Tectonic and Structural Controls on Geothermal Systems

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Abstract

Although heat source, in terms of enthalpy and/or temperature, is usually considered the prime factor driving geothermal systems and has traditionally been the basis of their classification, this approach undermines the importance of the tectonic settings they exist in. The tectonic setting defines the regional stress regime that controls the permeability structure and determines the nature of heat source—magmatic or non-magmatic—and thermal regime—convective or conductive or a combination of the two, and also prevailing geothermal gradient and heat flow. Moreover, despite the tectonic setting being favorable, the local stress regime may make a geothermal system either highly productive or uneconomic, depending upon whether it aids the fluid circulation or not. Understanding the tectonic setting and local structural conditions may help enhance a low-performing exiting geothermal system's performance or even create a new one by developing artificial fractures to facilitate fluid circulation if a heat source—magmatic or non-magmatic (viz. radioactive)—is available.

Keywords

Geothermal fluid • Tectonics • Structure • Faults • Basement • Play type • Heat source • Magmatic • Non-magmatic • Convection-dominated • Conduction-dominated • Stress • Strain

1 Introduction

A geothermal system has been defined quite precisely by Boden (2016) as a combination of processes involving heat transfer to the Earth's surface through convection or conduction from source to sink, which envisages its present-day understanding and application. This can be elaborated further as follows, encompassing the processes and practices involved therein. A geothermal system is a combination of favorable physicochemical conditions within a confined volume of the Earth's crust, viz. heat and mass transfer, temperature and pressure gradients, permeability (natural, artificial, or enhanced), hydrology and fluid flow, fluid composition, and mineral equilibria that together transfer heat through convection and/or conduction) from a heat source (magmatic or non-magmatic) to a heat sink, usually in the form of surface manifestations (viz. hot springs, fumaroles, geysers, mud pools) or wells for extracting hot fluids—vapor or liquid—for direct (i.e., using heat itself) or indirect (i.e., using heat for some process, viz. electricity generation) use.

Furthermore, a geothermal system is referred to as (i) blind or hidden in the absence of any surface manifestation and (ii) enhanced or engineered, when the flow rate of the hot fluids (which is naturally very low or virtually absent due to low permeability) is increased to a rate sufficient for economical use (viz. district heating, electricity generation) by augmenting the natural permeability through technological solutions, viz. hydraulic fracturing, stimulation or fracking. In enhanced geothermal systems (EGS), the circulating fluid can be natural hot fluid if a low permeability formation hosts a hydrothermal system; or it can be an artificially injected fluid if the formation of the geothermal system does not contain enough fluid volume for heat extraction, and these are referred as Hot Dry Rock (HDR) systems. Aforesaid physicochemical conditions and processes are controlled by prevailing regional and local stress regimes, which will be discussed in this chapter.

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There are efforts to denominate geothermal systems based on fundamental geological parameters that dictate geothermal resources' characteristics, the tectonic environment in which they were formed, and how those resources might best be explored and developed (e.g., Walker et al. 2005; Erdlac et al. 2008; Moeck 2014; Boden 2016). This approach is significantly different from the traditional classification of geothermal systems based on temperature and enthalpy—low, medium, and high (e.g., Muffler 1979; Muffler and Cataldi 1977; Hochstein 1988; Benderitter and Cormy 1990; Haenel et al. 1988; Sanyal 2005). Several workers considered these classifications inconsistent and insufficient to categorize a geothermal system from the development point of view (e.g., Lee 2011; Moeck 2014).

2 Importance of Structural and Tectonic Controls

The tectonic setting and geologic structure of geothermal systems, which are the results of the prevailing regional and local stress regimes, control the physicochemical characteristics of geothermal resources. An understanding of these controls not only provides valuable input for geothermal resource development but has also been the basis of a relatively new classification of geothermal systems as a catalog of "play types" similar to the oil industry (Moeck 2014). This catalog has evolved from the early ideas of Muffler (1973, 1976) to their refinement by Walker et al. (2005), Erdlac et al. (2008), and Boden (2016). This classification or cataloging is from a geothermal developer's perspective, based either on the geographical extent of favorable settings (Philips et al. 2013) or repeating sets of prospects with common characteristics defining a "play type" (King et al. 2013).

On the other hand, this classification is governed by the plate tectonic setting on a large scale and structural elements (local stress fields, rock mechanics, fracture systems) on a smaller scale. For example, whether a geothermal system or "play type" is related to convection or conduction-dominated heat transfer and/or is magmatic or non-magmatic depends on its tectonic setting. On the other hand, the local stress field's orientation controls fluid flow along the faults, and rock mechanics defines permeability anisotropy of the fractured reservoirs. Moeck's (2014) cataloging of geothermal "play types" has been used here as the basis for discussing the tectonic controls on the geothermal systems.

Since EGS and HDR development involves the creation of new fractures to increase permeability and the orientation as well as the growth of these fractures are strongly controlled by the stress field and rock mechanics, comprehending the stress fields and rock mechanics is particularly important for (i) designing the stimulation process and define injection rates for creating these artificial fractures (Moeck et al. 2009), (ii) keeping the induced fractures open during production and subsequent formation pressure drop (Moeck 2014) and (iii) risk assessment during injection through fault reactivation potential analysis using the slip and dilation tendency technique (Moeck et al. 2009), which also includes reinjection (Moeck and Backers 2011). This is achieved through 3D structural geological modeling, stress field analysis, and fault stress modeling during all the three stages of geothermal field development—exploration, drilling, and reservoir engineering (Moeck et al. 2009; Moeck 2014).

Understanding and characterizing the tectonic and structural controls on geothermal systems has been an ongoing focus on different scales, from plate tectonics (e.g., Muffler 1973, 1976; Heicken 1982) to the local structural regime (e.g., Rowland and Sibson 2004; Faulds et al. 2010a 2010b 2010c, Rowland and Simmons 2012). In short, it is the geologic setting that constrains the temperature, fluid composition, and reservoir characteristics of a geothermal system and establishes whether it is a convective or conductive system (Moeck 2014). The locations of geothermal fields worldwide are invariably tectonically controlled. They are often associated with block faulting, grabens or rifting, and collapsed caldera structures, with reservoir depths of around 1–3 km (Nicholson 1993). Typical settings are around active plate margins such as subduction zones (e.g., Pacific Rim), spreading ridges (e.g., Mid-Atlantic), rift zones (e.g., East Africa, Central India), and orogenic belts (e.g., Himalayas, Mediterranean). The characteristics of a geothermal system (viz. its thermal and hydrological regimes, fluid chemistry and dynamics, faults and fractures, stress regime, regional heat flow, lithology, rheology) are controlled by its tectonic framework.

Moreover, locally, faults can act not just as fluid conduits for a geothermal system but quite often form barriers for fluid circulation. In some cases, such barriers may lead to a segmentation of the system, some parts being more productive than others and some even unproductive. Mariposa geothermal field in south-central Chile has two lobes like wings of a butterfly (*mariposa* in Spanish), having different characteristics. This difference is evident from the magnetotelluric survey and structural studies for an ongoing geothermal development project (Hickson et al. 2011, 2012; Fox Hodgson 2012). Moreover, cap rocks that contain the heat of the geothermal system and overlie the reservoir may be breached due to high rates of uplift and ensuing erosion, e.g., in the Andes (Coolbaugh et al. 2015).

3 Convective and Conductive Heat Transfer

The heat transfer mechanism of a geothermal system is seldom either convective or conductive. It is instead a combination of both, dominated by one of them. Moreover, convection-dominated systems could be (i) magmatic, induced by a magmatic heat source (Fig. 1), or (ii) non-magmatic, in which the geothermal fluid originates as meteoric water that gains heat through the circulation to depth within a fault zone (Fig. 2). Consequently, structural controls have a significant effect on fluid flow pathways in such systems. Thus, besides a high-temperature gradient, high permeability is necessary to facilitate convection. The high geothermal gradient, natural fluid flow, and fluid dynamics are the characteristics of convection-dominated geothermal systems (Moeck 2014). So they host high enthalpy resources at shallow depth, making them most attractive for geothermal developers and investors.

On the other hand, conduction-dominated geothermal systems are marked by the absence of or limited convective flow of fluids and form low to medium enthalpy resources. Such systems are usually located in passive tectonic settings, with the geothermal gradient being average. They are less attractive for the developers and even more so for the investors, as the exploitable reservoirs are deeper (often more than 2 km) than convection-dominated geothermal systems. Moreover, conduction-dominated geothermal systems are hosted in low permeability rocks, viz. compacted sandstones, carbonates, massive granites. They need to create fractures to develop or enhance permeability to facilitate fluid circulation using EGS technology to make such "deep geothermal" resources economical (Moeck 2014). The conduction-dominated geothermal systems could be of magmatic or non-magmatic origin.

Magmatic activities can give rise to both conduction and convection-dominated geothermal systems, depending primarily upon the prevailing permeability conditions and the heat source. It has been reported that conduction-dominated systems hosted in or above igneous bodies are related to high radiogenic heat production, e.g., high heat-producing radioactive elements rich granites (e.g., Singh et al. 2018). Such systems occur in areas with no active volcanism and without or absence of present day tectonic activities. Moreover, the lack of large volumes of natural fluids marks the conduction-dominated magmatic geothermal systems due to low permeability conditions. As a result, such fluid-less systems require EGS technology, i.e., hydraulic fracturing and injection-induced circulation of fluids to transfer heat from depth to surface, for their development. On the other hand, convection-dominated magmatic geothermal systems require a magma chamber as the heat source in tectonically active areas.

In convection-dominated geothermal systems, whether magmatic or non-magmatic, fluids transport heat from the reservoir to the surface. As a result, temperature and volume of the fluids—vapor, brine, or both—that can be extracted from the reservoir, and the depth of the latter determines whether a geothermal resource is economical, as drilling of geothermal wells is expensive and risky, and even more so if the reservoirs are deeper. Moreover, to make a geothermal system sustainable, optimization between production and injection of the remnant thermal fluid is vital, extending a geothermal development project's life. Nonetheless, for the latter, origin of the reservoir fluid and their chemistry, recharge characteristics are also important but are beyond this chapter's scope.

Another aspect related to structural controls is the effect of topography on geothermal systems, the influence of steep topography, particularly highlighted by Hochstein (1988), which causes large volumes of meteoric water recharge convective geothermal systems through high infiltration rates. However, the effect of steep terrain on the hydraulic head is not limited to the convection-dominated geothermal systems in volcanic areas, as envisaged by Hochstein (1988). It applies to the sedimentary basins as well, e.g., Alberta Basin in Canada with a low enthalpy conduction-dominated systems hosted in carbonate and compacted sandstone reservoirs (Bachu 1995; Weides et al. 2012). Thus, steep

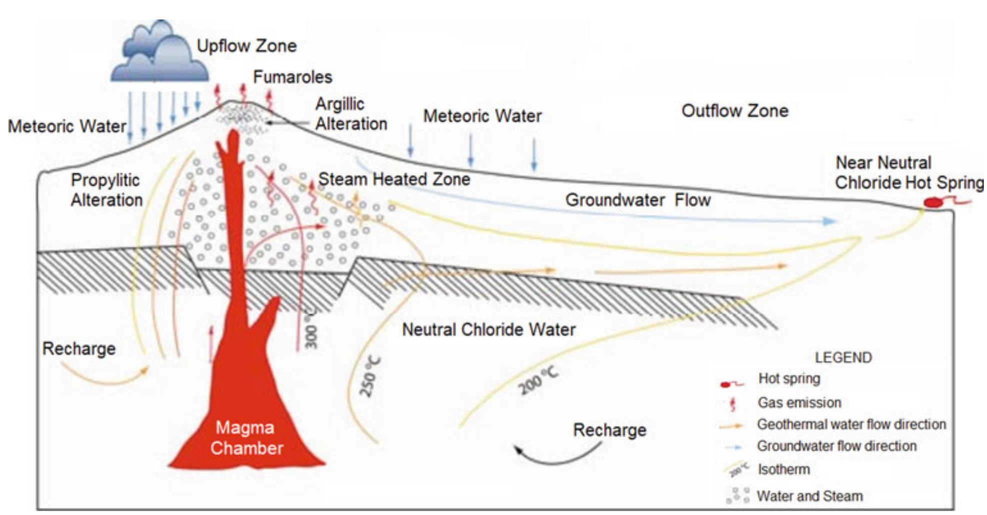


Fig. 1 Magmatic convection-dominated geothermal system (modified from Henley and Ellis 1983)

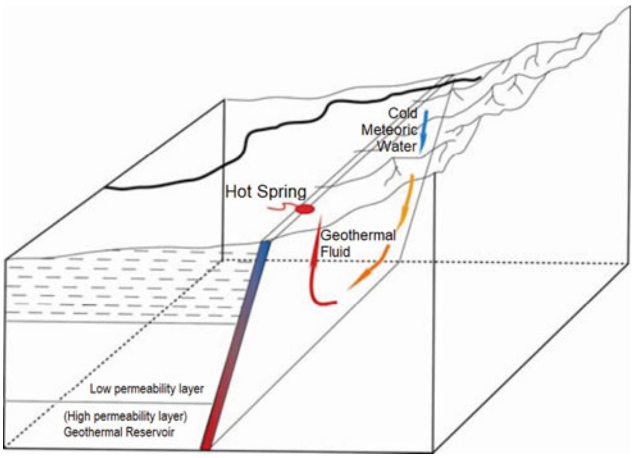


Fig. 2 Non-magmatic convection-dominated geothermal system (modified from Reed 1983)

terrain can affect infiltration in both high and low enthalpy systems.

Most of the geothermal power plant complexes across the world have convection-dominated high enthalpy geothermal systems with shallow magma chambers as the heat source. They are transacted by faults facilitating the rapid circulation of geothermal fluids and recharging the reservoir (Boden 2016). Considering that deep faults and active volcanism are associated with active plate tectonic margins, understanding dynamic tectonic processes at different scales is crucial to characterize convection-dominated high enthalpy geothermal resources. On the other hand, for conduction-dominated low enthalpy geothermal systems, understanding the present-day stress field particularly and the geodynamic evolution process is essential to develop or enhance permeability by creating new fractures. For the reasons and observations described above, modern cataloging of geothermal systems (e.g., Moeck 2014) is based on the plate tectonic setting (Fig. 3), a heat source (magmatic or non-magmatic), and local geologic controls on heat transport mechanism, storage system, and permeability structure. The following section presents some salient examples of major tectonic controls on geothermal systems.

4 Tectonic Controls on Geothermal Systems

4.1 Controls on Heat Transfer

4.1.1 Convection-dominated Geothermal Systems

Mantle convection-driven plate tectonic processes and associated volcanism at active plate margins presents favorable conditions for high enthalpy, convection-dominated geothermal system (Moeck 2014), viz. (i) magmatic arcs above subduction zones at convergent plate margins (e.g., the Andean Volcanic Arc), (ii) divergent margins—intra-oceanic (e.g., Mid-Atlantic Ridge) and intra-continental (e.g., East African Rift, Central Indian Rift), (iii) transform plate margins with strike-slip faults (e.g., the San Andreas) and (iv) intraplate ocean islands formed by hot spot magmatism (e.g., Hawaii). Major fault zones (e.g., Liquiñe Ofqui Fault Zone in Chile, Alam et al. 2010) can act as principal fluid conduits that lead to crustal regions of elevated heat flow. Upwelling asthenosphere and asthenospheric bulge may cause such high heat flow at subduction zones and beneath the rifts, respectively. Rapid exhumation and tectonic denudation in extensional terrains may also lead to increased heat flow, e.g., Rechnitz Window, Eastern Alps (Dunkl et al. 1998).

In convection-dominated geothermal systems, the circulating fluids transport heat from depth to shallow reservoirs or discharge to the surface. The igneous activity (e.g., magma chamber) in volcanic areas, faults in extensional terrains, or both (e.g., intrusive bodies at fault zones) control such systems. The circulating geothermal fluids, originating from infiltration of meteoric water from a high elevation, may receive contributions from the magmatic fluids in such systems (Moeck et al. 2014).

Magmatic Convection-dominated Geothermal Systems
Such systems can be associated with volcanism—(i) basaltic (e.g., divergent plate margins, as in Iceland, Arnórsson et al. 2008), (ii) basaltic to andesitic (e.g., island arcs, as in Java, Pambudi 2018), and (iii) andesitic to dacitic (e.g., convergent margin, as in Chile, Alam et al. 2010)—or plutonism (e.g., continent–continent convergent margins, as in Tuscany, Italy, Bertini et al. 2006). Magma chambers, the heat source for such systems in volcanic areas, control their fluid chemistry, which depends on the parental melt, magmatic recharge, and crystallization in the magma chamber. Volcanic convection-dominated geothermal systems can be separated into upflow and outflow zones (Hochstein 1988), the former being directly above the magma chamber (Fig. 2) and hosts the primary reservoir, which is the target for large-scale power production (Moeck 2014).

In contrast, the outflow zone is distal to the heat source and generally associated with a secondary reservoir (of medium to low enthalpy) and can be utilized for small power plants if the flow rate is sufficient (Hochstein 1988). The temperature gradient at the outflow zone typically increases at shallow depth and declines below the outflow layer (see the isotherms in Fig. 1), thus do not reflect a high-temperature geothermal reservoir directly beneath the

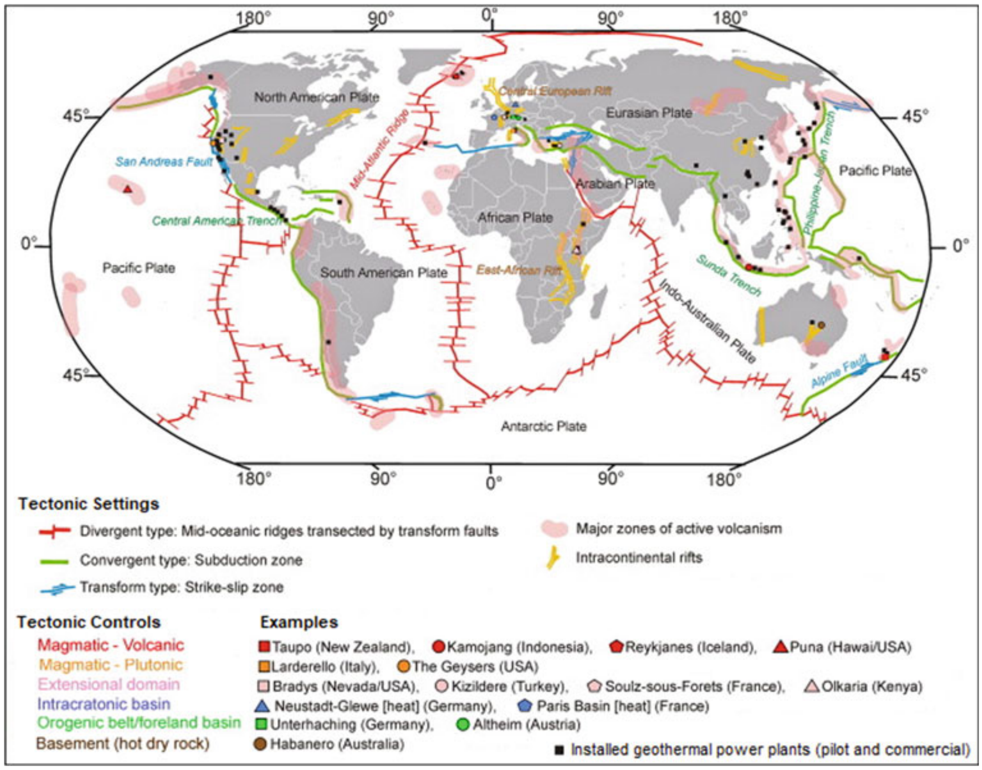


Fig. 3 Tectonic controls on geothermal systems with examples of major fields across the world and location of installed power plants (modified from Moeck 2014)

surface discharges, which are typically springs with or without travertine deposits. On the other hand, geothermal manifestations in the upflow zone are acidic springs with varying degrees of altered rock-forming alteration clays, indicating high-temperature reservoirs immediately underlying them. The upflow zone often has a vapor-dominated part above a liquid-dominated part.

In volcanic areas, the condensate layers in steep terrains can conceal high-temperature reservoirs formed by upwelling acidic fluids that get condensed and neutralized at a shallow depth above the heat source (Fig. 1). The surface discharges in the outflow zone from such condensate layers acquire the condensate layer's cation content. Thus the geochemistry of such discharges are different from that of the original vapor (Schubert et al. 1980; Hochstein 1988). For the formation of condensate layers, a low permeability domain at a depth of the steam-water boundary in vapor-dominated systems is necessary (Schubert et al. 1980).

A geothermal system with a cooling pluton or even an extinct one with radioactive elements as a heat source, on the other hand, can have extensions varying from a few hundred square meters to several square kilometers, depending upon the size of the batholith and other associated intrusive bodies, viz. stocks, dikes, sills, laccoliths, and lopoliths. Such systems are highly dependent on the age and size of emplacement of the intrusive bodies and/or the presence of radioactive elements, e.g., large-scale granitic bodies supplying remnant and radioactive heat to an overlying geothermal system. Geothermal systems associated with plutonism could be related to recent plutonism and extension, as in Laderello, Italy (Minissale 1991), or active volcanism, typically at magmatic arcs along convergent margins, as in Java (Indonesia) or mid-oceanic ridge settings at divergent plate margins, as in Iceland. At Laderello, young (0.3–0.2 Ma) magmatism related to granite intrusions generates a fluid-dominated layer above the granite and a vapor-dominated layer above the former. Moreover, Pliocene extension associated with magmatic rocks' emplacement generates low-angle normal faults that control meteoric water's recharge (Bertini et al. 2006).

On the other hand, at the Geysers, a large felsic pluton is the heat source for a vapor-dominated fluid in a porous meta-sedimentary reservoir capped by a low permeability serpentinite mélange and meta-greywacke (Ingebritsen and Sorey 1988). These low permeability lithologies impede the meteoric recharge of the geothermal system. They have necessitated the injection of treated sewage water to sustain the system and keep the heat recovery at optimum level (Majer and Peterson 2007).

Non-magmatic Convection-dominated Geothermal Systems

Non-magmatic convection-dominated geothermal systems are typically fault-controlled, wherein convection occurs along the faults, combined with infiltration of meteoric water along the faults (Reed 1983; Fig. 1). Sometimes, the fluids may leak from the fault into a concealed permeable layer, and in turn, fluids can move from such permeable layer into the fault zone and from there to the surface as hot springs. This sub-type of fault-controlled geothermal systems is fault-leakage controlled (Moeck 2014). The Great Basin (United States), which is part of the northern Basin and Range Province, is perhaps the most cited example of fault-controlled geothermal systems (e.g., Faulds et al. 2010a 2010b 2010c). Other salient examples of such extensional geothermal systems are from Western Turkey (e.g., Faulds et al. 2009) and tectonically active intracontinental rift grabens, viz. East African Rift, Upper Rhine Graben in Central Europe, SONATA (Son-Narmada-Tapi; Verma 1991; Minissale et al. 2000) rift system in Central India.

An increase in bicarbonate and magnesium, coupled with a decrease in boron, sulfate, and chloride contents, typically indicates near-surface mixing of upwelling geothermal water with groundwater or meteoric water (Flynn and Ghusn 1983; Nicholson 1993). In such systems, the stress regime controls the fluid circulation along the faults. Consequently, stress modeling helps identify the faults that favor the circulation of the geothermal systems in a complex fault system (Faulds et al. 2010b, Moeck et al. 2010), e.g., Bruhn et al. (2010) and Jolie et al. (2010) found the dilational or shear dilation faults the most favorable. Moreover, proper reinjection and maintenance of reservoir pressures are crucial to the management of geothermal field with the presence of fossil geothermal fluids, e.g., at Great Basin (Faulds et al. 2010b), for which a comprehensive study of the fault systems is vital. The reinjection well-sites for a fault-controlled geothermal system must avoid thermal breakthrough of the injected cooled water along permeable faults to the production wells. To ensure this, injection and production wells should not be along the same or interconnected fault(s) in the same fault block.

4.1.2 Conduction-dominated Geothermal Systems

Geothermal systems located at passive plate tectonic settings (e.g., passive continental margins and intracontinental tectonically inactive areas) are mostly conduction-dominated in the absence of asthenospheric anomalies. The conductive settings of sedimentary basins exemplify them. In such systems, a near-normal heat flow heats the deep reservoirs. In the basement or crystalline igneous rocks, heat originates from granites, leading to a significant positive thermal anomaly, e.g., at the EGS reservoir in granitic rock at Soultz in France (Genter et al. 2000). HDR type EGS resources lack producible natural thermal fluids and require fluid injection through artificial fracture networks. Due to the new developments in EGS technologies, conduction-dominated geothermal systems have become quite significant. Thus naturally non-existent essential conditions associated with a geothermal system can be generated (as in HDR systems) by creating fractured reservoirs in crystalline rocks. An existing system can also be improved through reservoir and fracture network enhancement in tight, i.e., low permeability rocks (Moeck 2014). Permeability anisotropy, predominantly controlled by faults and/or lithology, characterizes such systems. They can be classified into three types: (i) the basement/crystalline rock type, (ii) the intracratonic basin, and (iii) the orogenic belt type (Moeck 2014).

Basement/crystalline Rock Type Systems

Although crystalline rocks (e.g., granites) in igneous provinces are potential heat sources themselves, these low intrinsic porosity and permeability rocks require reservoir development, i.e., enhancing permeability through stimulation techniques. This measure facilitates circulation between the injector and producer wells in HDR systems, wherein the rock mass acts as the heat exchanger (Moeck 2014). The main challenge of EGS development in crystalline rocks is creating an augmented permeability structure between the two wells, which is overcome by a thorough understanding of the stress field, e.g., the magnitude of the intermediate principal stress, as it controls the in situ stress regime. Further considerations include geomechanical parameters and failure models of the reservoir rocks under stimulation conditions (discussed in Sect. 5).

Intracratonic Basins and Orogenic Belts Type Systems

The conduction-dominated geothermal systems may occur in different geologic settings where there is no active igneous activity. In such cases, the tectonic activity within the geothermal system is low, feeble or absent. Such systems could be (i) within intracratonic basins, (ii) orogenic belts, and associated foreland basins. In the first setting, a near-normal heat flow heats deep aquifers in the sedimentary basins at great depth (>3 km, e.g., Québec, Eastern Canada, Majorowicz and Minea 2015). In the second setting, advective heat transport plays a key role. High permeability domains and deep-rooted faults allow deep circulation of meteoric water, often associated with the subsequent formation of hot springs (e.g., Manikaran in the Himalayas, Alam et al. 2004; Chandrasekharam et al. 2005). Two basin types hosting geothermal systems (Moeck 2014) are (i) extensional or lithospheric subsidence basins, viz. Central European Basin (e.g., Scheck-Wenderoth et al. 2014), and (ii) foreland basins within orogenic belts, viz. the Molasse Basin of the Alps (e.g., Chelle-Michou et al. 2017), Western Canada Sedimentary Basin associated with the Rocky Mountains (Higley et al. 2005).

4.2 Major Tectonic Settings for Geothermal Systems

The tectonic setting of geothermal systems firmly controls their thermal regime and chemistry. Primary tectonic settings and associated geothermal systems are described here.

4.2.1 Divergent Boundaries

Midoceanic Ridges

Such boundaries have exceptionally high volcanism rates along the ridge or plate boundary because seafloor-spreading concurs with a geologic hot spot or mantle plume in this case. The examples include Iceland, which lies across the Mid-Atlantic Ridge (MAR), a divergent boundary marking the eastern edge of the North American tectonic plate and the western edge of the Eurasian plate (Arnórsson 1995a, b, Arnórsson et al. 2008; Boden 2016).

Continental Rifts

Continental rifting marks significant crustal extension, which in due course of geologic time gives rise to a new ocean basin, e.g., the formation of the Red Sea about 25 million years since the rifting started about 30 million years ago (Boden 2016). Two main types of continental rifts can be associated with the geothermal systems, with (e.g., Olkaria, Menengai, Longonot, and Eburru in the East African Rift) or without (e.g., Dixie Valley in the Basin and Range Province of the USA, Tattapani, Bakreswar, Tantloi, Surajkund, Rajgir, Munger in SONATA geothermal province in Central India) magmatic heat source.

4.2.2 Convergent Boundaries—Continental and Island Volcanic Arcs

The subduction of the oceanic lithosphere leads to partial melting of the overlying mantle rocks. The volatiles released from the former aid in lowering the melting point of the latter. The generated magma rises to the upper crust and produces volcanoes. The subduction of the oceanic lithosphere beneath the continental lithosphere gives rise to continental volcanic arcs (e.g., Andes, Ramos 2009). Relatively older and colder oceanic lithosphere subducting beneath younger and warmer oceanic lithosphere gives rise to the island arcs with volcanic island chains (e.g., Japan, Philippines, Marianas, Condie 2011, 2016). Shallow magma chambers in continental and island volcanic arcs serve as heat sources for the overlying geothermal systems; e.g., producing geothermal fields at Los Azufres and Los Humeros in Mexico (Martinez 2013; Elders et al. 2014), Miravalles in Costa Rica (Ruiz 2013), San Jacinto Tizate in Nicaragua (Chin et al. 2013), and Berlin and Ahuachapan in El Salvador (Herrera et al. 2010). The strain in volcanic arcs may vary from compressional to extensional in a direct (head-on) convergence; e.g., subduction zone rollback at Taupo Volcanic Zone (Seebeck et al. 2014; Villamor et al. 2017). However, it may be transtensional or transpressional with oblique convergence; e.g., Andean Volcanic Zone (Dewey and Lamb 1992; Dewey et al. 1998; Cembrano and Lara 2009; Sielfeld et al. 2019).

Although producing geothermal systems occur in both compressional and extensional strain conditions in the volcanic arcs, Wilmarth and Stimac (2014 2015) found that those associated with the arcs having complex structural settings induced by either oblique convergence, involving transtension in particular (e.g., Salak, Indonesia; Aprilina et al. 2015) or intra-arc rift-related extension (e.g., Wairakei, New Zealand; Villamor et al. 2017) are more favorable from a geothermal development point of view (Boden 2016). Where convergence is oblique, strike-slip faults can form in the overlying plate; e.g., Liquiñe Ofqui Fault Zone in south-central Chile (Alam et al. 2010). In areas of fault step-overs, zones of transtension can occur, forming possible pull-apart basins that can foster crustal dilation (improved permeability for convection of geothermal fluids) and the rise of magma into the upper crust forming heat source (Boden 2016). Such high-level intrusions of magma can also thermally weaken overlying rocks, leading to gravitational collapse and dilation, generally orthogonal to the direction of plate convergence (Downs et al. 2014; Holden et al. 2015; Boden 2016). As a result, a series of extensional basins or grabens can form that, in association with heat from volcanism, can help create highly productive geothermal systems, viz. Los Humeros and Los Azufres in the Trans-Mexican Volcanic Belt, Miravalles in Costa Rica (e.g., González-Partida 2005; Bernard-Romero et al. 2010; Bernard et al. 2011).

The producing geothermal field at Miravalles (DiPippo 2012), an excellent example of a geothermal system developed in a local zone of extension (transtension) in a continental volcanic arc, is located within an N-NW trending graben on the southwest flank of the Quaternary Miravalles stratovolcano (Chavarría-Rojas 2003; Chavarría-Rojas 2003; Boden 2016). This graben may be related to possible left steps in W-NW striking left-lateral faults (transtension) arising from the oblique (left-lateral) convergence between the downgoing Cocos plate and overriding Caribbean plate (e.g., DeMets 2001; Symithe et al. 2015; Rosas et al. 2016; Boden 2016). The graben's bounding and internal faults have produced secondary fracture permeability of the volcanic rocks, facilitating convection of geothermal fluids and developing this field as a significant power producer (Boden 2016).

A highly productive Hatchobaru–Otake geothermal field (e.g., Taguchi and Nakamura 1991; Taguchi 2001), located on the flanks of an active Mt. Kuju volcano, lies within the arc-parallel, E-NE trending Beppu–Shimbara graben (Ehara 1989). This graben transects the island, and its N-NW-directed extension reflects slab rollback of the subducting Philippine oceanic plate (Ehara 1989). Moreover, both northeast-striking graben-parallel faults and northwest-striking normal faults control the flow of geothermal fluids at Hatchobaru (Momita et al. 2000). The northwest-striking normal faults may be a consequence of strike-slip motion on the northeast-striking graben-parallel faults, resulting in local NE-directed extension in areas where northeast faults stepover (Momita et al. 2000).

4.2.3 Convergent Boundaries—Back-Arc or Intra-Arc Extension

In this setting, the extension is oriented perpendicular to the arc, resulting in elongated grabens that run parallel for a significant part of the arc (Feuillet et al. 2002; Sdrolias and Müller 2006; Arai et al. 2018; Magni 2019). Such dilation is more extensive than the localized extension related to transtension in the case of oblique convergence, as discussed in the previous section. The development of back-arc or intra-arc spreading is more common in island arcs than continental arcs (Sdrolias and Müller 2006). Back-arc or intra-arc spreading (Feuillet et al. 2002; Sdrolias and Müller 2006; Condie 2011, 2016; Boutelier and Cruden 2013; Nakakuki and Mura 2013; Arai et al. 2018; Magni 2019) are significant because the extensional forces promote secondary rock permeability and crustal dilation (Boden 2016). This further aids intrusion of magma to high crustal levels that can serve as a heat source and development of overlying convecting and potentially exploitable geothermal systems (Boden 2016); viz. rift-related systems at Wairakei and Rotokawa in the Taupo Volcanic Zone, New Zealand (Seebeck et al. 2014; Villamor et al. 2017).

4.2.4 Convergent Boundaries—Continental Convergence

Due to the continental convergence, viz. at the Himalayas (Kious and Tilling 1996), the crust is over-thickened, viz. Tibetan plateau, which lowers the geothermal gradient (Vanderhaeghe et al. 2003; Beaumont et al. 2006; Shi et al. 2017). Moreover, the resultant compressional stress decreases the fractures and faults' dilation, reducing permeability (Rogers 2003). Furthermore, in the absence of subduction, as the continental plates merely colloid, there is no magma generation as in the case of convergent continental and island arcs, so a magmatic heat source is absent (Fucheng et al. 2018). However, high temperature and pressure metamorphism due to continental collision can cause partial melting of rocks, resulting in magma forming a heat source to drive a geothermal system (Bea 2012).

Moreover, buried radiogenic granitic and high-grade metamorphic rocks (Bea 2012) can also be the source of heating up of deeply circulating meteoric water. Geothermal systems in such a setting could be associated with localized extension zones within an otherwise compressional stress regime due to fault-valve action (Sibson 2020). In the absence of subduction, the crust is thickened during the continental collision, and extensional forces are orthogonal to the main compression direction (Mo et al. 2007). This occurs because the uplifted and thickened crust commences to collapse or buckle under its own weight. Depending on the rock type and strength, normal faults can develop in localized extensional zones and bind the horsts and grabens perpendicular to the central mountain range (Brun 2002). Such extensional normal faults and associated dilation are responsible for permeability in this otherwise tight zone. This helps the circulation of meteoric water circulate at great depths. This water is heated and then returns to the surface due to buoyancy through other normal faults to form hot springs and shallow geothermal reservoirs, e.g., Yangbajing and Yangyi geothermal fields, both in Tibet (Boden 2016).

4.2.5 Transform Boundaries

Transform boundaries are typically found in the ocean floor associated with mid-ocean ridge spreading zones. However, some well-known ones on the continent, e.g., San Andreas Fault extending from northwestern Baja Mexico to northwestern, California Anatolian Fault of northern Turkey on land, the Alpine Fault on the South Island of New Zealand (Boden 2016). Geothermal systems associated with transform boundaries are generally limited. However, some critical exceptions occur when extension with or without magmatism occurs together with the transform motion (Boden 2016).

For example, there is a zone of discontinuous north to northwest-striking right-lateral faults constituting the Eastern California Shear Zone and Walker Lane of western Nevada within the San Andreas Fault (Faulds et al. 2005a, b). Thus, the southern part of the San Andreas Fault (Fig. 4) has two favorable conditions for hosting geothermal systems, evident by geothermal fields of Cerro Prieto (Mexico), the Salton Sea (USA), and the Imperial Valley (USA). The northeast-striking normal faults, developed due to the local transtensional regime, provide conduits for geothermal fluids

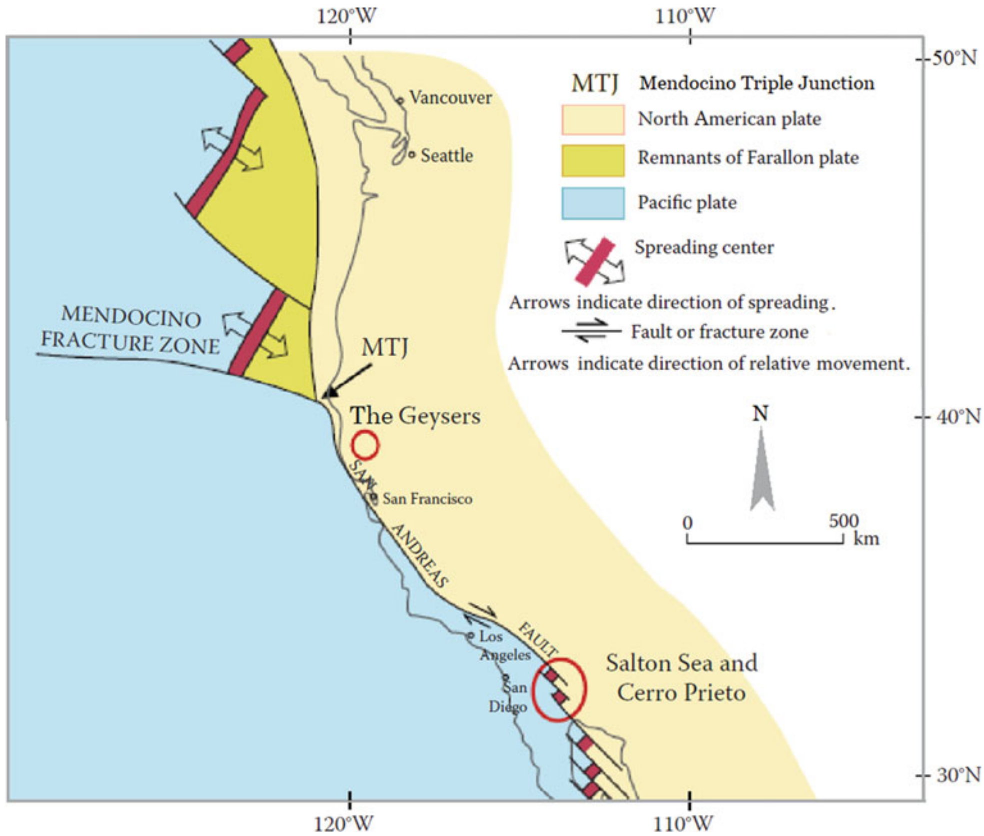


Fig. 4 Tectonic setting of the San Andreas Fault with the location of important geothermal systems located on its either end (modified from Boden 2016)

circulation (Bennett 2011). Another favorable condition is the presence of shallow magmatic heat sources (Bennett 2011). This is because of the buried spreading ridge segments and thinning of the crust due to the extension leading to the lowering of pressure, which induces the melting of the heated rocks (Bennett 2011).

Although the source of heat at The Geysers, located near the north end of the San Andreas Fault (Fig. 4), is magmatic, unrelated to spreading or extension, as is the case at the southern end of the San Andreas Fault. Instead, it is related to the transition from a previous subduction margin of the now-extinct Farallon plate to the current transform margin (Boden 2016).

4.2.6 Hot Spots

Upwelling of relatively stationary mantle plumes constitutes hot spots, where the ascent of mantle material and lowering of pressure together induce partial melting and subsequent rise of magma to form volcanic centers. Then, the tectonic plate's movement over a mantle plume gives rise to a series of volcanic centers that may lead to a series of volcanic islands, e.g., the Hawaiian Islands within the Pacific tectonic plate (e.g., White 2016). Since the Big Island of Hawaii currently lies over the plume, it hosts five active volcanoes. Kilauea is the most active one in Hawaii for being directly above the plume, close to which Puna geothermal field is located (Boden 2016).

4.2.7 Stable Cratons

Located away from plate boundaries within the continents, they are geologically stable, as evident from the absence or rare occurrence of seismic events or volcanism. They include: (i) deep sedimentary basins often with producing oil and gas fields (e.g., Teapot Dome oil field in Wyoming, Curry 1977; Williston Basin in North Dakota, Drake II et al. 2017), (ii) deep (3 to 5 km), old but still hot, granites due to radiogenic decay of U and Th (e.g., Cooper Basin in Australia, Holl 2015; Peninsular India, Singh et al. 2014). Although there is a general lack of interconnected fracture systems in such a setting, they are potential areas for developing HDR systems due to enough heat (usually > 200 °C).

5 Structural and Other Local Factors Controlling Geothermal Systems

Apart from the tectonic factors described in previous sections, local factors also play an essential role in defining geothermal systems. Sometimes, despite being located in a favorable tectonic setting, the local structural regime may make a geothermal system unproductive or uneconomical for development. Some of these factors are discussed here.

5.1 Pressure Difference between Hot and Cold Hydrostatic Head

Geothermal fluid flow can be described using Darcy's law, which quantifies the effect of a pressure differential on fluid flow through porous media. For example, the pressure difference between the hot and cold hydrostatic head of ∼ 10 MPa is the primary control on large-scale fluid flow in the Taupo Volcanic Zone (TVZ, Rowland and Sibson 2004; Grant and Bixley 2011).

5.2 Permeability Structure

Permeability, which may vary over several orders of magnitude depending on the rock type, is the most critical variable in dictating the extent of fluid flow. The minimum bulk permeability required for convection is 10^{-16} m^2 (Elder 1981; Henley and Ellis 1983; Cathles et al. 1997); however, the actual permeability structure is far more complex than estimated bulk permeability (Rowland and Simmons 2012). Moreover, the total rock mass directly involved with the transmission of fluid is minimal compared to the bulk volume of rock which hosts the flow network, wherein high-flux conduits occur under exceptional situations (Donaldson and Grant 1981; Elder 1981; Donaldson 1982; Grant et al. 1982; Hanano 2004).

5.3 Crustal Heterogeneity and Anisotropy

The heterogeneous assemblage of lithologic units results in significant lateral and vertical variations in permeability. For example, the stratified Quaternary volcanic sequence approximating a layered medium induces a strong contrast between layer parallel and layer perpendicular permeability at TVZ (Manning and Ingebritsen 1999; Rowland and Simmons 2012). The latter is controlled by the low-permeability layers (Manning and Ingebritsen 1999; Rowland and Simmons 2012). The vertical to horizontal permeability ratio is estimated as 1:40 for the volcanic stratigraphy at 3 km depth based on numerical reservoir modeling (Mannington et al. 2004; Rowland and Simmons 2012). Despite the abundance of granular layers conducive to diffuse flow within this Quaternary volcanic sequence at TVZ, macroscopic faults and fractures must comprise an essential component of crustal-scale permeability for two reasons: (i) intergranular porosity, and thus permeability, decreases with depth in granular materials (pyroclastic and sedimentary rocks) as a consequence of diagenetic processes (Björlykke 1997; Stimac et al. 2004, 2008; Rowland and Simmons 2012), and (ii) metasedimentary rocks, andesitic lavas and/or welded or silicified ignimbrites, if present within the convective regime (e.g., TVZ, Stern and Davey 1987; Broadlands-Ohaaki (Rowland and Simmons 2012), have insufficient permeability to sustain geothermal production, except where drilling has intercepted hydraulically conductive faults and fractures (Wood et al. 2001; Rowland and Simmons 2012). Thus, convective flow through any of these rock types requires fault and/or fracture-controlled permeability (Rowland and Sibson 2004; Rowland and Simmons 2012).

Moreover, the rock mass near the magmatic heat source, where the heat transfer is mostly through conduction, is considered impermeable for advective fluid flow unless permeability is developed through fracturing, microfracturing, or cavitation (Cox et al. 2001; Cox 2005; Micklethwaite et al. 2010; Rowland and Simmons 2012). In addition to their association with large earthquakes, shear zones and creeping faults are likely to be an essential means of channeling liquids from deep sources to the seismogenic zone base (Cox et al. 2001). Moreover, seismic events may cause episodic rupturing of the brittle-ductile transition processes near the magmatic source at the bottom of high-temperature convection cells, allowing magmatic fluids' entrainment into the meteoric convection regime (Rowland and Simmons 2012).

5.4 Hydrothermal Alteration and Mineral Deposition

Permeability, porosity, and rock strength that control mineral dissolution, transformation, and precipitation are continually modified as a function of time and space (e.g., Browne and Ellis 1970; Hedenquist and Browne 1989; Simmons and Browne 2000; Rowland and Simmons 2012). In a particularly evolved (also referred as long-lived, > 50Ky) geothermal systems, three types of alteration effects have been be recognized: (i) clay-rich alteration that forms in shallow steam-heated aquifers and on the periphery of the upflow zone reduces permeability as well as rock strength by increasing the proportion of clay minerals that replace volcanic glass and feldspars (e.g., Hedenquist and Browne 1989; Hedenquist 1990; Simmons and Browne 2000; Rowland and Simmons 2012), (ii) silicification and K-metasomatism due to deposition of quartz and adularia from rising and cooling chloride waters reduce porosity and permeability but increase rock strength (e.g., Ohakuri; Henneberger and Browne 1988), which may in turn enhance the development of fault-fracture–related permeability (e.g., Broadlands-Ohaaki, Simmons and Browne 2000; Rowland and Simmons 2012), and (iii) mineral deposition can line and therefore isolate high-permeability pathways from incursion of fluid from the surrounding country rock (Rowland and Simmons 2012). Thus, in the upflow zone, fracture permeability becomes increasingly important with the evolution of a geothermal system. Intergranular permeability is gradually decreased by pervasive silicification and mineral deposition, reflected in pore fluid pressure fluctuations in producing geothermal fields (Rowland and Simmons 2012).

5.5 Brittle Deformation and Conditions for the Development of High-Flux Fluid Conduits

The fluid flow in a geothermal system depends on fractures' distribution—fracture network—within the upper crust. However, characterizing fracture networks' three-dimensional geometry and defining a quantifiable connection between geometrical and hydraulic connectivity to understand fractured media flow is quite complex (Berkowitz 2002; Rowland and Simmons 2012). To deal with this complexity, Rowland and Simmons (2012) suggested a qualitative approach for understanding the role of various factors essential for developing high-flux conduits. This approach considers the macroscopic mode of brittle failure, lithology, stratigraphy, the seismic cycle, and the degree of sealing through hydrothermal cementation (Sibson 2000). Brittle structures rarely form in isolation, and their cumulative hydrologic effect additionally exerts an essential control on permeability heterogeneity and anisotropy (Rowland and Sibson 2004). Three macroscopic modes of brittle failure (Fig. 5, Table 1; Sibson 1998, 2004, 2020) are possible: shear failure (faulting, i.e., displacement parallel to the fracture surface), extensional failure (generation of dilational fractures perpendicular to the least principal stress, σ3), and hybrid extensional-shear failure (involving components of shear and dilation). The mode of brittle failure depends upon pore fluid pressure, P_f, differential stress defined by the difference between the greatest and least principal stresses ($\sigma_1 - \sigma_3$), and tensile strength, T, of the deforming rock volume, modulated by hydrothermal alteration and mineral deposition (Rowland and Simmons 2012). In tectonically active regimes (Fig. 6) that promote fluid flow, P_f, ($\sigma_1 - \sigma_3$) and T vary temporally and spatially. The effect on a failure mode is illustrated in the pore fluid factor (Fig. 7) and differential stress space (Fig. 8), where the pore fluid factor, λ_v, is the ratio between fluid pressure and overburden σ_v (Cox 2010).

5.6 Permeability in Fault Zones

Normal faulting with a minor strike-slip component is a favored brittle failure mode that exerts the principal structural influence on fluid redistribution, as in TVZ, where faulting and subsidence of the graywacke basement played a prominent role in controlling the structural development in the cover sequence (Rowland and Sibson 2001; Acocella et al. 2003; Rowland and Simmons 2012).

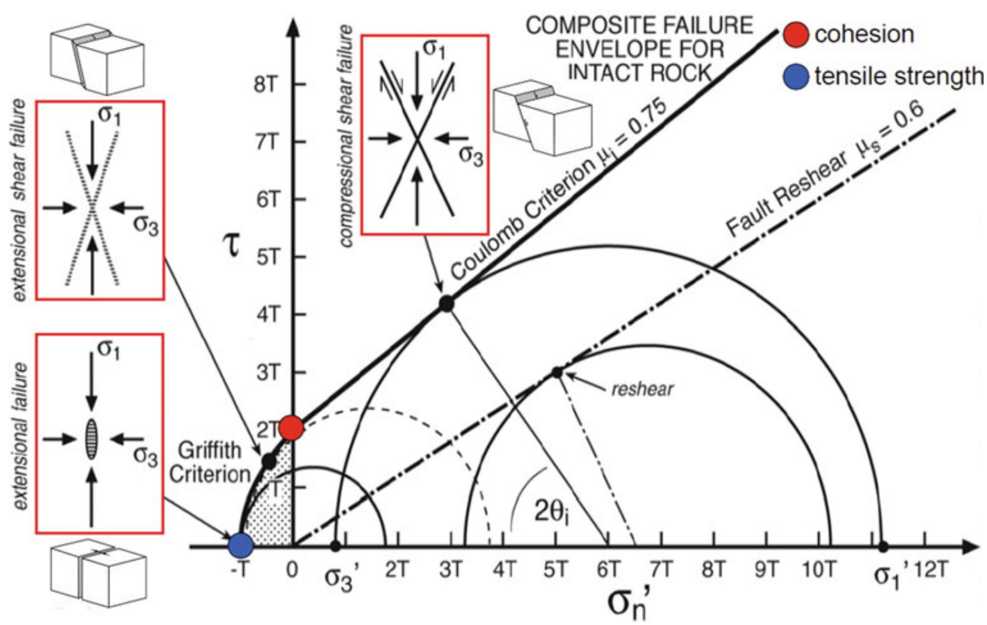


Fig. 5 Three macroscopic modes of brittle failure (modified from Sibson 2004)

Table 1 Relevant failure criteria, assumptions, and limitations in pore fluid factor-differential stress space (Sibson 1998; Cox 2010; Rowland and Simmons 2012; Ferrill et al. 2020; also see Figs. 4, 5, 6, 7)

Macroscopic mode of brittle failure	Failure criteria	Assumptions	Limitations
Brittle shear failure on optimally oriented faults in intact rock	Coulomb criterion $\tau = C + \mu(\sigma_n - P_f)$, where τ = shear stress, C = cohesive strength, μ = coefficient of friction in an isotropic rock mass, σ_n = normal stress, P_f = pore fluid pressure; for a fault inclined θ_{opt} to σ_1 $\lambda_v = [4C - \sigma_1 + 4\sigma_3]/3\sigma_v$, where σ_v = overburden, pore fluid factor $\lambda_v = P_f/\sigma_v$ and $\theta_{opt} = ½ \tan^{-1}\mu^{-1}$	Plane strain (σ_2 lies in the fault plane) $\sigma_v = \sigma_1$ assuming an Andersonian extensional stress regime $\mu = 0.75$ $C \sim 2T$	$(\sigma_1 - \sigma_3) \geq 5.66\,T$
Extension failure	$\lambda_v = (\sigma_3 + T)/\sigma_v$		$(\sigma_1 - \sigma_3) < 4\,T$, where T = tensile strength $P_f = \sigma_3 + T$
Hybrid extensional-shear failure	Griffith criterion $(\sigma_1 - \sigma_3)2 = 8\,T(\sigma'_1 + \sigma'_3)$, where $\sigma'_1 = (\sigma_1 - P_f)$ and $\sigma'_3 = (\sigma_3 - P_f)$ $\lambda_v = [8\,T(\sigma_1 + \sigma_3) - (\sigma_1 - \sigma_3)^2]/16T\sigma_v$		$4\,T < (\sigma_1 - \sigma_3) \leq 5.66\,T$

Fig. 6 Tectonically driven pathways to failure

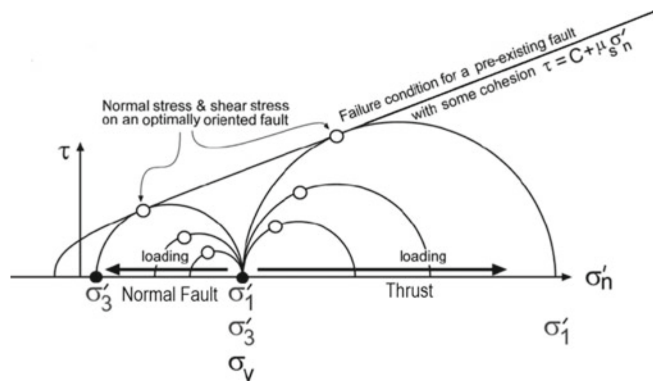
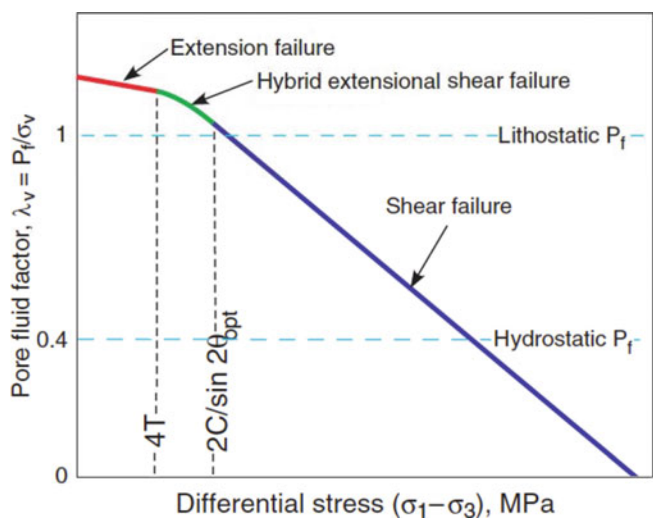


Fig. 7 Fluid driven pathways to failure



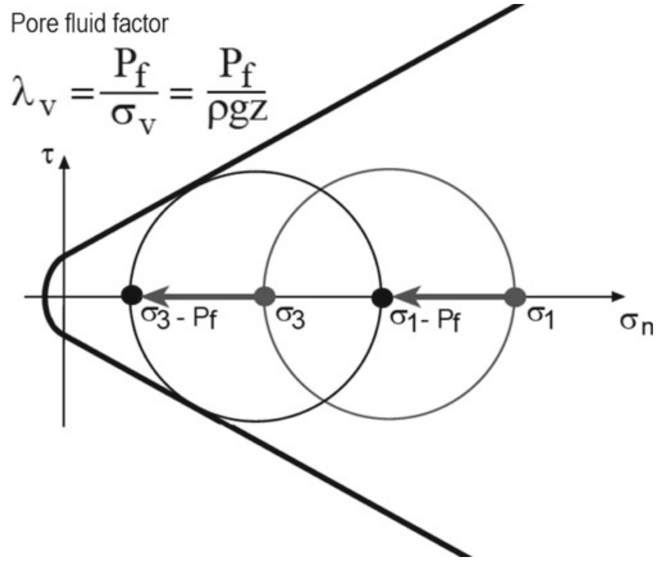


Fig. 8 Generic failure curve at some depth in the crust (modified from Cox 2010)

5.7 Fault Zone Complexity, Rift Architecture, and Directional Permeability

The permeability distribution within fault zones is affected by fault growth and interaction (Curewitz and Karson 1997) and structural overprinting (e.g., Berger et al. 2003). Geothermal fluid flows through interconnected fault-fracture networks within spatially more extensive fault and/or fracture systems (Rowland and Simmons 2012). Fault irregularities in the direction of slip can cause the development of highly permeable zones due to dilation within extensional stepovers and jogs (Sibson 2001). In extensional settings, jogs direct the fluid flow along strike (parallel to σ_2), while in strike-slip settings, jogs form subvertical pipes through which fluid flows (Sibson 2000). Moreover, such piped zones of enhanced permeability can also be developed due to normal fault growth and linkage, leading to dilated subvertical zones in some extensional settings (e.g., Nortje et al. 2006). Thus fault growth and linkage generate subvertical zones of enhanced permeability between normal fault segments, transfer fault intersections with rift faults, and lateral fault tips (Curewitz and Karson 1997; Micklethwaite 2009; Rowland and Simmons 2012; Fig. 9).

Additionally, rift architecture may lead to directional permeability. Moreover, extension in an area may cause the formation of segmented blocks with subparallel arrays of normal faults, which might host dyke swarms in the presence of magma (e.g., Dabbahu Rift in Afar, Rowland et al. 2007). In general, segmentation scales with the thickness of the mechanical layer that is breaking (Ebinger et al. 1999). The displacement between resulting segments must be accommodated, i.e., it should be coupled or linked with another movement. Accordingly, these segments are either "hard-linked" (Gibbs 1984) with transfer faults oriented at a high angle to the axis of rifting or "soft-linkage" (Rosendahl et al. 1986; Morley et al. 1990) with distributed deformation and small-scale faulting within the blocks between adjacent segments. The bulk permeability is a function of the cumulative effect of subparallel faults and fractures in the former and structurally favorable sites for enhanced vertical permeability in the latter (Rowland and Simmons 2012). Its value in rift segments contrasts with that in accommodation zones (Rowland and Sibson 2004). An array of faults and fractures has the same effect on bulk permeability as stratigraphic layering: permeability across the strike of the array is lower relative to other directions, regardless of whether faults behave as conduits or baffles to flow (Rowland and Simmons 2012). When superimposed upon a layered sequence, the combined effect reduces vertical and across-strike permeability relative to along-strike permeability (Fig. 9). In contrast, all favorable structural sites for focused vertical flow occur within accommodation zones (e.g., rift fault-transfer fault intersections, lateral fault tips on first-order structures, and linkage zones between first-order structures, Fig. 9). Thus, rift architecture may modulate fluid-flow paths such that upflow zones are favored in accommodation zones, and recharge and axial flow are selected in rift segments (Rowland and Sibson 2004).

The tendency of geothermal fields to occur around the basin's margin (e.g., Broadlands-Ohaaki, Rotokawa, Waiotapu, Te Kopia, Orakeikorako) in accommodation zones may produce basement highs (Wairakei-Tauhara, Ngatamariki), which suggests deep-seated control by inherited basement faults. These structures are of particular importance, because they are likely rooted in potentially permeable shear zones within the ductile lower crust, thus channeling geothermal fluids across the brittle-ductile transition zone (Cox et al. 2001).

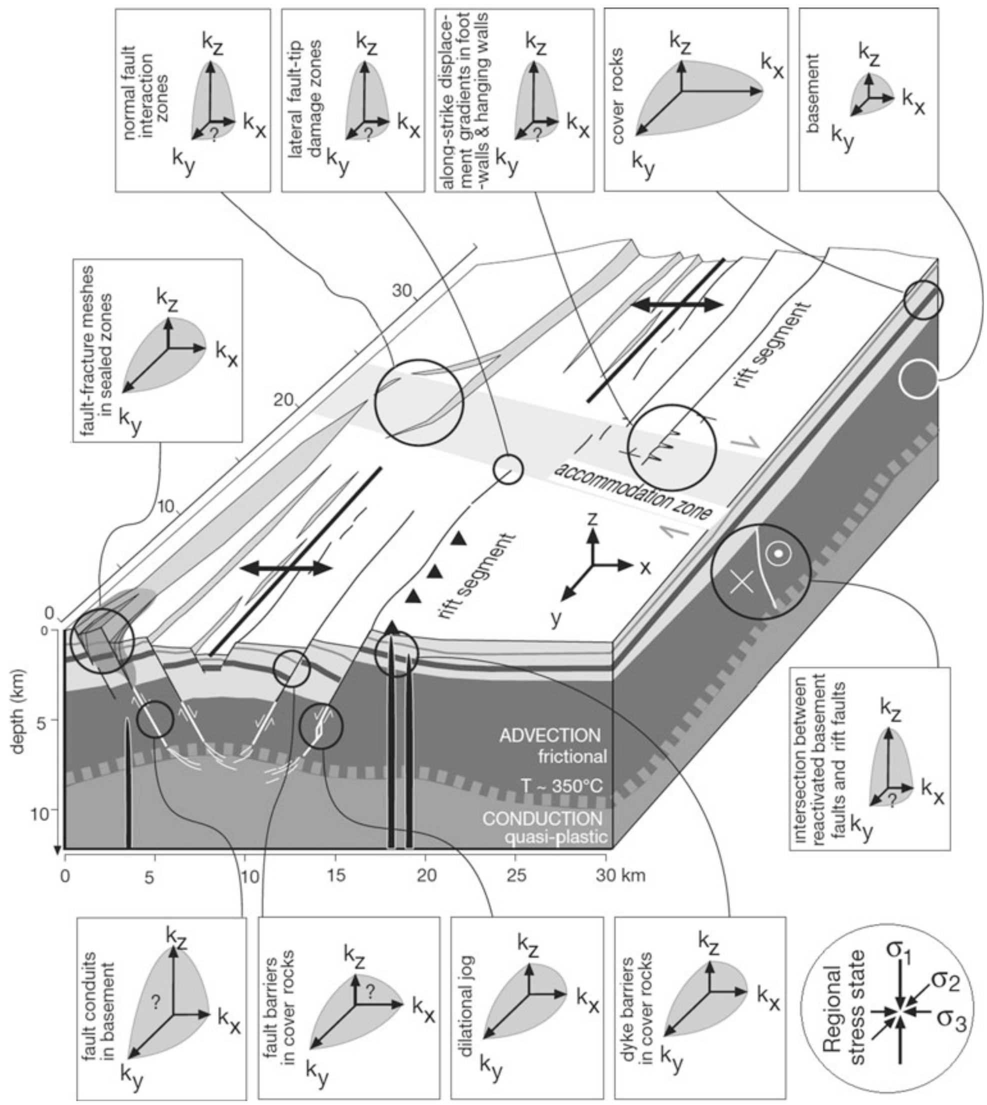


Fig. 9 Effect of structures on permeability anisotropy in the central Taupo Volcanic Zone, assuming a simple layered cover sequence overlying a competent basement with low intrinsic permeability (modified from Rowland and Sibson 2004, Rowland and Simmons 2012) [Notes: (i) Three-dimensional permeability diagrams depict indicative relative magnitudes of mutually perpendicular across strike (x), along strike (y), and vertical (z) permeability for different structural settings within the rift system. (ii) Question mark between arrows indicates uncertainty in the relative magnitude of adjacent directional permeability values. (iii) Settings with enhanced and localized vertical permeability are favored in accommodation zones.]

Thus, the main factors constraining fluid flow are (i) magmatic intrusion, which mainly supplies heat and energy, but which can also facilitate fracture extension at the tip of a dike; (ii) proximity to the brittle-ductile transition, which limits the downward flow of water; (iii) the tensile strength of host rocks which is generally high; and (iv) hydrostatic fluid pressure, which is maintained because brittle failure relieves the build-up of fluid overpressures.

Lastly, it is essential to distinguish between paleostress and recent stress since an area's overall structuration is primarily controlled by paleostresses. The overprint of local and regional current stresses can modify the permeability of the existing fracture system. For example, fractures aligned along the SHmax are more likely to keep themselves “open,” and the ones perpendicular to it would most likely be “closed.” This underscores the importance of dealing with tectonic and structural controls together for the geothermal systems, not just tectonic or structural controls alone or separately.

6 Conclusions

Tectonic and structural settings control the inherent variability in the nature and evolution of a geothermal system, both of them being dynamic. A combination of these in time and space being unique makes each system unique. Apart from controlling hydrological conditions, viz. permeability structure governs the fluid—both recharging cold fluid and hot geothermal fluid—flow, pressure head difference between cold and hot water flow that makes a system sustainable establish whether the thermal regime is conductive or convective. Despite being complex and difficult to predict, the fluid-flow paths in geothermal systems are quite self-regulating, as evidenced by long-lived (>50,000 years) systems worldwide. This is particularly true for the systems under natural conditions with recharge and discharges in harmony and the natural stimulation of fluid conduits through seismic events, creating new fractures and reopening fractures clogged by hydrothermal deposits.

In the case of geothermal fields under production, besides optimizing the geothermal discharge from wells, proper care should be taken for ensuring the continuance of the permeability structure through geophysical studies, viz. microseismic, to make the geothermal systems sustainable.

The conditions for the development of permeability structure that facilitates geothermal fluid flow and circulation, viz. dilation of the conduits in the fault zones, accommodation zones that transfer extension between rift segments, can be summarized as follows. Geothermal fluid flow is primarily controlled by (i) heat source—magmatic or non-magmatic—that drives convective circulation, (ii) intergranular porosity and permeability, (iii) permeability due to fault-fracture network produced by tectonism or magmatism (volcanism and/or plutonism), (iv) conduits of volcanic and hydrothermal eruptions, and (v) hydrothermal alteration and mineral deposition causing porosity and permeability heterogeneity in the geothermal reservoir and conduits.

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