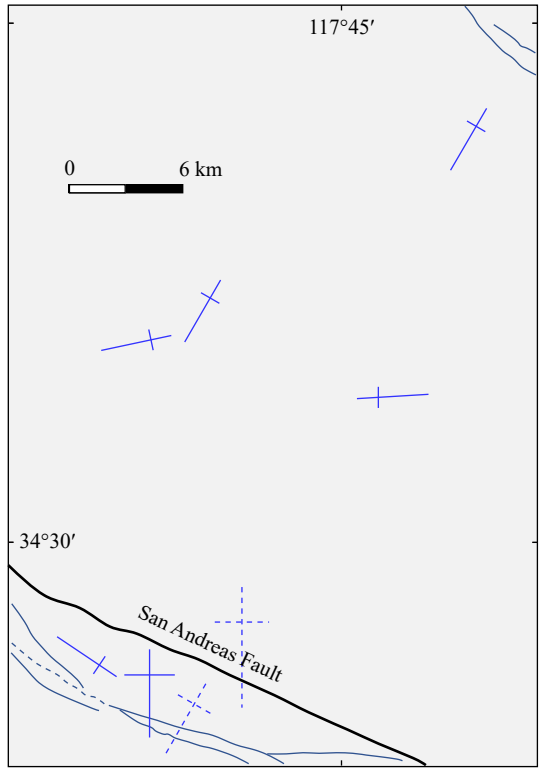

Fig. 5. Stress pattern close to the San Andreas Fault (after Sbar *et al.* [45]). M.L. Sbar, T. Engelder, R. Plumb, and S. Marshak, *J. Geophys. Res. Solid Earth*, 84, 156-164(1979) [45], Copyright Wiley-VCH Verlag GmbH & Co. KGaA. Reproduced with permission.

Afterward, Lin *et al.* [54] observed from various stress measurement results of boreholes penetrating the north section of the Chelungpu fault that in a narrow depth interval of 1133 m depth in hole B, the stress orientation (averaging 212°) is about 90° different from that at other depths. Peng *et al.* [55] performed stress measurements in three boreholes near the middle segment of the Qingchuan fault using the HF and overcoring methods and observed that the maximum horizontal principal compressive stress direction adjacent to the fault was NE–NEE, which was somewhat different from the that of the regional tectonic stress oriented in SN–NNE. Similarly, the HF stress measurement results near the Xishan–Wanyaogou fault zone in Urumqi demonstrated that the maximum horizontal principal compressive stress orientation close to the fault is NE–NEE, which was not in harmony with the azimuth of the regional tectonic stress of SN–NNE [56]. The stress measurements of hole B in the scientific drilling of the Chelunpu fault in Taiwan Province implied that the maximum principal stress direction was deflected by about 20°–50° at a depth of about 975–980 m [36]. Information

about the maximum principal stress direction in the Wenchuan Scientific Drilling hole near the Longmenshan fault from 950 to 1175 m suggested that the stress azimuth deviated from the regional stress field by 16°–22° [57]. Moreover, Tan *et al.* [58] measured the *in situ* stress at depths of greater than 1000 m via the HF method at six positions near faults in an exploration area and found that the direction of *in situ* stress significantly deviated near faults, but the direction and degree of stress deflection were related to their location. That is, the stress direction would deflect toward the direction close to the fault strike at the ends, intersection, and footwall of faults and between faults, while it would deflect toward the direction away from the fault strike at the footwall. These results indicate that the stress direction changes observed in different areas are diverse but emphasize that stress direction fluctuations frequently occur in the vicinity of faults.

Furthermore, Li *et al.* [59] collated previously published stress data and investigated the overall stress pattern of the west and east sides of the Yishu fault zone and found that the σ_H direction in the west side of the fault zone is predominantly oriented in the NEE–SWW orientation, while the σ_H orientation in the east side is dominantly in the NWW–SEE orientation (Fig. 6). As an ultra-crustal fault, the motion nature of the Yishu fault zone directly reflects the regional tectonic stress field. Different survey results demonstrated that the current tectonism of the fault zone principally possesses compressional and dextral strike-slip features [60], which is highly supported by the distinct prevailing σ_H orientations on the two sides. The difference in the stress direction near the fault zone may reflect the different structural properties of regional geological units on the two sides. It is also related to fault morphology and its development and evolution. Moreover, four abnormal or non-abnormal modes of stress direction were found near faults and fractures in the Taiwan Chelungpu-fault Drilling Project (TCDP) hole B [36]: (1) abrupt (discontinuous) rotation near faults or fractures; (2) gradual rotation; (3) suppression of breakouts at faults, fractures, or lithologic boundaries; and (4) no abnormality in the stress direction. The above cases imply that faults play an important role in controlling the local stress field near faults, and the stress state near the faults has obviously changed compared with the regional stress field due to the influence of fault activities. Essentially, the fault causes local heterogeneities in the crust, thereby resulting in the refraction and/or rotation of the stress orientation.

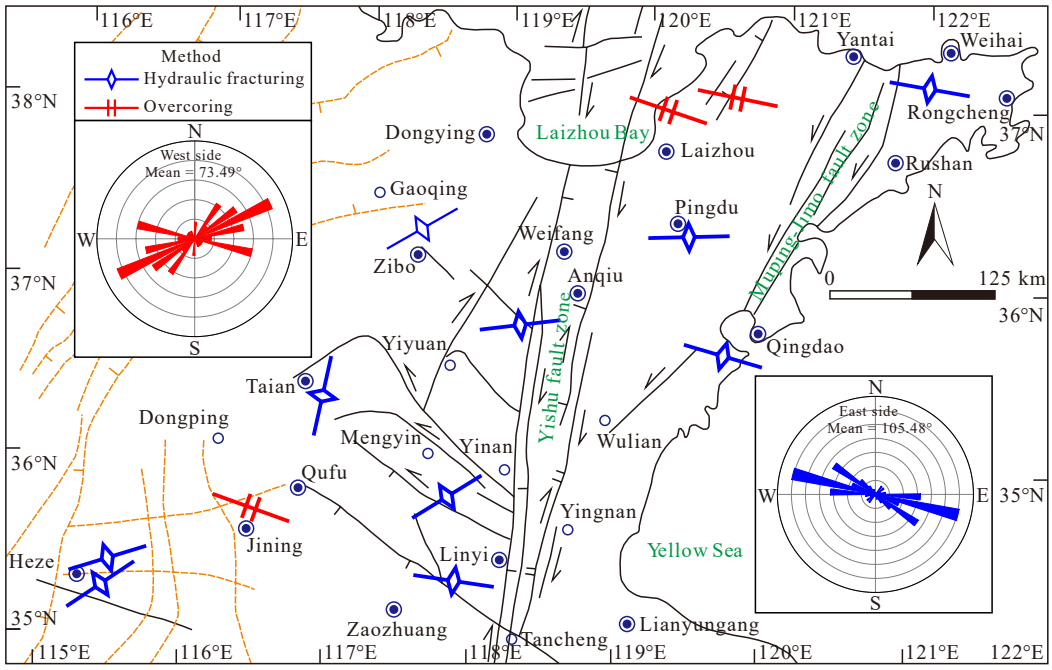

Fig. 6. Average maximum horizontal principal stress orientation (the data were ranked as categories B, C, and D based on the World Stress Map quality ranking system) near the Yishu fault zone [59]. Reprinted from *Rock Mech. Rock Eng.*, 52, P. Li, M.F. Cai, S.J. Miao, and Q.F. Guo, New insights into the current stress field around the Yishu fault zone, Eastern China, 4133-4145, Copyright 2019, with permission from Springer Nature. N, E, S, W denotes the north, east, south, and west direction, respectively.

Through the above analysis, the distribution of the principal stress orientation near faults is not always uniform and shows a fairly complicated nature. These examples indicate that stress orientation is influenced by the fault structure and how that fault was developed. These anomalous stress directions are the result of the fault structure and/or lateral changes in the rock's elastic properties. Compared with the regional principal stress direction, the stress orientation adjacent to the fault changes in different degrees, which is inconsistent with the regional stress direction and forms a certain angle with the fault, and this change is mainly limited to a certain distance near the fault. The possibility of stress direction variations with length scales of a few centimeters to tens or even hundreds of kilometers is superimposed on the remote stresses [36,53]. Away from the fault area, the principal stress orientation gradually tends to be consistent with the regional

stress orientation. Moreover, the principal stress orientations not only vary greatly near the fault but also appear different between the hanging and foot walls of the fault sometimes. In some cases, the principal stress orientations of the two walls can differ by 90°.

In summary, the above studies strongly prove that faults have a significant impact on the stress state with respect to the stress magnitude and orientation. The influence of faults on the *in situ* stress can be regarded as the influence of the directional fracture system on the stress transmission, resulting in a complex spatial distribution pattern of the stress state distribution. The complexity of fault development is closely associated with the change amplitude of stresses. The more a fault develops, the greater the change amplitude of the stress magnitude and direction, and the stress direction is extremely dispersed in the area where the fault is extremely developed, and the fault pattern is extremely complex [61]. Furthermore, the end of a fault and the inflection point of its geometric shape are often high-stress concentration areas, and the stress direction near the fault will be disturbed to varying degrees, especially at the end of the fault. Generally, faults have two basic elements, namely, the fault plane and surrounding rocks on both sides. Although the movement of a fault occurs along the fault plane, it is essentially the relative motion of blocks on the two sides. Therefore, the stress field near the fault should be determined by the physical and mechanical properties of the fault itself and the rocks on the two sides. Moreover, numerous potential sources of stress exist in a fault area that could disturb the stress field. Hence, these factors need to be deeply studied to reveal the mechanism of the stress anomaly or decoupling in fault areas. By contrast, the special tectonic conditions in fault areas can greatly reduce the reliability of stress measurement results, which may represent the disturbed local stress field rather than the real regional stress state. In addition, one or two stress measurements within a fault area may not be sufficient to define the distribution and heterogeneity of stresses within the structural formation. Several stress measurement activities are necessary. In addition to field measurements, the application of general numerical models is also of great value in analyzing stress disturbances in fault areas [5].

Although the stress state is relatively stable in a certain period, it is dynamic and changeable. Any stress measurement result is only a transient state at a specific position and time. In particular, the activities of active faults in the geological period are staged, sometimes calm, and sometimes intensified. The fault is deformed under the action of regional tectonic stress, which leads to stress concentration or release in the fault area. This is a cyclic process, and it is not surprising that different stress measurement results in the fault area appear.

The regional stress field features are not always exactly the same as those in fault areas. In particular, when the fault structural conditions or local rock mechanical properties change, the stress field can have a unique distribution feature. Although the regional stress field is inconsistent with the stress field in fault areas, they appear to be in harmony with each other in terms of the formation mechanism and activity of the tectonic system, showing a certain genetic relationship. In general, a preliminary qualitative and quantitative assessment of stress variations in fault areas has been made, which is an important step forward for us to understand the stress field in the faulted rocks and the formation mechanism of stress anomalies and helps to update our knowledge and methodologies to build a deterministic model of how faults and geological heterogeneities interfere with the stress state [40]. This aspect needs to be further studied in the future to analyze the influence mechanism of faults on the regional stress field and stress measurement results.

3. Effect of stress changes on faulting

The cause of fault reactivation is correlated with various endogenous and/or exogenous factors, and efforts are being made to study the influence of various factors on the slip of faults, especially active faults. In recent years, numerous theoretical studies and field test results indicate that the change in the stress state is the key reason triggering the reactivation of faults. The preparation and nucleation of fault activities are essentially a process of rock stress accumulation and mechanical instability [62–64]. Under a specific regional structure setting, with the enhancement of the stress field, the stress acting on a fault accumulates enough to exceed the frictional strength of faulted rocks, resulting in the reactivation and sliding of the fault [22]. Hence, the change in stress plays an important role in faulting.

In different geographical, geological, and structural areas, the stress magnitude and orientation are considerably different. As early as 1951, Anderson [65] had built three different *in situ* stress regimes based on the magnitude relationship among σ_H, σ_h, and σ_v (vertical stress) and combined them with different fault types [3]. That is, (1) $\sigma_H > \sigma_h > \sigma_v$ denotes a thrust faulting stress regime, which is beneficial to the activities of thrust faults; (2) $\sigma_H > \sigma_v > \sigma_h$ indicates a strike-slip faulting stress regime, which favors the activities of strike-slip faults; and (3) $\sigma_v > \sigma_H > \sigma_h$ represents a normal faulting stress regime, which favors the activities of normal faults. This phenomenon is the famous Anderson's theory of faulting [65], which assumes that the fault is generated by shear failures caused by *in situ* stress. This theory reveals the relationship between fault properties and stress states and is the basic theory for analyzing the nature and strength of modern tectonic movements. The present-day stress condition may differ from the fault types observed in the crust, especially for thrust faulting. For instance, the existence of a thrust fault does not necessarily represent the thrust faulting in the current stress state [66]. The reason may be that the paleostress in this case is probably in the thrust faulting. However, it was an unstable stress state and more likely to be converted to a strike-slip or normal faulting due to stress relaxation or other reasons. Although the crustal stress state does not completely conform to Anderson's theory of fault-

ing, it is feasible to adopt Anderson's fault classification principle in most cases.

The continuous or rapid increase of the stress value in fault areas leads to high-stress concentration (relative to the background stress), which can easily induce a fault slip. The effect of stress changes on fault reactivation or seismic activity has attracted extensive attention. For example, Stein *et al.* [67] studied the stress transfer of 10 earthquakes with Ms ≥ 6.7 from 1939 to 1992 on the North Anatolian fault in Turkey and stated that nine earthquakes were driven by the change of the Coulomb failure stress. The typical value of the Coulomb failure stress increase is 0.1–1 MPa, which is equivalent to the long-term stress loading effect of 3–30 years. Papadimitriou *et al.* [68] investigated the corresponding relationship between the evolution of the Coulomb failure stress and the occurrence of earthquakes in the Xianshuihe fault zone based on the historical earthquake data in western Sichuan of China and found that most strong earthquakes in this fault zone have occurred in the area where the Coulomb failure stress has increased since 1895. For the Ms 8.1 Kunlun earthquake in 2001 caused by the reactivation of the East Kunlun fault zone in the northern part of the Qinghai–Tibet Plateau, the stress magnitude near the fault zone measured prior to the event is 4–6 times the average level [69], which indicates that the stress is highly concentrated before the fault slip. Similarly, the stress measurements indicate that prior to the Ms 8.0 Wenchuan earthquake in 2008, the stress on both ends of its seismogenic fault (i.e., Longmenshan fault) in the northwest edge of the Sichuan Basin was obviously greater than the background stress, and the principal stress obtained at the approximately 400 m subsurface in the footwall of the seismogenic fault rose to approximately 2–3 times that at the same depth in the footwall of the surrounding faults [19]. In addition, before the Ms 5.6 Dongco earthquake in 2004, Ms 6.6 Nimu earthquake in 2008, Ms 7.1 Yushu earthquake in 2010, and Ms 7.0 Lushan earthquake in 2013, the local principal stress along the seismogenic fault rose to 2–5 times the background stress [70], and the resulting high-stress triggered fault activities and subsequent earthquakes.

In the above cases, the high-stress anomalies caused by the stress increase in the fault areas appear to be the direct cause of fault slips or earthquakes. Thus, a stress increase is probably a feasible precursor to predicting fault reactivation. This finding, however, needs to be verified by more strong evidence and cases, but it at least allows us to further understand the close relationship between stress conditions and fault activity. The stress change reflects the current tectonic movement of the crust. When the stress change occurs on the active fault plane, it may be correlated with the fault slip and the seismicity along the fault. Even small stress fluctuations, such as those caused by reservoir impoundment, are sometimes sufficient to activate faults under specific tectonic conditions. In addition, the long-term stress monitoring results confirmed that crustal stress is dramatically variable, primarily showing stress fluctuations and stress jumps [71]. Stress anomalies in diverse times and ways probably indicate the source stress field prior to fault reactivation in distinct developmental stages.

Moreover, the spatial relationship between stress orientation and fault strike has a certain influence on fault stability. In general, as the angle between the principal stress direction and fault strike is in an orientation conducive to slips, the fault may be activated as it is interfered with by a small external force. Under the disturbance of certain conditions, the direction of tectonic stress in fault areas can show temporary changes over a certain period. For example, driven by plate movements, earthquakes, or local force sources, the stress direction in the fault area may be deflected in a two-dimensional horizontal plane or rotated in a two-dimensional longitudinal section [22]. Once the stress direction deflects to the position that causes the fault to be out of equilibrium, a fault slip may occur. As a consequence, identifying the transformation in stress orientation is one of the purposes of fault reactivation prediction. In addition, the intersection region of the stress direction is often the strain concentration area and a dangerous region of fault activities. Notably, the stress orientation in fault areas will be influenced by other factors, so it is difficult to accurately grasp the fault activity only from the perspective of the stress direction.

Under the action of plate movements, when the strain energy stored in the geological body reaches a certain degree and the tectonic stress level outweighs the fault friction strength, fault sliding may occur along the optimally oriented plane [59,72]. The critically stressed faults can move in various forms. Faults can rupture and slide within a few seconds, releasing stored energy rapidly to induce earthquakes, or they can slowly release energy in the form of slow earthquakes for a long period (ranging from a few minutes to several months) or release energy with imperceptible relative creep. The first situation is the most well-known manifestation. Whether the fault moves rapidly or slowly mainly depends on the stress conditions of the fault and the rock strength of the geological body in which it is located [26,73–74]. Generally, when the faulted rock strength is high, the fault has a strong locking ability, and the fault slides intermittently and suddenly under the action of tectonic stress, often accompanied by seismic activity. Such faults are called stick-slip active faults, which require special attention. By contrast, when the fault is in the soft rock with a low rock strength, the locking ability of the fault is weak, and the rock layers on both sides of the fault slide continuously and slowly along the fault plane when disturbed by excess stress, usually without earthquakes, sometimes accompanied by small earthquakes. This type of fault is called an active creeping fault. The slip mechanism of the two types of faults is fairly different, and the possible abnormal phenomena before reactivation are also different.

Through the above analysis, the change in local stress conditions induces the uneven distribution of stress accumulation, thus changing the stress state of the fault. Thus, the occurrence of a fault slip is the process of long-term accumulation, concentration, and strengthening of the *in situ* stress in its key structural parts, which ultimately causes the sudden

release of a large amount of strain energy [75–77]. As an important trigger factor for fault reactivation, stress variation is generally associated with fault occurrence, fault properties, and frictional strength [78]. Therefore, how to capture the stress conditions and stress change characteristics in fault areas is the key to exploring the structural characteristics of faults and the gestation and occurrence process of fault reactivation. If the measuring sites are fully distributed and the stress state in the fault area is repeatedly measured at certain intervals, it is possible to estimate the risk of fault reactivation and predict the occurrence of subsequent events (e.g., engineering instability and earthquake). In addition, a large enough relative stress monitoring network can be established near active fault zones for intensive monitoring and to obtain a dynamic stress evolution map of the zone, which can help capture the stress anomaly of the entire fault zone. The monitoring results from sufficient locations in diverse orientations and distances of the monitoring networks can afford useful information for deducing the possibility and time of reactivation of the target fault in the future through systematic analyses.

4. Fault reactivation assessment based on stress conditions

4.1. Mohr–Coulomb criterion

The nature of fault plane dislocation is that stress weak points appear in the rock, and shear failure occurs at the critical slip surface. The slip tendency of a critically stressed fault can be estimated by the limited frictional coefficient, and the crustal stress state determined from the field measurement is consistent with the failure equilibrium [79]. The stresses acting on a rock mass are specified with respect to three orthogonal principal stresses, i.e., maximum (σ_1), intermediate (σ_2), and minimum (σ_3) principal stresses. Compressional failure will occur along a plane containing the intermediate stress, σ_2, and will be independent of the σ_2 value [39,80]. Hence, to assess the failure criteria, only σ_1 and σ_3 should be considered and resolved into a shear component (τ) and a normal component (σ_n) along a plane other than incorporating σ_1 and σ_3 (Fig. 7(a)). According to the Mohr–Coulomb criterion [81], considering the influence of pore pressure, the shear stress on the failure surface is mainly composed of the cohesion of the fault against sliding and the friction force generated on the surface:

$$\tau = \tau_0 + \mu(\sigma_n - P_0) \quad (1)$$

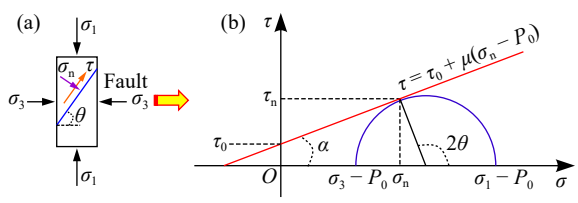

Fig. 7. Schematic diagram of the stress state acting on a fault (a) and the corresponding Mohr–Coulomb failure criterion (b), where τ_n is the shear stress across the plane and α is the internal friction angle.

where τ is the shear stress required for failure on a specific plane; τ_0 is the cohesion; σ_n is the normal stress across the plane; P_0 is the pore pressure; and μ is the frictional coefficient on the plane.

Mohr's circle is a convenient method to decompose principal stresses into normal and shear components. In general, drawing a circle uses the $(\sigma_1 - P_0)$ and $(\sigma_3 - P_0)$ values to define the diameter along the normal stress axis (Fig. 7(b)). The circle is the locus of normal and shear stress components on planes at an angle θ to σ_3. The Coulomb failure criterion is added to Mohr's circle diagram to define the slip condition. A fault will slip when the circle becomes tangent to the failure envelope. The point of a tangential intersection determines the angle 2θ, which indicates the orientation of the plane on which a fault slip will occur.

As plotted in Fig. 7(b), if σ_1 and σ_3 acting on the fault make Mohr's circle not intersect with the failure envelope of the rock, fault reactivation will not occur. Evidently, changing the magnitude or ratio of the difference between σ_1 and σ_3 may lead to slippage. Some of the conditions under which a fault slip can be triggered were described by Bell [39]. When the rock in the fault area bears a high horizontal stress (σ_H) and a minimum principal stress in the vertical direction (σ_v), the removal of overburden through some human activities (e.g., quarrying and mining) will reduce σ_v, and the difference between σ_H and σ_v will increase accordingly. This condition will enlarge the diameter of Mohr's circle, such that the circle may intercept the failure curve of the rock. If the vertical stress decreases so much, the shear slip in the form of thrust faulting is likely to occur. By contrast, the impounding bodies of water in surface reservoirs and dammed lake reservoirs can cause an increase in vertical stress, thus increasing the difference between the vertical and minimum horizontal principal stresses, which will also expand the diameter of Mohr's circle [39]. Consequently, the circle can intercept the failure curve of the rock, causing a shear slip in the form of normal faulting. Such an induced slippage preferentially occurs where σ_1 is vertical. In addition, natural rainfall or injection of fluid into the fault area in any stress regime can increase the reservoir pressure, thus reducing the effective stress. If the reservoir pressure rises fast enough to cover the pore elastic effect, a shear slip may occur in the intermediate principal stress plane [39]. In the stress environment compatible with this mechanism, fault slips and seismic events have been recorded in many fault regions. For example, during geothermal power production, shale gas exploitation, carbon dioxide sequestration, or enhanced oil recovery, water is usually injected into the geological structure near the target area, and the local strong fluid pressure disturbance reduces the effective stresses and can reactivate faults, which may induce earthquakes. In fact, some seismic events, such as the Ms 5.8 earthquake in Oklahoma, USA, in 2016 [82], Ms 5.5 earthquake in Pohang, South Korea, in 2017 [83], and the Ms 5.7 earthquake in 2018 and Ms 5.3 earthquake in 2019 in the southern Sichuan Basin in China [84] have been proven to be triggered by the injection of fluids into the subsurface during the operation of geothermal or

shale gas projects. Notably, if the removal of fluid significantly changes the horizontal stress, then a fault slip may also occur, which has been explained by the theoretical model [85].

A naturally occurring fault slip involves the rapid destruction of the fault surface, thus releasing the crustal stress. Therefore, changing the pre-existing stress state can induce such events. The conditions of the fault slip are summarized by invoking the famous Mohr's circle diagram. The induced fault slip is most likely triggered by the high-stress level and principal stress anisotropy, so relatively small stress changes will promote reactivation. Furthermore, if the applied load or fluid pressure changes and the original stress state changes to a failure mode, then fault reactivation will occur. Altogether, the effect of stress conditions on the fault slip and related events highlights the significance of understanding controls on the crustal stress in risking engineering and natural hazards.

4.2. Characteristic parameters derived from *in situ* stresses

The stability of a fault depends on the inherent friction strength of the fractured rock and pore pressure [86]. *In situ* stress information provides an unparalleled opportunity to gain new insights into the fault strength and laws that control fault reactivation. The shear and effective normal stresses acting on a fault can be expressed by principal stresses, which is a function of the direction of the fault plane relative to the stress field [8], i.e.,

$$\sigma' = l^2(\sigma_1 - P_0) + m^2(\sigma_2 - P_0) + n^2(\sigma_3 - P_0) \quad (2)$$

$$\tau = \sqrt{l^2(\sigma_1 - P_0)^2 + m^2(\sigma_2 - P_0)^2 + n^2(\sigma_3 - P_0)^2 - \sigma'^2} \quad (3)$$

where σ' is the effective normal stress; l, m, and n are the directional cosines of the fault normal with respect to the principal axes.

The value of cohesion is virtually negligible [87]. For a cohesionless fault, the slip tendency of the fault can be judged according to the ratio of shear stress to the effective normal stress of the fault plane:

$$\mu = \tau/\sigma' \quad (4)$$

According to the stress value and direction of the region of interest and the occurrence of nearby faults, the μ value can be computed by Eq. (4). Jamison and Cook [87] contrasted the relationship between stresses and different types of faults and found that the fitting slope on the curve of half the maximum stress difference as a function of half the sum of the maximum and minimum principal stresses interpreted as a Coulomb frictional coefficient. Moreover, they observed that various types of faults possess distinct friction coefficients, which actually mirror the differences between faults. Thus, fault stability could be assessed by the limited friction coefficient [9]. Based on Mohr–Coulomb failure criterion, Byerlee [88] conducted laboratory tests on many different types of rocks and determined that the frictional coefficient of most rocks was in the range of 0.6–1.0, which signifies the well-known Byerlee's law. Su *et al.* [89] stated that the upper and lower limits of the μ value are 1.1 and 0.65, respectively, averaging 0.85, when the stress magnitude is 150–250 MPa and proposed that μ of 0.6–1.0 is reasonable to evaluate the fault stability in the upper crust. Moreover, some scholars have analyzed the measured stresses and proved that the frictional coefficient obtained through several laboratory tests is suitable for fault slip analysis [3,87]. Hence, μ in the range of 0.6–1.0 has been widely used as the basic standard to determine fault activities [23,72].

The μ value of a fault obtained based on laboratory or field tests is compared with the frictional coefficient of 0.6–1.0 to assess the slip tendency of the fault. The lower limit value of the frictional coefficient of 0.6 is taken as the critical frictional coefficient to evaluate the slip tendency of the fault plane. When the frictional coefficient of the local position of the fault plane is higher than 0.6, the fault is considered to be extremely unstable and may slide on the fault plane with proper orientation. The fault plane with a suitable orientation refers to the plane where the normal direction of the plane intersects the σ_H orientation. If the slip range of the fault extends to the surface, it is likely to trigger an earthquake with a large-scale surface rupture [17]. In another case (i.e., $\mu < 0.6$), the mechanical state of the fault is evaluated as stable.

Meanwhile, when the stress accumulation in the crust reaches a certain degree, the stress will be released by means of fault movements to maintain the crust stability. Based on the measured *in situ* stress data, Townend and Zoback [75], Zoback *et al.* [90], and Wang *et al.* [19] used the ratio of the maximum shear stress to the effective average principal stress as the stress accumulation index μ_m to represent the regional stress accumulation level in the shallow crust. μ_m is expressed as [81]

$$\mu_m = \frac{(\sigma_1 - \sigma_3)/2}{(\sigma_1 + \sigma_3)/2 - P_0} \quad (5)$$

μ_m and μ have similar physical meanings because they are derived from the same theory, and their mathematical relationship is given by

$$\mu_m = \frac{\mu}{\sqrt{1+\mu^2}} \quad (6)$$

Tanaka *et al.* [91] found that μ_m can be employed to evaluate the effect of the *in situ* stress on the fault stability, and μ_m will become smaller after sliding. μ_m is positively correlated with the degree of regional stress accumulation; that is, the greater the μ_m, the higher the stress buildup level is, and the greater the possibility of fault sliding. Notably, μ_m is independent of the direction of stress. In general, the value of μ_m in the crust should not exceed 0.5–0.7; otherwise, shear failure may occur [92]. According to Townend and Zoback [75], when μ_m is less than 0.3, the stress accumulation level is low; when μ_m approaches or exceeds 0.5, the stress buildup degree is large or in the limit state. μ_m can indicate the friction properties of faults from the perspective of regional stress enhancement. If the stress accumulation exceeds the limit level, some well-oriented faults may slip theoretically. Therefore, many researchers [12,18–19,59,74] have employed μ_m calculated by stress measurements to

quantify the risk of fault sliding and contrast the stress conditions in different areas. For example, the results of multistage *in situ* stress measurements near the epicenter of the Ms 7.2 Hyogo Prefecture earthquake in Japan show that μ_m values increased from 0.16 to 0.53 prior to the event and then recovered to approximately 0.2 after the event [91]. Furthermore, the seismogenic fault of the Wenchuan Ms 8.0 earthquake in 2008 is considered the Longmenshan fault. After averaging all the μ_m values computed from the measured stresses before and after the event, the average μ_m values of the Longmenshan fault before and after sliding are 0.46 and 0.36, respectively, and the declining percentage of the average μ_m values after sliding is 22% [19]. Apparently, the variation in the μ_m value of the fault is correlated to the fault stability. Therefore, repeated stress measurement activities at the same position at a certain time interval can identify whether μ_m changes with time, thereby speculating the occurrence time of fault reactivation.

Nonetheless, although the stress buildup in the shallow crust is consistent with the scope defined by Byerlee's law in most cases [12,47,62,75,93], more and more empirical evidence suggests that there is a big disparity between the actual μ (or μ_m) value obtained from the measured stresses in different locations and the static frictional coefficient determined in the laboratory [7,8,19,59,87,94–95]. During the stress accumulation in the crust, the μ (or μ_m) value in the deep part cannot easily reach the range of 0.6–1.0 (or 0.5–0.7) or even as low as approximately 0.1–0.2 [96], which is quite different from the shallow measurements. Thus, the real stress buildup ability in the shallow crust is likely to be much smaller than Byerlee's scope. Li *et al.* [59] believed that μ_m of 0.5–0.6 can be considered an index of a high-stress buildup. This condition implies that a faulting activity probably occurs in the high-stress accumulation zone, and the range of 0.2–0.5 of μ_m can be used for the lower and upper limits regarding fault activation prediction. Moreover, the above analysis shows that regardless of the lithology and roughness of the fault plane, the fault slip occurs in a specific friction coefficient range. However, the friction coefficient is largely governed by fault gouges, which contain clay and other substances and appears to degrade the friction coefficient and associated friction strength [78]. Some test results [94,97–99] have indicated that the friction coefficients of fault gouges in various regions are remarkably smaller than those obtained by Byerlee. Therefore, the existence of fault gouges and their friction features should be considered to make a complex analysis of the mechanical stability of different types of faults. In addition, fault instability is related to the change in stress orientation, the increase in differential stress close to the fault, and the angle between the σ_H direction and a normal fault. Thus, the simple increase in the μ_m value does not necessarily mean that fault reactivation will take place. Of course, the discussion about the link between the frictional coefficient and Byerlee's law is preliminary, and there is still a fierce debate on the frictional coefficient of laboratory-scale samples and the whole fault. Furthermore, the dynamic friction coefficient that should be used for fault activity analysis is usually lower than the static friction coefficient when two rocks move relatively [100–101], so the dynamic friction coefficient of the fault needs to be further studied.

In addition, according to Chang *et al.* [8], the parameter R can reflect the best-fit stress model that makes the fault slip and is calculated as

$$R = \frac{\sigma_1 - \sigma_2}{\sigma_1 - \sigma_3} \tag{7}$$

Essentially, R reflects whether the correlation of σ_1, σ_2, and σ_3 is beneficial to the fault slip. For the best-fit stress model, the R value is estimated to be approximately 0.5; that is, the value of σ_2 is close to the average value of σ_1 and σ_3, which may lead to a fault slip [2,7–8]. The histogram distribution of the size of the parameter R has been counted by some researchers to evaluate the activity of faults [2,7–8,102]. The evaluation result of the parameter R is equivalent to those of the parameters μ and μ_m.

Through the above analysis, it can be appreciated readily that understanding the stress accumulation level and slip tendency of a fault under a given stress condition is crucially important to evaluating fault reactivation and associated geological disasters. Monitoring the stress in fault areas, observing the dynamic change in the stress state, and then investigating its relationship with the tectonic setting, as well as the preparation and occurrence of fault activity, are the keys to detecting precursor information and the dynamic mechanism of fault activities. Thus, the potential correlation between crustal stress and fault reactivation can be explored according to the stress measurement results. Nevertheless, many factors cause the change in the frictional coefficient in the evaluation of fault reactivation. In addition to stress conditions, fault stability is highly dependent on material properties, fluid environment, and loading rate at the fault. These factors exhibit significant spatial heterogeneity, which makes it more difficult to grasp the stress state acting on the fault from the current survey [22]. Hence, the analysis of fault slips only from the perspective of *in situ* stresses inevitably leads to some limitations. In addition, if only the shallow stress state is available, it is difficult to accurately assess the fault sliding without the knowledge of the correlation between the shallow and deep stress conditions in fault areas. Consequently, there are still some controversies about whether it is appropriate to use the Mohr–Coulomb criterion and *in situ* stress measurements to evaluate fault activities. Notably, these methods are actually qualitative or semi-quantitative analysis methods, and they do not contain information about the stress direction, which can only reflect the magnitude relationship of the principal stresses. To fully understand the mechanism of fault activation and the characteristics of precursory anomalies, further research should refer to related fields, including seismology, geophysics, and plate tectonics [2]. Therefore, these factors need to be comprehensively taken into account to improve the reliability of stress variation analysis so as to estimate the risk of fault reactivation objectively and accurately. Essentially, the fault activity

parameters and fault activity intensity index established by Liu *et al.* [103–104] are very promising and can quantitatively characterize fault activities. Furthermore, they provided a numerical simulation method using paleostress and strain based on geomechanical modeling, which can gain insights into the fault activity characteristics and genetic mechanism.

5. Correlation between *in situ* stresses and fault properties

The stress state in fault areas is influenced by various geological factors, among which fault properties, such as the fault strength (frictional coefficient of the fault), composition of fault gouges, and fault density, exhibit a considerable influence. This is a fundamental problem in fault mechanics.

The fault strength can be analyzed using information related to the *in situ* stresses in a fault area. The correlation between *in situ* stresses and fault strength can be explicitly illustrated by Mohr's circles. The linear Mohr–Coulomb failure envelope and maximum shear stress (τ_{max}) along a fault are given by [66,78]

$$\tau = \mu\sigma' \tag{8}$$

$$\tau_{max} = \frac{\sigma'_H - \sigma'_h}{2} \tag{9}$$

where σ'_H and σ'_h are the maximum and minimum effective stresses, respectively; and τ_{max} is the maximum shear stress.

Zhang and Zhang [105], Zhang [66], and Li *et al.* [78] studied the interaction of the frictional coefficients of faults, shear stresses, and effective stresses in a critically stressed condition. Fig. 8 shows three Mohr circles incorporating linear Mohr–Coulomb failure envelopes derived from three frictional coefficients of μ = 0.2, 0.4, and 0.6 [78]. μ = 0.6, 0.4, and 0.2 indicate strong, moderate, and weak faults, respectively. As plotted in Fig. 8, under the critically stressed condition (before shear failure), a strong fault corresponds to a larger Mohr's circle with a smaller effective minimum stress (σ'_{hs}) and larger shear stress (τ_{smax}). A moderate fault is characterized by a medium Mohr's circle with a medium effective minimum stress (σ'_{hm}) and medium shear stress (τ_{mmax}). However, a frictionally weak fault that defines the smaller Mohr's circle possesses a lower shear stress (τ_{wmax}) and higher effective minimum stress (σ'_{hw}) to maintain fault stability. This theoretical understanding may potentially explain the small shear stress inferred along deep faults, such as the Tan–Lu fault zone [27] and San Andreas Fault zone [93,106]. Moreover, the correlation between frictional strength and horizontal principal stress imposes an important influence on the formation process of fault structures in the crust and slip tendency. For instance, to prevent a fault with weak friction from sliding, a higher minimum stress (or smaller maximum shear stress) is needed [105] to be favorable for bearing a high-stress environment. Meanwhile, for a fault with a certain strength, under different stress conditions, the fault can be stable without spontaneous sliding transient or unstable with spontaneous failure. As a fault is unstable, sliding may destroy part or the entire fault [107]. In some cases, faults alternate between these behaviors in multiple cycles.

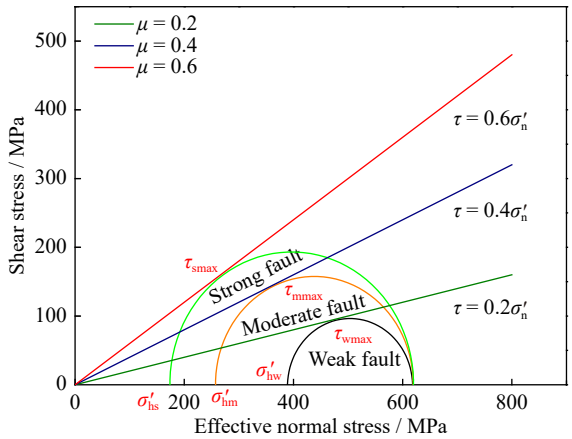

Fig. 8. Interaction of the frictional coefficients of the fault, shear stresses, and effective stresses in a critically stressed state [78]. Reprinted from *Bull. Eng. Geol. Environ.*, 81, P. Li, Q.F. Guo, and M.F. Cai, Contemporary stress field in and around a gold mine area adjacent to the Bohai Sea, China, and its seismological implications, art. No. 86, Copyright 2022, with permission from Springer Nature.

From another point of view, the level of stress that can be accumulated in a fault is determined by the fault strength. The larger the fault strength, the greater the stress level accumulated in the fault. When a fault cannot withstand the excessive accumulated stress, the fault will slide and release the energy, and the stress will be reduced accordingly, thereby forming the control of the stress field. This condition is in accordance with the concept that stress is constrained by the frictional strength of pre-existing well-oriented faults [108]. The heterogeneity of the present-day stress state could be regarded as the consequence of the heterogeneity of fault strength; that is, overall, the stress is smaller in the weak fault region and larger in the strong fault area [8]. Before the occurrence of a fault slip, stress accumulates at the fault and its surroundings. The development of stresses leading to faulting nucleation with time is schematically provided by Kanamori and Brodsky [109]. Theoretically, the fault strength and stress accumulation rate are constant. The basic process of faulting nucleation can be well understood, but its details are quite complicated. In fact, the fault strength and stress accumulation rate are not uniform with time. The activation of one fault segment will statically or dynamically change the stress on the adjacent fault segment and accelerate or decelerate the fault activity according to the geometry of the fault [109]. Furthermore, the strength of the crust is not constant with time. The migrating fluid may significantly weaken the crust, thus changing the time of fault activity. The stress drop during fault activities may also vary from event to event. These complex factors and their influence on the fault reactivation interval are not very clear, which makes it extremely difficult to accurately predict fault reactivation.

Meanwhile, friction rate dependence is a crucial parameter to control fault stability. Based on laboratory experimental

evidence provided by Ikari *et al.* [110], this parameter has been proven to have a systematic relationship with frictional strength, which applies to various mineral components related to natural faults. In general, fault zones are composed of a mixture of materials with contrasting strengths, which is likely to affect the diversity of sliding behaviors observed in natural faults. Fault gouges with weak friction are generally composed of phyllosilicate minerals, and their friction increases with the sliding velocity (i.e., velocity-strengthening behavior), which inhibits frictional instability [110]. On the contrary, fault gouges with large frictional strength show velocity-weakening and velocity-strengthening frictional behaviors. These substances are mainly quartzofeldspathic in composition, but in some cases, they comprise some phyllosilicate-rich gouges with large frictional coefficients. Accordingly, under certain stress conditions, for an unstable natural fault, the compositions of fault gouges require to be modified to maintain their stability. Hence, fault stability highly depends on the fault strength and *in situ* stress conditions.

In addition, a certain quantitative correlation seems to exist between the fault density and *in situ* stresses, but this relationship is usually vague because the actual fault density is usually scale-related [111]. According to Chang *et al.* [8], with the increasing cumulative fault length, the magnitudes of stresses expressed by stress ratios decrease. Although the population of faults may not be the only factor, it is probably related to the current stress magnitudes, so the low-stress area is characterized by a relatively high fault density. In this regard, a question that needs to be clarified is whether faults in low-stress areas can really help to release stress [8].

The correlation between *in situ* stresses and fault properties allows us to believe that the contemporary stress background may be controlled by faults in the tectonic stable area and that the current stress field is retained in the form of residual stress through stress relaxation caused by fault slips. Townend and Zoback [75] also proved the concept of "stable crust" from the point of view of *in situ* stresses and faulting, and they believed that the stable intraplate crust would suffer continuous small-scale damage. Furthermore, this correlation enhances the ability to predict fault slip tendency via stress measurements, which can be adopted to further refine the assessment of fault reactivation risk in the future.

6. Conclusion and prospect

An interaction occurs between the *in situ* stress state and faults, which leads to complex geological phenomena and has been widely concerned. Substantial progress has been made in this interaction. Generally, the stress state, including magnitude and direction near a fault, changes significantly locally, and stress jumps or discontinuities across faults, resulting in stress anomalies. Conversely, the variation in the stress state may also cause the mechanical instability of faults. Different stress states often trigger fault activities with different properties. Hence, it is necessary to fully understand the stress condition of a fault area for distinguishing the frictional strength of faults, predicting the sliding tendency of faults, and evaluating the risk of fault reactivation, which is of great engineering and scientific significance.

Qualitative fault stability analysis using some characteristic parameters under the contemporary stress environment provides a method to assess the relative risk of fault sliding. However, in an actual geological environment, the coupling of endogenous and exogenous geological processes makes the instability mechanism and sliding process of faults extremely complicated. Fault activities may involve a complex stress history, and the quantitative relationship between the stress change and fault instability is almost unknown. In addition, it is still difficult to accurately estimate fault activities, especially long-term activities. To further understand the coordination between the current stress state and fault activities, it is critically important to determine the stress state close to faults and key structural parts and evaluate its change with time. Hence, in the future, a large number of stress measurement campaigns, especially measurements at greater depths, will be essential. Meanwhile, long-term continuous real-time stress monitoring of the target position on a fault should be conducted. In addition, the relationship between the stress condition on the fault and the specific structure and seismic background should be comprehensively analyzed based on the independent seismic geological observation results in the field and the circumstantial evidence provided by the laboratory tests and numerical simulations to identify the key reasons affecting the stress changes around faults. Accordingly, the spatial and temporal stress changes near/on potentially dangerous faults can be explained, and the evolution of crustal rock strain and the mechanical process during fault activities can be revealed. More importantly, much attention shall be paid to distinguishing the genetic mechanisms of abnormal stress states and the type and scale of stress variations and exploring the mechanisms of pre-faulting anomalies and fault reactivation.

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Conflict of Interest

The authors declare that they have no conflicts of interest.

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