Spatial and temporal changes in saltwater wedges in coastal karst aquifers are still poorly understood, largely due to complex mixing processes in these heterogeneous environments, but also due to anthropogenic forcing such as pumping, which commonly affect natural variations in wedges. The purpose of this study was first to characterize the hydrodynamic functioning of a karst aquifer in an oceanic temperate climate with little anthropogenic pressure but strongly influenced by a high tidal range and second, to evaluate the extent and movements of a saltwater wedge influenced by both the tide and the natural recharge of the aquifer. Variations in specific conductivity combined with water chemistry results from six boreholes and two lakes located in the Bell Harbour catchment (western Ireland) enabled us to assess the extent of the intrusion of the saltwater wedge into the aquifer as a function of both karst recharge and tidal movements at high/low and neap/spring tidal cycles. The marked spatial disparity of the saltwater wedge was analysed as a function of both the hydrodynamic and the structural properties of the karst aquifer. Results showed that the extent of the saltwater wedge depended not only on the intrinsic properties of the aquifer but also on the relative influence of the recharge and the tide on groundwater levels, which have opposite effects. Recharge in the Burren area throughout the year is large enough to prevent saltwater intruding more than about one kilometre from the shore. A strong tidal amplitude seems to be the motor of sudden saltwater intrusion observed in the aquifer near the shore while the position of the groundwater level seems to influence the intensity of the salinity increase. Competition between recharge and the tide thus controls the seawater inputs, hence explaining temporal and spatial changes in the saltwater wedge in this coastal karst aquifer.

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1. Introduction

Understanding the interaction between freshwater and seawater in coastal karst aquifers is critical for current and future water management, especially in the context of various EU Directives (Water Framework Directive, Floods Directive, Marine Strategy Framework Directive) and given the potential impacts of changing weather patterns and sea levels (Ferguson and Gleeson, 2012). The natural equilibrium between seawater and freshwater is still poorly understood, largely due to the complex mixing processes in these heterogeneous environments (Dörfliger and Willwer, 2005). The presence of karst conduits that allow discharge of freshwater and ingress of sea water make it difficult to predict spatial and temporal changes in the saltwater wedge (Bakalowicz, 2005). In these aquifers, the extent of the saltwater wedge thus depends on the flow path network, resulting in irregular patterns of saltwater intrusion into the aquifer (Arfib et al., 2002). Previous work on saltwater wedges in karst aquifers show the effect of areas of high hydraulic conductivity on the distribution of the mixing zone (Moore et al., 1982; Stringfield and LeGrand, 1971). The morphology and geometry of conduits have been shown to be first order controls on groundwater temperature and specific conductivity (SpC) gradients in the aquifer (Beddows et al., 2007). Knowledge of the spatial variations in the hydrodynamic properties of the aquifer is therefore required to understand the extent of the saltwater wedge. Hydraulic diffusivities can be estimated through groundwater responses to tides. Jacob (1950) and Ferris and Branch (1952) provided analytical solutions to assess hydraulic diffusivity from changes in groundwater levels, while considering a one-dimensional (1D), homogeneous, isotropic, confined, and semi-infinite aquifer with a sharp boundary condition subject to oscillating forcing. In this study, analyses of water table fluctuations in boreholes highlighted three types of hydrodynamic response that were confirmed with the Jacob–Ferris method.

Previous studies showed that saltwater intrusion is driven by the hydraulic gradient, which in turn, is controlled by the difference in hydraulic heads between the sea and the aquifer (Arfib et al., 2002; Blavoux et al., 2004; Bonacci and Roje-Bonacci, 1997; Cooper, 1959; Henry, 1959; Herzberg, 1901; Hubbert, 1940). The hydraulic gradient depends on (i) tidal fluctuations, (ii) natural recharge of the aquifer and (iii) spatial heterogeneities of its hydrodynamic properties. Tidal fluctuations may be a major factor in determining variations in boundary conditions. For example, Beddows et al. (2007) showed that variation in pressure heads driven by the tide can increase the rate of mixing. Martin et al. (2012) quantified the exchanges between the matrix and conduits as a function of tidal amplitude. However to date, no study has explained the spatial and temporal changes in the saltwater wedge in a karst aquifer taking into account both water chemistry (major cation/anion analysis) and tidal influence (at high/low tide and neap/spring tide scales) on variations in SpC in an aquifer with very low anthropogenic forcing. Natural recharge of the aquifer may also be a major factor driving variations in inland hydraulic heads: the extension of the saltwater wedge over time could depend on the equilibrium between groundwater recharge and the influence of tides that modify the boundary conditions. While a lot of published studies (Arfib et al., 2002; Bayari et al., 2010; Fleury et al., 2007; Maramathas et al., 2003; Tulipano, 2005) focus on coastal karst aquifers in arid to semi-arid areas such as the Mediterranean region, this study focuses on the Bell Harbour karst catchment, which is characterised by a strong tidal range, a consistently high recharge throughout the year, and a low population density in a rural agricultural setting. Because of the low water demand in the Burren area, this zone is very well suited to study the saltwater intrusion mechanisms in a close to natural context (little anthropogenic forcing). The Intergovernmental Panel on Climate Change (IPCC) forecast global sea level rise of up to 1 m by 2100, and this will directly affect the hydrogeological behaviour of the karst aquifer. These changes are examined at the light of the presented results.

The aim of this study was to test the hypothesis that recharge and the tide control spatial and temporal changes in the saltwater intrusion. To this end, the intrinsic properties of the karst aquifer (hydrodynamic properties, structure) were characterised, and then an attempt was made to link these properties and spatial and tem-
poral changes in the saltwater intrusion into the karst system. The relative influence of recharge and of the tide on the water level (WL) and variations in SpC measured in six boreholes and two lakes is discussed and a general conceptual model explaining the extent and movements of the saltwater wedge in a karst aquifer little affected by anthropogenic pressure is proposed.

2. Study area

2.1. Location and climatic conditions

The Bell Harbour catchment, covering an area of around 50 km² in County Clare, is located in a large karst area called the Burren on
the west coast of Ireland and is defined by a valley surrounded by upland areas to the west, south and east that reach an altitude of about 300 m. Discharge from the catchment is to the north into Bell Harbour, which opens onto Galway Bay (Fig. 1). The Burren has an oceanic climate, with annual precipitation averaging 1500 mm and annual effective rainfall estimated at 980 mm (Walsh, 2012). The rainfall occurs throughout the year, although the spring and early summer months tend to be drier.

2.2. Geology

The Burren region is one of the most extensive limestone karst areas in north-western Europe (600 km$^2$) (Gallagher et al., 2006) dominated by massive or bedded Carboniferous limestone of several hundred metres thickness and bounded to the east by limestones of the Gort Lowlands and to the south by Namurian sandstones and shales (Fig. 1b). In the Bell Harbour catchment, gently dipping (2–3° to the south) pure-bedded limestones of the Burren Formation underlie the entire area. These rocks are characterised by pale grey and thickly to massively bedded limestone with occasional cherty intervals of shale and dolomite horizons (Pracht et al., 2004). These limestones are underlain by impure limestones (which are not expressed at the surface) (~400 m in thickness), which overlie Devonian Old Red Sandstones and the Galway Granite pluton (Fig. 2). In the Burren, faults can be identified both on the surface (Simms, 2001) or underground (Judd and Mullan, 1994): only two major faults are mapped in the area, one of which is in the western part of the Bell Harbour catchment study area (Fig. 1c). This fault, known as MacDermott’s Fault, runs approximately north-south and shows a slight sinistral displacement of members of the Burren Formation. Pracht et al. (2004) assume the apparent displacement on the fault is minimal, less than 200 m. Geophysical investigation of this fault is currently being completed, using electrical resistivity tomography, as part of this ongoing research. Initial indications suggest that the fault extends beyond its current mapped extent running under Bell Harbour bay (O’Connell, 2012, personal communication). Joints and veins are extensive throughout the area (Gillespie et al., 2001).

2.3. Hydrogeology

The landscape in the Burren region (and in the study area) is among the best examples of karst landscape in Europe and is the finest example of karst terrain in Ireland (Drew, 2001). The pure bedded limestones are exposed to weathering and are susceptible to dissolution, leading to the formation of a well-developed shallow epikarst of 5–10 m thickness, and distinctive karst topographic features such as swallow holes, sinking streams, limestone pavements, caves and large springs. There are three main groundwater flow systems operating in this region (from the shallowest to the deepest): (1) a strongly altered, rapidly draining shallow epikarstic system present in the upper 5–10 m; (2) an unsaturated (or infiltration) zone where vertical flows pass through the rock via system of fissures or joints; (3) a phreatic zone where three distinct compartments are linked together: conduits which allow rapid groundwater flows, fissures which allow relatively rapid flows, and the limestone matrix which allows slow movement of groundwater. Flow paths through the unsaturated zone may also be along similar flow paths.

Diffuse recharge dominates in the Bell Harbour catchment across the large area of exposed limestone pavement in the southern (higher elevation) portion of the catchment. The main flow path of the water in the limestone is controlled by (i) the horizontal bedding partings (Fig. 2) as the impermeable shale layers or chert lenses limit the vertical movement of the groundwater (Drew, 2003), and by (ii) the hydraulic head boundary conditions (groundwater basin divide to the south, and the boundary imposed by the tide in Bell Harbour). Water drains almost wholly by underground conduits directly into Bell Harbour via submarine or littoral diffuse springs (Fig. 2). Groundwater flow rates in the range 50–150 m/h have been recorded from water tracer tests under base flow conditions in the Burren, and flow velocities are assumed to increase up to four fold in flood conditions and to halve under very low flow conditions (Drew and Daly, 1993). Aquifer storage is low and during extended wet periods the conduit systems can back up, leading to the development of two temporary lakes or turloughs (Luirk and Gortboyheen lakes), which either drain or feed the karst.
aquifer, and may persist on the landscape for weeks or even months (Fig. 1c).

2.4. Recent hydrogeological results

At the start of this study, field investigations were undertaken to locate karst features and to assess and select measurement locations (boreholes, springs etc.). Six private boreholes were selected for the study. Five intertidal springs were identified along the eastern shore of Bell Harbour and there are two turloughs and one permanent lake (Muckinish Lake) in the catchment (Fig. 1). Discrete SpC measurements at the five intertidal springs were done at low tide (LT), and ranged from 40 mS/cm to 400 mS/cm depending on the tidal stage and the hydrological period. A LIDAR map of the area showed circular features on the bed of the Bay which are aligned with MacDermott’s fault. These were assumed to be either submarine springs discharging fresh or brackish water, or swallow holes allowing seawater inflow into the aquifer (Fig. 1c). These features have not previously been described; a salinity survey of the bay performed during this project, as well as continuous SpC data collected at different depths above one of these circular features, suggest that brackish water outflow from the aquifer only occurs during large recharge events. Brackish water outflows and saltwater inflows may therefore alternate, depending on the hydraulic gradient within the aquifer, similar to the karst system of the Betic Cordilleras in Southern Spain (Fleury et al., 2008).

2.5. Field sampling and data available

This project is the first detailed groundwater study to be undertaken in this catchment. Temperature, SpC and water level (WL) were recorded in six boreholes (B03, B05, B08, B15, B57 and B59) and two lakes (Muckinish lake, L01 and Luirk lake, L02) at 15-min intervals from September 2010 to September 2011 and at five-minute intervals from December 2011 to January 2013 (Fig. 1c). Data were not recorded at every measurement point for the full periods. In Situ Aqua TROLL 200 loggers were installed in the six boreholes and CTD-Divers were sited at two lakes. The In Situ loggers were connected to a vented cable to account for atmospheric pressure. A Baro-Diver installed adjacent to borehole B05 allowed for compensation of the pressure measurements from the CTD-Divers. An RTK (Real Time Kinematic) GPS survey of all boreholes and collection points was undertaken to relate all WL data collected to mean sea level (MSL). In addition, vertical profiles of salinity and temperature were recorded on June 6, 2012 in four boreholes: B03, B05, B08 and B57 during a rainy period. Data on tide height were provided by the Marine Institute and were recorded at Galway Port station at a six minute time interval. This data set was then converted into a five minute step by linear interpolation to enable direct comparison with data collected in the boreholes and lakes. Continuous measurements of the WL taken over a five month period at an intertidal spring in Bell Harbour allowed estimation of a tidal time lag of 1 h in the bay (which was used in this study), compared to the tide height recorded at Galway Port station. Rainfall data were measured at National University of Ireland, Galway (NUIG) using a daily read rain gauge: readings are totals for the 24-h period from 9 am to 9 am the next day. The hydraulic gradient was calculated to estimate the potential outflow at each measurement point, distinguishing the load variation in the aquifer from the load variation at the limit (tidal variations in the bay). Tidal height was subtracted from the WL and the result divided by the linear distance between the measurement point and the shore. An example of the data from B05 is shown in Fig. 3. Additionally, discrete water chemical measurements were performed: water samples were collected for major cation/anion analysis from four boreholes (B05, B08, B15, B57).
and two of the turloughs (L01, L02) on March 21, 2011 during a relatively dry period. Before raw groundwater samples were collected from existing water taps connected directly to the pumps in boreholes, field parameters were monitored to indicate geochemical stabilization (\(\Delta t\) = 10 min).

3. Assessment of karst hydrodynamic properties

WL variations were analysed first in response to precipitation and second in response to tidal fluctuations to better understand the hydrodynamic behaviour of boreholes and the overall system. However, even though the analysis of the functioning of the aquifer using recharge events implies consideration of direct infiltration through sinkhole and the pathway between the surface and the borehole, most of the analysis relies on the interpretation of WL recession curves, that immediately follow the recharge event. These WL recession curves are driven by both the hydrodynamic properties and the connectivity of the saturated zone, allowing comparison with the analysis of WL variations in boreholes due to the effect of tide.

Moreover, these two methods were used during different periods: (i) during “high flow periods”, when the main process driving WL variations is recharge as the tidal variations have little effect on WL; (ii) during “low flow periods” when the main process driving WL variations is the tide as there is no recharge. These two different methods were used to ascertain the coherency between the various types of hydrodynamic environment estimated from hydrodynamic responses to recharge events and the different hydraulic diffusivities assessed using the Jacob–Ferris equation for each borehole.

3.1. Hydrodynamic response to recharge events

Observations of changes in the WL in each of the six boreholes during and after a recharge event enabled determination of the typical hydrodynamic behaviour of each borehole using the intensity of the variation in WL (amplitude and period of influence \((P)\)) and the slope \((S)\) on the WL recession curve. The steepness of the slope was assumed to reflect the drainage of the karst system (Powers and Shevenell, 2000; Shevenell, 1996): a steep slope represents the dominant effects of the larger karst features (i.e. conduit/large fissures) on drainage, but also includes the effects of the other regimes. A break and decrease in the slope illustrates the storage of the aquifer that results in the emptying of well-connected karstified fissures and/or the matrix. Three segments in a borehole recession curve in a multiple porosity karst system can thus be interpreted as the consequence of three types of flow occurring in: (i) conduits/large fissures; (ii) fissures; and (iii) the matrix. One example of a typical hydrograph recession curve was selected for each of the six boreholes and classification was based on the range of responses starting from the borehole with the highest amplitude and steepest slope (when several segments were present, only the steepest one was used for the classification) (D1) to the borehole with the lowest amplitude and flattest slope (D6) (Fig. 4).
the boreholes but not to the lakes (L01, L02). (Changes in the WL of the lakes are strongly dampened by the capacity of the lake and consequently cannot be compared to changes in WL in the boreholes). For example, the hydrodynamic response of B57 exhibits a high amplitude peak (>10 m) and a steep recession curve \(S_y = 6.3 \text{ m/d} \) (Fig. 4). Moreover, its period of influence was short \((P_T = 1.6 \text{ days})\). This indicates a large value of hydraulic diffusivity \((D_1)\) and hence high connectivity of this borehole with the karst drainage network, due to well-connected fissures.

### 3.2. Hydrodynamic response to tidal variations

The Jacob–Ferris method (Ferris and Branch, 1952; Jacob, 1950) for tidal propagation through a coastal aquifer was used to estimate hydraulic diffusivity \((D)\) values, although a variety of more sophisticated analytical solutions exist for one and two-dimensional systems (Li et al., 2000; Townley, 1995; Trefry, 1999) and for vertical section systems (Guo et al., 2010; Li and Jiao, 2002). The relatively simple Jacob–Ferris equation remains useful for estimating aquifer hydraulic properties from measured groundwater level variations (Rotzoll et al., 2013), even though problems can arise from the application of this simple homogeneous tidal propagation models. Differences in hydraulic diffusivities calculated from either tidal efficiency values \((D_e/\text{ratio of amplitude between the tidal level in the bay and the WL data collected at the monitoring point})\) or delay values \((D_d/\text{time lag between the tidal level in the bay and the WL data collected at the monitoring point})\) have been noted in numerous studies (Drogue et al., 1984; Erskine, 1991; Ferris and Branch, 1952; Smith, 1999). Normally, for a confined homogenous aquifer where head losses are isotropic, \(D_d/D_e\) value should be 1 (Ferris and Branch, 1952). These inconsistencies of observed hydraulic diffusivities (tidal efficiency-delay inconsistencies) have only been partially explained because of the various influences at play: vertical flows, phreatic influences (Jha et al., 2003), geometric effects (e.g. variable aquifer thickness), non linearity associated with capillarity and density-driven flow and spatial heterogeneity (e.g. horizontal layering) (Trefry and Bekele, 2004). These tidal efficiency-delay inconsistencies were also identified in this study but the assessed hydraulic diffusivities are in agreement with the spatial distribution of the hydraulic diffusivity field at the catchment scale.

The estimated values correspond to the hydraulic diffusivity of an equivalent porous medium between the sea and the borehole, which may be representative of either one of these hydrodynamic environment (conduit, fracture and matrix), or more likely a combination of these environments. Use of the Jacob–Ferris equation generally assumes a confined aquifer with homogeneous hydrodynamic properties and considered the propagation along the aquifer of sinusoidal oscillations of pressure in one-dimensional flow (implying neither vertical nor flow parallel to the shoreline). However, we can assume that some karst features (especially conduits), where pressure variation propagates between the bay and the aquifer, are confined; for unconfined karst features the range of variation of the water table level (in response to the tidal amplitude) is small compared to the saturated thickness of the aquifer, which makes it possible to use the Jacob–Ferris equation to assess its hydrodynamic properties (Erskine, 1991), which corresponds to Eq. (1):

\[
h = h_0 e^{-x\sqrt{s/\pi S/10T}} \sin \left(2\pi t_0 - x\sqrt{s/\pi S/10T}\right)
\]

where \(h\) is WL (m) above MSL; \(x\) is distance from the sea (m); \(t\) is time (d); \(t_0\) is the period of tidal oscillation (d) which for the west of Ireland is 0.52 days; \(h_0\) is the amplitude of tidal oscillation (m); \(T\) is the transmissivity of the aquifer \((m^2/d)\); and \(S\) is aquifer storage.

Tidal oscillations remain sinusoidal with a time lag \((delay = d)\) and decrease exponentially in amplitude with distance from the sea (tidal efficiency factor = \(f\)). An analytical solution to Eq. (1) provides expressions for \(d\) and \(f\):

\[
d = x\sqrt{S_{0}/\pi T}
\]

(2a)

\[
f = e^{-x}\sqrt{s_{0}^{2}/\pi T}
\]

(2b)

These solutions can be rearranged into expressions for aquifer hydraulic diffusivity \((D)\) in terms of the delay \((D_d)\) and tidal efficiency \((D_t)\):

\[
D_t = \frac{T}{S} = \frac{(x_{t_0})}{(4\pi d_{t_0})}
\]

(3a)

\[
D_t = \frac{T}{S} = \frac{x_{t_0}}{(\ln f_{t_0})d_{t_0}}
\]

(3b)

The delay is the time lag between the tidal signal and WL signal and the tidal efficiency factor corresponds to the ratio of the tidal signal with the amplitude of the WL. For each monitoring point, the delay and the tidal efficiency factor were calculated for separate 15-day periods without the influence of recharge and for different tidal amplitudes during spring tide (ST) and neap tide (NT). The average hydraulic diffusivity was then deduced: \(D_t\) from the tidal efficiency factor and \(D_d\) from the delay factor. This method was not used for B15 as no tidal influence was detected in this borehole, probably due to its low hydraulic diffusivity and its distance from the shore.

### 3.3. Results

The boreholes can be ranked on their hydraulic diffusivities (from highest to lowest) based on their hydrodynamic response to recharge (Fig. 4 and Table 1):

### Table 1

<table>
<thead>
<tr>
<th>Borehole</th>
<th>x (m)</th>
<th>Typology</th>
<th>Amplitude (m)</th>
<th>Period of influence (P^t) (d)</th>
<th>Slope of recession curve (S) (m/d)</th>
<th>Type of hydrodynamic environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>B57</td>
<td>2400</td>
<td>D1</td>
<td>10.4</td>
<td>1.6</td>
<td>6.3</td>
<td>Conduits</td>
</tr>
<tr>
<td>B08</td>
<td>1065</td>
<td>D2</td>
<td>6.25</td>
<td>4.3</td>
<td>2.6</td>
<td>Conduits/large fissures</td>
</tr>
<tr>
<td>B59</td>
<td>2300</td>
<td>D3</td>
<td>5</td>
<td>&gt;10.4</td>
<td>0.9</td>
<td>Fissures and matrix</td>
</tr>
<tr>
<td>B05</td>
<td>265</td>
<td>D4</td>
<td>3.3</td>
<td>4.5</td>
<td>0.9</td>
<td>Small fissures</td>
</tr>
<tr>
<td>B15</td>
<td>4490</td>
<td>D5</td>
<td>4.4</td>
<td>6.4</td>
<td>0.9</td>
<td>Matrix</td>
</tr>
<tr>
<td>B03</td>
<td>560</td>
<td>D6</td>
<td>3.75</td>
<td>9.1</td>
<td>0.9</td>
<td>Matrix</td>
</tr>
</tbody>
</table>
• B57 is located in a zone of high hydraulic diffusivity (D1) (Section 3.1), representative of a conduit-dominated flow environment.

• The amplitude (6.25 m) and the slope (S08 = 2.6 m/d) of the recession curve at B08 were lower than B57 but the slope remained relatively steep in comparison with those of other boreholes. B08 is located in a zone of relatively high hydraulic diffusivity (D2), representative of conduits or large fissures-dominated flow environment.

• B59 showed three different slopes on its recession curve representative of an environment with different hydraulic diffusivities. The steepest one (S09a = 1.5 m/d) was however lower than at B08 (S08 = 2.6 m/d) and the two others were particularly low and almost equal (S09b = 0.5 and S09c = 0.4 m/d) suggesting that borehole hydrodynamics are characteristic of both fissures and matrix-dominated flow environments (D3). These three successive slopes on its recession curve for which two are representative of a poorly hydraulic diffusive zone explain the particularly long observed period of influence P (P29 > 10.4 d).

• D4, D5 and D6 have been assigned to B05, B15 and B03 respectively: despite lower amplitude observed at B05 than at B15 and B03, it has been assumed its hydraulic diffusivity is higher as the slope of the recession curve (S05 = 0.9 m/d) was steeper than the ones at B15 (S15a = 0.7 and S15b = 0.3 m/d) and B03 (S03a = 0.7 and S03b = 1.0 m/d), and its period of influence much shorter (similar to the one of B08). B05 is therefore considered to be moderately connected with the karst drainage network (D4): B05 hydrodynamics would be considered as small fissures-dominated flow environment.

• B15 and B03 both have smaller slopes on their recession curve which are representative of an environment with different hydraulic diffusivities. Their recession curves were however very flat and expanded (P15 = 6.4 and P03 = 9.1 d). B15 and B03 are assumed to be poorly connected with the main drain karst system (D5 and D6) and representative of a matrix-dominated flow environment.

The outputs from the Jacob–Ferris equations are given in Table 2. Boreholes and lakes are ranked from highest hydraulic diffusivities (B57) (only D5 has been taken account; the reasons are explained in the discussion below) to lowest hydraulic diffusivities (B03).

3.4. Discussion: Consequences in terms of intrinsic functioning of the karst aquifer

The hydraulic diffusivities calculated using the efficiency factor and the delay factor are largely similar (Tables 1 and 2). However, as noted previously, important differences have been observed between the hydraulic diffusivity estimated on the basis of the tidal efficiency factor (Df) and hydraulic diffusivity calculated by the delay equation (Dd) for B59, B15, B03, L01 and L02, showing a ratio Df/Dd decreasing with hydraulic diffusivity values (Table 2). From Eq. (1) it is assumed that head losses can affect the tidal efficiency factor (f) or the delay (d). For a confined homogenous aquifer where head losses are isotropic, Df/Dd value should be 1 (Ferris and Branch, 1952): a difference between Df and Dd at the same distance from the shore would indicate heterogeneity of the environment. Head losses in the matrix are supposed to be higher than in the conduits: at each tidal cycle, the tidal wave enters the aquifer preferentially via the conduits that feed the matrix. Head losses occur in the conduits but additional head losses arise when the tidal signal propagates from the conduit through the matrix; in turn, this leads to a larger dampening of the sinusoidal wave and a higher delay of the tidal signal recorded in the matrix than in the conduits. However, in the matrix environment, the dampening of the tidal signal amplitude seems comparatively much more important than the increase of the delay. As a consequence, in a matrix environment, the relative change of the tidal efficiency factor (f) differs from the relative change of the delay (d). In this case, two boreholes located at the same distance from the shore, one drilled in a matrix-dominated flow environment (e.g. B59 located 2300 m from the shore), and the other in conduit-dominated flow environment (e.g. B57 located 2400 m from the shore), would have the ratio f/ Daneffectmatrix = 1 and the ratio d/ Ddmat = 1: e.g. fB57/fB08 = 29 and dB57/dB08 = 0.4. There is a difference of two orders of magnitude between these two ratios. Nonetheless, the Jacob–Ferris equation accounts for these differences in estimated hydraulic diffusivities by attenuating them: e.g. ratio Df/Ddmat < 1 or the delay (d). Df thus appears to be more sensitive to the heterogeneity of the aquifer than Dd. The latter may give a mean hydraulic diffusivity value of the area around the borehole and the aquifer between the monitoring point and the shore by taking into account all flow environments encountered (matrix, fissure and conduit); while Df may give a hydraulic diffusivity value more representative of low flow dominated environment such matrix or floor (if they are encountered). Thus, where Df is not affected by either matrix or fissure flow environments, Df and Dd determined from the hydraulic signal in boreholes with high hydraulic diffusivities (e.g., B57 and B08 supposed to be located in conduit-dominated flow environment) are similar. The Dd value is considered as the hydraulic diffusivity reference, because it is more representative of the overall hydraulic diffusivity of the aquifer between the monitoring point and the shore than Df.

Strong propagation biases have been also noticed in unconfined aquifers (Jha et al., 2003) or layered aquifers when the lower layer is more conductive than the upper layer (Trefry and Bekele, 2004) and both concluded that Df is more reliable than Dd. The Bell Harbour aquifer may be compared to these aquifers as it is also a layered aquifer; however, the presence of drainage system conduit/matrix makes it confined through the saturated conduits and unconfined in the surrounding matrix. Before it can reach a borehole located

| Table 2 | Hydrodynamic response of the borehole to tidal variations. Hydraulic diffusivities (Df and Dd in metres square/second) were calculated using tidal efficiency factor (f) and delay factor (d, in hours). |
|---|---|---|---|---|---|
| Borehole | x (m) | f | Df (m²/s) | d (h) | Dd (m²/s) | Ratio Df/Dd |
| B57 | 2400 | 0.131 | 97.3 | 3h35 | 124 | 0.8 |
| B08 | 1065 | 0.4036 | 98.2 | 1h55 | 91.5 | 1.07 |
| B59 | 2300 | 0.0045 | 12.7 | 8h4 | 19.1 | 0.65 |
| L01 | 485 | 0.0637 | 2.8 | 2h20 | 18.2 | 0.15 |
| L02 | 900 | 0.0052 | 2 | 1h55 | 14.6 | 0.14 |
| B05 | 265 | 0.4 | 5.8 | 1h25 | 12.4 | 0.47 |
| B03 | 560 | 0.0040 | 2.1 | 3h30 | 7.1 | 0.3 |

Hydraulic diffusivity values in bold (Df) are assumed to be more reliable than Dd values. Values at L01 and L02 are in italics as they are not comparable with borehole values.
in unconfined matrix flow environment, the tidal flow firstly propagates through saturated conduits, before it can be expressed in the unconfined matrix. This mix of confined/unconfined systems and the fact that head losses in conduit flow environment differ from head losses in matrix flow environment, makes it difficult to compare karst aquifers with more heterogeneous systems. In this case \( D_\text{f} \) is not necessarily more reliable than \( D_\phi \).

From these observations, the \( D_\text{f} \) value is considered as representative of the aquifer diffusivity, as it varies more homogeneously than \( D_\phi \); furthermore, the ratio between \( D_\phi \) and \( D_\text{f} \) allows characterising the type of hydrodynamic environment at each measurement point. The lower the \( D_\phi/D_\text{f} \) ratio the greater the component of matrix flow environment is present in the aquifer system between the point measurement and the shore.

Consequently, when the \( D_\text{f} \) value is larger than the \( D_\phi \) this suggests matrix flow dominates over conduit flow; in this case, borehole hydrodynamics are assumed to be characteristic of a *matrix-dominated flow environment*, such as B03 (Table 2); B15 is considered to be characteristic of the same environment as B03 due to the similarity of their recession curves. Conversely, when \( D_\phi \) and \( D_\text{f} \) values area similar, this means that water flows preferentially through conduits and consequently that the connection (between either the borehole or the lake) and the karst drainage network corresponds to conduits (e.g., B57 and B08; Table 2). The *fissure-dominated flow environment* is intermediate between conduit and matrix-dominated flow environments and can be illustrated by interim \( D_\phi/D_\text{f} \) ratios such as those observed for B05 and B59. These hydraulic diffusivity values are therefore consistent with the previous interpretation based on the typology of hydrodynamic responses to recharge events (Tables 1 and 2). From these results, generalised karst features (matrix, fissure and conduit) have been associated with each borehole and are shown in Fig. 5.

For the two lakes L01 and L02, the \( D_\phi/D_\text{f} \) ratios were 0.15 and 0.14 respectively, which can be attributed to head losses due to exchanges between the karst aquifer and the lake; the capacity of the lake cushions the tidal amplitude and may be added to the dampening effect of conduit/matrix on tidal amplitude explained above. The \( D_\phi \) inferred from the lakes’ hydrodynamics and recession curves are therefore not comparable with the \( D_\text{f} \) inferred from borehole hydrodynamics. The delay does not appear to be affected by the capacity of the lake and \( D_\phi \) estimated for the lakes can therefore be compared with the \( D_\text{f} \) observed in the boreholes (Table 2). However, the calculated \( D_\phi \) value was considered to be insufficient to assess the type of hydrodynamic environment for the two lakes.

Fig. 5 shows a hydrogeological profile of the catchment where depths of the boreholes and the loggers and their karst features associated are shown. WLs at HTs and LTs on 20th of February 2011 are shown in Fig. 5. This figure shows that WL in boreholes located close to the shore in an environment of a high to intermediate hydraulic diffusivity (B05 and B08) can differ considerably between HT and LT. These differences are less obvious in a matrix environment (B03), in areas well inland from the coast (B59) or where lake effects occur (L01). The WL at B57 was low compared to the other boreholes and was only characteristic of a conduit flow environment after a recharge event: at this location, water from the recharge was flushed very rapidly through the conduits towards the bay.

4. Spatial and temporal variability of the saltwater wedge

4.1. Spatial extent of the saltwater wedge

The water chemical data collected in boreholes and lakes in the study site showed that the groundwater was characterised by a Ca–HCO₃ facies representative of water flowing within limestone except at L01, where the water contained a significantly high proportion of Na, Cl, SO₄, Mg and K, i.e. representative of a seawater signature (Fig. 6). Water sampled from B05, B08 and L02 are all freshwater. Slightly higher concentrations of Na, Cl, SO₄ and K were recorded in B05, B08 and L02 (Fig. 6 and Table 3) compared to the
The proportions of these ions are representative of a seawater signature; indeed, Na/Cl ratios of all the samples follow the freshwater-seawater mixing line confirming that the main contribution of the salinity is due to a saltwater intrusion (Pulido-Leboeuf, 2004).

The minimum, maximum, and average SpC values recorded by the loggers at the monitoring locations are shown in Figs. 1c and 5 and confirm the inferences from the water chemical analysis: the high average SpC value at L01 (13,000 μS/cm) representative of brackish water with variations up to ~28,600 μS/cm, confirms the presence of the saltwater wedge at L01 most of the time. As described above, there is a strong hydraulic connection between the lake and the bay that may be due to the development of preferential conduits in the weathered zone parallel to the fault core of MacDermott’s fault (Bense et al., 2013).

In the other boreholes and in L02, the average SpC varied between 600 and 720 μS/cm except at B15 (430 μS/cm) which is situated furthest inland. Water chemistry data suggested that B05, B08 and potentially L02 can possibly be slightly influenced by seawater, and this was confirmed by maximum SpC values that reached 1240 μS/cm at B05, 1290 μS/cm at B08 and 830 μS/cm at L02 (the lower maximum value at L02 may be due to the dilution effect of the lake). The values at B05 and B08 are two to three time higher than the SpC at B57, B59 and B15 (maximum values of 720, 780 and 550 μS/cm respectively), considered as the background SpC in the area (Fig. 1c). B05 and B08 are assumed to be located in a variable mixing zone (Fig. 1c). The lack of water chemistry data from B03 and the largest recorded SpC value of 800 μS/cm from this borehole make it difficult to estimate if B03 is influenced by the saltwater wedge.

The vertical variation in the SpC recorded at B05 and B03 (Fig. 5) made it possible to determine the vertical variation in the haloclines: 10 m below MSL, the increase in the SpC from 500 μS/cm described above, there is a strong hydraulic connection between the lake and the bay that may be due to the development of preferential conduits in the weathered zone parallel to the fault core of MacDermott’s fault (Bense et al., 2013).

In the other boreholes and in L02, the average SpC varied between 600 and 720 μS/cm except at B15 (430 μS/cm) which is situated furthest inland. Water chemistry data suggested that B05, B08 and potentially L02 can possibly be slightly influenced by seawater, and this was confirmed by maximum SpC values that reached 1240 μS/cm at B05, 1290 μS/cm at B08 and 830 μS/cm at L02 (the lower maximum value at L02 may be due to the dilution effect of the lake). The values at B05 and B08 are two to three time higher than the SpC at B57, B59 and B15 (maximum values of 720, 780 and 550 μS/cm respectively), considered as the background SpC in the area (Fig. 1c). B05 and B08 are assumed to be located in a variable mixing zone (Fig. 1c). The lack of water chemistry data from B03 and the largest recorded SpC value of 800 μS/cm from this borehole make it difficult to estimate if B03 is influenced by the saltwater wedge.

The vertical variation in the SpC recorded at B05 and B03 (Fig. 5) made it possible to determine the vertical variation in the haloclines: 10 m below MSL, the increase in the SpC from 500 μS/cm

Table 3
Concentration of Na, Cl, SO4, Mg and K measured at the three monitoring points slightly influenced by seawater, at L01, which is strongly influenced by seawater, and at B15, considered as Ca–HCO3 background water.

<table>
<thead>
<tr>
<th>Monitoring point</th>
<th>Na+ (mg/l)</th>
<th>Cl (mg/l)</th>
<th>SO4²⁻ (mg/l)</th>
<th>Mg²⁺ (mg/l)</th>
<th>K⁺ (mg/l)</th>
</tr>
</thead>
<tbody>
<tr>
<td>L01</td>
<td>1380</td>
<td>4874</td>
<td>848</td>
<td>136</td>
<td>56</td>
</tr>
<tr>
<td>B05</td>
<td>10</td>
<td>29</td>
<td>9</td>
<td>6</td>
<td>5</td>
</tr>
<tr>
<td>B08</td>
<td>8</td>
<td>214</td>
<td>9.4</td>
<td>6</td>
<td>3</td>
</tr>
<tr>
<td>L02</td>
<td>7</td>
<td>16.4</td>
<td>&lt;5</td>
<td>3</td>
<td>2</td>
</tr>
<tr>
<td>B15</td>
<td>7</td>
<td>14</td>
<td>&lt;5</td>
<td>2</td>
<td>1</td>
</tr>
</tbody>
</table>

Ca–HCO₃ background water. The proportions of these ions are representative of a seawater signature; indeed, Na/Cl ratios of all the samples follow the freshwater-seawater mixing line confirming that the main contribution of the salinity is due to a saltwater intrusion (Pulido-Leboeuf, 2004).
to 1000 µS/cm at B05 and from 500 µS/cm to 750 µS/cm at B03 indicates the presence of the mixing zone at B05 below 10 m below MSL; despite an increase of 250 µS/cm at 10 m below MSL in B03, the maximum SpC values are well below a saltwater influence signal. Lower SpC values observed close to the base of B05 (850 µS/cm) are assumed to be due to a karst conduit where freshwater flows occur (Fig. 5). The vertical salinity profile at B08 showed only a slight increase in SpC close to the surface (1 m above MSL) from 450 to 550 µS/cm, probably due to the epikarst, which may allow rapid drainage of recharge water into the borehole (Fig. 5). This borehole is relatively shallow (depth is to 9 m below MSL) and it may not be deep enough to reflect the halocline. The profile at B57 exhibited a stable SpC value of 720 µS/cm throughout the vertical profile, and this SpC is representative of typical groundwater in the Burren karst aquifer not affected by saltwater wedge (Fig. 5).

4.2. Temporal variability of the extent of the saltwater wedge

4.2.1. Temporal variation in saltwater wedge at high/low and neap/spring tidal scales

The impact of high tide/low tide fluctuations on the SpC was assessed for different tidal amplitude periods to evaluate how far the saltwater wedge moved at each tidal cycle. This analysis was performed on B05, B08 and L01 data sets which are the only monitoring points where tidal variations of SpC were observed. The range of the SpC variations observed during strong spring tide (SST) periods (tidal amplitude >3.8 m), spring tide (ST) periods (tidal amplitude between 2.5 m and 3.8 m) and neap tide (NT) periods (tidal amplitude <2.5 m) as well as the time lag between the tidal signal in the bay and the highest WL (or SpC) data observed for one tidal cycle.

Tidal SpC variations during NT and some ST periods may not be observed due to this inertia effect. Tidal SpC variations were higher in B05 than in B08, suggesting that the impact of the tidal variation and of the saltwater wedge might be stronger in B05 than in B08, which is located farther from the bay (Table 4). The time lag at B05 varied with the tidal amplitude unlike at B08 for which the time lag stayed constant for all tidal amplitude periods.

To identify the cause of the differences in time lags, the WL and SpC values at B05 versus the tide levels in Bell Harbour were analysed during one tidal cycle for each NT, ST and SST periods. Notable hysteretic cycles were observed caused by the remnant effect of the tidal level on the WL and SpC values, confirming the tidal impact on WL and SpC values which increased and decreased with the rising and falling tide respectively (Fig. 7a and b). Differences in
the time lag have been observed in hysteresis cycles of SpC between different tidal amplitude periods while the time lag of WL versus tidal level in Bell Harbour remains constant over time (Fig. 7a and b). Hysteresis cycles of SpC are different according to the tide periods:

- During ST or NT periods, SpC values started to increase slowly at rising tide in B05: the SpC was highest at the beginning of the following rising tide, which explained the long time lag observed between the high tide in the bay and the highest SpC value of a tidal cycle at B05 (11 h). SpC values are higher during rising rather than falling tides which explains the clockwise direction of the loops. The SpC decreased rapidly at HT and continued to decrease slowly during the falling tide (Fig. 7b).
- During the SST period, SpC values increased rapidly at HT resulting in the short time lag observed under these conditions (2 h). SpC values are lower during rising rather than falling tides and the loop changes to an anticlockwise direction. SpC also decreased rapidly at low tide (LT) and values remained relatively stable during rising and falling tides (Fig. 7b).

These cycles highlighted the rate of variation in the SpC depending on the stage of the tide (for a high/low tidal cycle); for example, the rate of increase in SpC was slower than the rate of decrease during ST or NT periods, and slower than the rate of increase during SST periods. However, some hysteresis cycles observed in SpC data from B05 showed a time lag of 11 h despite SST periods.

By comparing SpC, WL, tide levels in the bay and hydraulic gradient (i) variations over time for several periods, all of the conditions required to get the shorter time lag (2 h) have been identified (Fig. 8):

- tidal amplitude (Tamilp) > ~3.8 m (SST periods),
- WL at HT > ~3.8 m above MSL, without being affected by the recharge.

If these conditions are met, the short time lag of 2 h remains between the tide in the bay and the tidal SpC variations even if the Tamilp in the bay decreases. However, even if the Tamilp is high, if the WL decreases below ~3.8 m at HT, the process stops and the time lag reverts to 11 h (Fig. 8).

Hysteresis cycles (SpC values versus the tide levels in the bay) were also observed at L01 and B08:

- At L01, they were only observed during some ST and SST periods when the tidal amplitude was sufficiently high; no time lags were observed in NT and some ST periods between the tide in the bay and SpC signal in the lake. When hysteresis cycles were observed, the SpC increased rapidly at HT and decreased slowly from the beginning of the rising tide.
- At B08, hysteresis cycles revealed no significant differences for the different periods of the tide confirming that the SpC at B08 was not influenced by tidal amplitude. SpC increased slowly during the falling tide and decreased slowly during the rising tide; the tidal influence was much more subtle at this monitoring point, due to its distance from the shore. Thus, hysteresis cycles allow direct observation of variations in SpC depending on the stage and the amplitude of the tide.

4.2.2. Conditions for larger intrusion of the saltwater wedge into the karst aquifer

Occasional, temporary and significant increases in SpC were recorded at B05 (Fig. 3) and L01. During these temporary larger saltwater wedge intrusions, the range of SpC values and variations were different at B05 and L01:

- SpC values were higher than 1000 µS/cm at B05 while they were higher than 15,000 µS/cm at L01; the mixing zone extended across the entire water column at B05 while L01 was still located in the saltwater wedge but with higher SpC values than normal.
- Variations of SpC were higher than 500 µS/cm at B05 whereas they were higher than 2000 µS/cm, and up to 15,000 µS/cm at L01.

As above, the conditions required to get more extensive intrusions of the saltwater wedge (and the mixing zone) have been identified by comparing SpC, WL, tidal levels in the bay and hydraulic gradient (i) variations over time for several periods at B05 and L01:

- SpC values were higher than 1000 µS/cm at B05 while they were higher than 15,000 µS/cm at L01; the mixing zone extended across the entire water column at B05 while L01 was still located in the saltwater wedge but with higher SpC values than normal.
- Variations of SpC were higher than 500 µS/cm at B05 whereas they were higher than 2000 µS/cm, and up to 15,000 µS/cm at L01.
The lowest hydraulic gradient observed during a low/high tidal cycle ($i_{\text{min}}$) was between 0% and 0.5% and occurred close to the HT (Fig. 9).

- At L01, WL at LT should be $< 1.8$ m above MSL and the tidal amplitude in the bay $> 3.5$ m (WL in the bay at HT is then $> 1.75$ m above MSL) to start the process (Fig. 10). $i_{\text{min}}$ was then close to 0 or slightly negative and occurred at HT (Fig. 10): an inversion of the hydraulic gradient occurred between the WL in the bay and L01 for a short period during the process.

Additionally, a large difference in the hydraulic gradient between HT and LT ($i_{\text{ampl}}$) has been observed at B05 over a number of HT/LT cycles before the increase of SpC ($i_{\text{ampl}} > 1\%$; Fig. 9). These variations in the hydraulic gradient between the aquifer and the bay could facilitate intrusion of the seawater and thus could drive the saltwater wedge further inland. The temporary increase of SpC ceased at B05 and L01 when the tidal amplitude in the bay decreases. The latter seems to be the motor of the process by inverting the hydraulic gradient at L01 and by inducing strong amplitudes in the hydraulic gradient ($i_{\text{ampl}} > 1\%$) at B05.

From the study of several temporary increases of SpC at L01 and B05, it appears that the intensity of SpC increases varies with the altitude of the WL at the monitoring point: the lower the WL, the higher the observed SpC value. The elevation of the WL in the borehole or the lake appears to have a large influence on the range of SpC increases.
4.3. Discussion on the temporal saltwater wedge variations

The maximal extent of the saltwater wedge and mixing zone intrusions into the aquifer is shown in Fig. 1. All the monitoring points affected by the saltwater wedge – tidally (L01, B05 and B08), continuously (L01) or temporarily (B05) – are located less than two kilometres from the shore (Fig. 1). A hierarchy of influence of the saltwater wedge at the monitoring points has been established from these observations and from the hysteresis cycles (Table 5). The saltwater wedge (> 10,000 μS/cm) was observed only at L01 (SpC values were mainly brackish, i.e. > 10,000 μS/cm). B05, and possibly B08, appear to be reflective of a temporary mixing zone (1000 μS/cm < SpC < 10,000 μS/cm) rather than directly by the saltwater wedge, because of the low observed maximum SpC values (<1500 μS/cm) (Fig. 1).

The main results of this study are summarised in Table 5 suggesting that for L01, B05 and B08:

- Tidal influences on the SpC values increased with the tidal amplitude.
- Hysteresis cycles of SpC versus tidal level in the bay have been observed:
  - At L01: for ST and SST periods: the increase of SpC due to the tide was rapid and large which confirms a high hydraulic diffusivity between the lake and the bay and the bay due almost certainly to the development of conduits along the large MacDermott’s fault.
  - At B05: the rate of variation in SpC with the tide was more rapid during some SST periods (time lag of 2 h) than during NT and ST periods (time lag of 11 h). Short time lags (2 h) and larger SpC changes (>100 μS/cm) at B05 are related to a large tidal amplitude in the bay when the WL in the borehole is above a given threshold during a period of no recharge. When water level is >3.8 m MSL at high tide this temporarily activates a more direct hydraulic path (probably a conduit) that allows seawater intrusion. The large amplitude of the tide may facilitate the seawater intrusion along this path when the hydraulic gradient (between the bay and B05) is > 0.5% at high tide (Fig. 9).
  - At B08: despite its high estimated hydraulic diffusivity (Table 2), the hysteresis cycles of SpC versus tidal level in the bay showed the increase in SpC due to the tide was slow and relatively weak for any tidal periods. The tidal influence of the saltwater wedge at B08 is thus less important than at L01 and B05 (Table 5).
- Larger saltwater wedge intrusions occurred:
  - During SST periods when WL at LT was <~1.8 m for L01 and <~2 m for B05. B08 is too shallow to record the direct influence of the saltwater wedge.
  - When the hydraulic gradient was very low or slightly reversed; this phenomenon has been already observed in other coastal karst aquifers (Bonacci and Roje-Bonacci, 1997; Drogue, 1989; Fleury et al., 2008).
  - Strong tidal amplitude seems to be the motor of the process by driving seawater into the aquifer, while the position of the WL in the aquifer may influence the intensity of the SpC increase.

The temporal extent of the saltwater wedge is therefore dependent on a balance between the recharge which controls the WL in the aquifer, and the tidal amplitude which controls the water level in the bay and the range of seawater tidal movements. Similar behaviours have been observed in a Turkish karst aquifer where SGD occurs essentially through the fracture system (Ozyurt, 2008).

The monitoring points affected by the saltwater wedge are those with a relatively high hydraulic diffusivity located less than 1000 m from the shore. L01 is assumed to be well connected with the bay through MacDermott’s Fault and B05 and B08 are considered to be characteristic of fissured and conduit hydrodynamic environments, respectively (Table 1). The extent of the intrusion of the saltwater wedge into the karst aquifer is therefore tightly linked with the geological structure (fault) of the aquifer and its intrinsic properties (hydraulic diffusivity).

In order to study saltwater wedge variations in coastal karst aquifers subject to an oceanic climate and a high tidal range like at Bell Harbour, it is therefore recommended firstly to focus on strong spring tide periods during periods of low or absent recharge near the shore and preferentially in high hydraulic diffusivity zones.

Temporal variations of the saltwater wedge and the mixing zone have been proposed through a conceptual cross-section model including B57, B08 and B05 (Fig. 11a and b). Only the most pertinent monitoring points have been shown for clarity:

- For a large flood event during NT period (Fig. 11a), conduits along the MacDermott’s Fault connected to B57 are only slightly affected by the mixing zone near the shore (<~500 m) from which freshwater to brackish water flows through submarine springs in the middle of the bay. The aquifer may only be slightly affected by the saltwater wedge located at least 20 m below MSL at the shore.
During low recharge and SST periods (Fig. 11b), the saltwater wedge may enter further into the aquifer, preferentially through the fissured environment via intertidal springs which become sinkholes, and through the conduit along the MacDermott’s Fault. Submarine springs in the middle of the bay behave like intertidal springs where seawater from the bay can intrude in the aquifer. In this case, the saltwater wedge may enter in the deep part of B05 located in a fissure-dominated flow environment at 265 m from the shore. The mixing zone would expand further and shallower in the aquifer and the halocline observed at 10 m below MSL at B05 may rise. It is assumed the mixing zone could expand at 20 m below MSL at 1000 m from the shore. Under these conditions it would not reach B08 (9 m below MSL) and its associated conduit(s) (Fig. 11b).

Submarine groundwater discharge in Bell Harbour occurs mostly in a fissure-dominated flow environment via intertidal springs and through conduits developed along MacDermott’s Fault only during large flood events (Fig. 11a). While the karst is well developed and the conduits open directly to the sea, the groundwater discharge is mainly fresh (<1000 μS/cm) and the saltwater wedge remains close to the shore (<500 m). Even during low recharge and SST periods, the mixing zone at 1000 m from the shore is only present at ~20 m below MSL (Fig. 11b) which differs greatly from other karst aquifers such as those found along the Mediterranean coast (e.g. extensive saline intrusions of more than several kilometres from the coast in Crete in Greece (Arfib et al., 2007); permanent brackish inland and/or submarine springs (Fleury et al., 2007); or continuous inflow of seawater observed simultaneously with outflow of brackish water (Drogue, 1989)).

5. Conclusion

In this non-anthropogenic coastal karst aquifer with a strong tidal influence the spatial variability of the saltwater wedge has been found to be tightly linked to the geological structure and intrinsic properties of the karst aquifer. Values of SpC (which reflected the extent of the saltwater wedge) were relatively low and stable throughout the catchment, except at L01. A large fault connecting the land and the bay underlies L01, explaining why the lake is within the saltwater wedge. Three types of hydrodynamic environment (conduits, fissures and matrix) were defined for each of the six boreholes by comparing two different methods: analysis of the recession curves of the WL data, and use of the Ferris equation to assess the impact of the tide on WL data. These approaches, however, did not work for the data collected from the lakes. Analysis of the impact of the tidal amplitude on SpC values measured at the monitoring points confirmed the influence of the tide on SpC values at B05, B08 and L01. Sudden temporary increases in SpC (>1000 μS/cm at B05 and >15,000 μS/cm at L01)
ments will be critical to predicting and managing these dynamic phenomena. Characterising coastal karst catchments and collection of sufficient data will be critical to predicting and managing these dynamic environments.

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References


Hubbert, M.K., 1940. The theory of ground-water motion. J. Geol., 785–944.


Pracht, M. et al., 2004. Geology of Galway Bay. A geological description to accompany the Bedrock Geology, 1(100,000).


