Evidence for seismogenic normal faults at shallow dips in continental rifts

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Abstract.

Several recent observations indicate that normal faulting earthquakes occasionally occur on faults dipping < 35°, dips often considered shallow. Most of these occur in the Woodlark and Aegean rifts. The Woodlark and Aegean rifts are found to generate significantly more earthquakes than others, and are the most rapidly extending, so display the widest variety of fault behavior. Even within the Woodlark rift system extension rates vary along strike, with the shallowest-dipping faults confined to the most rapidly rifting segment. Here, several events (MW 6.0 – 6.8) feature nodal planes dipping 23-35°. These planes are subparallel to shear zones bounding nearby metamorphic core complexes, including one imaged to 8-9 km depth by seismic reflection profiling. In the western Gulf of Corinth at least one large event (Mw 6.4) occurred on a fault dipping ~33°. As the Woodlark example, this rift segment exhibits a high opening rate (10-20 mm/yr). Several other cases elsewhere, based on older historic data, microseismicity, or geologic inference suggest seismic slip at similar or shallower dips. Still, no documented large earthquake exhibits seismic slip on sub-horizontal surfaces (dip <10-15°). Stress rotation may explain the 23-35° dips, but thus far no realistic mechanism has been found. More likely, these faults represent surfaces somewhat weaker than surrounding rock, through some combination of modest cohesion of the surrounding rock and slightly lower frictional coefficients on the fault. Such weakening may be a consequence of high slip rates, which rapidly generate large offsets and mature fault systems.
Although fault mechanics appears to preclude brittle slip on low-angle normal faults, dipping less than 30° (e.g. Anderson 1951), several structures found in the geologic record appear to suggest such slip is possible (e.g. Wernicke 1995). The seismic record, until recently, was absent of any well-documented normal-faulting earthquakes with such dips (Jackson & White 1989), in part prompting the development of theories for generating low-angle faults without requiring brittle slip at those dips (e.g. Buck 1988; Wernicke & Axen 1988). In many of these theoretical studies, dips of the active portion of normal faults are limited to ≥45° (e.g. Lavier et al. 1999). However, in recent years several examples of normal faulting earthquakes have been documented for which dips less than 35° have been claimed, some of which may show dips < 30° (Abers 1991; Aber 1997; Bernard et al. 1997; Doser 1987; Johnson & Loy 1992; Miller & John 1999; Rietbrock et al. 1996; Rigol et al. 1996). This evidence has been used as demonstration that low-angle normal faulting can occur seismically.

The core of this paper is a review of these observations and the settings in which they occur, with focus on the Woodlark Rift of Papua New Guinea. A global analysis of seismicity data shows that background seismicity rates in rifts correlate well with extension rates, so it should come as no surprise that most evidence for these “unusual” earthquakes comes from those rifts most rapidly extending (Woodlark and the Aegean). Such a correlation may reflect sampling bias (Wernicke 1995), or may reflect more favorable conditions (fault weakening, block rotation, etc.) at high extension rates. The focal mechanisms for the potential low-angle earthquakes reveal two things. First, several large earthquakes clearly rupture fault planes dipping 23-35°, so models of fault generation requiring brittle slip at dips of ≥45° require revision. The 23-35° dips are consistent with frictional properties expected for mature faults with well-developed gouge zones (e.g. Byerlee & Savage 1992), and unusual mechanics may not be needed. Second, at present, no large earthquake has been found showing slip on a fault unequivocally dipping less than 20°, although such slip can be found in microseismicity. However, the record of reliable focal mechanisms is no more than 35 years long, much less than large-earthquake repeat times, even in the few rifts where 25-35° dips are found. Thus, the presence of low-angle normal faults in the geologic record remains enigmatic – we cannot yet conclude whether shallower-dipping faults are truly aseismic, or merely occur infrequently.

The use of the term “low-angle” varies considerably; it most often refers to normal faults dipping less than 30° but is sometimes applied to a wider range of dips. Although the 30° dip threshold has mechanical significance, it lies in the middle of the uncertainty range of dips for many of the earthquake-generating faults discussed here, so use of the term “low-angle” may confuse. The events discussed here are interesting because they show faulting at dips close to or less than 30° or because they have been claimed to be “low-angle”; in any case, the faults slip despite the relatively high shear stresses required for neighboring rock. An alternative term is needed to describe these earthquakes. Rather than use the term “low-angle”, I will describe these fault geometries as “shallow-dipping”, referring to specific dip ranges wherever possible, and hopefully will offend fewer readers.

### Table 1. Active Continental Rift Systems

<table>
<thead>
<tr>
<th>Rift</th>
<th>Length, km</th>
<th>Width, km</th>
<th>Opening Rate, V mm/yr</th>
<th>Maximum Earthquake Depth, km</th>
<th>Extrapolated number of events N with ( M &gt; 4 ) per year, per ( LVH \times 10^7 m^3 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Woodlark</td>
<td>600</td>
<td>100</td>
<td>25-40</td>
<td>7-9</td>
<td>6.8 ± 0.9, 2.6 ± 0.7</td>
</tr>
<tr>
<td>Aegae</td>
<td>800</td>
<td>500</td>
<td>35</td>
<td>10-15</td>
<td>12.8 ± 0.2, 2.9 ± 0.5</td>
</tr>
<tr>
<td>Basin-Range</td>
<td>1200</td>
<td>600</td>
<td>10</td>
<td>15</td>
<td>1.7 ± 0.1, 1.2 ± 0.3</td>
</tr>
<tr>
<td>Red Sea/Suez</td>
<td>2200</td>
<td>250</td>
<td>7-12</td>
<td>5-10</td>
<td>1.2 ± 0.1, 1.7 ± 0.6</td>
</tr>
<tr>
<td>Baikal</td>
<td>1500</td>
<td>100</td>
<td>4.5</td>
<td>25</td>
<td>1.4 ± 0.2, 1.2 ± 0.7</td>
</tr>
<tr>
<td>East Africa</td>
<td>3000</td>
<td>100</td>
<td>&lt;3</td>
<td>35</td>
<td>2.0 ± 0.1, 3.8 ± 3.1</td>
</tr>
<tr>
<td>Rhine</td>
<td>400</td>
<td>100</td>
<td>0.5-1</td>
<td>20</td>
<td>0.09, 0.6 ± 0.3</td>
</tr>
<tr>
<td>RioGrande</td>
<td>1000</td>
<td>100</td>
<td>&lt;3</td>
<td>15</td>
<td>0.1 ± 0.1, 0.5 ± 0.4</td>
</tr>
</tbody>
</table>

Descriptive data from Ruppel (1995) and other sources as discussed in text. Opening rates and lengths refer to continental segments only. Annual earthquake rates extrapolated from maximum-likelihood regressions (Figure 1), normalized to along-strike length of rift, or to product of \( L \), seismogenic zone depth \( H \), and opening rate \( V \). Uncertainties are 1-σ, for \( N \) from regression; propagated uncertainties in \( V \) and \( H \) assumed uniformly distributed at 5 mm/yr and 2.5 km, respectively, unless range is shown. Seismicity data for Rhine Graben from Ahorner (1983), ISC catalog elsewhere.

### Seismicity in Rifts

As in other tectonic settings, seismicity in rifts is controlled by a combination of plate motion rates, seismic coupling, fault geometry, and the short duration of instrumental records of earthquakes. To illustrate this, a comparison of the seismicity rates for several of the world’s rifts to known extension velocities shows good correlation (Figure 1, Table 1). Other compilations have primarily concentrated on the largest earthquakes by cataloging fault plane solutions (e.g. Jackson & White 1989) or cumulative seismic moment (Jackson & McKenzie 1988), and so provide useful insights into extension rates, as the largest earthquakes constitute the majority of the extension ac-
commodated by earthquakes. However, such studies are often hampered by the relatively short period of instrumental recording compared to the recurrence time of most faults, making the present comparison a useful complement.

Table 1 compares seismicity rates with rift opening rates for the several rifts studied; Ruppel (1995) discusses other parameters. Seismicity rates calculated here are based on seismic moment estimates derived from standard catalogs (Figure 1), normalized by the along-strike length of the rifts. These data provide both seismicity rates as a function of moment-magnitude and estimates of detection thresholds in different rift. Again, the seismicity rates do not provide estimates of overall extension rates in a region, which is dominated by the occurrence of infrequent, large earthquakes, but describe the background seismicity that makes up most of the earthquake record.

Rift opening rates are compiled from a number of sources, with preference given to geodetic rates in the Basin and Range (Dixon et al. 1995), Aegean (Le Pichon et al. 1995), the Rio Grande (Savage et al. 1985), Baikal (Calais et al. 1998), and the Rhine Graben (Ahorn 1983). Elsewhere, opening rates are based on magnetic lineations in the Woodlark Rift (Taylor et al. 1995) and the Red Sea/Gulf of Suez. Finally, an upper limit of extension rates for East Africa can be gained by closure of global plate velocity circuits, following DeMets et al. (1990). Uncertainties typically lie in the range of a few mm/yr, so differences between fast and slow rifts are significant.

These compilations illustrate several trends in the rift-seismicity record. First, seismicity rates correlate well to plate opening rates. When normalized to extension rates (F) and depth of seismogenic zone (H), seismicity rates show remarkable constancy (Table 1). This suggests that large variations in seismic coupling sometimes inferred from seismic moment release rate may be partly attributable to the short historical record for large earthquakes, as may be true for subduction zones (McCaffrey 1997). Second, detection of earthquakes is poor in many but not all rifts. For example, the Woodlark Rift shows evidence of incomplete sampling at $M_W < 5.2$ and almost no earthquakes in the ISC catalog with $M_W < 4.8$, while earthquakes as small as $M_W = 3.0$ are routinely reported for the Aegean and Basin and Range.

After correcting for sampling, two rifts stand out as having high seismicity, the Woodlark and Aegean (Table 1). These are the two most likely places to see rift-related earthquakes, and they are also regions where extensional geometries that accommodate large amounts of extension are favored.

Large Earthquakes

The Woodlark Rift, Papua New Guinea

Setting. The Woodlark Rift (Figure 2) shows a full transition from sea-floor spreading in the east to extension in quasi-continental crust farther west (e.g. Mutter et al. 1996; Taylor et al. 1999). The continental section features development of several metamorphic core complexes along the rift (Davies & Warren 1988). Extension takes place about a pole somewhere west of 147°E, with rates in the continental section ranging from 10 to 40 mm/yr (Taylor et al. 1999). The easternmost extensional feature, Moresby “Seamount”, lies immediately west of the oceanic rift tip.
and also shows a north-dipping master fault zone that exposes basement rocks. The westernmost identified core complex, the Dayman Dome, lies on the Papuan Peninsula near 149°E longitude (Davies & Warren 1988) and appears to be active (Ollier & Pain 1980). Although the fault systems connecting these structures are poorly known except by seismicity, all show bounding shear zones dipping to the north and appear to lie in a single extensional domain. Subordinate deformation, reflected in seismicity and subsurface faults south of the D’Entrecasteaux Islands, suggests that some deformation may occur there west of 150.5°E (e.g., Mutter et al. 1996); its kinematic relationship to the main deformation zone is unclear.

![Image](image.png)

**Fig. 2.** Tectonic setting and seismicity of the Woodlark Rift, relocated, from Abers et al. (1997). Large circles: events for which waveform-derived focal mechanisms are available; small circles: others. The D’Entrecasteaux Islands are Goodenough, Fergusson, and Normanby, labeled G, F, and N respectively. Other labels, M: Moresby “Seamount”; D: Dayman Dome, a core complex on the Papuan Peninsula. Thick black lines show spreading center in eastern oceanic half of the rift (Taylor et al. 1999), or faults bounding core complexes in western half with barbs on hanging walls (Davies & Warren 1988). Dashed line surrounding white region denotes new (< 5 Ma) sea floor.

The large core complexes of the D’Entrecasteaux Islands (Figure 2) appear very young, and likely are still in the process of being exhumed. Despite tropical erosion conditions, the bounding shear zones still control the shapes of land surfaces and in many places appear fresh near their base, which along with other geomorphic observations suggest continued activity (Ollier & Pain 1980). Their footwall rocks show some of the most recent rapid uplift known, with footwall rocks experiencing 700-900°C and 5-6 kbar conditions as recently as 3-4 Ma, and 4-5 kbar at similar temperatures at 1.5 – 2 Ma (Baldwin et al. 1993; Hill & Baldwin 1993), requiring unroofing from midcrustal depths within the last 2 Ma. Late cooling recorded by apatite fission track ages on shear-zone rocks occurred at 0.4-1.0 Ma (Baldwin et al. 1993). Although such dates reveal little about the last 0.5 Ma history, the continued structural control of land surface morphology suggests that the rapid Quaternary exhumation continues through the present (e.g. Senior & Billington 1987).

The very rapid unroofing of these rocks correlates well with the rapid extension here, which ranges from 20 to 40 mm/yr across the core complex belt over the ~6 Ma history of extension. The dominant shear zones bounding the D’Entrecasteaux Islands core complexes dip consistently to the north at 20-25°, although some evidence of more complex faulting patterns exists (Hill et al. 1992).

**Earthquake locations.** Seismicity has been reprocessed and relocated utilizing three-dimensional velocity models and fully nonlinear relocation procedures, as reported elsewhere (Abers et al. 1997). The arrival times constraining the locations largely come from teleseismic catalogs, limiting earthquakes to magnitudes greater than ~4.8 (Figure 1). The resulting epicenters (Figure 2) show seismicity to be localized in a ~25 km wide band and confined to the north side of the rift zone near 9.5°S, at least east of 150°E. Most of these epicenters lie near the major fault zones bounding the core complexes of the Fergusson and Goodenough Islands and the Dayman Dome, and lie just north of Moresby Seamount (e.g. Mutter et al. 1996; Taylor et al. 1999). Hence, much of the seismicity appears to follow a contiguous system of faults along strike, and is generally consistent with movement on faults that lie on the north side of the core complexes. Such a pattern would be expected were the faults responsible for core complex exhumation presently active.
Earthquake fault planes. First-motion focal mechanisms (Ripper 1982; Weissel et al. 1982) document the extensional nature of the Woodlark-D’Entrecasteaux Islands region. These methods can have relatively large uncertainties, as the ray geometries are rarely ideal. Studies of Woodlark Rift earthquakes based on waveform inversion provide much clearer constraints on the geometry of faulting, and are discussed here (Abers 1991; Abers et al. 1997).

Overall, focal mechanisms reveal north-south extension (Figure 3), consistent with magnetic lineations (Taylor et al. 1995; 1999; Weissel et al. 1982). Normal faulting extends west to 148°W, the central Papuan Peninsula, throughout the region of metamorphic core complex development; the pole of opening must lie farther to the west. Waveform modeling shows source depths consistently less than 8-9 km below sea level (Abers et al. 1997); this cut-off depth is somewhat shallower than seen at other continental rifts (e.g. Jackson & White 1989) and is consistent with elevated temperatures during rapid rifting.

The most unusual aspect of faulting here is the abundance of normal faulting mechanisms with dips of 20-35° near the oceanic rift tip (151.7°E; Figure 3). Between 150.5°E and 153°E, the dips of the shallower nodal planes for normal-faulting earthquakes all lie between 23° and 36°. By contrast, shallow planes for earthquakes in the Papuan Peninsula rift segment (147°-150.3°E) all dip between 38° and 51°. This variation correlates with extension rates: rates are 20-40 mm/yr between 150.5°E and 153°E, and lower to the west. In at least two of the eastern cases, discussed below, there is good evidence that the shallower plane is the fault plane. Hence, the tendency toward shallow-dipping fault planes appears to correlate with extension rate.

1985 Gameta Earthquake. The largest event ($M_w = 6.8$; #1 on Figure 3) occurred 29 October 1985, 10-40 km east of the eastern core complex on Fergusson Island as determined by relocation, directly along-strike from the dominant shear zone that bounds the Oiatabu core complex of northeast Fergusson Island (Abers 1991). The focal mechanism shows one plane dipping NNW at 23°; waveforms rule out dips for that plane as steep as 30° (Abers 1991). Unfortunately, no direct method exists to choose the fault plane from among the two nodal planes, but comparison with the nearby shear zone favors the shallow-dipping plane. First, a north-dipping plane would be expected were the shear-zone bounding faults active in their present orientation. The shear zone dips 20-25° to the north. Second, the north-dipping plane is consistent with uplift of the core complex while the south dipping plane suggests that the island is submerging (Figure 4). The for-
mer scenario is preferred, since the core complexes appear to be actively exhuming, as discussed above. The alternative, that major faulting occurs today on south-dipping faults, suggests a radical (and ad hoc) termination to several million years of core complex exhumation in the last 0.4 Ma, while the geomorphic expression (e.g. Ollier & Pain 1980) continues to show core complex exhumation.

**Moresby Seamount Earthquakes.** A normal fault dipping ~30° seems likely along the north-dipping fault that bounds Moresby Seamount, the easternmost active fault block (Abers et al. 1997; Mutter et al. 1996; Taylor et al. 1995). Here, several earthquakes show normal-faulting focal mechanisms with one plane dipping near 30° (Figure 3). Multi-channel seismic lines reveal the presence of a planar north-dipping surface here that appears to be a downdip extension of the northern flank of the seamount. From an on-axis line this feature’s dip is constrained via dip-moveout techniques to lie between 25° and 35° (Abers et al. 1997), and parallel off-axis seismic lines have been used to infer dips closer to 25° (Taylor et al. 1995). One significant earthquake, the closest to the central seismic profile, shows a north dipping fault plane dipping 25-35° or nearly coincident with the imaged surface (Figure 5). This earthquake suggests that the feature imaged on the seismic line is a fault, and one that exhibits brittle failure (i.e. earthquakes) along its surface. This interpretation was confirmed by ODP Leg 180 drilling, which sampled the exposed fault surface at Site 1117 (arrow on Figure 5). Undeformed rock (gabbro) at ~100 m below sea floor grades upward toward the fault surface into increasingly sheared breccias and mylonites; the upper 4 meters below sea floor are interpreted as fault gouge of brittle origin (Shipboard Scientific Party 1999).

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The water depth overlying the Moresby Seamount earthquake ($M_H = 6.2$; #8 on Figure 3) is confirmed by the presence of prominent water multiples in its seismograms, and places tight constraints on the epicenter of this event (Abers et al. 1997). The period of these reverberations is exceedingly sensitive to the exact water depth overlying the source, and here requires a water depth of $3.0 \pm 0.2$ km, placing it under the deepest part of the basin (Figure 5). There, only the shallow-north dipping plane shows prominence in the seismic reflection record; the conjugate south-dipping fault system is overlain by only ~2 km of water. Hence the choice of fault plane is clear, and it seems likely that this event occurred on a normal fault dipping near 30°.

**Gulf of Corinth**

The Gulf of Corinth (Figure 6) accommodates a large fraction of extension in the Aegean, perhaps 15 mm/yr of the ~35 mm/yr total (Clarke et al. 1998). Recent GPS geodetic measurements in the western Gulf, east of Patras, have shown that ~12 mm/yr extension is localized across the single fault system underlying the Gulf (Clarke et al. 1997), accompanied by abundant microseismicity in the Patras area (Rigo et al. 1996). Utilizing waveform cross-correlation methods, Rietbrock et al. (1996) were able to reduce hypocentral uncertainties for the microseismicity to a few tens of meters, and show that most events lay in a plane dipping north at 12-20°. Although focal mechanisms of some microearthquakes show a nodal plane of similar dip (Rigo et al. 1996) reanalysis of an expanded data set shows that the north-dipping planes most often dip at ~30° at depth in the western Gulf (Hatzfeld et al. 2000). Perhaps, the plane of seismicity reflects the brittle-ductile transition rather than a fault plane (Hatzfeld et al. 2000).
crease to ~6 mm/yr (Clarke et al. 1996). These show focal mechanisms consistent with dips of 10-25° to the N (12-20° in the relocations of Rietbrock et al. 1996). Star shows focus of 1995 Aigion mainshock and heavy line shows fault plane inferred from seismic waveforms and geodetic observations (Bernard et al. 1997). N-S section transects western Gulf near 22.0° E longitude. The “A” marks the town of Aigion, which was heavily damaged.

**Peruvian Andes**

In the Peruvian High Andes, normal faulting occurs at high elevations over an otherwise convergent plate boundary (e.g. Dalmayrac & Molnar 1981). Much of the evidence for extension comes from Quaternary fault scarps, as earthquakes are relatively rare. The largest normal-fault earthquake in the Andes occurred near Ancash, Peru, in 1946, and produced extensive surface rupture along the west-dipping Quiches fault (Bellier et al. 1991). First motions and waveform modeling show the fault plane dips 30° for this event, with a source depth of 15-17 km and an Mw of 6.8 (Doser 1987); a 10° uncertainty in dip is claimed by the author. The actual uncertainties could be larger, as instrumentation in 1946 was both poorly distributed and of uncertain calibration. The seismic moment and surface rupture give a source dimension of many tens of kilometers; a slightly larger seismic moment (Mw = 7.0) was estimated from fault offset and length (Bellier et al. 1991). Trenching across the surface trace reveals 1.7-3 m of offset across a fault, which together with one previous post-glacial rupture imply a vertical displacement rate less than 0.25 mm/yr. Hence, the Ancash earthquake represents a third region in which moderately low-angle normal faulting may have produced an earthquake. Unlike the previous two examples, it occurred in a setting where the long-term extension rate is fairly low; geodetic data limit extension across the Altiplano to less than a few mm/yr (Norabuena et al. 1998).

**Pre-Historic Examples**

The geologic record contains many examples of normal faults that appear to have moved at shallow dips (Axen 1992; Wernicke 1995). In the two examples below, some evidence exists for seismic slip or surface offset; low-angle faulting is claimed to occur via earthquakes.

Johnson and Loy (1992) showed that late Quaternary fault scarps along the Santa Rita fault, southern Arizona, extend to depth as planar features dipping 20° on seismic
reflection sections. While no earthquakes have been observed on this fault, geomorphic evidence suggests movement within the last 100,000 years, over an area corresponding to an earthquake with $M_c = 6.7-7.6$. Although seismic slip is not confirmed, this fault likely moved at depths normally considered within the brittle regime and at a dip near 20°.

Near the Chemehuevi-Sacramento Mountains, depositional patterns led Miller and John (1999) to infer earthquakes along a low-angle fault (dip ≤ 30°), active between 23 and 12 Ma. They argue on the basis of the geometry and distribution of coarse-grain strata that sediments were deposited directly upon a detachment fault while that fault lay at shallow dip (< 30°) and was active. Megabreccias represent catastrophic depositional events, which they argue were triggered by earthquakes with magnitude ≥ 6.0, the only clear source for which is the detachment fault itself. Fault slip rates here were fairly rapid, 7-8 mm/yr during peak extension.

**Discussion**

These observations suggest that earthquakes do occasionally occur on normal faults that dip near 30°, the low end of the range found by Jackson and White (1989). However, their occurrence is rare and perhaps tied to special geologic circumstances. In most cases such faulting is found in fault systems of high slip rate (>10 mm/yr), the one exception being the Ancash, Peru event, on which dip constraints are relatively weak. Furthermore, except for microseismicity the record does not yet include an earthquake on a low-angle fault dipping <20°. Whether that absence reflects a fundamental physical limit of normal faulting or merely a longer recurrence time for such events (Wernicke 1995) remains to be seen.

Explanations for such faulting generally fall into two categories, those that rely upon basal tractions (e.g. Melosh 1990; Yin 1989) or magmatic intrusion (Parsons & Thompson 1993) to rotate stress into a favorable geometry, and those that appeal to fault weakening (e.g. Axen 1992). While the former approach has many attractions, it has been shown that most stress-rotation models do not generate sufficient shear stresses to produce brittle sliding or (in the magmatic intrusion case) promote low-angle faulting in very small areas (Wills & Buck 1997). Some situations with extreme flow rates at the base of a brittle layer may produce the sufficient differential stresses (Westaway 1999). Such a situation may indeed be a consequence of rapid extension, particularly if extension is driven from below in some way (e.g. Abers et al. 1997). Still, without constraints on the temperature, flow and rheology of the lower crust, these models are difficult to test and remain speculative.

**Failure requirements**

**Analytical estimates.** Do the observed dips actually require very weak faults? Here, we calculate the stress conditions observed on faults for the dips observed, for simple (i.e. Anderson-like) driving forces, and find that only slight deviations from Byerlee’s law are necessary to explain the observed fault geometries. Figure 8 illustrates the calculation: a fault with dip $\delta$ is cohesionless with static frictional coefficient $F$, while the surrounding medium is cohesive. At the relatively shallow depths of faulting observed (Figures 4-7) the effective maximum compressive (vertical) stress $\sigma_y'$ is low and the tensile strength $T$ limits rock strength in extension (Figure 8b). The minimum effective stress during faulting ($\sigma_y'$) may be negative in this situation: as long as $\sigma_y' > -T$, tensile failure or hydraulic fracture does not occur. To estimate this limit, parabolic Griffith – Coulomb failure criterion is assumed for the intact rock (Sibson 1998), while the fault is cohesionless, with stresses calculated as in Figure 8. In this scenario, faulting continues on the cohesionless fault at non-optimal angles until either Coulomb (shear) fracture or tensile (hydraulic) fracture occurs in the surrounding rock. While the true stress state is likely affected by bending and topography (e.g. Buck 1993; Forsyth 1992), this calculation provides a first-order estimate of the physical limits to fault dip, with few assumptions of plate rheology.

Below sea floor, effective vertical stress is $\sigma_y' = \rho g (z - z_w) (1 - \lambda)$, where $\rho$ is the rock density, $g$ is gravitational acceleration, $z - z_w$ is the depth below the sea floor, and $\lambda$ is the ratio of pore pressure to lithostatic pressure (~ 0.4 for hydrostatic pore pressure) (e.g. Brace & Kohlstedt 1980). If the fault slides frictionally at coefficient of friction $F$ and dip $\delta$, a relation exists between $\sigma_y'$, $\sigma_y''$, $\delta$, and $F$ on the fault, so that a non-dimensional horizontal stress can be defined by:

$$S = \frac{-\sigma_y'}{\rho g (z - z_w)} = (1 - \lambda) \left( 1 - F \cot \delta \right) \frac{1}{1 + F \tan \delta}$$

(following Sibson 1985). The hydrofracture condition, $-\sigma_y' < T$, can then be recast in terms of dimensionless parameters such that hydrofracture does not occur so long as

$$S < \frac{T}{\rho g (z - z_w)^3} = T'.$$

Figure 8c shows values of $S$ vs. $\delta$ for several appropriate values of $\lambda$ and $F$. For typical rocks with simple faults $F \sim 0.6$ (e.g. Brace & Kohlstedt 1980), but in well-developed fault zones with thick gouge layers, $F \sim 0.5$ may be more appropriate (see review of Lockner 1995). Many scenarios (and some direct evidence) call for super-hydrostatic conditions with $\lambda > 0.4$ (e.g. Axen 1992; Hickman et al. 1995) although that may not be necessary here. The hydrofracture (tensile failure) condition is less clear; $T$ in crystalline rock varies from 10 to 40 MPa with values of 25 MPa typical for mafic rocks (Lockner 1995), comparable to cohesion in intact rock assumed in more complex faulting models (Buck 1993; Forsyth 1992; Lavier et al. 1999). In the Woodlark Rift, earthquakes do not exceed 8 km depth and water depth varies from 0 to 3 km, so $\rho g (z - z_w) \sim 130 - 210$ MPa, giving a hydrofracture threshold of

$$T' = 0.11 - 0.19$$

at the base of the seismogenic zone as indicated on Figure 8c. $T'$ could be considerably larger at shallower depths. Additional calculations show that extensional failure should be reached before shear (Coulomb) fracture as long as $T' > -0.25(1-\lambda)$, for reasonable internal friction coefficients, over the relevant depth range, so Coulomb fracture can be ignored. Thus, the limitations set by extensional failure...
(hydrofracture) control the minimum allowed fault dip. A slight weakening and small cohesion can be very important in this scenario, as effective stresses are small.

\[ \sigma_1' = \sigma_{1'} + \Delta \sigma \]

\[ \delta \]

\[ r \]

\[ F = \tan \phi \]

\[ 2 \tau \]

\[ \text{Hydrofracture limit} \]

The stress calculation just described assumes simple stress geometry, in particular one with no stress rotations. Topographic variations could produce large changes in stress fields at least near the surface (e.g. McTigue & Mei 1981). One example is given in Figure 9, based on topographic effects caused by Moresby Seamount (calculation described in Abers et al. 1997). Here, tectonic stresses are applied by stretching the crust until frictional failure at optimal angles \( F = 0.7 \) is achieved somewhere at each depth. To isolate the topographic effect, cohesion is ignored and hydrostatic pore pressure is assumed. The faulting region is in a state of decreased shear stress owing to a stress shadowing effect, but stresses are rotated slightly to favor low-angle faulting. The net result is that failure is possible on the fault dipping 25° at \( F \) near 0.3. The added effects of elevated pore pressure or cohesion should allow sliding at \( F \) closer to 0.5 - 0.6, as suggested in the previous section. Similar topographic stresses are expected for the Gulf of Corinth, where topography likewise defines a narrow rift basin between tall mountains.

**Fig. 8.** Relationship between fault dip (\( \delta \)), rock tensile strength (\( T \)), and effective principal stresses (\( \sigma_{1'} > \sigma_{3'} \)) required for low-angle normal faulting when \(-T < \sigma_{1'} < 0\). (A) Assumed normal fault geometry in cross-section; fault has frictional coefficient \( F \) and no cohesion. (B) Mohr circle diagram showing relation between shear stresses \( (\tau_{0}, \sigma_{0}) \) and effective normal stresses \( (\sigma_{3'}, \sigma_{1'}) \). Stresses on the fault lie on the circle; the line labeled “friction” is the frictional failure envelope \( (F = 0.6) \) and that labeled “failure” is the assumed Coulomb-Griffith fracture envelope for intact surrounding rock with tensile strength \( T \) (after Sibson 1998). Hydraulic (tensile) fracture occurs if \( \sigma_{1'} \) becomes less than \(-T\). (C) Calculated stress ratio \( S \), defined in text, vs. fault dip % for various assumptions of \( F \) and \( \lambda \) (pore pressure ratio). Gray bar shows the range of hydrofracture limits, \( T' \), estimated in text for the base of the seismogenic zone in the Woodlark Rift.

**Interpretation.** Figure 8c shows that faults dipping as low as 25° should slip without hydraulic fracture for all conditions, and this limit may be decreased to 15-20° with modest decreases in strength or increases in pore pressure. Normal faults rapidly enter the hydraulic fracture regime as dips decrease below 10-15°, and slip on such faults would require other physical mechanisms (e.g. Axen 1992; Wernicke 1995). All observed large earthquakes occur with dips above this limit, and mechanical problems exist only if the microseismicity evidence for dips <20° (e.g. Rietbrock et al. 1996; Rigo et al. 1996) reflects large-scale stresses. The large event with the shallowest dip, the Gameta earthquake, has a centroid depth of 3 km (Abers 1991). At these depths the hydrofracture limit is not reached until \( S = 0.3 \), a condition easily reached on faults dipping near 20° provided cohesion remains high. Thus, we conclude that observed normal faults could be understood without requiring unusual frictional conditions.

**Topographic effects.** The stress calculation just described assumes simple stress geometry, in particular one with no stress rotations. Topographic variations could produce large changes in stress fields at least near the surface (e.g. McTigue & Mei 1981). One example is given in Figure 9, based on topographic effects caused by Moresby Seamount (calculation described in Abers et al. 1997). Here, tectonic stresses are applied by stretching the crust until frictional failure at optimal angles \( F = 0.7 \) is achieved somewhere at each depth. To isolate the topographic effect, cohesion is ignored and hydrostatic pore pressure is assumed. The faulting region is in a state of decreased shear stress owing to a stress shadowing effect, but stresses are rotated slightly to favor low-angle faulting. The net result is that failure is possible on the fault dipping 25° at \( F \) near 0.3. The added effects of elevated pore pressure or cohesion should allow sliding at \( F \) closer to 0.5 - 0.6, as suggested in the previous section. Similar topographic stresses are expected for the Gulf of Corinth, where topography likewise defines a narrow rift basin between tall mountains.

**Fig. 9.** Stresses predicted on the Moresby Seamount Fault, Papua New Guinea, in the presence of both topographic and extensional forces, from Abers et al. (1997). Fault is assumed to be planar and dip ~25°, as Fig. 5. Extension drives region to frictional failure beneath its weakest point (beneath the “seamount” massif), assuming a coefficient of friction of 0.7 and hydrostatic pore pressure beneath sea floor. Additional forces are not considered, such as basal tractions or transient effects, and material is assumed cohesionless.

These two sets of calculations are not exhaustive, and many alternative mechanical calculations can be found in
the literature. Nevertheless, they show that reasonable assumptions about material strength and stress can account for normal-faulting earthquakes with dips of 20-35°.

Conclusions

Seismicity in rifts tends to correlate with overall rate of extension. This observation explains the tendency of unusual behavior such as low-angle faulting to be reported most commonly in the Woodlark Rift and Aegean, the two most rapidly opening rifts. In these locales, normal faults dipping 23-35° are observed in the regions of fastest extension: 25-40 mm/yr near the rift tip in the Woodlark-D’Entrecasteaux region, and 10-20 mm/yr across the western Gulf of Corinth. These observations only slightly extend the range of previously-reported dips for normal faults, but demonstrate that the low end of the dip range is commonly achieved; models of faulting that restrict dips to ≥ 45° can not explain many earthquakes. Several other examples of potentially seismogenic low-angle normal faults have been reported in the literature but none as well constrained. These observations suggest that processes associated with high extension rates (and high overall extension) on individual fault systems may produce the conditions needed for low-angle faulting. For example, well-worn faults should achieve effective frictional coefficients near 0.5. Such coefficients do not require very unusual materials but may be expected in faults which achieve large offsets in short time spans, so can develop mature, uncremented gouge zones.

The physical conditions for such faulting remain poorly understood. Except for the Aegean case, little is known locally about the geometry of faulting, and neither the conditions within the fault zones nor the behavior of the deeper ductile crust are well known. In particular, the role of the lower crust in these processes may be critical, yet few direct observations exist on its state. Also, direct measures of the cohesion, mechanical strength and permeability structure of these active fault zones would allow several models to be tested. Future experiments and observations aimed at resolving these ambiguities may be necessary.

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