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Mesodolomite formation as a result of layer parallel shortening and secondary porosity development by cooling of formation waters or cooling due to thrust emplacement: an example from the Canadian Foreland Fold and Thrust Belt

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Introduction

In major parts of the Canadian Foreland Fold and Thrust Belt (FFTB), some of the best reservoir units occur in Devonian and Mississippian dolomitised strata. However, reservoir characteristics are highly variable in the dolomites, and depend on crystal fabric (in relation to recrystallisation of mesodolomite replacement) and secondary porosity development. In this study, results from a transect from Banff to Calgary will be presented.

Dolomite types

In general three major types of dolomite have been recognised (table 1), namely:

- 1) pervasive fine to medium crystalline dolomites. This dolomite often develops in the upper part of shallowing upward cycles, where it is often associated with chickenwire anhydrite nodules (which may be pseudomorphosed) (Al-Aasm and Lu, 1994; Cioppa et al., 2000; Al-Aasm and Packard, 2000) reflecting upper intertidal to supratidal settings (Lonnee and Al-Aasm, 2000). However it also forms massive units in subtidal strata. These dolomites are considered to be eogenetic as is also supported by the crosscutting relationships with compactional stylolites. $\delta^{13}\text{C}$ -values normally plot in the range of marine related carbonates. $\delta^{18}\text{O}$ -values of 0‰ VPDB and even enriched values have been encountered in the fine crystalline dolomites (Al-Aasm and Packard, 2000; Cioppa et al., 2000), however, more often these dolomites possess depleted values varying around -4 to -6‰ VPDB (e.g. Durocher and Al-Aasm, 1997) and lower (Lonnee and Al-Aasm, 2000). The first values are in support of a marine origin, or indicate even the involvement of hypersaline solutions. The depleted values are explained by recrystallisation of originally marine-related dolomites, as attested by mottled luminescence fabrics. In the case of recrystallisation, the $\text{Sr}^{87}/\text{Sr}^{86}$ ratio is often slightly higher than the marine signature (Cioppa et al., 2000), otherwise their $\text{Sr}^{87}/\text{Sr}^{86}$ ratio is close to Mississippian or Devonian seawater composition respectively. More radiogenic dolomites often correspond with the more $\delta^{18}\text{O}$ -depleted dolomites. Intercrystalline porosity in the medium crystalline dolomites can be important (15-20%). It is secondary in origin, due to the dissolution of micrite matrix.

Type 1	fine to medium crystalline	5 - 100 μ m	sucrosic, fabric destructive non ferroan, non- or reddish to lilac luminescence, planar-e to -s textures no fluid inclusions
Type 2	Meso-dolomite	50-1000 μ m	planar-s (finer crystalline) and non- planar (coarser crystalline), non-ferroan uniform lilac to dull luminescence, with brighter vug-rimming phases
Type 3	Zebra-dolomite	100-5000 μ m	diagenetic repetition of bands (a) & (b), dull red luminescent with wispy textures outer crystal borders zoned no fluid inclusions

Table 1: Dolomite types and major characteristics

- 2) Mesodolomite is a general term to describe medium to coarse crystalline dolomite lithologies. These dolomites are clearly less porous, with the exception of lithologies where a secondary macroporosity exists. Typically an increase in non-planar crystal boundaries with increasing crystal size occurs. Based on their extinction pattern under crossed polarizers, the internal structure of the mesodolomite crystals seems to be composed of different sectors, explaining the mozaic extinction pattern. Locally some smaller sucrosic dolomite crystals are floating within the mesodolomite crystals. These fabrics point towards a recrystallisation origin of the mesodolomites, as advocated by many authors. However mesodolomite fabrics which are replacive in origin are attested by the fact that relict textures of the original strata (e.g. oolites, bioclasts, ...) have been recognised. In most of these dolomites, carbon isotope ratios appear to be buffered by the precursor carbonates (Al-Aasm and Packard, 2000; Cioppa et al., 2000). $\delta^{18}\text{O}$ -values are always rather depleted and often plot below -8‰ VPDB. Often the coarsest fabrics possess the most depleted stable oxygen signatures. The isotopic variations of mesodolomite often show a covariant trend between oxygen and carbon isotopes (Al-Aasm et al., 1996; Cioppa et al., 2000). Mesodolomites show higher $\text{Sr}^{87}/\text{Sr}^{86}$ isotope ratios than coeval seawater. This dolomite possesses a Late Cretaceous to Paleocene paleomagnetic remanence, pointing towards a dolomitisation event in relation to the Laramide deformation. Cioppa et al. (2000) noticed that a correlation of magnetic grain size and isotope values exist in the mesodolomites.
- 3) Zebra dolomites are characterised by a repetition of 1-10 mm thick dark grey (a) and white (b) dolomite bands, building up abbabba-sequences. The (b)-bands can make up between 50 to 65% of the zebra pattern. A central cavity, up to several mm across and few cm long, may separate two adjacent (b)-sheets, giving an overall porosity of about 3%, but porosity is extremely variable and varies from 0 to 10%. The zebra dolomite patterns seem to develop parallel to cleavage planes. The transition to the host limestone can be very sharp and is locally bordered by subvertical joints. It is likely that there is a tectonic control on the development of this dolomite. This is also attested by the fact that mm- to cm-scale subvertical displacements of the zebra bands occur. The degree of displacement gradually evolves to a maximum before it fades back to zero over a distance of a few centimeter. Typical for these lithologies is also the small scale development of stylolites within the dark grey bands or at the contact between the (a) and (b) bands. The dark grey (a) zebra dolomite

bands are composed of medium to coarse crystalline (100-200 μm), inclusion-rich turbid non-planar xenotopic-A interlocking dolomite crystals. Between the crystals, detrital clay occurs as well as some authigenic albite crystals. The contact between the dark grey (a) and white (b) zebra bands is gradational and characterised by an increase in crystal size towards the b-sheets. Dolomite crystals in the latter are coarse crystalline (0.5-5 mm) and non-planar to planar-c. Because of the presence of numerous inclusions, the (b)-crystals are generally milky white. Next to the cavities, transparent and almost inclusion-free crystal rims develop. The (b)-crystals display sweeping extinction and curved planes typical of saddle dolomite. Here, it was impossible to detect fluid inclusions, while in the transparent rim fluid inclusions are $<3\mu\text{m}$ in size. They are, however, too small to allow to see clear phase transitions. Therefore no fluid inclusion data are yet available for the studied zebra dolomites, but in general T_H values of reported zebra dolomites worldwide are $>80^\circ\text{C}$ (Nielsen et al., 1998; Boni et al., 2000), a temperature which is in line with the development of saddle dolomites. These dolomites are extremely depleted in $\delta^{18}\text{O}$ with most values plotting between -16 and -19.4‰ VPDB and for carbon between -0.3 and -1.9‰ VPDB. No difference exists between the (a) and (b) bands. Both bands are more or less stoichiometric.

Other types of dolomites (pseudomorphic, saddle, fracture lining, ...) have been reported but can generally be assigned to one of the three types described above.

Discussion

Origin of the fine to medium crystalline matrix dolomites

With the risk to over-generalize, most fine to medium crystalline Mississippian matrix dolomites display features typical for dolomitisation by marine derived fluids. Locally involvement of meteoric water, or burial fluids has been invoked (e.g. Al-Aasm and Packard, 2000; Al-Aasm and Lu, 1994; Durocher and Al-Aasm, 1997). However, the depleted stable oxygen isotopic compositions and/or high temperature fluid inclusions can equally be explained by recrystallisation. As working hypothesis these dolomites are assigned to the family of eogenetic marine derived dolomites similar to time equivalent dolomites from all over the world

Origin of the mesodolomite

Whether the mesodolomite should be explained by wholesale replacement of a pre-existing dolomite (i. e. matrix dolomites) or corresponds to a replacement of a previously non-dolomitised limestone is a matter of debate. Some replacement clearly occurs, but based on the observations made in this study only $<10\%$ is of this types, i.e. possesses relict textures of the precursor limestones. Since the pervasive matrix dolomite is fabric selective these relict textures clearly support a replacive origin. However, a major Mg mass-balance problem occurs if the other $>90\%$ are explained in a similar way. Therefore, a dominant replacement origin of a pre-existing dolomite is favoured (see also Al-Aasm and Packard, 2000). A puzzling feature in all the fluid flow models proposed (Deming et al., 1990; Ge and Garven, 1992, 1994) is that relatively small volumes of fluid from an external source flowing through a limited rock volume or compared to the existing connate water in the rock volume are postulated. This is also advocated by many authors studying cement types and dolomitisation in Fold and Thrust Belt systems (Mucchez et al., 1994; Machel and Cassey, 1999; etc. ...). We propose a model where layer parallel shortening (LPS) is the driving force inducing large scale recrystallisation, i.e. mesodolomite formation. According to Averbuch et al. (1992) and Frizon de la Motte (1997), LPS stylolitic planes develop immediately prior to thrust emplacement. Since the mesodolomites are affected by these bed perpendicular stylolites, we advocate that

recrystallisation of matrix dolomite and mesodolomite formation occurred just before thrust emplacement when tectonic compression was highest. In our concept we thus envisage a caterpillar-like displacement of the recrystallisation front during thrust emplacement. In this scenario there will be a resetting of the stable oxygen isotope composition and recrystallisation temperatures will be recorded in the newly formed fluid inclusions. Values however will differ from place to place in function of the position in the tectogene. Carbon values will be rock buffered as long as no additional carbon source becomes available. The fact that the latter has not been dramatically reset also indicates that recrystallisation precedes hydrocarbon expulsion. The slight increase in $\text{Sr}^{87}/\text{Sr}^{86}$ isotopic composition most likely relates to radiogenic Sr derived from intrabasinal shales, however it should be investigated whether intrastratal clay interaction in the dolomites itself is not another likely candidate. This would also explain the authigenic quartz which locally has been observed in the mesodolomites and the slight higher Fe contents in the mesodolomites. It is well-known that dolomite has a high recrystallisation potential and that normally recrystallised dolomites are more stoichiometric than non-recrystallised (Gregg and Shelton, 1990; Montañez and Read, 1993; Smith and Dorobek, 1993). This seems however not to be the case in the mesodolomites studied in the Alberta Basin. This is explained in the model by recrystallisation in a tectonic strain field with low water circulation.

Origin of the Zebra dolomites

The formation of zebra dolomites is related to the expulsion of suprahydrostatic fluids during a tectonic compressional phase, as was proposed for similar zebra dolomites by Nielsen et al. (1998). Dolomitisation may have been caused by hydrothermal brines that originated within basinal areas of the orogenic pile circulating deep within the crust, and ascended in their setting in the External Zone. In the case of the studied Canadian zebra dolomites the position of the zebra dolomitisation seems to be controlled by the former paleogeographical transition from Cambrian platform to basinal deposits. Noteworthy is that the zebra dolomites are non-ferroan. A relationship of expulsion of compactional fluids derived from basinal shales seems less likely at least in the scenario where fluids expulsion is concomitant with smectite to illite diagenesis.

Porosity development

In the matrix dolomites some intracrystalline porosity occurs due to the preferential dissolution of relict calcite (micrite phases). Also in some mesodolomites of reef debris talus deposits, some macroporosity occurs due to the dissolution of non-dolomitised stromatoporoids and tabulates. This porosity is thought to have developed due to “cooling of formation waters” or in relation to “host rock cooling”. In the first scenario (e.g. Giles and de Boer, 1990) it is assumed that fluids which eventually were in equilibrium with calcite, when channelled along faults and subsequently entering a calcite-bearing dolomites, will have the potential to dissolve calcite and thus create porosity. This is under the assumption that the fluids do not react with carbonate bearing rock along their ascent. The model of “host rock cooling” relies on the same physical constraints as the previous scenario, however, here it is assumed that the cooling is due to the uplift of a structural element during thrust emplacement. This uplift is often occurring at the middle stage of thrusting and would cool down the entire rock unit, its formation waters inclusive. This would generate an important potential of dissolving some calcite. The question why the dolomite did not dissolve remains unsolved. However as reported by Giles and de Boer (1989) the degree of saturation with respect to calcite (and dolomite) of cooled formation waters is strongly pH dependant and thus it is not unlikely that a scenario exists where calcite might dissolve while dolomite is not affected. However, some modelling should be carried out to confirm this scenario.

Another scenario that could be invoked is the dissolution of calcite by acidic fluids generated during the maturation of hydrocarbons (Giles and Marshall, 1986), however, this scenario is unlikely since one would expect to see some involvement of depleted CO₂ in the subsequent cementation of the calcites, which occur in the macropores.