Petrogenesis of the Tikorangi Formation fracture reservoir, Waihapa-Ngaere Field, Taranaki Basin

SD Hood, CS Nelson and PJJ Kamp

Department of Earth Sciences, University of Waikato, Private Bag 3105, Hamilton, Email s.hood@waikato.ac.nz

Abstract

The subsurface mid-Tertiary Tikorangi Formation is the sole limestone and the only fracture-producing hydrocarbon reservoir within Taranaki Basin. This study, based on core material from seven wells in the onshore Waihapa/Ngaere Field, uses a range of petrographic (standard, CL, UV, SEM) and geochemical techniques (stable isotope, trace element data, XRD) to unravel a complex diagenetic history for the Tikorangi Formation. A series of eight major geological-diagenetic events for the host rock and fracture systems have been established, ranging from burial cementation through to hydrocarbon emplacement within mineralised fractures. For each diagenetic event a probable temperature field has been identified which, combined with a geohistory plot, has enabled the timing of events to be determined.

This study has shown that the Tikorangi Formation comprises a complex mixed siliciclastic-carbonate-rich sequence of rocks that exhibit generally tight, pressure-dissolved, and well cemented fabrics with negligible porosity and permeability other than in fractures. Burial cementation of the host rocks occurred at temperatures of 27-37°C from about 0.5-1.0 km burial depths. Partial replacement dolomitisation occurred during late burial diagenesis at temperatures of 36-50°C and at burial depths of about 1.0 km, without any secondary porosity development. Fracturing occurred after dolomitisation and was associated with compression and thrusting on the Taranaki Fault. The location of more carbonate/dolomite-rich units may have implications for the location of better-developed fracture network systems and for hydrocarbon prospectivity and production. Hydrocarbon productivity has been ultimately determined by original depositional facies, diagenesis, and deformation.

Within the fracture systems, a complex suite of vein calcite, dolomite, quartzine, and celestite minerals has been precipitated prior to hydrocarbon emplacement, which have substantially healed and reduced fracture porosities and permeabilities. The occurrence of multiple vein mineral phases, collectively forming a calcite/dolomite-celestite-quartzine mineral assemblage, points to fluid compositions varying both spatially and temporally. The fluids responsible for vein mineralisation in the Tikorangi Formation probably involved waters of diverse origins and compositions. Vein mineralisation records a history of changing pore fluid chemistry and heating during burial, punctuated by changes in the relative input and mixing of downward circulating meteoric and upwelling basinal fluids. A sequence of mineralisation events and their probable burial depth/temperature fields have been defined, ranging from temperatures of 50-80°C and burial depths of 1.0-2.3 km. Hydrocarbon emplacement has occurred over the last 6 m.y. following the vein mineralisation events. The Tikorangi Formation must continue to be viewed as a potential fracture reservoir play within Taranaki Basin.

Introduction

The discovery in 1988 of the Waihapa oil field in southern onshore Taranaki within the fractured carbonate-dominated Tikorangi Formation opened up a new play in Taranaki Basin (Fig. 1 A, B) (King & Thrasher 1996). The earliest Miocene limestone-rich reservoir (Fig. 2) hosts hydrocarbons in open, but often extensively mineralised fracture systems (Plate 1A, B) (Kamp & Hood 1994) formed during Early Miocene overthrusting. Until 1988, all Taranaki Basin production had come from siliciclastic sandstone (King & Thrasher 1996; Smale et al. 1999). The fractured limestone play is dominantly an oil-producer (Fig. 1C). The Waihapa-Ngaere Field is located within the Tarata Thrust Zone in eastern

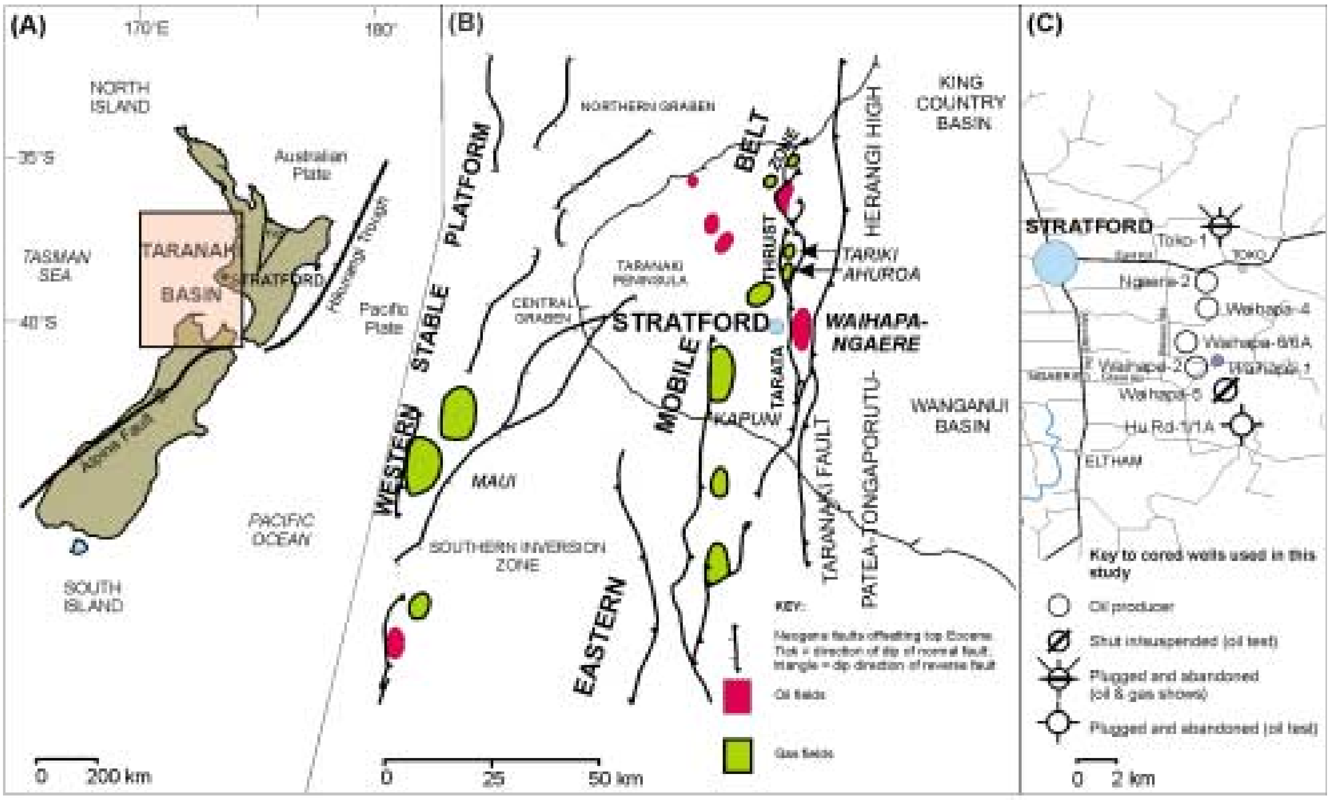

Figure 1: (A) Location of Taranaki Basin in New Zealand. TVZ, Taupo Volcanic Zone. (B) Major structural and tectonic elements and main oil/condensate accumulations within Taranaki Basin, including the Waihapa-Ngaere field (after King & Thrasher 1996). (C) Location and current status of the seven onshore wells providing core from Tikorangi Formation within the Waihapa-Ngaere Field that form the basis for this study.

Taranaki Peninsula (Fig. 1B, C) (King & Thrasher 1996). The Tikorangi Formation continues to attract both recent and on-going exploration activity in new prospecting areas outside the currently producing fields.

The objectives of this study have been to unravel the diagenetic history of the Tikorangi Formation, and in particular the vein mineral paragenesis, using core samples from the Waihapa-2, -4, -5, -6 and Ngaere-2 wells (Fig. 1C). Particular emphasis is placed on the identification of late diagenetic mineral phases which occur as precipitates in fractures and as replacements of other minerals, including their paragenesis. Eight geological-diagenetic events for the Tikorangi Formation host rock and fracture systems have been established ranging from initial cementation of the deposits through to hydrocarbon emplacement into the fractures. For each diagenetic event a probable temperature field has been identified which, combined with a geohistory plot, has enabled the timing of events to be determined. A variety of analytical techniques have been applied, including

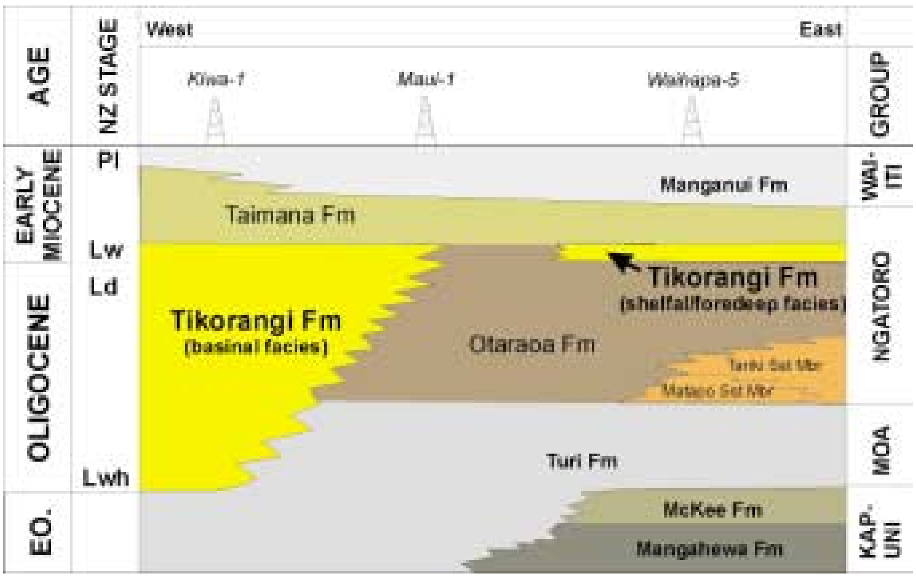

Figure 2: Schematic lithostratigraphy for Tikorangi Formation in Taranaki Basin showing age of the Oligocene to earliest Miocene (Lwh-Lw) basinal facies and earliest Miocene (Lw) foredeep and shelfal facies. Note the occurrence of limestone-dominated Tikorangi Formation amongst siliciclastic-dominated facies. New Zealand Stages are: Ar, Runangan; Lwh, Whaingaroan; Ld, Duntroonian; Lw, Waitakian; Pl, Altonian.

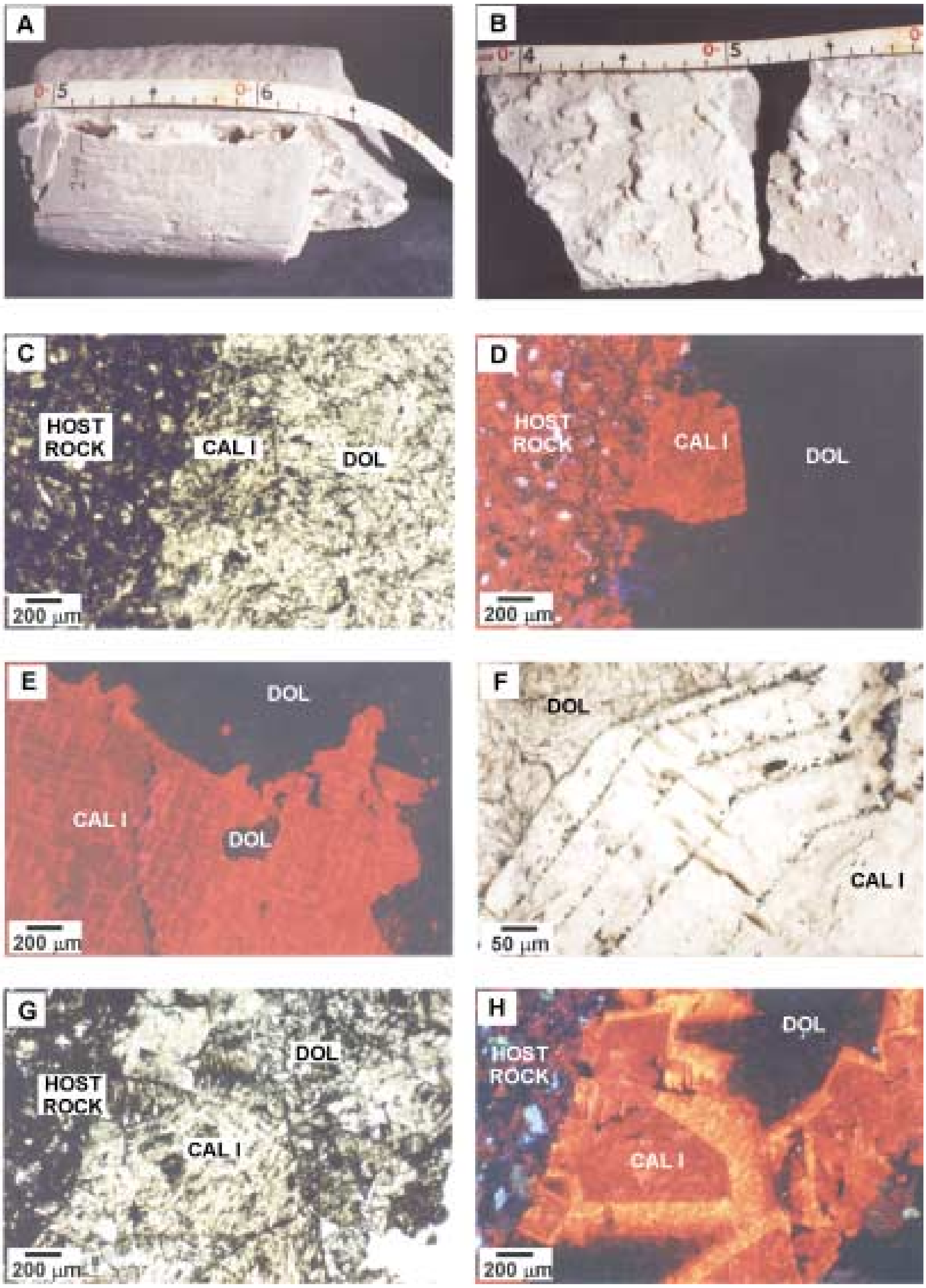

Plate 1: Core and thin section (PPL and CL) photographs of the major fracture porosity occluding carbonate phases, first generation calcite and baroque dolomite, in the Tikorangi Formation. (A) Fracture partially healed by coarse first generation calcite coated by fine baroque dolomite producing mammillary habit (sample W4.7.2B). (B) Fracture surface supporting large equant first generation calcite coated by very fine sucrosic dolomite (sample W4.7.2A). (C, D) Moderately ferroan dull luminescent first generation calcite cement precipitated at host rock boundary has been overgrown, but also partially replaced, by non-luminescent (black) ferroan baroque dolomite crust (evident in A, B). Note the very dirty appearance of calcite and generally cleaner dolomite (sample W2.9.12). (E) Dull luminescent first generation calcite showing evidence of partial dolomitisation by non-luminescent ferroan dolomite (sample W4.7.2B). (F) First generation calcite, observed in PPL containing primary fluid inclusions, overgrown by dolomite. (G, H) Zoned ferroan first generation calcite lining a vein with central porosity occluded by non-luminescent baroque dolomite (sample NG2.4.1B). CAL I, first generation calcite; DOL, dolomite.

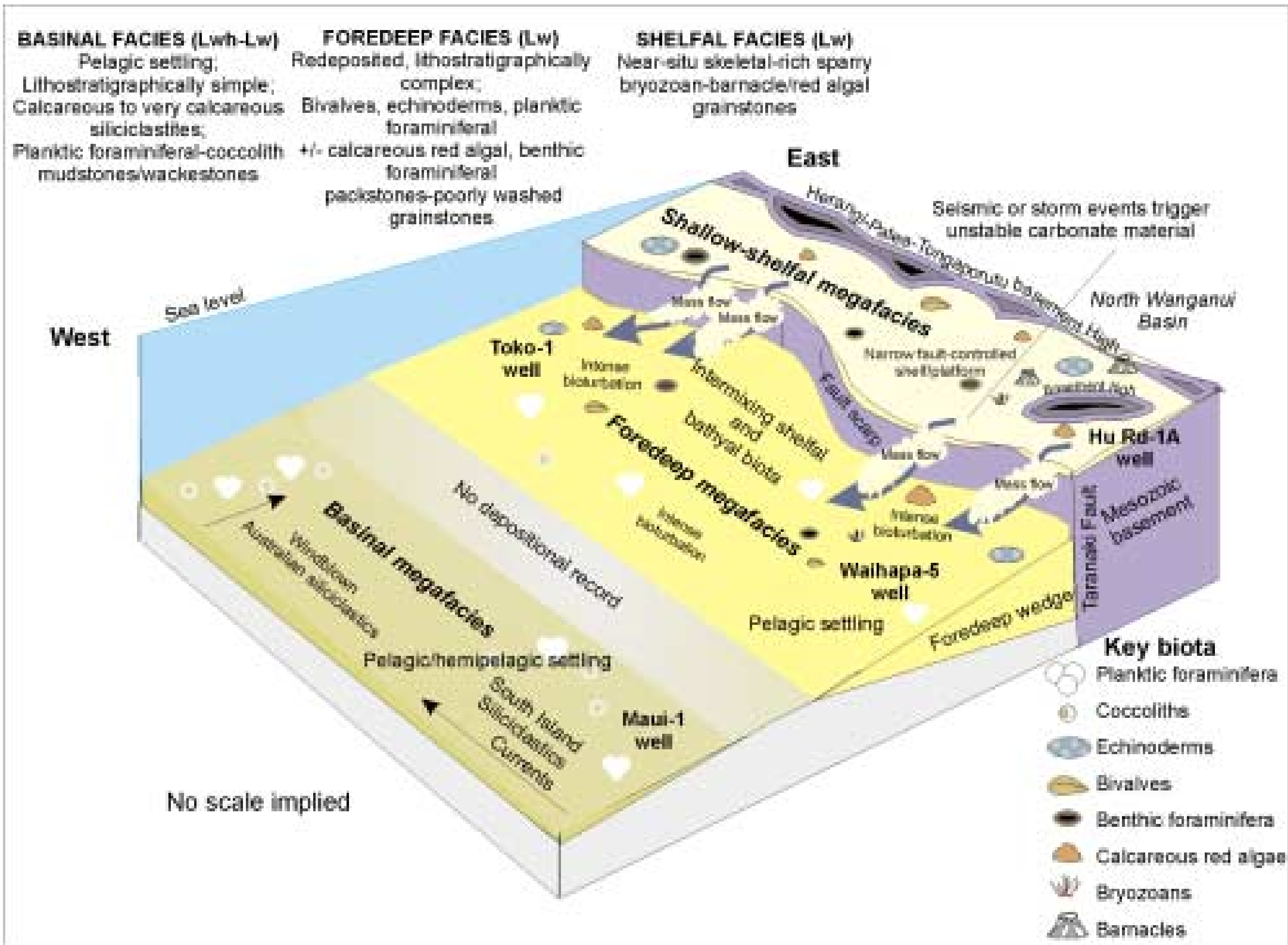

Figure 3: Tikorangi Formation depositional model showing the location and major characteristics of the three megafacies in eastern Taranaki Basin during earliest Miocene (Lw) time (from Hood 2000). Lwh, Whaingaroan; Lw, Waitakian.

standard, cathodoluminescence, and ultra-violet fluorescence microscopy, trace element and stable oxygen and carbon isotope geochemistry, and fluid inclusion geothermometry (see Hood (2000) for details). Of special interest are the environmental parameters that combined to produce a co-existing calcite-dolomite-quartzine-celestite vein mineral assemblage that variably healed the fractures, thereby influencing hydrocarbon flow rates.

Geological background

Taranaki Basin is largely a subsurface basin off the west coast of North Island, New Zealand (Fig. 1A, B). Its geology and petroleum systems have been described by King & Thrasher (1996). The basin has had distinct phases of evolution conducive to the generation and entrapment of hydrocarbons. The sedimentary fill records a Cretaceous through Cenozoic depositional megacycle involving transgression from the Late Cretaceous to Early Miocene, followed by regression since then.

A passive margin setting changed in the Late Oligocene – Early Miocene with the development of the Australia-Pacific convergent plate boundary through New Zealand. Rapid subsidence started at about 35 Ma in northern areas of the basin and moved southwards and eastwards reflecting progressive foundering of the paleo-shelf. The Ngatoro Group, a carbonate-rich sequence of Oligocene to Early Miocene age (Fig. 2), was deposited into a rapidly subsiding foredeep. The carbonate-dominated Tikorangi Formation within this group includes three major megafacies: shelfal, foredeep, and basinal (Fig. 3; Hood 2000). A phase of overthrusting began about 22–21 Ma when basement was thrust westwards along the Taranaki Fault, truncating and overriding the Tikorangi Formation (King & Thrasher 1996). During this compressive phase, the basin was partitioned into actively (Eastern Mobile Belt) and passively subsiding (Western Stable Platform) sectors (Fig. 1B).

The Eastern Mobile Belt is a broad region of Neogene tectonic deformation including the Tarata Thrust Zone, which has experienced up to 7 km of contraction (Fig. 1B) (King & Thrasher 1996). The Tarata Thrust Zone delineates a zone of locally intense deformation in which several stratigraphic traps, including the Waihapa-Ngaere Field structure, are located. In detail the zone comprises a series of imbricate en echelon thrust faults (thin-skinned overthrusting) and associated anticlines which contain most of the petroleum occurrences discovered to date.

Paleosalinity and paleothermometry determinations

Homogenisation temperatures (T_h) and freezing temperatures (T_m) of fluid inclusions in different vein mineral phases were measured using a USGS gas-flow heating/freezing system (see Table 1 for details). Following Goldstein & Reynolds (1994), T_m (temperature of final melting of ice) values were used to derive the salinity of precipitating waters, where NaCl (ppt) = $0.17 - (19.22 \times T_m) - (0.93 \times T_m^2) - (0.34 \times T_m^3$

Paleotemperatures of mineral precipitates have been derived from the $d^{18}O$ isotope data shown in Table 2 and Fig. 4. Values have been calculated using a range of water (SMOW) values (-1‰, 0‰, +1‰ PDB) for slightly meteorically influenced through to slightly enriched marine. For calcitic samples the paleotemperature equation of Shackleton (1967) was used: $T(°C) = 16.9 - 4.38(d^{18}Oc\text{-}w) + 0.1(d^{18}Oc\text{-}w)^2$ where $d^{18}Oc\text{-}w$ is the isotope composition of calcite minus that of water. Paleotemperatures for dolomite were calculated from the equation of Fritz & Smith (1970): $T(°C) = 31.9 - 5.55(d^{18}Od\text{-}w) + 0.17(d^{18}Od\text{-}w)^2$ where $d^{18}Od\text{-}w$ is the isotope composition of dolomite minus that of water. Burial depths of vein mineral formation have been estimated assuming a starting depositional bottom-water temperature of 8°C (Head & Nelson 1994) and an inferred geothermal gradient of 29°C/km, both comparable to the modern situation (Armstrong et al. 1996).

Diagenetic events

The following section outlines the eight major diagenetic events identified for the Tikorangi Formation host rock and associated fracture systems. These events are summarised in a geohistory plot (Fig. 5), in a generalised schematic paragenetic sequence (Fig. 6), and in a more detailed sequential model of vein mineralisation (Fig. 7). The key event data are summarised in Table 3, and the petrographic details of the vein minerals in Tables 4 and 5, with illustrations in Plates 1 and 2.

Event 1 – Host rock cementation (23.5-22.5 Ma; Fig. 6C)

The Tikorangi Formation host rocks (Plate 1A, C, D) are mixed siliciclastic-carbonate deposits. Using Dunham's (1962) petrographic classification, the rocks range from variably siliciclastic calcareous mudstones to wackestones to packstones to clean grainstones (Hood & Nelson in prep.; Hood et al. in prep.) These deposits were cemented during burial diagenesis by ferroan calcite microspar and micrite producing tight strongly pressure-dissolved fabrics with essentially zero porosity (Fig. 6C; Hood 2000). The ferroan nature of the low-Mg calcite cements is consistent with derivation from burial fluids, having been precipitated under reducing conditions from Fe-rich pore fluids sourced from siliciclastic-rich interbeds, a situation typical of other calcitic mid-Tertiary New Zealand limestones (Nelson et al. 1988; Hood & Nelson 1996). Cementation of the limestone-dominated sequence may have commenced at temperatures as low as about 20°C, or 0.5 km burial depth, and was complete by about 1 km burial depth at temperatures up to about 37°C (Table 3).

Event 2 – Host rock dolomitisation
(23-20 Ma; Fig. 6D)

Following lithification, partial burial dolomitisation of the Tikorangi Formation rocks occurred by fabric-selective

Table 1 Fluid inclusion data and nature of geological fluids for Tikorangi Formation vein minerals.

Mineral	Number measured	Total occurrence	Occurrence liquid+vapour[1]	Th[2]	Tm_3	Salinity (ppt)	Fluid type	Fluid(s) origins	Hydro-carbons
Calcite - first generation	52	Many	Rare	90+	-0.5 to -4 (-1.5)	9 - 84 (28) saline depleted to saline enriched with av. <normal marine	Mixed: meteoric> shallow burial>deep burial	Largely influenced from meteoric fluids entering through cascade zones	Absent
Dolomite	18	Rare	Very rare	88+	–	26 - 47 (33) Slightly saline depleted to saline enriched. Av. ~normal marine	Mixed: deep burial (basinal)> shallow burial> meteoric	Influx of basinal brines precusory to hydrocarbon emplacement	Absent
Celestite	19	Common	Rare	110+	-1.4 to -2.5 (1.8)	26 - 47 (33) Slightly saline depleted to saline enriched. Av. ~normal marine	Mixed: deep burial (basinal)> shallow burial> meteoric	Influx of basinal brines precursory to hydrocarbon emplacement	Absent
Calcite - second generation	52	Many	Some	101+	-1 to -4 (-1.9)	2-84 (35) Slightly saline depleted to saline enriched. Av. normal marine	Mixed meteoric> shallow burial- deep burial	Mix of basinal brines, hydrocarbon-bearing and meteoric fluids	Present

[1] Geothermometry enables the minimum temperature of fluid entrapment in a two phase liquid+vapour fluid inclusion (FI) to be determined, as well as the salinity of the fluid.
[2] Th or temperature of homogenisation is obtained by heating a two-phase FI from room temperature until the vapour phase disappears (i.e., liquid fills the FI).
[3] Tm, or temperature of final melting, is observed upon reheating a frozen FI and is recorded when all ice has returned to a liquid state. The freezing point depression is then related to the salinity of the FI (Shepherd 1985; Goldstein & Reynolds 1994).

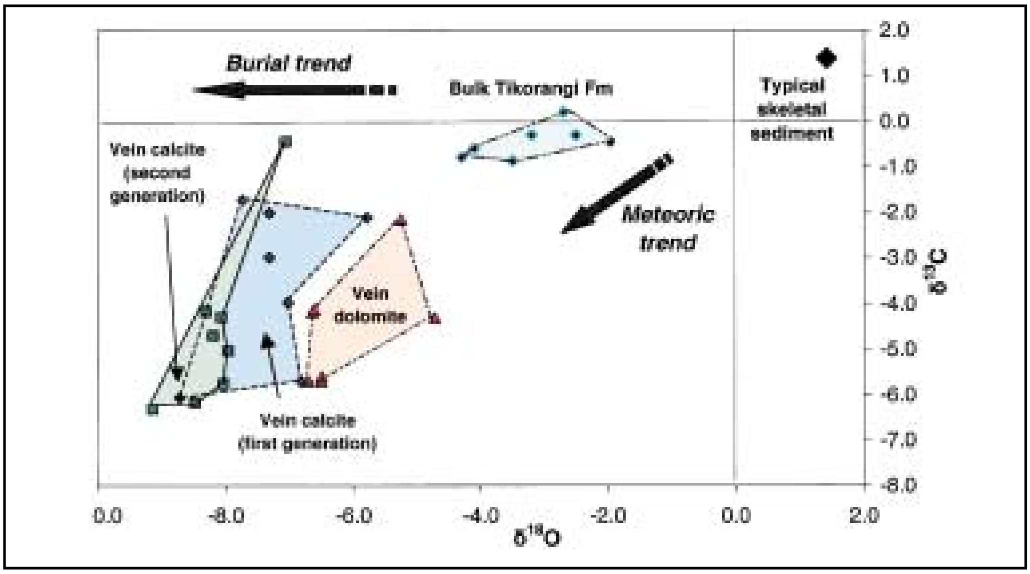

Figure 4: $d^{18}O$-$d^{13}C$ plot and generalised isotope fields for the host Tikorangi Formation and carbonate vein mineral phases. A typical New Zealand skeletal sediment value has been plotted as a reference (after Nelson & Smith 1996). The $d^{18}O$ data have been used to generate the paleotemperatures in Table 2 (see text).

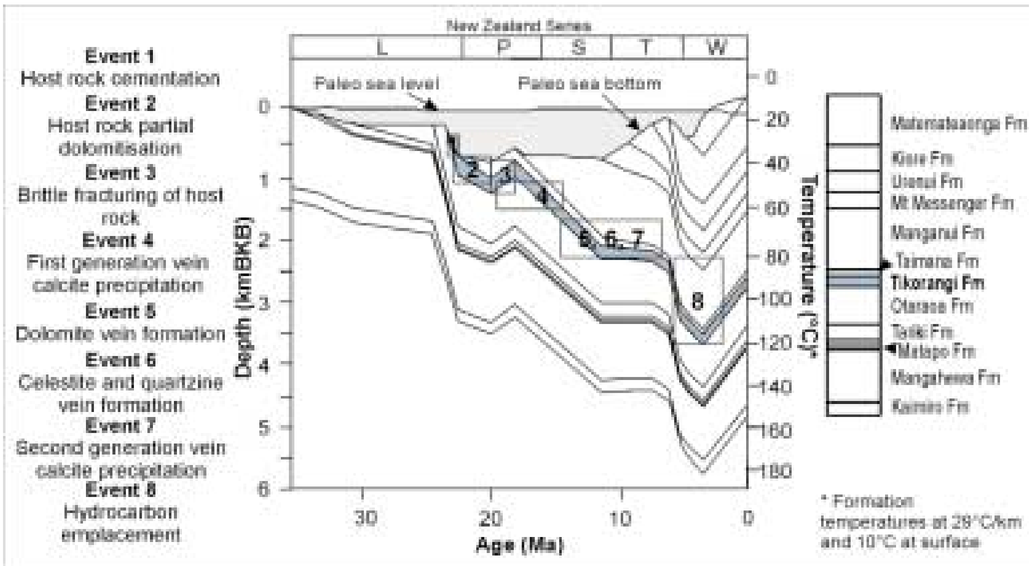

Figure 5: Schematic summary showing the timing of major diagenetic events within the Tikorangi Formation and associated fracture systems based on a geohistory plot determined in-house for Waihapa-1 (see Figure 1C for location). Event windows based on mean isotope temperatures. kmBKB, kilometres below kelly bushing. New Zealand Series are: L, Landon; P, Pareora; S, Southland; T, Taranaki; W, Wanganui.

replacement of interparticle, and rarely intraparticle, micritic/clay-rich matrix by ubiquitous but generally small quantities (typically <20%, rarely up to 50%) of unimodal, very fine (20-90 µm), scattered free-floating euhedral dolomite rhombs (Fig. 6D). Rhombs have dull luminescent Fe-rich cores, and often oscillatory bright and dull concentric outer zones. The dolomites are invariably poorly-ordered, non-stoichiometric calcian-rich (av. 58 mol% $CaCO_3$) and highly ferroan (av. 13 mol% $FeCO_3$) varieties (Hood 2000).

The nonstoichiometric Ca- and Fe-rich nature of the dolomite is suggestive of a late burial replacive diagenetic origin (Hood 2000), over a temperature range of possibly 40-50°C (Table 3). Fe was largely supplied from the siliciclastic-rich units. Mg sources were probably from often bioturbated clay-rich micrite, the release of Mg from clay mineral structures, and the pressure-dissolution of low- and intermediate-Mg calcite skeletons. Dissolution during dolomitisation more or less balanced precipitation, preventing development of secondary porosity and permeability enhancement, thus limiting the Tikorangi Formation's ability to act directly as a hydrocarbon reservoir outside of fracture porosity (Hood 2000).

Event 3 – Creation of fracture porosity

(20-18 Ma; Fig. 6E, F, 7)

Following burial lithification and partial dolomitisation, the Tikorangi Formation underwent significant horizontal compression when the Taranaki Fault and associated Tarata Thrust Zone became the mobile western edge of a large area of contraction that formed in the foreland behind the propagating subduction zone (Fig. 6E; King & Thrasher 1996). This Early Miocene faulting and folding created

Table 2 Stable oxygen isotope paleotemperature calculations for Tikorangi Formation host rock and vein minerals using a range of SMOW values and an average geothermal gradient of 29°C/km.

Mineral phase	Stable Isotopes						Isotope T (°C)			Isotope T (°C)			Isotope T (°C)			Paleodepth (km)			Paleodepth (km)			Paleodepth (km)		
	$\delta^{13}C$			$\delta^{18}O$			$\delta^{18}O_{SMOW} = -1$			$\delta^{18}O_{SMOW} = 0$			$\delta^{18}O_{SMOW} = +1$			$\delta^{18}O_{SMOW} = -1$			$\delta^{18}O_{SMOW} = 0$			$\delta^{18}O_{SMOW} = +1$		
	Min.	Max.	Mean	Min.	Max.	Mean	Min.	Max.	Mean	Min.	Max.	Mean	Min.	Max.	Mean	Min.	Max.	Mean	Min.	Max.	Mean	Min.	Max.	Mean
Host rock (*7) (Tikorangi Formation)	2.0	-0.9	-0.5	-2.0	-4.3	-3.2	21	32	27	26	38	32	31	43	37	0.5	0.8	0.7	0.6	1.0	0.8	0.8	1.2	1.0
Vein calcite - first generation (8)	-1.7	-6.2	-3.8	-5.8	-8.7	-7.4	40	57	49	46	63	55	51	69	61	1.1	1.7	1.4	1.3	1.9	1.6	1.5	2.1	1.8
Vein calcite - second generation (8)	-0.5	-6.3	-4.6	-7.1	-9.1	-8.2	47	59	53	53	65	59	59	72	65	1.3	1.8	1.6	1.5	2.0	1.8	1.7	2.2	2.0
Vein dolomite (7)	-2.2	-5.7	-4.5	-4.7	-6.7	-6.1	55	69	65	62	77	72	69	85	80	1.6	2.1	2.0	1.9	2.4	2.2	2.1	2.6	2.5

Note: Bracketed ( ) value denotes number of samples.* Denotes inclusion of data for 5 samples from Smale et al. (1999).

Table 3 Major diagenetic events occurring within the Tikorangi Formation.

Diagenetic event		Description	Isotope temp. (°C mean range)	Timing (Ma)	Timing (NZ Stage[1])
1	Host rock	Tikorangi Formation host rock cementation	27 - 37	23.5 - 22.5	Lw
2		Host rock partial dolomitisation	37 - 49	23 - 20	Lw
3	Fracture systems	Creation of fracture porosity	38 - 53	20 - 18	Po
4		First generation vein calcite precipitation	49 - 61	19 - 14.5	Pl-Sl
5		Vein dolomite formation	65 - 80	14.5 - c.12	Sl-Sw
6		Vein celestite and quartzine formation	65 - 80	14.5 - c.10	Sl - e.Tt
7		Second generation vein calcite precipitation	53 - 65	c.10 - c.7	Tt
8		Hydrocarbon emplacement into fracture networks	80 - 120	<6	Tk-?Wn

[1] Lw, Waitakian; Po, Otaian; Pl, Altonian; Sl, Lillburnian; Sw, Waiauan; Tt, Tongaporutuan; Tk, Kapitean; Wn, Nukumaruan. E, early.

extensive fracture systems within the Tikorangi Formation in the vicinity of the Tarata Thrust Zone (Fig. 1B, 6F; Plate 1A). Dolomite-rich units would have been more susceptible to brittle fracture than non-dolomite-bearing units because of greater intact rock strength and therefore brittleness (e.g., Martindale & Boreen 1997). This could have implications for the location of extensive networks of fracture systems and for hydrocarbon prospectivity and production in the Tikorangi Formation.

The key role of fracturing in the Tikorangi Formation has been the creation of fracture porosity and permeability enhancement in otherwise essentially non-porous and impermeable host limestones. For the fractures to have remained open, the Tikorangi Formation rocks must have had sufficient shear strength to resist the outwards horizontal stress (Bjørlykke 1994) resulting from burial diagenesis. Brittle failure was enhanced by the rapid onset of a compressive stress regime, and by the fine-grained and dolomitic nature of the limestones. Fractures may have opened and closed several times by a process of seismic pumping (e.g., Mann 1994). Open fractures became avenues of enhanced permeability with an ensuing history of ongoing displacement and episodic mineralisation prior to hydrocarbon emplacement. Fracturing is likely to have quickly connected the deep-burial environment with meteoric fluids in the shallower subsurface, abruptly reducing the pore fluid pressure gradient into the fractures, and potentially drawing down meteoric fluids (e.g., Choquette & James 1990).

The distribution and productivity of fractured Tikorangi Formation, while ultimately determined by tectonic configuration, is equally dependent upon the original depositional facies and the subsequent diagenetic processes affecting the rocks (e.g., Martindale & Boreen 1997). Fracture occurrence and spacing were determined by variability in rock composition, bed thickness, and porosity, while the predominantly north-south fracture orientation relates to the general direction of east-west compression. Highest fracture densities are suggested to occur in the tight, low porosity, well-cemented dolomitic-rich carbonates, Hood's (2000) facies C and D rocks, with possibly lower fracture densities in the more siliciclastic-rich facies A and B rocks. The facies C and D limestones have the lowest gamma ray and highest sonic velocities of all Tikorangi Formation samples, indicative of their dense, well-lithified nature (Hood 2000).

Event 4 – First generation vein calcite precipitation (19-14.5 Ma; Fig. 7)

Ferroan calcite (Table 4; Plate 1C-H) is a late diagenetic vein mineral formed after fracturing of the Tikorangi

Table 4 Petrography of carbonate vein minerals in fractures within Tikorangi Formation.

Carbonate phases	Crystal habit	Crystal size	Occurrence	Relative abundance	Staining (after Dickson	Crystal clarity	CL signature	Intercrystalline porosity	Petrographic features	Timing/ origin
Calcite ($CaCO_3$)										
Ferroan calcite - first generation	Dog-tooth- drusy equant euhedral- subhedral	Commonly 0.5 - 7mm (rarely to 20 mm)	Major vein fill, thick coatings on fracture surfaces	Very common	Dull-deep blue, slightly-highly ferroan	Typically dirty brown inclusion rich	Rarely concentrically zoned, patchy, non-dull red/orange	Common where associated with recrystalisation	Strain recrystallisation, sheared, slicken- sided, partially dolomitised	Phase 1 precipitate
Ferroan calcite- second generation	Coarse equant- dogtooth- platey, clear-	2 - 5mm, rarely>20 mm	Single crystals/clusters on: dolomite; slickensided	Rare to some	Light blue moderately ferroan	Clear to dusty, primary inclusion trains	Unzoned, patchy non-dull red/orange	Absent	Clean, clear sparry, course	Phase II Primary precipitate
Dolomite [$CaMg(CO_3)_2$]										
Baroque (ferroan) dolomite	Sucrosic, subhedral to typically anhedral xenotopic (nonplanar) fabrics	Coarse up to 1mm	Typically surface crust - coats phase I calcite - also vein fill	Common	Rarely unstained - typically deep green/blue, May ferroan outwards	Clean white sparry to pale brown	Dull - typically non (black)	Substantial porosity within central vein systems, minor resulting from dolomitisation	Partial phase I calcite, undulose extinction, some warped crystal faces, caclite inclusions	Phase III commonly a precipitate, common calcite phase I partial replacement

Table 5 Petrography of non-carbonate vein minerals in fractures within Tikorangi Formation.

Non-carbonate mineral phases	Crystal habit	Crystal size (mm)	Occurrence	Relative abundance	Staining (after Dickson	Crystal clarity	CL signature	Intercrystalline porosity	Petrographic features	Timing/ origin
Celestite ($SrSO_4$)										
Tabular celestite	Tabular plate- like anhedral to euhedral crystals, rarely prismatic	0.5 - 2.0mm	Fracture fill	Some - locally common	N/A	Very dirty inclusion rich, often pale brown to yellow discolouration	Dull - very dull purple/blue	Some intercrystalline porosity	High relief, fluid calcite inclusions, twinned	Phase IV Primary and replacement - some carbonate mineral effects
Acicular	Single needles, rarely inter- grown radiating crystals	0.2 - 0.5mm	Fracture fill	Rare	N/A	Clear-dirty	Very dull - dull purple/blue	Absent	Often ghost-like forms	Phase IV Replacement calcite ghosts
Authigenic quartz (siO_2)										
Fibrous	Fine, elongated fibres, some bundles	~0.1mm	Fracture fill	Very rare	N/A	Dirty, micritic appearance	Moderate- bright blue	Absent	Often ghost-like forms associated micritic-like calcite, felted	Phase ?IV Calcite replacement
Spherulitic quartzine	Irregular mass, containing scattered aggregated or spherulitic centres	Amorphous irregular patches in CL; pin-point sized crystals in CPL	Fracture fill	Some	N/A	Yellow/brown clean to dirty and inclusion- rich	Amorphous moderate/ bright vivid to darker blue	Absent	?Organic- bearing, relict inclusions, often dolomite/host rock contact	Phase ?IV Secondary - carbonate mineral relicts - commonly associated with dolomite

Formation at significant burial depths and elevated temperatures (Fig. 6G, 7). First generation calcite is variably textured, dog-tooth, drusy equant, white to grey calcite which constitutes the bulk of the vein filling. This calcite is ferroan and may exhibit a pale yellow discolouration resulting from oil staining. Crystal textures range from fine (0.5-1 mm) to coarse (3-7 mm), rarely >20 mm size (Plate 1B). Crystal growths are often multi-directional while multiple phases of movement and crystal growth are evident. Calcite crystals commonly exhibit an unzoned (Plate 1D), non- to dull red/orange luminescence. A general lack of zoning and dull- to non-luminescence are characteristic of Fe^{2+}-rich burial-derived cements (Nelson et al. 1988; Hood & Nelson 1996). Rarely, coarse slightly ferroan drusy calcite may show a concentric zonation (Plate 1H).

First generation vein calcite temperatures of formation range from about 50-60°C and probably occurred at approximately <19-14.5 Ma (Table 3; Fig. 7). Slickensiding of first generation ferroan calcite provides evidence of ongoing movement during precipitation (Plate 2E, F). Summary trace elemental matrices and schematic pie diagrams used to suggest the relative influence of meteoric, shallow-burial, and deep burial pore fluids during mineral precipitation have been constructed for both calcite and dolomite by Hood 2000. These data, as well as fluid inclusion data (Table 1), suggest a relatively strong meteoric signature associated with calcite vein mineralisation (Fig. 7). Episodic movement/ fracturing is likely to have been a key to facilitating fluid migration (Parnell 1994). Evidence suggests that fluid flow in fault-related fracture systems is episodic where faults, such as the Taranaki Fault marking the eastern boundary of the Tarata Thrust Zone, draw fluids into the crust (Parnell 1994). This may have been a mechanism for introducing meteoric fluids at depth. Faulted reservoir rocks such as the Tikorangi Formation are often characterised by oilfield waters deficient in salinity (North 1985). Studies have documented evidence for calcite and dolomite cements in environments where mixing of waters of widely varying salinity has occurred (Feng & Meyers 1998). Various mixing models have proposed that mixtures of meteoric water and sea water could be effective solutions for producing dolomite. The key to this type of model is that such mixing can result in undersaturation with respect to calcite but supersaturation with respect to dolomite (Morse & Mackenzie 1990). Modelling suggests that meteoric water flow may reach 2-3

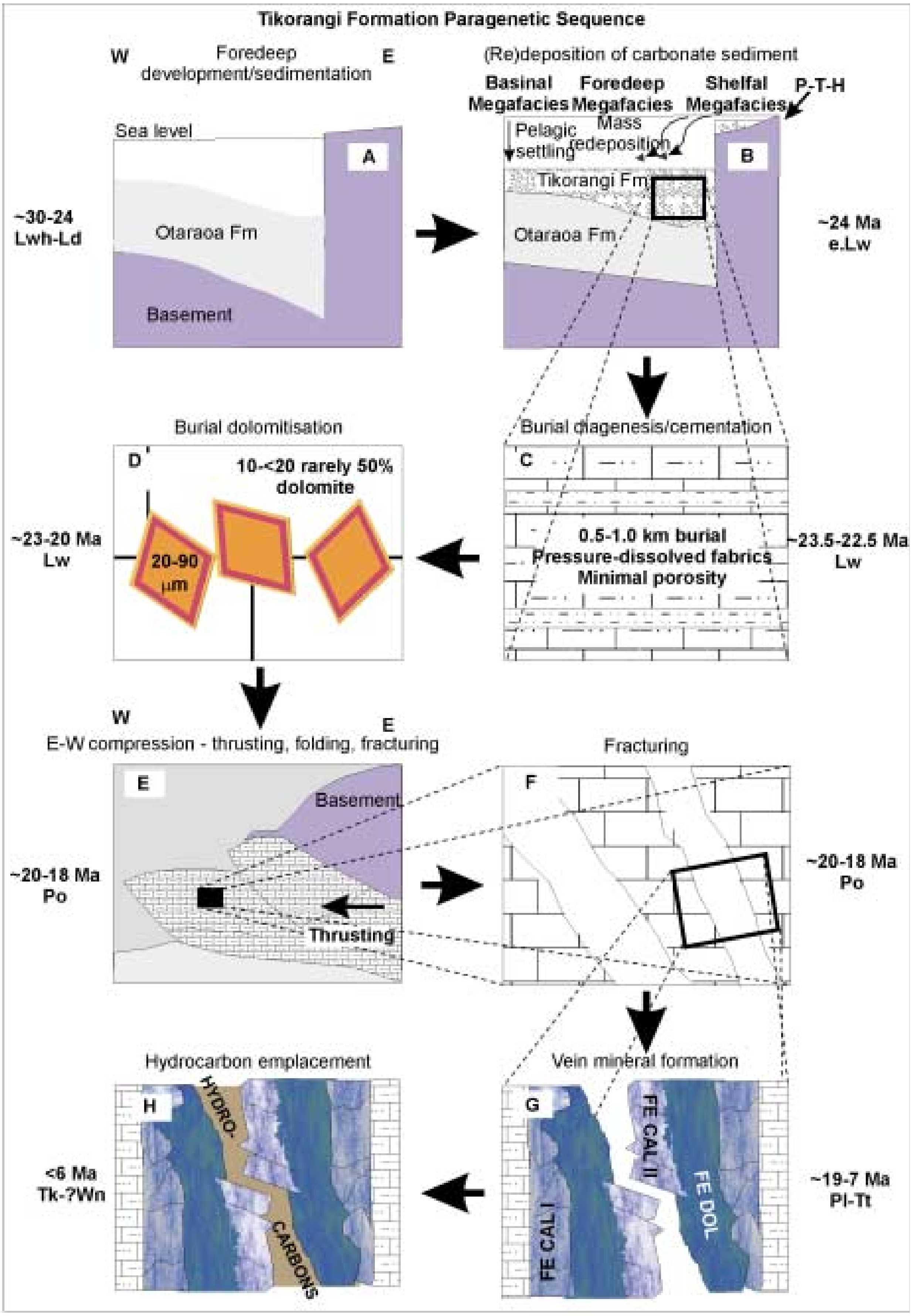

Figure 6: Schematic summary paragenetic sequence for the carbonate-dominated Tikorangi Formation from deposition to hydrocarbon emplacement. Timing of events is shown in approximate Ma together with the corresponding New Zealand Stages: Lwh, Whaingaroan; Ld, Duntroonian; Lw, Waitakian; Po, Otaian; Pl, Altonian; Tt, Tongaporutuan; Tk, Kapitean; Wn, Nukumaruan. P-T-H, Patea-Tongaporutu-Herangi High; FE CAL I, first generation ferroan calcite; FE CAL II, second generation ferroan calcite; FE DOL, ferroan (baroque) dolomite.

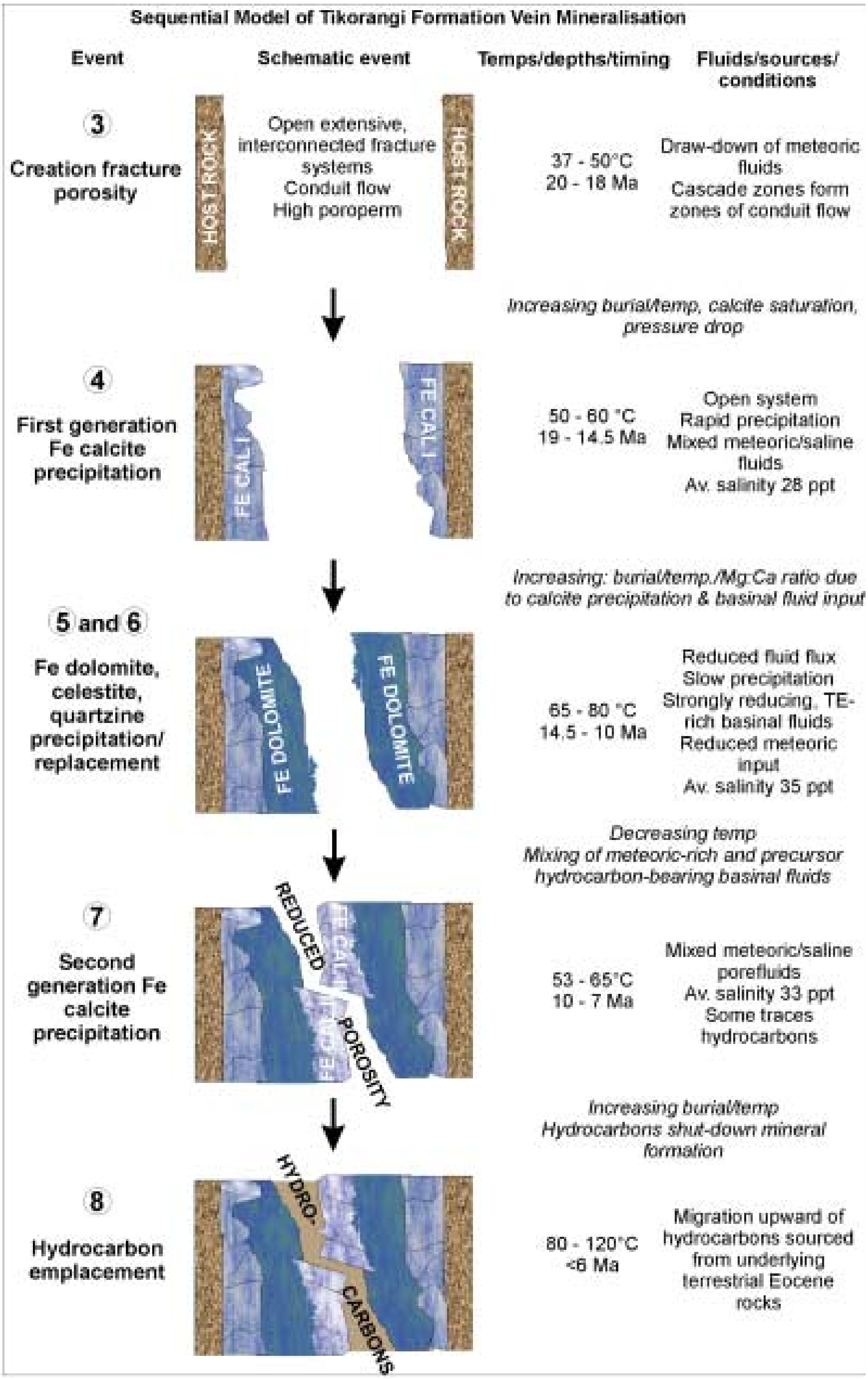

Figure 7: Schematic model summarising major vein mineralisation events in the Tikorangi Formation. FE CAL I, first generation ferroan calcite; FE CAL II, second generation ferroan calcite; TE, trace element.

km into basins like the Gulf of Mexico (Bjørlykke 1994). The question then is how deep can meteoric water penetrate beneath onshore Taranaki Basin? Allis et al. (1997) have noted that sandstones at 2 km depth could be conduits for deeply circulating groundwater on a timescale in the order of 10^4 to 10^5 years.

Event 5 – Vein dolomite formation
(14.5-c.12 Ma; Fig. 7)

Vein dolomite (Table 4) formed after first generation calcite, and is the next most voluminous vein mineral type, occurring as both a primary precipitate (Plate 1C, D, F-H) and as a replacement phase of calcite (Plate 2A, B). A mammillary texture results from a thin dolomite crust (up to 2.5 mm), comprising fine crystals (0.1-1 mm long), coating much of the first generation calcite on fracture surfaces (Plate 1B). The dolomite may be stained yellow/brown by hydrocarbons. Crystals often have a characteristic sweeping undulose extinction with curved or saddle-like shapes, a feature of baroque dolomite (Gregg 1988; Tucker & Wright 1990). Late saddle dolomites are commonly Fe-rich (Tucker & Wright 1990) like those in the Tikorangi Formation.

Baroque dolomite (Table 4; Plate 1C-H, 2A-D) potentially formed as a primary precipitate and less commonly as a replacement of earlier first generation calcite at temperatures averaging 65-80°C (Fig. 6G, 7). This lies within the oil window temperature of 60-150°C, a range in which baroque dolomite is commonly formed in association with hydrocarbons (Tucker & Wright 1990). Poorly ordered Ca-rich baroque ferroan dolomite, also reflecting the large amount of iron present in the mixed carbonate-siliciclastic system (Land 1983), is typical of a very late burial diagenetic origin (Zenger & Dunham 1988). Baroque dolomite is suggested to form at elevated temperatures coincident with the oil window, from about 60-150°C, and is often associated with sulfate-bearing carbonates and hydrocarbons (Morrow 1990), which has implications for celestite formation and the timing of hydrocarbon emplacement.

Baroque dolomite can be of replacement or, most commonly, very late pore-fill precipitative origin (Tucker & Wright 1990). A dominantly precipitative origin for the Tikorangi vein dolomite is favoured and is consistent with the isopachous coating it provides to the first generation vein calcite. Other petrographic evidence also supports a less voluminous replacement origin in the form of calcite relicts poikilitically enclosed within dolomite; preferential replacement of finer calcite crystals; and partially dolomitised coarser calcite crystals (Hood 2000). Relict calcite inclusions are common-place in dolomite that have replaced calcite. The absence of fluid inclusions in the dolomite is indicative of slow crystal growth (Goldstein & Reynolds 1994).

For all except Ca, the trace element concentrations in the Tikorangi Formation vein dolomites show increasing values down-section (Hood 2000). This is suggestive of dolomitising fluids being sourced generally from below the formation, and becoming more dilute as they move up-section. Vein dolomite shows a summary elemental matrix dominated by deep-burial and shallow-burial signatures, and only relatively minor meteoric influence (Hood 2000). Thus fluids are thought to have been of a burial/basinal type and Mg- and Fe-rich (Table 1; Fig. 7). Allis et al. (1997) have suggested that presently the Tikorangi Formation is charged by hydrocarbon and associated saline formation water leaking from sources within an underlying overpressured zone. Fluids may have imported Mg from underlying shale sequences, partially dissolving the precursor calcite phase, precipitating dolomite, and exporting Ca and Sr. Dolomite formation is suggested to have been terminated by a change in environmental conditions resulting from an increase in input of cool, relatively Mg-poor meteoric-derived fluids after which the precipitation of second generation calcite (event 7) occurred. Precipitation of carbonate minerals was eventually terminated by the introduction of hydrocarbon-bearing fluids.

Event 6 – Vein celestite and quartzine formation
(c.12-c.10 Ma; Fig. 7)

Celestite ($SrSO_4$) occurs as small patches of crystals in close association with the vein carbonate phases. Crystals are tabular, plate-like anhedral to euhedral crystals (Table 5; Plate 2E, F), or prismatic with a high relief and an often dirty pale brown to yellow discolouration. The celestite can poikilitically enclose relict inclusions of ferroan calcite and rarely dolomite rhombs, so that, much like dolomite, it is considered to have both a primary cement and secondary replacement origin. Celestite replacement can be confined to discrete thin calcite laminae separated by slickensided surfaces (Plate 2E, F). Celestite is clearly identified under CL by its usually very dull but characteristic purple/blue luminescence. Formation of vein celestite (Table 5) is likely to have resulted from the interaction of heated Sr-rich formation waters enriched in sulfate, possibly related to hydrocarbon-bearing fluids. At relatively modest burial depths (3,000 m or less) there is definite potential for highly variable sulfate concentrations in subsurface waters (Land 1997) conducive to celestite formation, for example as has been reported from DSDP cores (Baker 1986).

Non-detrital quartz (quartzine) (Table 5) appears in a few samples and may occur in two crystal forms. The first is fine, “length-slow chalcedony”, forming bundles of fibres/needles which may radiate from a common point. While characteristic of a void-filling cement it may also be a replacement fabric (Hesse 1990a). The second form is spherulitic quartzine (Plate 2A, B). Areas with a pale brown discolouration and inclusion-rich appearance may include organic matter related to hydrocarbons. Spherulitic quartzine crystals display a spectacularly vivid, generally moderately bright blue luminescence (Plate 2B).

Chertification within the Tikorangi Formation veins may have involved some precipitation of primary precipitates, but especially the replacement of both calcite and dolomite as evidenced by the poikilitic inclusion of carbonate relicts. The small crystal size of quartzine (Table 5) suggests rapid and homogenous nucleation, while replacement fabrics suggest relatively low silica concentrations. Quartzine often occurs at the host rock/vein contact and may be a result of the weakening and fracturing at this contact allowing Si-rich fluid migration. Evidence of a relatively late stage origin is provided by dirty areas, thought to be organic-rich inclusions related to precursor hydrocarbon-bearing fluids, occurring after first generation calcite and the majority of dolomite carbonate mineralisation (cf. Hesse 1990b). Formation would have been promoted in the presence of Mg and high alkalinity, conditions similarly favourable for dolomitisation (Hesse 1990a). The close association of dolomite and quartzine phases is suggestive of formation within a similar time frame, and they are not considered to be mutually exclusive. Elevated heat flow, generally regarded as necessary for a replacement origin, would have necessitated temperatures in the range of 60-100°C.

Event 7 – Second generation vein calcite precipitation (c.10-c.7 Ma; Fig. 7)

This second phase of ferroan calcite post-dates the earlier first generation calcite mineralisation, and also the thin isopachous coatings of dolomite mineralisation (Fig. 6G, 7). It comprises large (2-22+ mm), single or clustered, clear/translucent "free growing" crystals that exhibit a range of equant, dog-tooth, and platey habits. These ferroan calcite crystals formed directly upon dolomite (Plate 2C, D). In cases where dolomite phases are absent, a similarly textured phase of large equant ferroan calcite crystals grew on the surface of the first generation calcite following a period of movement and slickensiding. Second generation calcite contains some evidence of hydrocarbon entrapment (Plate 2G, H). Rarely, inclusions may show evidence of heterogeneous entrapment and segregation of hydrocarbon and water phases. Aqueous inclusions containing high-pressure bubbles of methane can be observed to have formed solid gas-hydrates (clathrates) (Plate 2H inset).

Second generation calcite (Table 4) formed when temperatures averaged 53-65°C, perhaps resulting from the introduction of cooler meteoric fluids from up-section. The presence of petroleum fluid inclusions in second generation calcite suggests in some cases precursory hydrocarbon-bearing fluids have migrated along with aqueous fluids from c.10 Ma.

Event 8 – Hydrocarbon emplacement
(<c.6 Ma; Fig. 7)

Hydrocarbon emplacement *per se* is inferred to have taken place during the past 6 m.y., post-second generation calcite formation, corresponding to burial temperatures and depths in the Tikorangi Formation fracture systems of 80-120°C and 2.0-3.5 km, respectively (Fig. 6H, 7). Armstrong et al. (1996) and King & Thrasher (1996) suggested that Eocene source-rocks may have begun expelling hydrocarbons in the later Miocene, while this study suggests hydrocarbons entered the Tikorangi fracture systems from the latest Miocene, since the New Zealand Kapitean (Tk) Stage (Table 3).

Conclusions

The distribution and productivity of hydrocarbon reservoirs, while often determined by tectonic configuration, are also dependent upon original depositional facies and diagenetic processes. The lithofacies diversity of the carbonate-dominated Tikorangi Formation, coupled with superimposed burial diagenesis and tectonic fracturing, have resulted in a complex reservoir rock. Post-lithification processes in the Waihapa-Ngaere Field involved substantial compressional tectonics and thrust faulting (in the Tarata Thrust Zone), and extensive brittle fracturing of carbonate strata and episodic fluid flow in response to tectonic loading. Fracture systems provided significant fracture-induced porosity into which a suite of calcite-dolomite-celestite-quartzine minerals formed prior to hydrocarbon emplacement, as summarised in Table 3 and the paragenetic model presented in Fig. 7. Mineralisation substantially reduced fracture porosity and permeability.

Petrographic, trace element, stable isotope, and fluid inclusion data record a varied and complex history of changing pore fluid chemistry and heating during burial, punctuated by changes in input of downward circulating meteoric and upwelling basinal fluids. The initial ferroan low-Mg vein calcite formed from Fe-rich, meteorically modified, saline fluids in which mixing occurred by percolation of deeply circulating groundwaters, particularly in cascade zones. Baroque dolomite subsequently formed both as a primary cement and as a calcite replacement following onset of higher temperatures (65-80°C) influenced by a change to more Mg-rich basinal fluid sources, precursory to hydrocarbon emplacement *per se*. Celestite and quartzine phases coincided with, or marginally post-dated dolomite, and formed as both spatially complex carbonate replacements and cements. Second generation ferroan low-Mg vein calcite later formed at cooler temperatures (53-65°C), perhaps resulting from the introduction of cooler meteoric fluids from up-section. The presence of petroleum fluid inclusions in the second generation calcite suggests in some cases precursory hydrocarbon-bearing fluids have migrated along with aqueous fluids from c.10 Ma, with hydrocarbon emplacement occurring in the last 6 m.y. Throughout this complex mineralisation history continuing vein movement is recorded by shearing, recrystallisation, pressure-dissolution, twinning, and slickensiding on mineral surfaces.

Acknowledgements

This paper derives from a PhD project and post-doctoral research made possible through funding by a University of Waikato Postgraduate Scholarship and funding from the Foundation for Research Science and Technology (UOW815). We thank Petrocorp Exploration Ltd, and especially Dr Alex McAlpine, for providing access to drill core and in-house petroleum reports. The support and hospitality of Gerry Spanninga (Petrocorp) is also acknowledged. We thank Ministry of Economic Development staff for assistance in their library and core store facilities in Wellington.

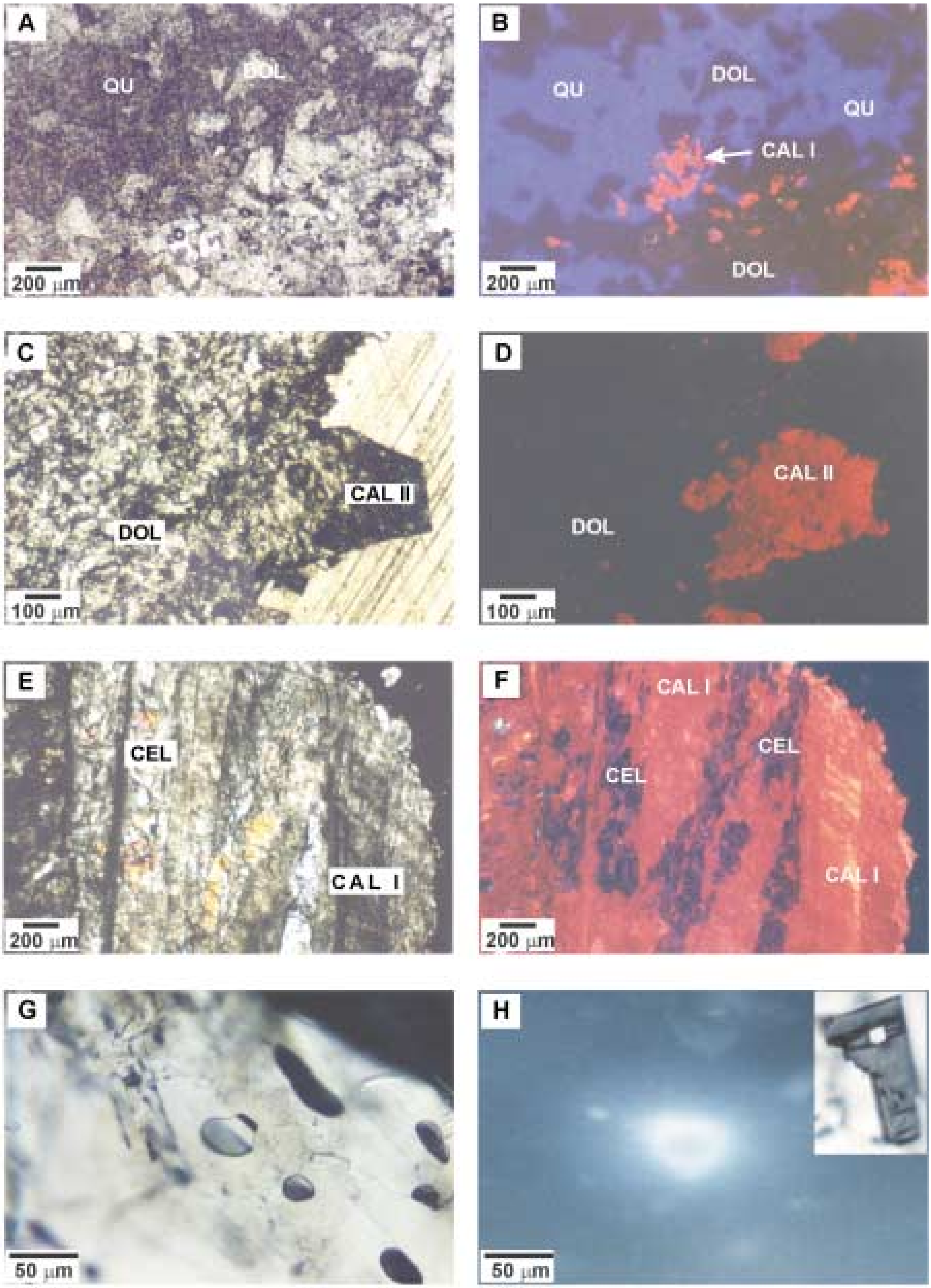

Plate 2 Photomicrographs showing several vein mineral types in the Tikorangi Formation as observed under PPL, CL, and UV. (A, B) Complex relationships between first generation calcite (dull red luminescent), and secondary (replacive) dolomite (non-luminescent) and quartzine (blue luminescent) phases (sample W4.7.4C, PPL & CL). (C, D) Second generation calcite has precipitated directly on dolomite phase (sample W2.9.12B, PPL & CL). (E, F) Celestite occurring as a replacement phase in first generation calcite exhibiting low first order interference colours and a dull purple/blue luminescence. Note the multiple generations of calcite separated by slickensided surfaces (sample NG2.4.3, CPL & CL). (G) Hydrocarbon-bearing fluid inclusions occurring within second generation calcite. (H) White luminescence is produced by hydrocarbon-bearing fluid inclusions under UV light. Inset shows inclusions containing high-pressure methane which has formed a solid gas-hydrate (clathrate), PPL (sample NG2.4.3). CAL I, first generation calcite; CAL II, second generation calcite; DOL, dolomite; CEL, celestite.

References

Allis, R. G.; Zhan, X.; Evans, C.; Kroopnick, P. 1997: Groundwater flow beneath Mt Taranaki, New Zealand, and implications for oil and gas migration. *New Zealand Journal of Geology and Geophysics 40*: 137-149.

Armstrong, P. A.; Chapman, D. S.; Funnell, R. H.; Allis, R. G.; Kamp, P. J. J. 1994: Thermal state, thermal modelling, and hydrocarbon generation in the Taranaki Basin, New Zealand. *In:* 1994 New Zealand oil exploration conference proceedings. Ministry of Commerce, Wellington. Pp. 289-30.

Baker, P. A. 1986: Pore-water chemistry of carbonate-rich sediments Lord Howe Rise, Southwest Pacific Ocean. *In*: Kennett, J. P.; von der Borch, C. C.; et al. Initial reports of the deep sea drilling project 90, vol. XC, Part 2. Washington. Pp. 1249-1256.

Bjørlykke, K: 1994: Fluid flow processes and diagenesis in sedimentary basins. *In*: Parnell, J. *ed*. Geofluids: origin, migration and evolution of fluids in sedimentary basins. *Geological Society Special Publication No. 78*, London. Pp. 127-140.

Choquette, P. W.; James, N. P. 1990: Limestones – the burial diagenetic environment. *In*: McIlreath, A.; Morrow, D. W. *ed*. Diagenesis. Geoscience Canada. Pp. 75-112.

Dunham, R. J. 1962: Classification of carbonate rocks according to depositional texture. *In*: Ham, E. D. *ed*. Classification of carbonate rocks-a symposium. American Association of Petroleum Geologists, Tulsa. Pp. 62-84.

Feng, H. L.; Meyers, W. J. 1998: Massive dolomitization of a Late Miocene carbonate platform: a case of mixed evaporative brines with meteoric water, Nijar, Spain. *Sedimentology 45*: 263-27.

Fritz, P.; Smith, D. G. W. 1970: The isotopic composition of secondary dolomite. *Geochimica et Cosmochimica Acta 34*: 1161-1173.

Goldstein, R. H.; Reynolds, T. J. 1994: Systematics of fluid inclusions in diagenetic minerals. Society of Economic Palaeontologists and Mineralogists short course 31, Tulsa.

Gregg, J. M. 1988: Origins of dolomite in the offshore facies of the Bonneterre Formation (Cambrian), southeast Missouri. *In*: Shukla, V.; Baker, P. A. *ed*. Sedimentology and geochemistry of dolostones. *Society of Economic Paleontologists and Mineralogists Special Publication No. 43*, Tulsa. Pp. 67-84.

Head, P. S.; Nelson, C. S. 1994: A high-resolution oxygen isotope record for the past 6.4 million years at DSDP Site 593, Challenger Plateau, southern Tasman Sea. *In:* van der Lingen, G. J.; Swanson, K. M; Muir, R. J. *ed*. Evolution of the Tasman Sea Basin: proceedings of the Tasman Sea conference, Christchurch New Zealand 1992. Balkema, Rotterdam. Pp. 159-179.

Hesse, R. 1990a: Origin of chert: diagenesis of biogenic siliceous sediments. *In:* McIlreath, A.; Morrow, D. W. *ed.* Diagenesis. Geoscience Canada. Pp. 227-252.

Hesse, R. 1990b: Silica diagenesis: origin of inorganic and replacement cherts. *In:* McIlreath, A.; Morrow, D. W. *ed.* Diagenesis. Geoscience Canada. Pp. 253-276.

Hood, S. D.; Nelson, C. S. in prep. A geophysical log-based lithostratigraphy and sedimentary evolution of the mid-Tertiary Tikorangi Formation limestone reservoir, Taranaki Basin, New Zealand.

Hood, S. D.; Nelson, C. S.; Kamp, P. J. J. in prep. Petrogenesis of mixed siliciclastic-carbonate shelf-to-basin megafacies in the cool-water mid-Tertiary Tikorangi Formation, Taranaki Basin, New Zealand.

Hood, S. D.; Nelson, C. S. 1996: Cementation scenarios for New Zealand Cenozoic nontropical limestones. *New Zealand Journal of Geology and Geophysics 39*: 109-122.

Hood, S. D. 2000: Subsurface stratigraphy and petrology of a cool-water carbonate fracture reservoir – the Mid-Tertiary Tikorangi Formation, Taranaki Basin, New Zealand. Unpublished PhD thesis, University of Waikato, Hamilton, New Zealand.

Kamp, P. J. J.; Hood, S. D. 1994: An investigation of the petrography and geochemistry of vein mineralisation within the Tikorangi Limestone, Taranaki Basin. A report prepared for Petrocorp Exploration, Earth Science Dept, University of Waikato, Hamilton. Unpublished open file petroleum report PR2016. Ministry of Commerce, Wellington.

King, P. R.; Thrasher, G. P. 1996: Cretaceous-Cenozoic geology and petroleum systems of the Taranaki Basin, New Zealand, Monograph 13. Institute of Geological and Nuclear Sciences, Lower Hutt.

Land, L. S. 1983: Dolomitization. Education course note series #24. 1982 AAPG Fall Education Conference in Denver, Colorado. American Association of Petroleum Geologists, Tulsa.

Land, L. S. 1997: Mass transfer during burial diagenesis in the Gulf of Mexico sedimentary basin: an overview. *In*: Montañez, I. P.; Gregg, J. M.; Shelton, K. L *ed*. Basin-wide diagenetic patterns: integrated petrologic, geochemical, and hydrologic considerations. *Society for Sedimentary Geology Special Publication No. 57*, Tulsa. Pp. 29-40.

Mann, U. 1994: An integrated approach to the study of primary petroleum migration. *In*: Parnell, J. *ed*. Geofluids: origin, migration and evolution of fluids in sedimentary basins. *Geological Society Special Publication No 78*, London. Pp. 233-260.

Martindale, W.; Boreen, T. D. 1997: Temperature-stratified Mississippian carbonates as hydrocarbon reservoirs - examples from the foothills of the Canadian Rockies. *In:*

James, N. P.; Clarke, J. A. D. *ed.* Cool-water carbonates. *Society for Sedimentary Geology Special Publication No. 56*, Tulsa. Pp. 391-410.

Morrow, D. W. 1990: Dolomite – Part 2: Dolomitization models and ancient dolostones. *In*: McIlreath, A.; Morrow, D. W. *ed*. Diagenesis. Geoscience Canada. Pp. 125-140.

Morse, J. W.; Mackenzie, F. T. 1990: Geochemistry of sedimentary carbonates. Developments in Sedimentology 48. Elsevier, Amsterdam.

Nelson, C. S.; Smith, A. M. 1996: Stable oxygen and carbon isotope compositional fields for skeletal diagenetic components in New Zealand Cenozoic nontropical carbonate sediments and limestones: a synthesis and review. *New Zealand Journal of Geology and Geophysics 39*: 93-10.

Nelson, C. S.; Harris, G. J.; Young, H. R. 1988: Burial-dominated cementation in non-tropical carbonates of the Oligocene Te Kuiti Group, New Zealand. *Sedimentary Geology 60*: 233-250.

North, F. K. 1985: Petroleum geology. Allen & Unwin, Boston.

Parnell, J. 1994: Geofluids: introduction. *In*: Parnell, J. *ed*. Geofluids: origin, migration and evolution of fluids in sedimentary basins. *Geological Society Special Publication No. 78*, London. Pp. vii-ix.

Shackleton, N. J. 1967: Oxygen isotope analyses and paleotemperatures reassessed. *Nature 215*: 15-1.

Shepherd, T. J. 1985: A practical guide to fluid inclusion studies. Blackie, Glasgow.

Smale, D.; Mauk, J. L.; Palmer, J.; Soong, R.; Blattner, P. 1999: Variations in sandstone diagenesis with depth, time, and space, onshore Taranaki wells, New Zealand. *New Zealand Journal of Geology and Geophysics 42*: 137-154.

Tucker, M. E.; Wright, V. P. 1990: Carbonate sedimentology. Blackwell Scientific Publications. Oxford.

Zenger, D. H., Dunham, J. B. 1988: Dolomitisation of Siluro-Devonian limestones in a deep core (5,350 M), southeastern New Mexico. *In:* Shukla, V.; Baker, P. A. *ed.* Sedimentology and geochemistry of dolostones. *Society of Economic Paleontologists and Mineralogists Special Publication No. 43*, Tulsa. Pp. 161-174.

Authors

STEVEN HOOD received BSc and MSc (Honours) degrees in Earth Sciences at the University of Waikato, and, while recipient of a University of Waikato Doctoral Scholarship, completed a PhD on the subsurface stratigraphy and petrology of the mid-Tertiary Tikorangi Formation fracture reservoir in Taranaki Basin in 2000. His research interests are currently focused on the petrology and diagenesis of cool-water carbonates with a particular focus on New Zealand Cenozoic carbonates. He is currently a Post-doctoral Research Fellow in the Department of Earth Sciences at the University of Waikato and is involved in undergraduate teaching.

CAM NELSON received BSc and BSc (Honours) degrees in geology at Victoria University. He lectured in the Department of Geology at the University of Auckland where he received his PhD before joining the Department of Earth Sciences at the University of Waikato in 1971 as its founding geological staff member. He was department chairperson from 1988 to 1996, and has been a professor since 1991. His research interests are in sedimentary and marine geology and stratigraphy, and Cenozoic paleo-oceanography and paleoclimatology of the southwest Pacific region. He is past president and office holder of the Geological Society of New Zealand, and was elected a fellow of the Royal Society of New Zealand in 1994.

PETER KAMP is professor of Earth Sciences in the Department of Earth Sciences at the University of Waikato and leader of the FRST-funded sedimentary basins research programme. Research activities include basin analysis, and fission track and U-Th/He thermochronology and thermochronometry applied to sedimentary basins and basement provinces.