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HYDROGEOLOGY AND HYDRODYNAMICS OF CORAL REEF PORE WATERS

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ABSTRACT

A wide variety of forces can produce head gradients that drive the flow and advective mixing of internal coral reef pore waters. Oscillatory gradients that produce mixing result from wave and tide action. Sustained gradients result from wave- and tide-induced setup and ponding, from currents impinging on the reef structure, from groundwater heads, and from density differences (temperature or salinity gradients). These gradients and the permeabilities and porosities of reef sediments are such that most macropore environments are dominated by advection rather than diffusion. However, diffusion-dominated micropores are so ubiquitous that biogeochemical processes are influenced by both the advective and diffusive domains. The various driving forces must be analyzed to determine the individual and combined magnitudes of their effects on a specific reef pore-water system. Pore-water movement controls sediment diagenesis, the exchange of nutrients between sediments and benthos, and coastal/island groundwater resources. Because of the complexity of forcing functions, their wide range of time and distance scales, and the uniqueness of their interactions with specific local reef environments, experimental studies require careful incorporation of these considerations into their design and interpretation.

INTRODUCTION

The internal pore water of coral reefs and related carbonate sediments exists in a hydrodynamically and biogeochemically active environment (Oberdorfer & Buddemeier 1985,1986; Sansone 1985; Buddemeier & Oberdorfer 1986; Parnell 1986; Roberts et al., this symp; Tribble et al., this symp.). An understanding of the mechanisms, rates, and pathways of water movement through these sediments is critically important to at least three major fields of study. First, reef diagenesis—where, how fast, and by what mechanisms it occurs—cannot be addressed quantitatively without knowledge of the fluxes of water and solutes through the reef framework and of their changes over time (Buddemeier & Oberdorfer 1986). Second, the mass transport of solutes (e.g., fixed and regenerated nutrients) between the pore waters, the benthic layer, and the overlying water mass is probably a major factor in controlling pore-water biological activity, and may be at least locally significant to benthic communities (Tribble et al., this symp.). Third, the quantity and quality of fresh groundwater resources on reef islands and reef-derived coastal plains will be determined by the interactions of the groundwater body with the surrounding marine pore-water environment (Herman et al. 1986).

This paper addresses the origins, magnitudes, and time scales of the various processes controlling the movement of pore water within reef sediments. Because of the variety of reef types and environments, we cannot hope to quantify or even describe every possible situation; rather, our objective is to develop some useful generalizations about possible and probable hydraulic influences on reef systems that can be evaluated for specific environments or experiments by individual investigators.

ADVECTIVE TRANSPORT PROCESSES

Darcy's Law is used as the basic formulation of the parameters governing flow through porous media. It is expressed as Eq. (1),

$$v_p = (K/n)(dh/dl) \quad , \tag{1}$$

where v_p is the pore velocity (m/s), n (dimensionless) is the effective porosity of the medium (drainable fraction of bulk volume), dh/dl (dimensionless) is the head gradient (difference in height h of the water column over the difference in distance l), and K is the hydraulic conductivity of the medium (m/s).

The mass transport version of the relationship is given by Eq. (2),

$$Q = KA(dh/dl) \quad , \tag{2}$$

in which Q is the mass flux (m^3/s), and A is the cross-sectional area (m^2) of the formation through which the flux occurs.

Hydraulic conductivity is a function of the intrinsic permeability of the formation and the fluid properties—in seawater at 25°C, a hydraulic conductivity of 1 m/s = a permeability of 1.03×10^{5} darcys. In this paper, we use the term permeability to refer qualitatively to the general phenomenon, and hydraulic conductivity when numerical values in m/s are cited. Effective porosity is the fraction of the sediment volume that drains under the influence of gravity (macropores) and is some fraction (often 0.3–0.5) of the total porosity. In the discussion that follows we use pore velocities; the darcy velocity, which is often cited in the literature, is the pore velocity times porosity, or K(dh/dl).

The above formulations of Darcy's Law have some limitations for application to reef systems. A minor concern is that they are valid only for laminar flow conditions; Enos & Sawatsky (1981) point out that the threshold for turbulent flow in coarse sands is in excess of 10^{-2} m/s, a value likely to be exceeded only in very limited and localized environments (e.g., surf zones or large solution cavities). A more significant problem is the implicit assumption of a homogeneous porous medium, since reef structures are well known to be both horizontally and vertically heterogeneous over a wide range of distance scales. The approximation is reasonably valid for system dimensions large compared to the scale of internal variability, but at the level of individual field observations it is quite likely that we are dealing with a heterogeneous medium that includes a component of cavernous flow conditions rather than with a homogeneous medium. Having stated that caveat, we shall proceed with the analysis, since there exist neither the field data nor the theoretical understanding required for a more rigorous approach to flow through heterogeneous coral reef media.

Laboratory permeametry of coral reef and island sediments yields values for K ranging from 10^{-8} to $> 10^{-2}$ m/s (Jacobson & Hill 1980; Enos & Sawatsky 1981; Wheatcraft & Buddemeier 1981; Oberdorfer & Buddemeier 1986), and field slug or pump tests produce values ranging from 6.5×10^{-4} to 2.2×10^{-2} m/s (Wheatcraft & Buddemeier 1981; Oberdorfer & Buddemeier 1986). The lowest permeametry values are typical of very fine unconsolidated sediments or highly lithified limestone; these sediments are not common in near-surface reef environments, and will be dominated by diffusive (see below) rather than advective pore-water transport. For our discussion of advective processes, we take 10^{-6} to 10^{-2} m/s as a typical range for most reef sediments, and we note that the upper end of this range probably represents determinations made on samples or systems for which Darcy's Law may not be strictly valid. Effective porosities for reef sediments commonly fall in the range of 0.15 to 0.35, although they may approach zero for highly lithified limestone and are virtually indefinable for systems with cavernous voids. In the discussion that follows, we shall adopt a value of 0.25 unless we specify otherwise. We also use the term sediment in its most general sense, without any implications about state of consolidation or diagenesis.

For ease of discussion, we choose to divide advective transport processes into two somewhat arbitrary classes: flow and advective mixing. Net flow occurs as a result of a net (or average), long-term, unidirectional head gradient across a volume of porous medium. For our present purposes, long-term will be considered anything significantly longer than a tidal period (i.e., $> 10^5$ s). The process that we term advective mixing is caused by more or less symmetrically oscillating head gradients (e.g., waves and tides) that induce short-term exchange across boundaries without necessarily creating a significant long-term net directional flow through them. Figure 1 presents a hypothetical scenario illustrating the distinction between flow and mixing as well as providing graphic examples of some of the forces discussed below.

In Fig. 1, wave setup and lagoon ponding provide long-term average gradients capable of driving a sustained flow, while waves create a force for vertical pumping in and out of the reef surface—a process of mixing overlying ocean water with the shallow reef pore

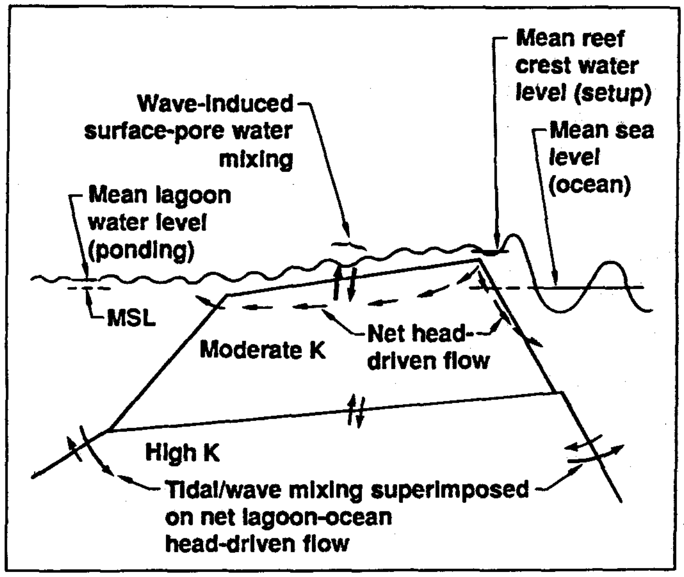

Figure 1. Idealized depiction of a windward reef showing some of the hydraulic heads and gradients and indicating possible pore-water advective flow and mixing processes. See text for discussion.

water. Wave setup onto a windward reef results in an elevated water surface on and behind the crest, with a resultant head gradient that will tend to drive shallow reef pore water lagoonward through the sediments (and possibly back out the shallow forereef slope as well). This gradient may be intermittent if the crest is exposed because of low tide and/or calm seas, but it will not reverse; therefore, any long-term net flow will be as shown. However, the pore waters of the shallow reef sediments will be subject to vertical mixing by wave (and possibly tidal) action, and the total vertical flux through the reef surface may far exceed the net head-driven horizontal flow within the reef. In this situation, any chemical signal associated with the horizontal transit time through the reef would probably be undetectable—erased by the vertical mixing along its path. A somewhat similar but reverse-flow situation occurs in the high permeability (e.g., Pleistocene) lower aquifer. Because of restricted outflow from the lagoon and wave setup across the reef, the lagoon water level is, on average, ponded above the mean ocean level (as with the wave setup case, the gradient may be intermittent but is virtually never reversed). This results in a net seaward flow through the lower aquifer that will be superimposed on an oscillating flow resulting from differences between tidal signal propagation in the ocean and in the lagoon.

HEAD GRADIENTS AND PORE VELOCITIES

In this section, we consider the forces that may generate head gradients in a coral reef system, and discuss the magnitude and time scales of the various gradients and resulting fluxes or velocities. Because of the diversity of reef structures and environments, we feel that there is no useful way to define average or typical values. However, because the upper limits for most of the parameter ranges are reasonably well established, we present as examples and order-of-magnitude estimates what we consider "maximum credible values" that are based on high-range estimates, but not on the compounding of absolute upper limits. This represents what can happen, and probably does in some locations, but it should not be considered applicable to reefs in general. In some situations, pore-water flow is driven by forces such as density gradients or pressure gradients that are not readily expressed as an equivalent head gradient. In these cases, or where the literature cited presents velocity data, estimates of pore velocities are easier to make than head gradient estimates. When that is the case, we discuss the pore velocities directly; for a given hydraulic conductivity these can be related to an equivalent head gradient by Eq. (1). Table 1 presents a summary compilation of the types and origins of the major hydraulic forcing functions, the environments most commonly subject to them, and the magnitudes (logarithms) of the maximum credible values of the effective distances, the head gradients, and/or the pore velocities associated with the various forces. Data are based on observations or on summaries of values published by various researchers; cases where the values cited are average rather than "maximum credible" are noted. The various categories are briefly discussed below.

Sustained gradients

Wave Setup. Two important examples of sustained head gradients have already been mentioned in connection with Fig. 1: these are wave setup and lagoon ponding. Both involve the wind-driven water transport onto and across the windward reef barrier. Under appropriate conditions of wind and tide, this setup on windward reef crests or island beaches may amount to several tenths of a meter, resulting in gradients from the point of setup to the lagoon that can range from $> 10^{-3}$ to 10^{-5}. Although setup conditions may be interrupted by calm weather or low tides, the long-term averaged gradients are probably not reduced by as much as an order of magnitude. A higher gradient than indicated in Table 1 may occur seaward of the crest between the setup point and the location on the forereef slope beneath the area just before the wave break. However, because the sediments in this high-energy reef-crest environment are commonly highly lithified, there may not be a significant path for pore-water flow in spite of the high gradients.

Ponding. If the lagoon behind the reef has restricted outflow, the excess water driven across the reef by wave setup can result in long-term ponding of the lagoon at heights above oceanic mean sea level. These effects have been demonstrated for Enewetak Atoll by Atkinson et al. (1981) and Buddemeier (1981). These lagoon-ocean gradients are probably capable of flushing high-permeability aquifers on time scales of years to decades. Reef-surface ponding in tide pools and reef moats may generate local head gradients orders of magnitude greater than the values tabulated for lagoon ponding (Scoffin & Stoddart 1978), but the effects are much more localized, and may persist for a smaller fraction of the tide cycle. Examples in this paper are discussed on a scale of reef dimensions; Mathews (1974) presents a qualitative discussion of tidally driven "ponding," both negative and positive, that can occur on carbonate platforms of larger dimensions. Negative ponding (e.g., a lagoon water level lower than the surrounding ocean) can also occur as a result of evaporation acting on confined bodies of water. In this situation, increased salinity will result in density-driven flow (see below) downward and out of the lagoon while the lower head will tend to produce head-driven flow into the lagoon.

Currents. When a current impinges on a coral reef, some of the kinetic energy of the moving water is converted to pressure on the face of the reef, resulting in an equivalent head capable of driving pore-water flow through the reef structure. The actual head gradient will be a complex result of the reef dimensions, the current velocities, and the pressure and flow fields around the structure. Andrews & Muller (1983) derive expressions for the internal flow velocities and patterns in a cylindrical patch reef as a function of the

Table 1. Origins, locations, and magnitudes of reef system heads, gradients, and pore velocities.

Head type: source	Typical environment	Upper end of typical value range: Log range (m); direction[a]	Log head gradient (dh/dl)	Log pore velocity (m/s)
Sustained:				
Wave setup	Windward shallow reefs, intertidal zones, beaches	1 to 2; v,h	−3	[−4][b]
Ponding	Lagoon/cross-reef, reef moats	2 to 3; v,h	−4 to −3	[−5 to −4][b]
Currents (all types)	System-wide	up to 3; h		up to −3
Groundwater	Near islands, coasts	1 to 3, v,h	(−2)	−6 to −5[c]
Thermal	General; near higher-temperature, non-reef substratum	3; v		−7[c]
Salinity	Beneath shallow or confined water bodies	up to 3; v		8[c]
Oscillating:				
Waves	Shallow reefs; sub/inter-tidal beach zones	≤ 1; v, (h)	−2 to −1	−2[d]
Tides (marine)	System wide; especially cross-reef	2 to 3; h, (v)	−4	[−5][b]
Tides (coastal)	Adjacent to/under islands, coasts, groundwater bodies	0 to 2; v,h	up to −1	−5

[a] v = vertical; h = horizontal; ( ) indicates lower significance or certainty.
[b] [ ] = calculated from head gradient and maximum K/n = 10^{-1}; averages lower.
[c] Average rather than maximum credible value; individual local values may be significantly greater.
[d] Peak instantaneous value; average high value ≧ −3.

impinging current and the permeability. Their solutions indicate that water enters the structure at 0° and 180° relative to the incident current direction, exiting at 90° and 270°. Their results compare satisfactorily with experimental observations on a highly permeable mid-lagoon patch reef, where a 10-cm/s tidal current generated a 0.2-cm/s internal flow within the reef. Because of the symmetry of the solution, the internal flow did not reverse when the direction of the external current reversed. If these results can be scaled up in size and down in permeability to the dimensions of atolls, the observations suggest that ocean currents may induce steady-state internal flows with pore velocities as high as 10^{-4} or 10^{-5} m/s within large reef structures with hydraulic conductivities on the order of a few tens of darcys.

Wave-induced pore-water flow. A source of reef pore-water flow that has not been discussed in the literature is a sort of corollary to cross-reef transport of surface water. Linear wave theory predicts no net mass transport, either in the wave or induced in underlying sediments, as a wave front passes. However, at depths comparable to the wavelength and especially near wave breaks, waves are not symmetrical and particle motions are better described by nonlinear wave theory, which results in a net mass transport as well as oscillatory motion. Because many windward reefs are characterized by unidirectional movement of waves and water across them, there will presumably be an induced net horizontal movement of pore water in addition to the oscillatory vertical flushing described below. Sleath (1984) alludes to this process briefly, but presents no equations directly applicable to the problem of reef sediments. We assume, primarily on the basis of intuition, that net fluxes from this source would be small relative to the wave-induced mixing. However, we believe that theoretical calculation of the fluxes to be expected from this source for various reef environments would be a worthwhile effort.

Groundwater head gradients. Groundwater bodies also provide a source of net flow. Pore velocites at the water table may be estimated from the recharge rate; for many reef islands, recharge rates of about one m/y are reasonable, which is equivalent to a pore velocity of roughly 10^{-7} m/s. The actual value is temporally variable: recharge during limited rainy seasons or episodic storms provides peak flows orders of magnitude higher than the mean, interspersed with periods of relative stagnation. More importantly, the flow velocities induced in the formation distant from the water table are very dependent on the permeability-controlled flow patterns. In the case where discharge is controlled largely by vertical mixing with a well-flushed underlying aquifer (Oberdorfer & Buddemeier, this symp.), the maximum velocities estimated from recharge will be applicable. However, where the more traditional concept of a groundwater lens is appropriate, water is constrained to exit the system through a narrow zone along the shallow subtidal/intertidal portion of the coastline. The fresh-water outflow induces a compensating inflow of underlying seawater and a mixing zone where the two layers combine. Because the area available for discharge is much smaller than the recharge area, discharge velocities may be amplified over recharge velocities by an order of magnitude or more, depending on the thickness of the mixing zone and the ratio of shore length to surface area (Johannes, 1980). Simms (1984) summarizes data indicating that coastal freshwater flows into the ocean will have pore velocities on the order of 10^{-6} m/s, and the seawater circulation induced by the freshwater outflow will be on the order of 10^{-7} m/s. Where there are large land masses or high relief, volume outflow and/or local head gradients may be greatly increased. This may combine with local geology to produce groundwater-induced flow through formations at some distance from the groundwater body and the shoreline, and may cause localized fluxes several orders of magnitude greater than those estimated above on the basis of homogeneous formations.

Density gradients. Density gradients may cause pore-water flow in two ways. Evaporation from shallow waters over banks and lagoons may produce high-salinity water that will sink through the underlying formation, displacing the lower-density sea water below it. The reverse process can occur if a geothermal heat source beneath a porous structure sets up convective flow, causing deep pore water to be warmed and rise within the formation as deep sea water moves into the formation to sustain the density-driven upward flow. Simms (1984) summarizes data from the Florida-Caribbean region indicating that reflux (salinity-driven) flows may have pore velocities in the range of 10^{-8} to 10^{-9} m/s, while thermally-driven flows approach 10^{-7} m/s. Thermal flows will be controlled both by the permeability of the formations and by the depth and temperature of the thermal source. In more geothermally active areas, thermal flows may be greater than those cited by Simms; Rougerie & Wauthy (1986) have suggested that the process, which they term "endo-upwelling," may be a significant source of nutrients to some of the reef and lagoon systems of atolls in French Polynesia. For reef systems where the turnover rate of oceanic water is high, significant thermal contributions to the total nutrient budget seem unlikely. However, the values cited by Simms could well produce a biogeochemically significant input to reef systems with some combination of smaller lagoon volumes, longer residence times (perhaps the result of smaller tidal fluxes), and moderately higher thermal fluxes. Regardless of the external effects, if such flows occur they are likely to be highly significant in terms of diagenetic processes.

Oscillatory gradients

Wave-induced gradients. Waves are the most obvious example of an oscillating head gradient, and are a force felt by almost all parts of a coral reef system. Sea waves and swells may have amplitudes (or maximum heads) of meters (neglecting storm waves, which are discussed elsewhere), periods of seconds to tens of seconds, and wavelengths of up to tens of meters. Effective head gradients resulting from the passage of individual waves decrease with increasing depth or wavelength and with decreasing amplitude. Head gradients may be as high as 0.1 in shallow, high-energy environments, but are defined only over distances of a few tens of meters at most and oscillate from positive to negative with the period of the wave. Equations have been derived to estimate wave-induced pore-water velocities for engineering purposes (Sleath 1970; Liu 1973), but we can use a much simpler approach to demonstrating their significance. If a point on the reef surface experiences an average vertical gradient of 10^{-2} over one half of a wave period and has a K/n value of 10^{-2} m/s, then wave periods on the order of 10 s will induce oscillatory movements of the pore water on the order of 10^{-3} m per cycle. A wave-induced vertical displacement of 1 mm (10^{-3} m) per cycle will, over the course of a day, move about 2 m^3 of water in and out of each m^2 of reef surface. It is flux and mixing of this sort, and probably of this magnitude, that accounts for the residence times of a few days reported by Tribble et al. (this symposium) for shallow pore water. In Fig. 1, this process is schematically represented by the large and equal arrows connecting the shallow subsurface of the reef with the overlying water. Both the fluxes involved and the depth-dependent profiles of pore-water residence times can presumably be modeled quantitatively as a one-dimensional dispersion problem with a time-dependent head gradient. However, we are not aware that any solutions have been developed for this particular application.

Tidal oscillations. While waves have periods and wavelengths that are short compared to most reef processes and structures, the length of a tide wave is far longer than the dimensions of any hydraulically connected reef structure. In spite of the fact that tide ranges (= total head change over time) may be as much as a few meters in some locations, the slowly changing water depths do not directly generate significant hydraulic gradients in submerged reef structures. Oscillatory gradients resulting from tidal signals are primarily interactive effects resulting from short-range perturbations of tide responses by the geometry of local reefs, coasts, or islands. For example, channel restrictions may damp the response of lagoon water level to ocean tide changes, causing gradients across reefs and channels that reverse over the tidal cycle. The resulting oscillating gradients, where they occur, will have values less than 10^{-3}. If we refer to Fig. 1 and postulate an average head difference between lagoon and ocean of 1 cm, reef dimensions on the order of 10^2 m, a maximum estimate of $K/n = 10^{-1}$ m/s, and a tidal cycle duration on the order of 10^4 s, then the linear displacement through the reef surface will be a significant fraction of a meter per day (or about 10^{-2} m^3/m^2 d). Although this is two orders of magnitude less than the flux estimated for the wave mixing, it is quite substantial on a geologic time scale and is particularly significant in view of the fact that it potentially affects the entire volume of high-permeability reef sediments and not just a layer immediately beneath the shallow reef flat. A study of the direct effects of tidal currents on a patch reef (Andrews & Muller 1983) was discussed above.

Tide-groundwater interactions. Tidal oscillations can result in significantly higher gradients in situations where they interact with interstitial fluids within an island, a coastal groundwater body, or a submerged aquifer. Damping of the tide signal within the aquifer will maintain the water table within a much narrower range than the tidal water levels, resulting in gradients in excess of 10^{-2} across the interface between the groundwater body and the open-water marine environment. As shown by Wheatcraft & Buddemeier (1981) and modeled by Herman et al. (1986) and Oberdorfer & Buddemeier (this symp.), this interface may be much more extensive than the intertidal shoreline. In the presence of a high permeability layer beneath the water-table aquifer, the mixing surface may have the same area as the island itself. Oberdorfer & Buddemeier (this symp.) have performed a calibrated numerical simulation of tidal mixing for an Enewetak Atoll groundwater case, and find oscillating, tidally driven pore-water velocites in the range 10^{-5} to 10^{-7} m/s. These mixing forces are important factors in determining the residence times and salinity distributions in groundwater bodies (Buddemeier 1981; Oberdorfer & Buddemeier, this symp.).

Other oscillating forces. Other periodic signals can be identified—internal waves, surf beat, seiches, etc. Although they should not be neglected in any detailed evaluation of a specific site, they are probably not in general very significant pore-water forcing functions because their maximum amplitudes are similar in magnitude to waves and tides and they occur far less frequently and consistently. Wavelengths are relatively long, so transient gradients would reach or exceed values of 10^{-4} only in unusual circumstances. When we consider the lower instantaneous gradients along with their reduced (relative to waves and tides) frequency and duration of occurrence, we see that they are commonly minor sources of energy for pore-water transport. In situations where they occur, the approaches used above for waves and tides may be used to scale and estimate their effects.

Transient events

Infrequent but high-energy events such as major storms, tsunamis, and earthquakes may result in head gradients orders of magnitude higher than those discussed above. However, durations will range from minutes to days (up to a maximum of 10^5 s), and for most reef areas, recurrence frequencies are on the order of 10^{-9} s^{-1} or less. On a long-term geologic scale, such events are unlikely to affect fluxes nearly as much as the more modest head gradients that are maintained 50–100% of the time (ponding, setup, groundwater, density flow, etc.). However, short-term impacts over small distances may be enormous, and there are at least two major effects that are beyond the scope of this discussion. One is alteration of the physical structure of the reef system so as to substantially change its hydrologic characteristics. Islands breached or flooded, channels blocked or reef rubble ramparts formed, sediment redistributed, biological communities killed—all may have long-term implications for pore-water flow patterns. A second effect is chemical: forces that inject large quantities of seawater into normally freshwater or brackish environments or vice versa can change pore-water chemistry in ways that will persist for long periods of time once more normal hydraulic conditions are re-established.

Diffusion

In the absence of advective transport, the distribution of dissolved solutes will be controlled by molecular (or ionic) diffusion according to Eq. (3),

$$F = -D(dC/dx) \quad , \tag{3}$$

where F is the flux of a given species per unit area per unit time, D is the diffusion coefficient of the species in the medium under consideration, and dC/dx is the concentration gradient of the species.

Li & Gregory (1974) have shown that the diffusion coefficients for the major ions in marine sediments are functions of the pore tortuosity and the measured aqueous diffusion coefficients. Their experimentally determined diffusion coefficients for non-reef sediments range from 10^{-10} to 10^{-9} m^2/s. In the discussion that follows, we take the upper end of that range as a reasonable single estimate of the magnitude of diffusion coefficients in reef sediments.

The Peclet number (P) is a dimensionless constant used in hydrology to assess the relative importance of advective and diffusive processes (Perkins & Johnston 1963). Where q is the pore diameter (m) and the other symbols are as defined above, the Peclet Number is given by Eq. (4):

$$P = v_p q/D \quad . \tag{4}$$

Substituting for v_p from Eq. (1), we obtain

$$P = q(K/n)(dh/dl)/D \quad . \tag{5}$$

Where $P \gg 1$, advection is dominant; where $P \ll 1$, diffusion dominates. Pore sizes in coral reef sediments range from $< 10^{-6}$ m (Enos & Sawatsky 1981) up to the order of meters in karst formations (although the latter dimension admittedly strains the normal concept of a "pore"). We have discussed the ranges of values for K, n, and D above; if we select a reasonable range of head gradient values, we can consider the controls on diffusive and advective processes over the possible range of reef pore-water environments.

For illustrative purposes, Fig. 2 plots the family of dh/dl values for P = 1 (approximate balance between advection and diffusion) as functions of log q (m) and log K/n (m/s); D is taken as 10^{-9} m^2/s (see above). The dh/dl values chosen (from unity to 10^{-5}) are a reasonable estimate of the range observed in reef environments. The upper left and lower right corners of the plot are dashed because it is unlikely that any significant pore-water environments occupy these spaces; since hydraulic conductivity and pore size are not completely independent variables, it is highly improbable, for example, that a system containing only very small pores would have a very high hydraulic conductivity.

In Fig. 2, the environments represented by the area below and to the left of a given dh/dl plot will be dominated by diffusion for all values of head gradient equal to or less than the indicated one. In the environments above and to the right of the line, advective motion will dominate. The plot indicates that in sediments with pore

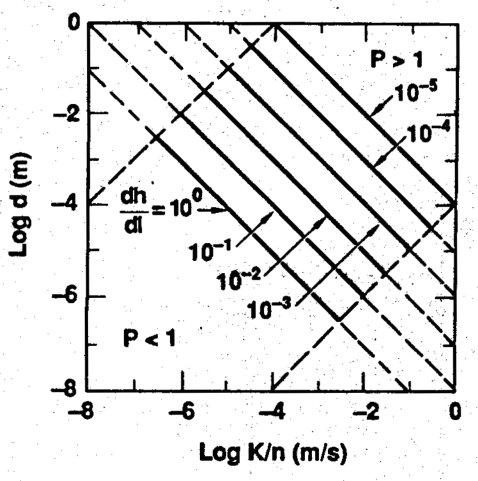

Figure 2. Relative importance of diffusion and advection as a function of pore diameter q (m), hydraulic conductivity divided by effective porosity K/n (m/s), and head gradient dh/dl. The lines for each value of head gradient plot the combinations of pore size and permeability/porosity for which the Peclet number P = 1. Advective motion dominates in environments where P > 1; diffusion dominates if P < 1. A diffusion coefficient of 10^{-9} m^2/s was used to calculate P.

diameters less than a micrometer and/or hydraulic conductivities less than about 10^{-5} m/s, diffusion will always be dominant. Similarly, at the upper end of the pore size and permeability ranges, long-term average gradients as low as a mm/km can cause advection to exceed diffusive transport. For most of the common reef sedimentary environments, both processes must be considered. If we consider the velocities required for P = 1, we see that micrometer pores require a high (but not impossible) velocity of 10^{-3} m/s, but in pores with dimensions of millimeters—and coral reef sediments may be much coarser than this—velocities of 10^{-6} m/s will compete with diffusive transport.

Further insights into the interaction between advection and diffusion may be gained by setting aside the assumption of homogeneous sediments with a narrow range of pore-size distributions upon which Fig. 2 was implicitly based. Since even relatively permeable unconsolidated reef sediments may have total porosities two to three times greater than their effective porosities, the difference betweeen the two porosities must represent the pore volume that exchanges its contents primarily by diffusion, independent of the head gradient or velocity of the water through the larger pores. Most sediments will therefore have significant volumes in which diffusion-controlled processes are occurring even if macroscopic hydrologic observations indicate that conditions are well within the advection-dominated domain. Because of the different time constants of the macropores and the micropores, heterogeneous sediments may have to be flushed with many macropore volumes before the contents of the micropores are exchanged even once.

DISCUSSION

The values derived and listed above are intended to provide a reasonable list of the factors influencing pore-water flow in reef systems and also estimates of the range of magnitudes exhibited by flow derived from the various sources. We have chosen as examples the "maximum credible flow"; locally, some environments may exhibit somewhat higher gradients and velocities, but in general somewhat lower values may be anticipated. The driving forces—tides, currents, wave energy, thermal flux, etc.—are geographically quite variable, and interact with the local reef structures in fashions that are, in detail, unique to each reef. One subject which we have completely disregarded is the relative systematics of permeability and head gradient distributions. For example, reef plate and forereef slope sediments tend to be highly lithified, while deep lagoon sediments are both unconsolidated and fine-grained. Both the physical and the hydrogeologic environments need to be considered in some detail when making interpretations or designing experiments based on pore-water dynamics.

We have discussed a wide—and probably not exhaustive—range of forcing functions. They cover a wide range of environments and time constants. For example, wave-induced mixing within a shallow reef affects only a limited environment, but results in a very rapid turnover of pore water and biogeochemically important solutes. At the other extreme, the slow, density-driven circulations may affect large volumes of sediment over a geologic time scale. Because of the side-by-side existence of macropores and micropores within reef sediments, most sedimentary environments will harbor diffusion-controlled processes coupled to the advectively flushed pore spaces. The overall characteristics of the system, especially with regard to chemical or microbiological processes, will be a complex result of the concurrent diffusive and advective processes.

When we restrict our consideration to the advective processes, we are struck by the fact that so many different processes can potentially result in head gradients (and velocities) of similar magnitudes, and by the fact that they are not mutually exclusive. The forcing functions may reinforce each other, be opposed, or be orthogonal. Individual investigations, including our own previous studies, have tended to focus on individual or a few mechanisms for pore-water displacement. Consider, however, the multiplicity of processes operating at a location like Enewetak. The studies cited have demonstrated groundwater effects, lagoon ponding, and wave setup; the atoll has a known subsurface thermal gradient, and since it is in the tradewind zone we can safely assume that it is subject to large-scale currents as well as to surficial wave mixing. This partial inventory serves to emphasize the difficulty in either assuming or measuring pore-water movement.

In both biological and geochemical fields, residence or flushing times are used to assess the importance of different processes. Specific types of diagenesis, for example, may be argued to require a certain number (e.g., 10^5) of pore-water replacement volumes flushed through the system (Enos & Sawatsky 1981). If something is known about the time available and the range of permeabilities, a minimum head gradient may be deduced. At the shorter end of the time scale, chemical fluxes can be used in combination with the hydrologic flux to evaluate chemical kinetics, microbial populations and/or metabolic rates, and the potential interactions between the pore-water environment and the benthic communities at the surface of the sediment (Johannes 1980). If such flushing times are to be useful, they must be inclusive; considering only a limited subset of processes may be seriously misleading. We also stress that the evaluation must be based on observed reality—if dolomite or a microbial population is present, demonstrations of the impossibility of that presence are useful only if pursued to the point of eliminating all but the actual causative mechanism(s) (Doyle 1887).

Figure 3 provides a family of flushing time curves. K/n and the time (seconds) required to flush 1 m of aquifer length are the axes; the curves plotted are for different values of head gradient. The curves' use can be illustrated by the following example. Consider a hypothetical process known to have recrystallized a formation 100 m thick over the course of 1 My. If the mean K/n is estimated to have been 10^{-3}, and 10^5 volumes of pore water were required by the stoichiometry of the process, what average gradient must have been present? One My is slightly more than 10^{13} s, and 10^5 iterations through 100 m

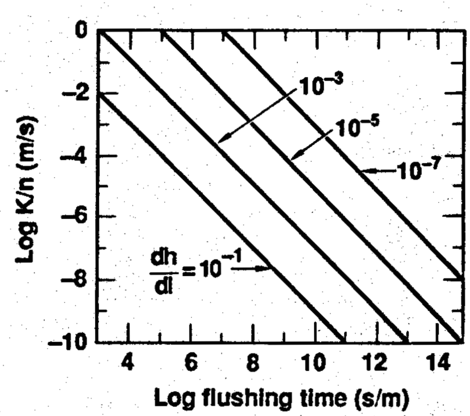

Figure 3. Pore-water flushing time (seconds per meter of path length) in relation to hydraulic conductivity divided by effective porosity (K/n) and the applied head gradient (dh/dl). Relationships are valid only for macropores (effective porosity) in a homogeneous medium, and not for cavernous voids or the diffusion-controlled micropore component of the total porosity.

is equivalent to 10^7 m of path length, so the time axis (necessary inverse velocity) is 10^6 s/m. This value and $K/n = 10^{-3}$ m/s intersect on the line for $dh/dl = 10^{-3}$; any average sum of gradients equal to or greater than this value will produce the required flux. Note that producing the observed effect within the Holocene (which provides only about 10^{11} s) would require an unrealistically high sustained gradient of 0.1—a finding that should provoke re-examination of the assumptions.

The scaling approach to pore-water hydrodynamics yields some important conclusions about experimental approaches to the assessment of pore-water processes. Stommel (1963) has commented on the dangers of using inappropriate scales of observation in oceanographic investigations; the problems are likely to be even greater in pore-water studies. The forces are numerous, variable, and of comparable magnitude, and the environments in which they occur are spatially and temporally heterogeneous. In addition, the signal (pore-water head or velocity) is often small relative to the noise (the pressures and flows generated in the open seawater), and also small in terms of the observational time scales and the instrumental sensitivities of the human observers. These constraints place a premium on thoughtful evaluation of the total system and on careful, complete, and interdisciplinary experimental design.

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