Predicting fluid pressure in sedimentary basins from seismic tomography

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SUMMARY

Gravitational compaction of thick (2–10 km) sediment accumulations in sedimentary basins is controlled by the interplay of mechanical and chemical processes that operate over many orders of magnitude in spatial scale. The compaction of sediments into rock typically involves a density increase of ≈ 500 to 1000 kg m^{-3} , occurring over a depth-scale of several kilometres. The volume decrease in the compacting sediments releases vast volumes of water, which plays an important part in the global hydrological cycle and also in tectonic and geochemical processes; including the formation of hydrocarbon and mineral deposits. This study utilizes recently developed tomographic seismic images from the Porcupine Basin, which lies in the deep-water North Atlantic Ocean. A generic method for predicting fluid pressure variations that are driven by gravitational compaction is developed over the scale of the entire sedimentary basin. The methodology is grounded upon both observational evidence and empirically based theories, relying on geophysical measurements and relationships between sediment porosities and densities. The method is based upon physical concepts that are widely used in the petroleum industry and applied extensively in models of overpressure development in sedimentary basins. Geological and geophysical data from exploration wells are used to test the predictions of the method at two locations within the basin and are found to be in good agreement with the theory.

Key words: Permeability and porosity; Europe; Seismic tomography; Sedimentary basin processes.

INTRODUCTION

Recent advances in the acquisition of multichannel seismic (MCS) data sets using long offset streamers has permitted the development of detailed seismic velocity models of the structure of sedimentary basins. This paper uses results from a previous study in the Porcupine Basin (Prada *et al.* 2018) to develop a methodology for predicting pore pressure variations at a basin-wide scale (10^2 km) from such tomographic models. The methods developed are generic and applicable to a variety of tectonic settings, where gravitational compaction of sediments into solid rock is important.

The developed methodology is based upon early work on the petrophysical properties (i.e. seismic velocities and bulk density) and the mechanical behaviour of compacting sediments over vertical scales of several kilometres (Athy 1930; Wyllie *et al.* 1956, 1958; Ludwig *et al.* 1970; Brocher 2005). These earlier studies were based on an experimental and empirical approach.

Over basin-wide scales, the degree of fluid overpressure is controlled by the balance between the creation of pore pressure by sediment loading and its dissipation, controlled by sediment permeability (Hantschel & Kauerauf 2009; Marin-Moreno *et al.* 2013a). This

is ultimately related to the development and breaching of lithological permeability barriers during basin development and sedimentation (Osborne & Swarbrick 1997; Swarbrick *et al.* 2002; Tingay *et al.* 2009; Prada *et al.* 2018).

The physical concepts used in this study are widely known and routinely applied in petroleum exploration and drilling technologies (Eaton 1975; Hantschel & Kauerauf 2009). These older concepts and axioms are used to build a *simple* predictive model for basin-wide variations in pore fluid pressure that is potentially testable with observational exploration borehole information.

More sophisticated geomechanical models that are constrained by seismic and exploration well data have been developed and applied to calculate overpressure due to disequilibrium compaction and thermal expansion in deep sedimentary basins, such as in the Black Sea Basin (Marin-Moreno *et al.* 2013b). The novelty of the simpler approach developed here lies in the use of high-resolution tomographic images of post-rift sedimentary seismic velocities for quantifying and predicting compaction and fluid pressure variations at the scale of an entire sedimentary basin.

The focus of this study is the Porcupine Basin, which lies in deep-water off the southwestern coast of Ireland (Fig. 1). This sedi-

Figure 1. Bathymetry map of the Porcupine Basin showing the location of the MCS line used in this study (black line). Only shot gathers along the red section were included in the traveltimes tomography. The white dots correspond to the two exploration wells used in this study. The blue outline around Well 1 indicates a region with 3-D seismic coverage. Bathymetry data set is extracted from Amante & Ekins (2009). The location of the Mesozoic basement fault system is also shown.

mentary basin is investigated, given the recent results on long-offset streamer traveltimes tomography of the post-rift sequence in the basin (Prada *et al.* 2018).

THE PORCUPINE BASIN

The Porcupine Basin formed in response to several rifting and subsidence phases during the Late Paleozoic to Cenozoic (Tate & Dobson 1988; Moore 1992; Croker & Shannon 1995; Naylor & Shannon 2011). Tectonic extension in the basin led to a dramatic crustal thinning that increases from north to south (Reston *et al.* 2004; O'Reilly *et al.* 2006; Watremez *et al.* 2016; Prada *et al.* 2017; Chen *et al.* 2018). A significant post-rift phase of extensional faulting and reactivation of syn-rift faults occurred during the Mid-Late Eocene (Worthington & Walsh 2016; Prada *et al.* 2018).

These recent results, based on the analysis of 2-D and 3-D MCS reflection data (Fig. 2a) revealed hitherto unrecognized phases of post-rift (i.e. Cretaceous and Cenozoic) normal faulting, which involved the reactivation of syn-rift faults. Palaeocene and Early Eocene sediments indicate an abrupt regression followed by regional subsidence from the Mid Eocene to Recent (McDonnell & Shannon 2001; Jones *et al.* 2001). This coincides with the reactivation of syn-rift basin-bounding faults during the Mid-Late Eocene times (Worthington & Walsh 2016).

Hemipelagic siliciclastic and calcareous sediments, interspersed with deep-water turbidite fan systems make up the predominant facies of this sediment-starved depositional environment. More details of the seismic stratigraphy and structural geology of the post-rift sequence (Fig. 2a) in the Porcupine Basin, relevant to the current paper, is given in Worthington & Walsh (2016).

Geophysical background

A detailed *P*-wave velocity tomographic image (Fig. 2b) of the shallow post-rift structure of the northern Porcupine Basin has been

derived using 10-km-long streamer data by Prada *et al.* (2018). This study demonstrated the application of quantitative seismic tomography of long-streamer controlled source data in providing insight into fluid and fault interaction within post-rift sedimentary systems. The results are particularly well constrained in three dimensions with the availability of a 3-D seismic data volume (blue polygon in Fig. 1), which samples the western basin-bounding fault zone and includes the seismic profile depicted in Fig. 2.

Prada *et al.* (2018) converted their seismic velocity model into density (Fig. 3a) using a velocity–density relationship derived from exploration wells in the area, providing estimates of the bulk sediment density (Fig. 3b). The vertical density and porosity variations, shown in Fig. 4, principally reflect the gravitational compaction of mud-prone sediments during burial. Lateral variations in velocity, density and porosity across the basin axis (Fig. 4) strongly correlate with a thicker Neogene and Quaternary post-rift sequence in the centre of the basin (Fig. 2).

These observations suggest fluid flow driven by differential compaction towards the margins of the basin with the western basin-bounding fault acting as a natural hydraulic barrier, leading to fluid pressure buildup in the turbidite sand sequences (Prada *et al.* 2018). In this paper, this interpretation is tested quantitatively, by calculating predictions of pore fluid pressure that can be compared to legacy exploration borehole fluid pressure data.

DENSITY AND SEISMIC VELOCITY

Since the 1950s, empirical relationships between seismic velocities and densities of sediments have been developed in the context of marine geophysical studies of sedimentary basins (Nafe & Drake 1957; Ludwig *et al.* 1970; Barton 1986) and hydrocarbon exploration (Gardner *et al.* 1974). A general review of these relationships with applications to seismic hazard assessment is given in Brocher (2005).

Empirical relationships between sediment densities (ρ_s) and compressional wave velocities (V_p) are usually cast in terms of polynomial functions of V_p . Examples are

$$\rho_s = 1.6612 V_p - 0.4721 V_p^2 + 0.0671 V_p^3 - 0.0043 V_p^4 + 0.000106 V_p^5, \quad (1)$$

—commonly known as the ‘Nafe–Drake relationship’ (Nafe & Drake 1957; Brocher 2005) and meant to be comprehensive; including unconsolidated sediment suspensions—and

$$\rho_s = 1.74 V_p^{\frac{1}{4}}, \quad (2)$$

—also commonly known as ‘Gardner’s rule’ (Gardner *et al.* 1974) and intended to apply to seismic velocities in boreholes, measured below the ‘mud line’, where a solid and coherent sediment grain framework is present.

With the availability of high resolution sonic velocity and electron density measurements from deep exploration wells, empirical relationships of this form can be formulated specifically using geophysical well-log data from any sedimentary basin. An example is from Hughes *et al.* (1998):

$$\rho_s = 0.295 + 1.337 V_p - 0.273 V_p^2 + 0.019 V_p^3. \quad (3)$$

This was obtained for the Shetland–Faeroe Basin, using exploration well-log data of this type. The same approach to predict sediment densities from seismic velocities is used in this study.

Figure 2. (a) Pre-stack time-migrated section showing the tectono-stratigraphic structure of the Porcupine Basin. Red lines are normal faults. Key horizons: C20, Early Miocene; C30, Late Eocene-Early Oligocene erosional surface; E2, Early-Mid Eocene; C40, base Cenozoic; BCU, Base Cretaceous Unconformity. The location of the two exploration wells is indicated as thick black lines. (b) Same seismic section combined with the tomographic model converted from depth to Two-Way-Time (TWT). Blue lines are iso-velocity contours of the V_p model (in km s^{-1}). Thick blue, red and orange arrows depict the location of the velocity–depth profiles shown in the lowermost right corner of the seismic section. See Prada *et al.* (2018) for details.

Application to the Porcupine Basin

To calculate sediment densities, the bulk density and sonic log measurements from exploration wells (Fig. 1) were used to build the following polynomial $V_p - \rho_s$ relationship:

$$\rho_s = 0.357 + 1.114 V_p - 0.182 V_p^2 + 0.010 V_p^3. \quad (4)$$

This relationship, which fits the data with a correlation coefficient $R \approx 0.86$, is valid for V_p between 1.8 and $\sim 6 \text{ km s}^{-1}$ (Fig. 3). See Prada *et al.* 2018, for more information on the methodology and model uncertainties. Moreover, the differences in the sediment densities that are obtained independently for eqs (3) and (4) are small (within 3 per cent). This provides supporting validation to the accuracy of the $V_p - \rho_s$ conversion, which is critical in the approach to fluid pressure estimation that is given later.

The 2-D density model resulting from the conversion of the V_p model using eq. (4) (Fig. 4a) shows increasing density with depth in the basin from 1800 to $\sim 2400 \text{ kg m}^{-3}$, while the base of the model displays densities of ~ 2400 to 2500 kg m^{-3} .

Porosities ϕ_s across the Porcupine Basin are derived from the calculated density variations using the following standard expression given in Wyllie *et al.* (1956, 1958):

$$\phi_s = \frac{\rho_{mx} - \rho_s}{\rho_{mx} - \rho_w}, \quad (5)$$

where ρ_{mx} is the matrix or grain density, ρ_w the density of the fluids occupying the pore space and ρ_s the bulk sediment density. The result of this calculation is shown in Fig. 4(b). Porosity measurements from drill-core samples in Well 1 and Well 2 (Fig. 1) compare favourably with predictions (see Fig. 4b and Prada *et al.* 2018, for details).

EQUATION OF SEDIMENT COMPACTION

To derive pressure estimations, it is assumed that the general equation for the compaction of sediment of lithological/mineral composition i at depth h is given by,

$$\frac{d(\rho_s - \rho_o)}{dh} = r_i \left(1 - \frac{\rho_s - \rho_o}{\rho_{mx} - \rho_o} \right), \quad (6)$$

where the variation of density with depth $d(\rho_s - \rho_o)/dh$ depends on the rate r_i at which the sediment with initial density ρ_o becomes compacted to bulk density ρ_s due to lithostatic (overburden) pressure, as it is buried to greater depths. As porosity is lost, sediment is gradually transformed into rock due to both mechanical compaction and diagenetic-chemical effects, until its density approaches that of constituent rock matrix minerals ρ_{mx} .

Assuming a constant compaction rate r_i at a point, the integration of this equation describes how density varies with depth for the lithology i at that point in space:

$$\rho_s = \rho_o + (\rho_{mx} - \rho_o) \left(1 - e^{-\frac{r_i(h-h_w)}{\rho_{mx} - \rho_o}} \right). \quad (7)$$

This relationship describes a curve in which the density ρ_s increases with depth h towards the limiting asymptotic value ρ_{mx} typical of metamorphic/igneous basement rock.

The value of the parameter r_i gives the rate at which these processes occur and is most easily related to eq. (7) by taking its log-transform:

$$|r_i| = \frac{\rho_{mx} - \rho_o}{h - h_w} \log_e \left(1 - \frac{\rho_s - \rho_o}{\rho_{mx} - \rho_o} \right). \quad (8)$$

Thus, the compaction rate r_i at any depth h can be determined directly from eq. (8) once the density variations ρ_s can be inferred

Figure 3. (a) Plot of sonic velocity and density borehole-log data from exploration wells presented in Prada *et al.* (2018), used to derive eq. (4). The blue band depicts the range of uncertainty of the polynomial relationship in eq. (4), attributed to the sparse distribution of samples. The red line is the polynomial relationship presented in Hughes *et al.* (1998). (b) Observed and predicted density variation in Well 1 using the $V_p - \rho_s$ calibration from Prada *et al.* (2018) (blue line), and from Hughes *et al.* (1998) (red line). The blue band depicts the degree of uncertainty derived from (a). (c) Corresponding observed and predicted densities for Well 2.

from seismic velocity measurements, using the empirical equations, described previously.

Empirical constraints

Eqs (6) and (7) imply a porosity–depth variation consistent with ‘Athys Law’ (Athys 1930). Expressing ρ_o and ρ_s in terms of porosities (ϕ_o , ϕ_s) and densities (ρ_w , ρ_{mx}) yields,

$$\rho_s = \rho_{mx}(1 - \phi_s) + \rho_w\phi_s \quad (9)$$

and

$$\rho_o = \rho_{mx}(1 - \phi_o) + \rho_w\phi_o. \quad (10)$$

Substituting eqs (9) and (10) into (7) implies

$$\phi_s = \phi_o e^{\frac{-r_i(h-h_w)}{(\rho_{mx}-\rho_o)}}, \quad (11)$$

which is ‘Athys Law’.

Over the depth range considered in this case (up to 3.5 km, see Fig. 3) eq. (7) describes the pathway by which density changes from ρ_o to ρ_{mx} for lithology i with increasing depth. An individual sediment lithology, at any point in space, may be characterised by a distinct (ρ_{mx} , ρ_o) couplet and a diagnostic (ρ_{mx} , ρ_o , r_i) triplet that determines its response to gravitational compaction.

The values of these parameters, which are widely used in studies of sedimentary basin subsidence (Steckler & Watts 1978; Sclater & Christie 1980; Kominz *et al.* 2016) may be highly correlated for a given sediment particle size and composition. There is however limited observational data to substantiate this or to resolve possible patterns (Kominz *et al.* 2011). Table 1 lists the parameter definitions used throughout this paper with their physical dimensions, while Fig. 5 defines these in space.

FLUID PRESSURE VARIATIONS

In the simple case of a gravitationally compacting sedimentary column that is in ideal hydrostatic equilibrium, the pore fluid pressure P_f of the sediments at depth h equals the hydrostatic pressure P_h and is defined as:

$$P_f = P_h = \rho_w g h, \quad (12)$$

with $g = 9.81 \text{ m s}^{-2}$ the gravitational acceleration and ρ_w the fluid density in the sediment pore spaces ($\rho_w \approx 1030 \text{ kg m}^{-3}$ for seawater).

The corresponding lithostatic pressure at depth h is given by the integral of the density distribution over the depth from the sea surface into the sedimentary column:

$$P_l = \rho_w g h_w + \int_{h_w}^h \rho_s g dh, \quad (13)$$

ρ_s is defined by eq. (7) and h_w is the ocean depth (Fig. 4). Substituting eqs (7) into (13) and assuming a constant rate $r_i = r_{amb}$ throughout the sedimentary column yields the lithostatic pressure for a sedimentary sequence in hydrostatic equilibrium:

$$P_l = \rho_w g h_w + \rho_{mx} g (h - h_w) - \frac{(\rho_{mx} - \rho_o)g}{R_{amb}} (1 - e^{-R_{amb}(h-h_w)}), \quad (14)$$

where $R_{amb} = r_{amb}/(\rho_{mx} - \rho_o)$ is the ambient porosity decay rate [equivalent to the compaction parameter (c) as defined in Sclater & Christie (1980)] for a column in perfect equilibrium (see Table 1 for parameter definitions). The equilibrium state (i.e. $R_i = R_{amb}$ everywhere in a hypothetical simple system) is reached when fluid pressure gradients decay, such that the fluid flux approaches zero over long geological timescales.

The non-equilibrium state

Where a sedimentary column is not in a state of ideal hydraulic equilibrium at any depth, the fluid pressure P_f must lie somewhere between P_h and P_l . The variation in P_f is computed from the spatial variations in the ‘observed’ values of R_i (Fig. 5) with respect to the equilibrium hydrostatic value (R_{amb}). The equilibrium value may show some dependence on sediment lithology: but this is unclear when considering the published compilations of the key parameters involved (Kominz *et al.* 2011, and Table 2).

In this case the fluid pressure (for the non-equilibrium state) with respect to the hydrostatic pressure at depth h is given by:

$$P_f = \rho_w g h + (h - h_w)(\rho_{sav} - \rho_w)g \left(\frac{e^{R_{amb}(h-h_w)} - e^{R_i(h-h_w)}}{e^{(R_{amb}+R_i)(h-h_w)}} \right), \quad (15)$$

Figure 4. (a) Bulk 2-D density model derived from the tomographic V_p model (Fig. 2) using eq. (4). (b) Fractional porosity across the Porcupine Basin derived from the bulk density model using eq. (5). Modified from Prada *et al.* (2018).

Table 1. List of parameter definitions with their SI units.

• h_w – ocean water depth (m)
• h – depth below sea surface (m)
• g – gravitational acceleration (9.81 m s^{-2})
• ρ_w – fluid density (kg m^{-3})
• ρ_{mx} – solid grain density (kg m^{-3})
• ρ_o – initial sediment density (kg m^{-3})
• ρ_s – sediment density at depth h (kg m^{-3})
• ϕ_o – initial sediment porosity (dimensionless) • ϕ_s – sediment porosity at depth (h) (dimensionless)
• r_i – sediment density decay rate of lithology (i) (kg m^{-4})
• r_{amb} – ambient equilibrium sediment density decay rate of lithology i (kg m^{-4}) – ($r_{amb} \geq r_i$)
• R_i – sediment porosity decay rate of lithology (i) (m^{-1}) – $R_i = \frac{r_i}{(\rho_{mx} - \rho_o)}$ • R_{amb} – ambient equilibrium sediment porosity decay rate of lithology i (m^{-1}) – $R_{amb} = \frac{r_{amb}}{(\rho_{mx} - \rho_o)}$ – ($R_{amb} \geq R_i$)
• P_h – hydrostatic pressure MPa ($\text{kg m}^{-1} \text{s}^{-1}$)
• P_l – lithostatic pressure MPa ($\text{kg m}^{-1} \text{s}^{-1}$)
• P_f – fluid pressure MPa ($\text{kg m}^{-1} \text{s}^{-1}$)

where the average density ρ_{sav} is determined by integration of eq. (7):

$$\rho_{sav} = \rho_{mx} - (\rho_{mx} - \rho_o) \frac{1 - e^{-R_{amb}(h-h_w)}}{R_{amb}(h-h_w)}. \quad (16)$$

Eq. (15) is fundamental to the fluid pressure calculations presented here.

It implies that where R_i approaches R_{amb} the fluid pressure approaches the hydrostatic value ($P_f \rightarrow P_h = \rho_w gh$). The condition $R_i \rightarrow 0$ as $h \rightarrow \infty$ becomes physically untenable, since the pore fluid overpressure eventually exceeds the strength of the compacting rock; leading to mechanical instability and fracturing, causing fluid pressure dissipation and a return to the equilibrium state.

The second term in eq. (15) gives the ‘kick’ in the hydrostatic pressure defined by the increase ΔP_f associated with the change $\Delta R = R_{amb} - R_i$, at depth h . Regions within the sedimentary basin where the sediments are significantly overpressured will be defined by low values of R_i .

THE APPLICATION OF THE METHOD

The application of the theory and method, given in the previous sections, first involves the use of eq. (8) for determining the distribution of R_i values across the tomographic models of physical parameters, derived for the Porcupine Basin (Figs 2 and 4). In order to apply this equation, ρ_{mx} and ρ_o are set equal to 2710 and 1710 kg m^{-3} , respectively, so that $(\rho_{mx} - \rho_o) = 1000 \text{ kg m}^{-3}$. This is in agreement with global observations from deep-sea mud rich sediments (see next section). The result of this calculation is given in Fig. 6.

Beneath the first 500 m of sediments, the value R_i shows little variation near the basin centre, suggesting that the sediments are near a state of hydrostatic equilibrium in this region; a state that is consistent with low rates of sedimentation during Neogene times (above the Early to Mid Eocene surface (C30) in Fig. 2). The lowest values of R_i are restricted to the eastern and western margin of the basin (at 95–110 and 160–180 km distance; Fig. 6), where it has been proposed that Recent Cenozoic faulting is related to overpressured zones (Prada *et al.* 2018).

Figure 5. Simple 1-D sketch showing the parameter definitions used in the physical model for gravitational compaction and hydrostatic pressure variation. h_w – ocean water depth, h – depth measured from ocean surface, ρ_w – seawater density, ρ_o – initial sediment density (density of seabed sediments), ρ_s – sediment density at any depth h , ρ_{mx} – sediment grain density (asymptotic density limit).

Defining the equilibrium value (R_{amb})

The ambient value R_{amb} is defined by the depth distribution of those R_i values within the centre of the basin (between 120 and 150 km of distance), where no observable normal faulting influences the tomographic results (Fig. 2) and sediments are likely near to a state of hydrostatic equilibrium. The depth distribution of R_i in Fig. 7 depicts an increase of R_i with depth down to 1.25 km beneath the seabed, where the value stabilises at a maximum value close to $0.60 \times 10^{-3} \text{ m}^{-1}$.

At near ocean floor depths (0–400 m), $R_{amb} < 0.4 \times 10^{-3} \text{ m}^{-1}$. These low apparent values may relate to a mechanical coupling between the formidable ocean-storm swell and the sediments on the ocean floor, which is related to the very frequent North Atlantic Ocean storms and hurricanes. This aspect is beyond the scope of the current paper; but this possibility will be considered elsewhere.

Based on these considerations and upon the results given in Prada *et al.* (2018), we estimate $R_{amb} = 0.60 \pm 0.02 \times 10^{-3} \text{ m}^{-1}$, with $\rho_{mx} = 2710 \text{ kg m}^{-3}$ and $\rho_o = 1710 \text{ kg m}^{-3}$. While ρ_o agrees with a recent global compilation of bulk density measurements of deep-water seafloor sediments from Deep Sea Drilling Project (DSDP) boreholes (Tenzer & Gladkikh 2014); ρ_{mx} is consistent with the abundance of mud-prone sediments in the Porcupine Basin (McDonnell & Shannon 2001) and also with the available petrophysical data (Prada *et al.* 2018) from the Porcupine Basin.

An important compilation from Ocean Drilling Project (ODP) data for the decay rates of surface porosity values from mud dominated clay sized clastic sediments between 175 and 1800 m burial depth (Kominz *et al.* 2011, 2016), independently yields $R_i = 0.598 \times 10^{-3} \text{ m}^{-1}$. This value lies exactly within our statistically determined value (Table 2 and Fig. 7). It is also close to the value obtained for a subset of ‘mixed data’ over the same depth interval ($R_i = 0.675 \times 10^{-3} \text{ m}^{-1}$); but this estimate comprises a larger

sample number, with more varied lithological composition (see Table 2).

PREDICTED FLUID PRESSURE VARIATIONS

The predicted fluid pressure (P_f) variations across the Porcupine Basin are shown in Fig. 8, both as absolute pressures (Fig. 8a) and the overpressure component (Fig. 8b). Significant overpressure magnitudes on the western margin of the basin (up to 5 MPa) corresponds to the low-velocity zone found by Prada *et al.* (2018).

Down-hole pressure measurements for the deep exploration boreholes (Wells 1 and 2 in Fig. 1) permit predictions to be tested against observations at two localities along the pressure model. Pressures were calculated from drilling mud specific gravity measurements obtained from well completion reports (yellow dots in Fig. 9). This information was provided by the Petroleum Affairs Division of the Department of Communications, Climate Action and Environment of Ireland; www.pad.gov.ie.

Measured pressure variations with depth in the exploration boreholes are generally predicted by the theoretical ones. For example, the value obtained at 1350 m is in very good agreement with the predicted value (Fig. 9a). A broad overpressured zone between 600 and 1500 m also corresponds closely to the stratigraphic interval, composed of Early to Mid Eocene turbidite sands, interbedded with deep-water mudstones.

These predictions support a fluid flow model for the Porcupine Basin, based on the earlier seismic observations (see fig. 12 in Prada *et al.* 2018) that suggested higher overpressured regions along the basin margins, related to gravitational compaction. The quantitative predictions of the current study confirms this earlier geological interpretation.

DISCUSSION

According to eq. (15) fluid overpressure depends solely on the decrease in the compaction factors (R_i) relative to an ideal ambient equilibrium value (R_{amb}) that is characteristic of lithology (i). The measured pressures encountered in the two exploration boreholes are in agreement with the modest predicted pressures (≈ 5 MPa): the highest overpressured region occurring along the margins of the Porcupine Basin (Fig. 8), associated with normal fault re-activation (Prada *et al.* 2018). Across the central region of the basin the fluid overpressure is low; consistent with gravitationally driven fluid flow towards the basin margins along high permeability sand-rich turbidite layers (Worthington & Walsh 2016).

Calibration and independent measures

The equilibrium value of this parameter (R_{amb}) derived with the method presented (Fig. 7) is identical to measurements from lithologically similar deep-sea sediments from ODP boreholes (Kominz *et al.* 2011). These sediment samples (Table 2) come from the same depth range as the resolution depth of the tomographic model (Fig. 2) and the parameters have been directly measured from sediments of similar sediment particle size and composition to those encountered in the exploration wells.

These observations suggest that the methodology used in this work measures R_{amb} on a basin-wide scale, where significant variation in bulk sediment properties (either inherent or due to chemical/diagenetic effects) is not important in defining it. This parameter,

Table 2. Sediment compaction parameters.

Sediment particle size	ϕ_o (initial porosity)	R_i (decay rate)	N
Clay/mud (175 to $\approx$ 1800 m)	0.614	0.598	71
mixed data (150 to $\approx$ 1800 m)	0.614	0.675	2463

Compaction parameters for clay and mixed sediment particle sizes measured and derived by Kominz *et al.* (2011) from a comprehensive compilation of ODP data. Data taken from Tables 3 and 4 in this publication with sample depths indicated. Depth range overlaps the resolution depth of the tomographic imaging technique used in this study (see Fig. 2 and Prada *et al.* 2018). N is the number of samples used to determine these parameters. The ‘mixed data’ subset of samples comprise the entire range of biogenic and siliciclastic sediments. See Kominz *et al.* (2011) and Kominz *et al.* (2016) for details on statistics. Units: $R_i \times 10^3 \text{ m}^{-1}$.

Figure 6. Variation in the calculated value of R_i across the Porcupine Basin derived from eq. (8). Note, that R_i shows less variation across the basin centre suggesting that the sediments are in a state near to hydrostatic equilibrium. Units: $R_i \times 10^3 \text{ m}^{-1}$.

Figure 7. 1-D depth distributions of R_i values across the centre (120–150 km of distance) of the Porcupine Basin (red profiles). Black line and error bars correspond to the average of the distribution and its standard deviation at each depth. The solid vertical line corresponds to $R_{amb} = 0.6$, used in the overpressure calculations. Units: $R_{amb} \times 10^3 \text{ m}^{-1}$.

which is equivalent to the compaction parameter (c)—as defined by Sclater & Christie (1980)—is essential in reconstructions of sedimentary basin subsidence; particularly in passive continental margin settings in the North Atlantic region (e.g. Kominz *et al.* 2016), where some of the earliest such studies were undertaken (Steckler & Watts 1978).

A key step in the pressure calculation (eq. 15) is the conversion of seismic P -wave velocities to bulk sediment densities. The relationship used to obtain this calibration (eq. 4) comes from the same exploration wells used to test the predicted fluid pressures. This means that the $V_p - \rho_s$ calibration and the measured pressures may not strictly be statistically independent.

However, eq. (3) was obtained independently by Hughes *et al.* (1998) from the Shetland-Faeroe Basin, where the geological setting and starved Cenozoic depositional environment are very similar to the Porcupine Basin. This equation yields an indistinguishable density distribution (Fig. 3) and hence leads to very similar predicted pressures. In this respect the calibration issue is not very significant in influencing the results presented in this study.

The role of chemical processes

Physical and chemical processes that accompany mechanical compaction are not explicitly included in the analysis: rather R_i is implicitly regarded as a bulk parameter that includes these correlated processes over the depth range investigated (Fig. 2). The importance of these processes vary with chemical composition, depth of burial, temperature and fluid flow regime (Marín-Moreno *et al.* 2013b; Colombo *et al.* 2017). For example, the breakdown of feldspar into clay minerals is likely to be important in arkosic quartz-sand sediments and is sensitive to temperature and fluid flow. Diagenetic (or

Figure 8. (a) Predicted fluid pressure (P_f) variations across the Porcupine Basin. (b) Predicted overpressure ($\Delta P_f = P_f - P_h$). Pressure variation is calculated from the values of R_i across the basin depicted in Fig. 6, using eq. (15).

Figure 9. Measured versus predicted pressure variations in Wells 1(a) and 2(b). Blue and red lines indicate hydrostatic and lithostatic trends, respectively. Yellow dots represent downhole pressures calculated from drilling mud specific gravity measurements. Black lines indicate predicted pressures from eq. (15). Red shaded region represents the total error resulting from the uncertainty in the tomographic model of Prada *et al.* (2018). The lithological column for Well 1 depicts a clear depth correlation between overpressured areas and Early-Mid Eocene sandstones (GR stands for natural gamma ray log).

low grade metamorphic) changes in the polycyclic clay minerals of mud-prone sediments in deep-water oceanic basins (Goldstein *et al.* 1984) are controlled by the same factors.

In contrast, sand-prone sediments have very different mechanical properties and undergo pressure solution with burial and increasing vertical stresses. Pressure solution is a chemical process involving the diffusion of quartz into the interstitial pore spaces between sand grains. For pure, well sorted, sands the few measurements indicate

the $R_{amb} \approx 0-0.2 \times 10^{-3} \text{ m}^{-1}$ over shallow depth ranges ($\leq 2 \text{ km}$) with the initial sediment fractional framework porosities of ≈ 0.4 , consistent with grain framework structure (Kominz *et al.* 2011, and references therein).

The limited available compilations appear to suggest that sediments comprised of predominantly clay-sized particles have R_{amb} values close to $0.6 \times 10^{-3} \text{ m}^{-1}$ (Kominz *et al.* 2011 and Table 2) over the depth, encountered in this study. This supports the notion that

physico-chemical effects are of secondary importance in mud-prone sediment mixtures; that R_{amb} may be considered as a bulk statistical parameter and also that the change $\Delta R = R_{amb} - R_i$ is principally due to fluid overpressure variation, caused by non-equilibrium compaction. The same is unlikely to be true at greater depths; as $\rho_s \rightarrow \rho_{mx}$ (eq. 7) and chemical/diagenetic processes, driven by higher pressures and temperature, gradually become significant.

Numerical models that involve a complete inclusion and description of chemical and physical phenomena, such as coupled thermal and fluid flow processes have been developed and calibrated with exploration well and seismic data (Marín-Moreno *et al.*, 2013b; Colombo *et al.* 2017). These models use empirical methods applied to wireline-logs such as Eaton's Method (Eaton 1975; van Ruth *et al.* 2004; Ramdhan & Goult 2018) to test model predictions, using advanced geophysical inversion techniques.

It would be interesting to apply these more sophisticated methods to test and compare with the simpler and more regional approach of the current study, which uses limited legacy exploration well data from the 1970s and 1980s. This requires a more rigorous study and analysis of specific, recently acquired, deep exploration well data from the Porcupine Basin. Such a study should include comprehensive pressure, temperature and petrological data, which are also colocated along the more recent long-streamer MCS lines in the basin (see www.pad.gov.ie).

CONCLUSIONS

(1) The agreement between the predicted and the measured fluid pressures at different depths is good; and the modest (≈ 5 MPa) elevated zones of overpressure are resolved in this case, along the marginal regions of the Porcupine Basin. Overpressure in the central region of the basin is lower, suggesting gravitationally induced flow towards the margins, as inferred in previous studies.

(2) A very high degree of internal consistency exists in this relatively simple method to measure overpressure. The method is founded on similar sediment compaction concepts as those used in more sophisticated approaches. Empirically derived bulk density changes in sediments, which are central to the method are similar to those measured from sedimentary basins with a similar geological environment.

(3) The predicted value of the equilibrium compaction factor R_{amb} agrees with independent measures from deep-sea sediments, compiled from ODP archives. This implies that it may be a gross diagnostic of bulk lithology.

(4) Further work is required to refine and validate the simple regional approach of this study, using recent deep borehole exploration and MCS data gathered in the Porcupine Basin. This would permit the inclusion of thermal, chemical and fluid flow processes and the usage of modern geophysical inversion techniques.

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