

Causes of underpressure in natural CO₂ reservoirs and implications for geological storage

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ABSTRACT

Geological carbon storage has the potential to reduce anthropogenic carbon dioxide emissions, if large volumes can be injected and securely retained. Storage capacity is limited by regional pressure buildup in the subsurface. However, natural CO₂ reservoirs in the United States are commonly underpressured, suggesting that natural processes reduce the pressure buildup over time and increase storage security. To identify these processes, we studied Bravo Dome natural CO₂ reservoir (New Mexico, USA), where the gas pressure is up to 6.4 MPa below the hydrostatic pressure, i.e., less than 30% of the expected pressure. Here, we show that the dissolution of CO₂ into the brine reduces the pressure by 1.02 ± 0.08 MPa, because Bravo Dome is isolated from the ambient hydrologic system. This challenges the assumption that the successful long-term storage of CO₂ is limited to open geological formations. We also show that the formation containing the reservoir was already 2.85 ± 2.02 MPa underpressured before CO₂ emplacement. This is likely due to the overlying evaporite layer, which prevents recharge. Similar underpressured formations below regional evaporites are widespread in the midcontinent of the United States. This suggests the existence of significant storage capacities with properties similar to Bravo Dome, which has contained large volumes of CO₂ over millennial time scales.

INTRODUCTION

Pilot projects have demonstrated the feasibility of geological carbon storage (GCS; Michael et al., 2010), and saline aquifers in the United States provide enough storage to stabilize CO₂ emissions at current levels for a century (Szulczewski et al., 2012). In addition, natural CO₂ reservoirs have stored large quantities of CO₂ on millennial time scales (Gilfillan et al., 2009; Sathaye et al., 2014). This suggests that GCS can make a significant contribution to CO₂ emissions reduction. However, concerns remain that the large-scale implementation of GCS can lead to pore-pressure buildup and induce seismicity, which could compromise storage security (Zoback and Gorelick, 2012). Although there is no evidence for this in pilot projects (Juanes et al., 2012), there is considerable concern due to the dramatic increase in seismicity associated with subsurface wastewater injection (Ellsworth, 2013). In addition, regional pressure buildup may also lead to the migration of formation brines into the potable aquifers near the storage site (Birkholzer et al., 2011; Chang et al., 2013).

Therefore, it is interesting to note that many natural CO₂ reservoirs in the United States are underpressured; i.e., they have gas pressures significantly below hydrostatic levels (Fig. 1). Although a recent global compilation of CO₂ reservoirs (Miocic et al., 2016) showed a wide range of pressures, it is important for long-term storage security to understand those processes that reduce the pore pressure in U.S. CO₂ reservoirs. Even after the pressure buildup due to CO₂ emplacement has dissipated, the gas pressures are expected to remain elevated relative to the brine due to capillary entry pressure (Lake, 1989). This suggests that there are natural processes that reduce CO₂ pressure over time. Here, we aim to identify these underlying mechanisms.

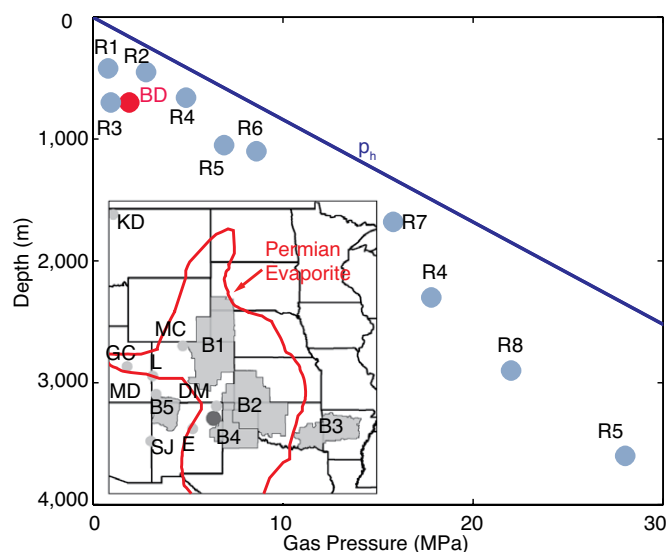


Figure 1. Underpressure in natural CO₂ reservoirs showing initial bottom hole pressures (IBHPs) from natural CO₂ reservoirs in Rocky Mountain and Colorado provinces (USA; Eppink et al., 2014). Insert shows locations of CO₂ reservoirs (R) relative to regional sedimentary basins (B) containing underpressured aquifers: B1—Denver, B2—Anadarko, B3—Arkoma, B4—Palo Duro, B5—San Juan, R1—St John's (SJ), R2—Estancia (E), R3—Des Moines (DM), R4—McElmo Dome (MD), R5—Gordon Creek (GC), R6—Kevin Dome (KD), R7—McCallum (MC), R8—Lisbon.

UNDERPRESSURE IN BRAVO DOME NATURAL CO₂ RESERVOIR

Here, we studied the Bravo Dome gas field, New Mexico, to understand the processes that contribute to underpressure in natural CO₂ reservoirs (see the GSA Data Repository¹ for Bravo Dome data). This formation is severely underpressured, and large amounts of data and previous work provide constraints on its pressure evolution (Fig. 2A). In particular, Bravo Dome offers detailed information about the distributions of the preproduction pressure, the pore space within the reservoir, and the magnitude of CO₂ dissolution.

Bravo Dome extends over an area of 3600 km² (Fig. 2B) and contains ~1.5 Gt of essentially pure volcanic CO₂ that was emplaced ca. 1.2–1.5 Ma (Johnson, 1983; Broadhead, 1990, 1993; Pearce et al., 1996; Gilfillan et al., 2009; Sathaye et al., 2014). The field is located on the Sierra Grande uplift and dips toward the Palo Duro, Tucumcari, and Anadarko Basins. The reservoir is 580–900 m deep and has formed in the Permian Tubb Sandstone, which overlies the Precambrian basement and is sealed by the overlying Cimarron Anhydrite (Fig. 2C). In the southeast, the Tubb Sandstone is mostly composed of sandstone with a few interbedded siltstones,

¹GSA Data Repository item 2017012, pre-production bottom hole pressure and bottom hole temperature data for all the measurement wells at Bravo Dome, is available online at <http://www.geosociety.org/pubs/ft2017.htm> or on request from editing@geosociety.org.

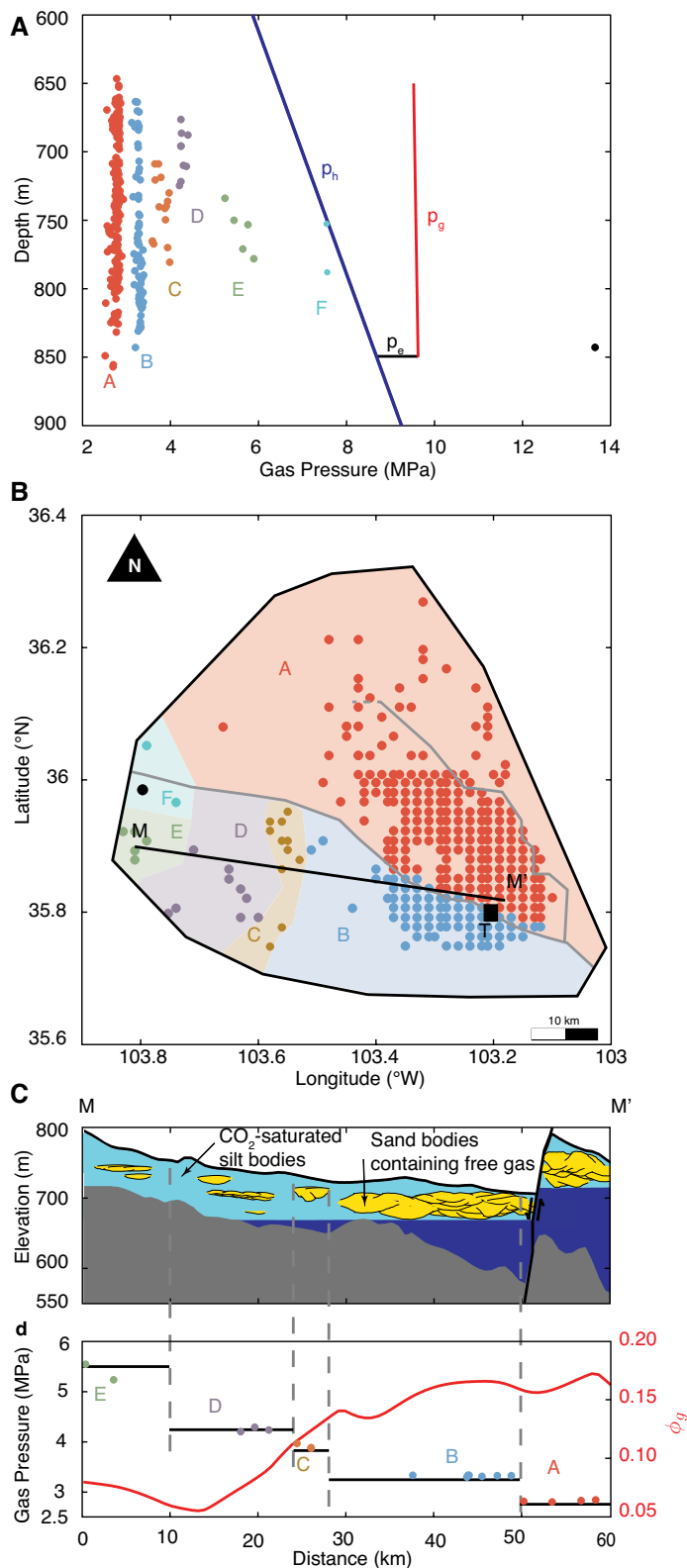


Figure 2. Pressures in Bravo Dome (New Mexico, USA) CO₂ reservoir. A: Initial bottom hole pressures (IBHPs) form multiple gas-static trends. p_h —hydrostatic pressure, p_e —capillary entry pressure, p_g —pre-production gas pressure. B: Map delineating hydraulically isolated compartments. Dots indicate well locations shown in A, gray lines indicate faults, and black square indicates location of apatite samples. C: Cross section along M-M', dashed line indicated in B. Fault offsetting the two main reservoirs is shown by black line. D: Gas pressures along cross section (shown as dots) and gas volume fraction (shown as red line).

but toward the northwest, the amount of siltstone increases and separates the individual sand bodies.

Given the mean permeability (42 mD) and the age of the reservoir, the underpressure should have dissipated due to inflow of brine from the surrounding aquifer. Significant CO₂ dissolution in the northeastern part of the reservoir indicates communication with the aquifer directly below the gas-water contact (Sathaye et al., 2014). However, the response of the reservoir to gas production beginning in 1981 suggests poor pressure communication with the far-field aquifer. The gas-water contact has remained unchanged by production, and instead the gas pressure in main part of the reservoir has dropped from 2.75 MPa to 0.4 MPa. Therefore, the reservoir is acting as a closed system with a constant gas volume on production time scales.

The preproduction gas pressures in the reservoir recorded multiple gas-static trends (Broadhead, 1993), indicating that the reservoir is divided into separate hydrologically isolated pressure compartments (Fig. 2A). Two well-defined main compartments in the east, labeled A and B in Figure 2B, contained 70% ± 14% of the CO₂ prior to commercial production in A.D. 1981. Toward the west, the compartments become smaller and less well defined, and the pressure increases (Fig. 2D). The two main compartments are clearly separated by a fault. The smaller compartments in the west could be bounded by small faults, or the sand bodies may have become disconnected and the CO₂ is entrapped by capillary entry pressure (Fig. 2C).

Both geochemical constraints and preproduction pressures indicate the reservoir filled from west to east (Gilfillan et al., 2008). Therefore, the compartments that are now isolated must have been connected during CO₂ emplacement. Currently, the highest gas pressures are close to 60% of the lithostatic stress, suggesting that hydraulic fracturing may have occurred (Zoback, 2007). These fractures must have subsequently sealed to maintain the pressure differences between the compartments. The compartmentalization of the reservoir and its response to production show that Bravo Dome has acted as a closed system for a significant part of its history.

PREVIOUSLY RECOGNIZED MECHANISMS GENERATING UNDERPRESSURE

Leakage is unlikely to contribute to underpressure at Bravo Dome because there is no evidence of CO₂ leakage to the surface, and only one small CO₂ accumulation in the overlying strata has been recognized (Broadhead, 1990; Fessenden et al., 2009). The expected gas pressure in the reservoir is 9.2 ± 0.2 MPa, based on the hydrostatic gradient and assuming brine density of 1000 kg/m³ and typical entry pressures in the Tubb Sandstone (Sathaye et al., 2014). However, the observed preproduction gas pressures in compartments A and B are only 30% and 35% of the expected value. The preproduction brine pressures in the surrounding basins indicate that formations of comparable age are on average 2.85 ± 2.02 MPa below hydrostatic pressure (Figs. 3A and 3B). This regional underpressure is likely due to the Cimarron Anhydrite, which isolates the underlying Permian rocks from topography-driven groundwater recharge (Belitz and Bredehoeft, 1988; Swarbrick and Osborne, 1998; Nelson and Gianoutsos, 2014). CO₂ was therefore emplaced into an underpressured aquifer, where the brine pressure was only 69% ± 22% of hydrostatic pressure. However, Bravo Dome is more underpressured than the surrounding regional aquifer. Therefore, additional mechanisms must have reduced the gas pressure after CO₂ emplacement.

Underpressure in sedimentary basins is commonly explained by erosional unloading (Russell, 1972; Neuzil and Pollock, 1983). Assuming a constant erosion rate, the maximum pressure drop is given by $\Delta P = \rho_b g m \Delta t$, where ρ_b is density of the eroded material, and m and Δt are the rate and duration of erosion, respectively. Since the emplacement of CO₂, the erosion rate in this area has been 3–19 m/m.y. (Nererson et al., 2013). Assuming the removed material was saturated soil with a density of 2700–3300 kg/m³, the maximum pressure drop due to erosion is 0.1–0.8 MPa, equivalent to 1%–12% of the total observed underpressure.

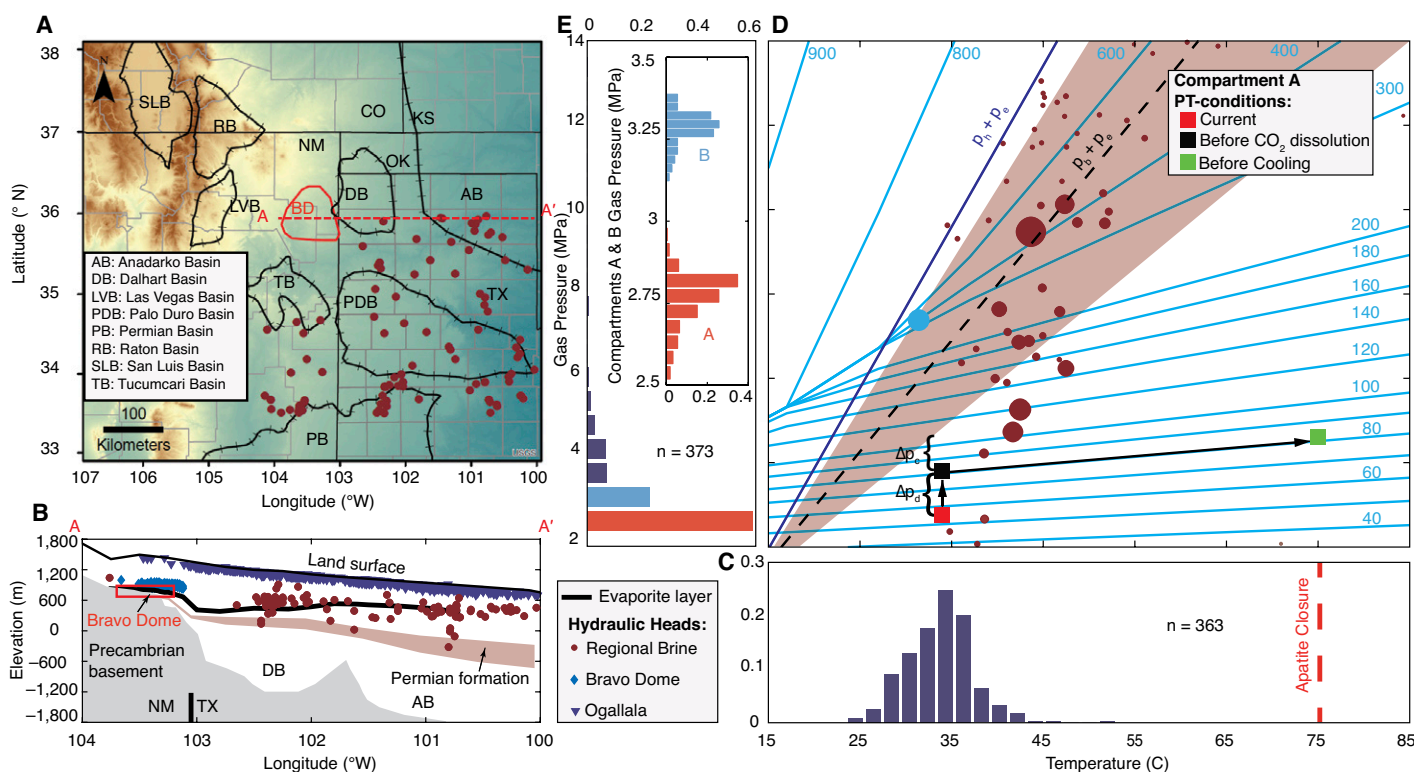


Figure 3. Contributions to underpressure in Bravo Dome (New Mexico, USA). **A:** Regional map of Bravo Dome and adjacent sedimentary basins. Well locations with preproduction brine pressure in Permian formations (Bair et al., 1985) are indicated by brown dots. **B:** Cross section along A-A', dashed line in A. **C:** Histogram of initial bottom hole temperature at Bravo Dome. **D:** Pressure-temperature-density phase diagram of CO₂ with isodensity contours. Light-brown cone shows range of estimated underpressure in Permian aquifers surrounding Bravo Dome. Brown circles are individual wells, and their size shows proximity to Bravo Dome. **E:** Histograms of initial bottom hole pressures (IBHPs) in Bravo Dome.

Erosion also leads to a reduction of the subsurface temperature. In isolated systems with constant volume, such as the Bravo Dome compartments, a reduction in temperature decreases the fluid pressure (Barker, 1972; Shi and Wang, 1986). At Bravo Dome, the cooling of magmatic CO₂ after emplacement could have led to a similar pressure drop. To evaluate this possibility, the maximum temperature of the reservoir after CO₂ emplacement has to be determined.

Thermochronology allows the determination of the time when a mineral was heated above its closure temperature. With a closure temperature of 75 °C, apatite provides constraints on the maximum temperatures that have been reached at Bravo Dome, and Figure 2B shows the location of a single well where several apatites have been dated by (U-Th)/He thermochronology (Sathaye et al., 2014). The apatite ages from this well do not record heating of the reservoir above the apatite closure temperature since the CO₂ emplacement at ca. 1.2–1.5 Ma. This suggests heating during the emplacement of the CO₂ was localized, and compartments A and B were never heated above 75 °C. Given the current reservoir temperatures 32–36 °C (Fig. 3C), the maximum temperature drop after CO₂ emplacement is less than 40 °C.

If the volume of these compartments and the mass of CO₂ within them remain constant during cooling, then the density of the gas is constant, and the associated pressure drop can be estimated using the pressure-temperature-density phase diagram of CO₂ shown in Figure 3D. The pressure drop can be inferred from the initial density and the change in temperature by following the corresponding isodensity line. The reservoir model for Bravo Dome allows us to estimate the initial density and hence the pressure drop due to cooling. Here, we focus on the pressure drop in the two main compartments. The volume of compartment A is 10.1 ± 2.2 km³, and it received 853 ± 170 Mt CO₂, resulting in an initial density of 81 ± 16 kg/m³. Similarly, compartment B has a volume of 3.1 ± 0.6 km³ and received 290 ± 60 Mt CO₂, resulting in an initial density of

93 ± 18 kg/m³. From the corresponding isodensity lines in Figure 3D, the maximum pressure drops due to cooling in compartments A and B are 0.78 ± 0.04 and 0.92 ± 0.05 MPa, respectively. Therefore, cooling of CO₂ after emplacement can account for at most 12% and 14% of the observed underpressure in each compartment.

CO₂ DISSOLUTION: A NEW MECHANISM?

Dissolution of CO₂ into brine is an additional mechanism that reduces the pressure at Bravo Dome, because the compartments are hydraulically isolated. Such pressure drops have been recognized theoretically (Steele-MacInnis et al., 2012), but they have never been identified in the field. Because thermal equilibration is faster than chemical equilibration, CO₂ dissolution is approximately isothermal. Therefore, the pressure drop can be inferred from Figure 3D, if the change in mass due to dissolution can be estimated using the geochemical characterization and the reservoir model (Gilfillan et al., 2008; Sathaye et al., 2014). In compartments A and B, 245 ± 49 and 70 ± 14 Mt CO₂ have dissolved into the brine, reducing the pressure by 1.02 ± 0.08 and 0.92 ± 0.05 MPa, respectively. This corresponds to 14% and 16% of the total observed underpressure.

DISCUSSION

The pressure drop due to CO₂ dissolution is comparable to that due to cooling and erosion. Given that the latter two are upper bounds, dissolution contributes the most to the post-emplacement pressure drop at Bravo Dome. These processes, together with the preexisting regional underpressure, account for 5.1 ± 2.5 MPa and provide an explanation for the observed low pressure, 6.4 MPa. The mechanisms discussed here may also provide an explanation for underpressure in other U.S. CO₂ reservoirs (Fig. 1).

The impact on GCS depends on the time scales over which CO₂ dissolution occurs. In high-permeability reservoirs, rates can be fast enough to dissolve significant amounts of CO₂ during injection (Neufeld et al.,

2010). In low-permeability reservoirs, the pressure drop due to dissolution is too slow to counteract the pressure buildup during injection, but it reduces CO₂ leakage and the displacement of formation brines, and it may contribute to the closing of hydrofractures.

Bravo Dome has stored a large amount of CO₂ for 1.5 m.y., but it does not correspond to the current conception of an ideal storage formation; it is neither highly permeable nor laterally open. This suggests that GCS may be possible in a broader range of formations than currently envisioned, increasing the storage capacity. In particular, large underpressured aquifers beneath regional evaporite layers have a proven seal and allow the injection of significant amounts of CO₂ without raising the pressure above hydrostatic. Such formations are widespread in the central United States (Fig. 1) and have previously been considered for hazardous waste injection (Puckette and Al-Shaieb, 2003).

CO₂ injection into these formations requires pressure management with brine extraction (Hosseini and Nicot, 2012) and hydraulic fracturing. However, if these challenges can be overcome, an additional 9.5 Gt CO₂ could be stored in the two main compartments at Bravo Dome, without exceeding hydrostatic pressure. This is comparable to the storage capacity that has been estimated for individual saline aquifers in the United States (Szulczewski et al., 2012).

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