

STUDIES IN GEOPHYSICS

The Role of Fluids in
Crustal Processes

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Degassing of Carbon Dioxide as a Possible Source of High Pore Pressures in the Crust

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INTRODUCTION

Increased pore pressures, especially pore pressures approaching lithostatic loads, change the state of effective stress and greatly reduce the work necessary for tectonic deformation (Hubbert and Rubey, 1959; Rubey and Hubbert, 1959). A number of mechanisms have been proposed that increase pore pressures (Hanshaw and Zen, 1965). One of the more interesting of the proposed mechanisms is the movement of carbon dioxide (CO₂) through the crust, suggested by Irwin and Barnes (Irwin and Barnes, 1975; Barnes *et al.*, 1978, 1984). The purpose of this chapter is to investigate the possible role of CO₂ as a source of high pore pressure by examining the following question: Given the current best estimate of the rate of CO₂ degassing, how low would the permeability have to be in order to generate pore pressures approaching lithostatic values?

Barnes *et al.* (1978, 1984) compiled worldwide data on the distribution of CO₂ discharge from the crust. They showed that most of the discharges are concentrated in two areas of the world: (1) a narrow circum-Pacific belt and (2) a broad mountainous area that extends across central and southern Europe and Asia Minor. They went on to note that the CO₂ discharge areas coincide with areas that are seismically active at the present time. Their clearest statement of the role of CO₂ in tectonic processes was presented in Irwin and Barnes (1975).

There is a continuing debate about the nature of volatiles in the mantle. Gold (see, e.g., Gold and Soter, 1980) argues that the carbon in the mantle is present largely as methane. On the other hand, most of the geological community reports the observations of carbon emanating from the mantle as CO₂ (see, e.g., Leavitt, 1982; Gerlach, 1988). The current consensus of the geological community seems to be that CO₂ is the more likely form of volatile carbon in the mantle. For the purpose of our simple experiment we have assumed a source of free CO₂ either in the mantle or within the crust.

Carbon dioxide is thought to come from three different sources: (1) organic material, (2) metamorphism of marine carbonate rock, and (3) degassing of the mantle (Barnes *et al.*, 1984). Each source is thought to have a different ratio of carbon isotopes. The evidence for the isotopic composition of mantle-derived CO₂ comes from analyses of fluid inclusions from volcanic rocks erupted along oceanic spreading ridges. Moore *et al.* (1977) reported $\delta^{13}\text{C}$ values ranging from -4.7 to -5.8 ‰ for CO₂ inclusions from basalts in the Pacific. Pineau *et al.* (1976) found similar values ($\delta^{13}\text{C}$ of -7.6 ± 0.5 ‰) for fluid inclusions in tholeiitic rocks from the mid-Atlantic ridge. Carbon dioxide with $\delta^{13}\text{C}$ in the range of -4.7 to -8.0 ‰ is thought to indicate mantle-derived CO₂, although the carbon isotope ratios alone do not provide an unambiguous indication of

a mantle derivation. $\delta^{13}\text{C}$ in the range of -4.7 to -8.0 ‰ can also be derived by mixing an organic CO_2 source, typically -20 ‰, with a marine carbonate CO_2 source, typically 0 ‰. However, in the case of the oceanic basalts a mantle source seems to be indicated as carbon isotope data is combined with other geologic information in order to interpret the source of the CO_2 .

Magmatic degassing may be a major source of free CO_2 within the crust. Harris (1981) showed that the solubility of CO_2 in tholeiitic basalts was strongly pressure dependent; thus, CO_2 that is in solution in magma at great depth will exsolve as the magma migrates upward in the crust and the pressure is reduced. Gerlach (1986, 1988) suggested that most of the CO_2 is degassed from plutons at depths of 5 to 10 or 12 km. Basaltic magmas associated with hot spots, such as Kilauea, have a higher CO_2 content than mid-ocean ridge basalts (Gerlach, 1988), so that they will tend to degas at higher pressures and greater depths.

GLOBAL FLUX OF CO_2

To estimate the permeability necessary to maintain near-lithostatic pore pressure, one must first estimate the flux of CO_2 . A number of investigators have made estimates of the CO_2 flux from deep in the Earth's interior to the surface. Leavitt (1982) summarized earlier estimates; we have extended Leavitt's summary to include more recent data. The various estimates are summarized in Table 11.1. They range from 1.7×10^{10} to 2×10^{13} moles of CO_2 per year; most estimates are in the range of 10^{12} moles per year. The scatter in the estimates is not as wide as one

might expect given the nature of the data.

If one assumes that the ultimate source of all carbon in the Earth is the mantle, one can place the current estimates of CO_2 flux in perspective by calculating a constant flux rate that would generate the known carbon reservoir. Current estimates of the global carbon reservoir are summarized by Sundquist (1985). One of these estimates (perhaps the best) is presented in Table 11.2. Assuming a constant rate of outgassing over 4.5 billion years yields a rate of outgassing of 2×10^{12} moles of CO_2 per year. These are grossly simple assumptions; however, the calculation suggests that the current estimated rate of CO_2 outgassing would approximately account for the global carbon reservoir. Other workers (e.g., Marty and Jambon, 1987; Gerlach, 1988) have also commented that the current rate is sufficient to generate the total carbon reservoir.

There are obvious complications, for example, differing rates of degassing in the geologic past and recycling of carbonate rocks in subduction zones. Des Marais (1985) noted that a rate of 10×10^{12} moles of CO_2 per year from the mantle would generate the global carbon inventory in approximately 700 million years and went on to point out the subduction tends to recycle carbon from the crust into the mantle. He estimates that perhaps half of the current flux is recycled carbon.

MODELING CO_2 FLUX IN THE CRUST

We have investigated the effects of a CO_2 flux at mid-crustal depths utilizing a numerical simulation model. The model simulates the simultaneous transport of mass and

TABLE 11.1 Reported CO_2 Inputs into the Atmosphere from the Earth's Interior

Reference	CO_2 Source	CO_2 Released (10^{12} moles/yr)
Borchert (1951)	Igneous and metamorphic	6.7
	Igneous only	1.3
Rubey (1951)	Total "excess CO_2 " since Earth's origin, including that from hot springs	0.5
Plass (1956)	CO_2 "released from the interior of the Earth"	2
Li (1972)	Total CO_2 at Earth's surface	1.0
Libby and Libby (1972)	Volcanic CO_2	0.017
Buddemeier and Puccetti (1974)	Hawaiian estimate	2.0
Anderson (1975)	Outgassing of oceanic crust	0.23
Baes <i>et al.</i> (1976)	Volcanoes, fumaroles, and hot springs	1.7 to 8.3
Leavitt (1982)	Volcanic eruptions	0.15
Javoy <i>et al.</i> (1982)	Mid-ocean ridges	20
Des Marais (1985)	Mid-ocean ridges	1 to 8
Marty and Jambon (1987)	C^3He , mid-ocean ridges	2.0
Gerlach (1988)	Mid-ocean ridges	0.3

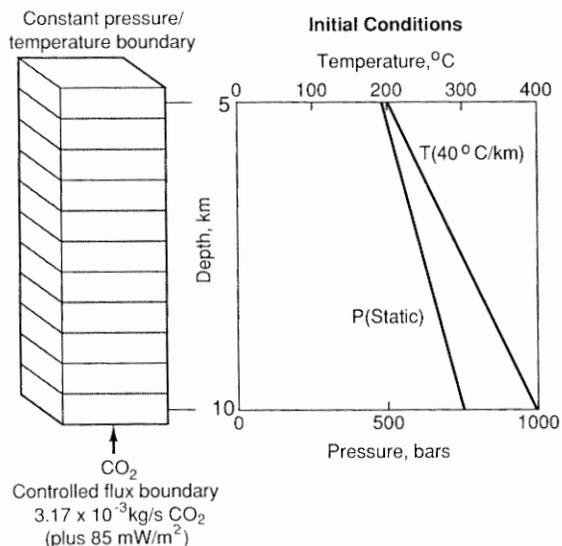


FIGURE 11.1 Schematic diagram of the one-dimensional model used to simulate CO_2 movement through the crust. Graph shows initial temperature and pressure profiles. The CO_2 flux at the lower boundary is derived by assuming that the global flux of CO_2 (about 3×10^{12} moles per year) is distributed over 1 percent of the Earth's surface area ($1.3 \times 10^6 \text{ km}^2$). The heat flow is typical of tectonically active areas.

heat by solving the appropriate set of coupled partial differential equations. It was originally developed to simulate geothermal reservoirs (Bodvarsson, 1982); we modified it to describe the properties of CO_2 rather than H_2O .

In our particular case we assumed one-dimensional flow vertically upward through a prism of crust, and we specify at the bottom of the prism what we believe to be a reasonable flux of CO_2 and heat. We then adjusted the permeability in the course of a set of numerical experiments. A schematic diagram of the model is shown in Figure 11.1. For the purposes of calculation we assumed that the pore fluid is pure CO_2 . The fluid properties of CO_2 were obtained from Kennedy and Holser (1966), Jacobs and Kerrick (1981), Vargaftik (1975), and Atkins (1978). The published density and viscosity data had to be extrapolated to higher pressures. Over the pressure and temperature ranges considered, CO_2 appears to have transport properties that are quite similar to water (Figure 11.2).

We assumed as the lower boundary condition a constant influx of CO_2 and heat. One can hypothesize other conditions at this lower boundary; however, for an initial calculation, constant flux provides insight. We have taken the flux as 3×10^{12} moles of CO_2 per year, which seems well within the estimates given in Table 11.2. This flux is distributed over the tectonically active area, which we have taken as 1 percent of the Earth's surface; it is then something of an average and may be substantially higher or lower locally.

Given this set of simple assumptions, we made a series of calculations to determine how low permeability would have to be in order to cause pore pressures approaching lithostatic conditions by simple permeation (i.e., without invoking enhanced transport or focusing effects). The results are summarized in Table 11.3. Since the model is run in a transient mode, the results are presented in terms of the time to reach lithostatic fluid pressure at 10 km depth, the lowermost cell of our simulated column of crust. At a permeability of 10^{-7} darcies, pore pressures do not approach lithostatic at steady state (infinite time). Given our assumed flux, the permeability must be on the order of 10^{-8} darcies or lower to generate pore pressures near lithostatic. Figure 11.3 shows pressure profiles at various times as the pressure builds to lithostatic for a case in which permeability is 10^{-9} darcies. In Figures 11.3 through 11.6, k is permeability, C_r is rock compressibility, and ϕ is porosity. Typical ranges of rock permeability are given by Brace (1980) and are discussed below (see Figures 11.7 and 11.8).

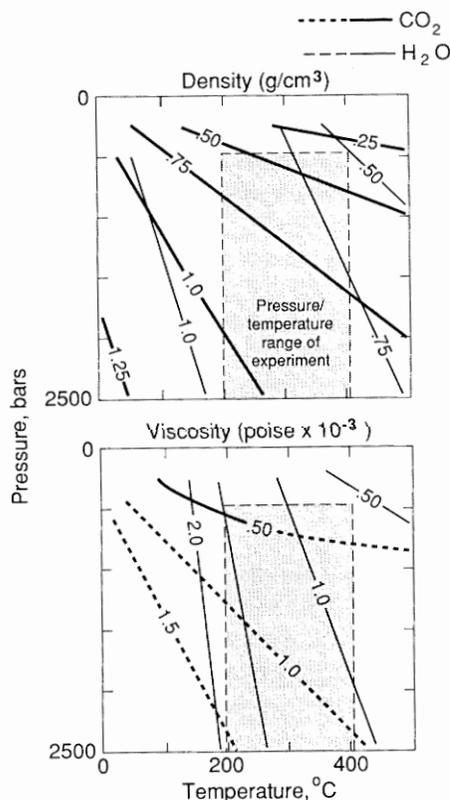


FIGURE 11.2 Density and viscosity of CO_2 and H_2O as functions of pressure and temperature. CO_2 properties (bold contours) are dashed where extrapolated from published data. H_2O properties are from Burnham *et al.* (1969), Keenan *et al.* (1969), and Haar *et al.* (1984). The pressure-temperature range of our experiment is shown by shading.

TABLE 11.2 Global Carbon Reservoir (from Sundquist, 1985)

Reservoir	Carbon (10^{15} tonnes)
Atmosphere	0.036
Oceans	6.4
Continents	7.3
Carbonate rocks	
Oceans	28
Continents	54
Metamorphic rocks	10
Total	106

TABLE 11.3. Time to Reach Lithostatic Pressure at a Depth of 10 km

Permeability (darcy)	Porosity	Time (10^6 yr)
10^{-9}	0.01	0.15
	0.02	0.30
10^{-8}	0.01	1.5
	0.02	3.0

HYDRAULIC FRACTURING

If the boundary condition at the base of a rock column of low permeability has a constant influx of CO_2 , it is possible to calculate arbitrarily high pore pressures. At some point increased fluid pressure will generate either a hydraulic fracture or a shear failure, which would increase the local permeability. The nature of the failure depends

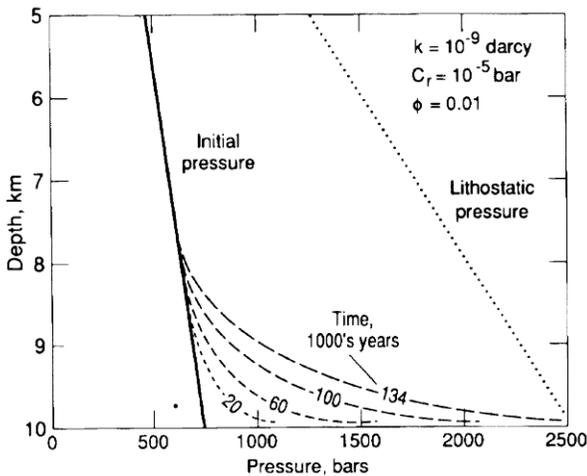


FIGURE 11.3 Pressure-depth profiles for various times after initiating flux of CO_2 into bottom of column (see Figure 11.4).

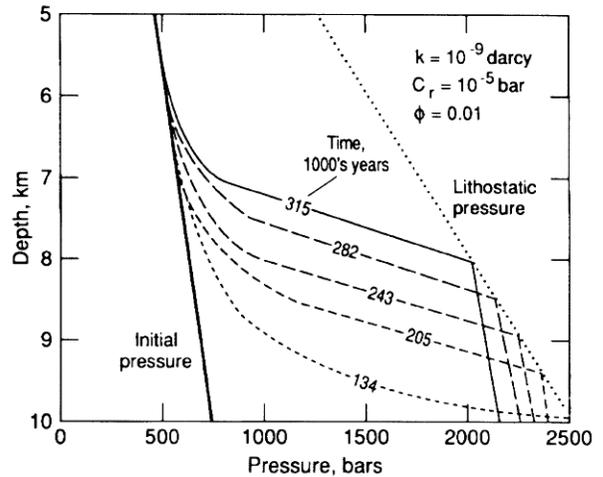


FIGURE 11.4 Pressure-depth profiles for various times. Once lithostatic pressure is reached, permeability is increased 1000 times, simulating the effect of fracturing.

on the local state of stress. In the case of a hydraulic fracture, a true tensional opening, the fracture will occur normal to the least principal stress. The fracture will occur when the pore pressure exceeds the sum of the least principal stress and the tensile stress of the rock. If the least principal stress is horizontal, the hydraulic fracture will tend to be a vertical opening. A variety of fracture orientations are possible depending on whether they represent shear or tensional openings and the local state of stress. When the vertical permeability is increased locally, pressure effects are distributed upward very quickly. The lithospheric load is a convenient upper bound for failure.

We have attempted to simulate a system in which hydraulic fractures are created. It seems that two possible processes can occur following fracturing. Once the pressure falls following the break, the fissure can (1) remain open, thereby increasing the local permeability or (2) seal itself, and return to something approaching its initial permeability. We have attempted to simulate both occurrences.

Figure 11.4 illustrates the pressure history in the lower portion of the column in the case where the fracture permeability remains high following a break. In this case permeability is increased 1000 times once the pressure within the simulated rock block reaches lithostatic. Note that lithostatic pore pressure migrates upward with time and that once breaks occur in the lower rock units they do not reach lithostatic pressures again. Figures 11.5 and 11.6 show what happens when the breaks are resealed, that is, permeability is returned to its original value following a break. The pressure once again builds to a lithostatic level at the bottom of the column, a second break occurs, and this sequence continues to repeat itself. As in the case

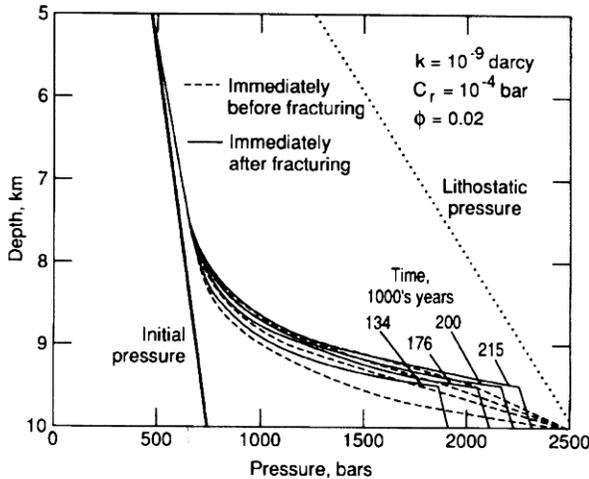


FIGURE 11.5 Pressure-depth profiles for various times for a system in which fracturing occurs once pressure reaches lithostatic and fractures quickly reseal themselves. In the simulation the following sequence is followed: (1) once pressure reaches lithostatic, permeability is temporarily increased 1000 times; (2) pressure drops quickly; and (3) once pressure drops, permeability is set back to its initial value, and the cycle repeats itself. There are two sets of profiles: the solid profiles are immediately before the fracturing and the dashed profiles immediately after. Pressure increases consistently with time within each set.

where the fissure remains open, the breaking will migrate upward (Figure 11.6). An interesting feature of this model is the pulsing nature of the pore pressure at a given depth (see Figures 11.5 and 11.6). Gold and Soter (1985) suggested a similar mechanism in considering the migration of fluids through the crust. Repeated episodes of fractur-

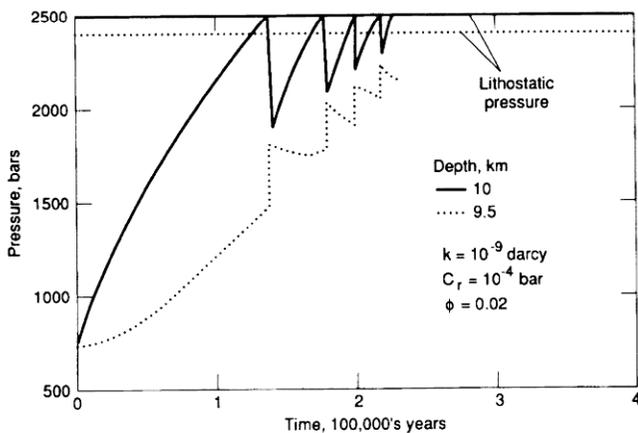


FIGURE 11.6 Pressure versus time for a system in which fracturing occurs once pressure reaches lithostatic and fractures quickly reseal themselves.

ing are also suggested by studies of hydrothermal ore deposits (see Titley, Chapter 3, this volume).

DISCUSSION

The results of our analysis suggest that degassing of CO₂ could be a source of high pore pressure, provided that the permeability of the rocks is sufficiently low. We must now consider whether such low permeabilities might reasonably be expected deep within the crust.

Brace (1980) has compiled both laboratory and field-measured permeability values for crystalline and argillaceous rocks. Figures 11.7 and 11.8 are adapted from Brace. Clearly, the values suggested by our simulation are in the lower range of what has been measured, both in the laboratory and in situ. If one examines only the in situ

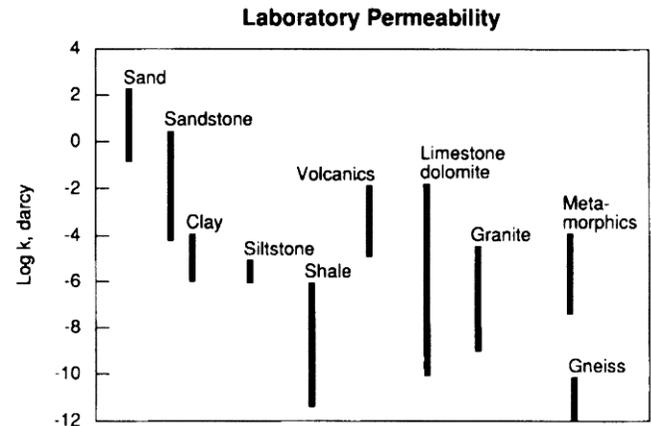


FIGURE 11.7 Range of laboratory permeabilities for different rock types (from Brace, 1980).

values, they are very near the low end of what has been measured. However, most of the in situ values have been measured near the Earth's surface, usually at depths above 500 m. It is our judgment that rock permeabilities in the range of 10⁻⁸ to 10⁻⁹ darcies are low but within the realm of expectation deep within the crust.

We have assumed in our calculations that the pore fluid is entirely CO₂. There is likely to be some H₂O present in the crust in the pressure-temperature range of the experiments. If significant amounts of both H₂O and CO₂ are present, the pore fluid would be a homogeneous single-phase mixture of H₂O and CO₂ at temperatures ≥300°C and a heterogenous two-phase mixture of H₂O-rich liquid and CO₂-rich vapor at lower temperatures. The presence of NaCl or other electrolytes would extend the two-phase region to higher temperatures (Bowers and Helgeson, 1983).

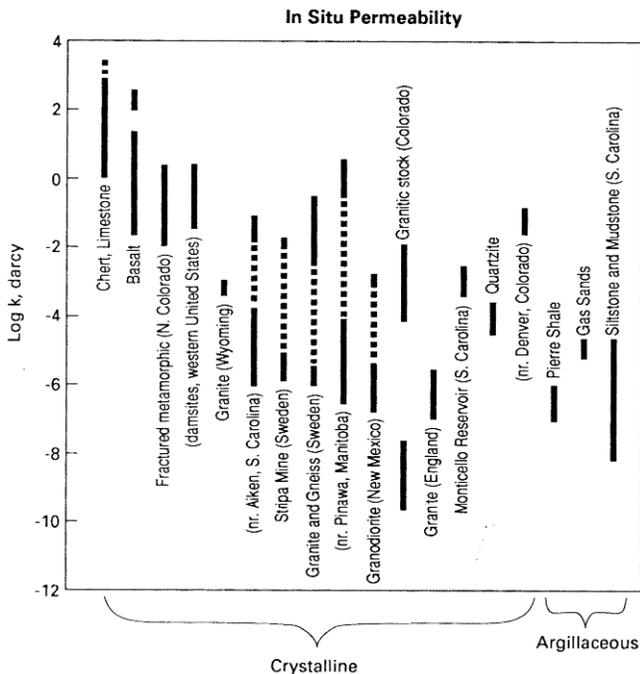


FIGURE 11.8 Range of in situ permeability measurements for crystalline and argillaceous rocks from various sites around the world (from Brace, 1980).

The limited solubility of CO_2 in H_2O in the pressure-temperature range considered (Takenouchi and Kennedy, 1964; Gehrig, 1980) implies that the volumetric flow rate required to transport a given flux of CO_2 in solution or in a two-phase mixture would generally be somewhat greater than the flow rate required to transport the same CO_2 as a separate phase. Since the density and viscosity of H_2O and CO_2 are comparable (Figure 11.2), the limiting permeabilities suggested here for anhydrous systems may be reasonably limiting values for hydrous systems as well.

Chemical controls on the buildup of CO_2 pressure may significantly constrain the applicability of our analyses. CO_2 pressures in natural systems at temperatures of 200° to 400°C will generally be limited by the reaction of Ca-Al-silicates, H_2O , and CO_2 to form calcite and mica and/or clay. At 300°C, for example, the partial pressure of CO_2 is likely to be on the order of tens of bars (Giggenbach, 1986). Where the available feldspar is converted to calcite and mica and/or clay, or the supply of H_2O is limited, the partial pressure of CO_2 is not fixed and may increase to greater values. This might occur where the flux of CO_2 is greatest (e.g., in fault zones and volcanic terranes). In general, CO_2 flux may be lower and/or permeabilities higher than those required to create high pore pressures; otherwise, we would expect most Ca-Al-silicates in the upper crust to be altered to calcite.

The question of fluid movement and pore pressure within the deep crust is complex. The intent of this chapter was a simple bounding calculation; we hope that it will provoke further thought, debate, and analysis. On balance it seems possible that fluid CO_2 migrating through the crust may yield high pore pressure where the rocks deep in the crust are sufficiently "tight" and the local geochemistry does not preclude high CO_2 fluid pressures.

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