Permeabilities, Fluid Pressures, and Flow Rates in the Barbados Ridge Complex

ELIZABETH J. SCREATON1, DENNIS R. WUTHRICH2, AND SHIRLEY J. DREISS

In accretionary complexes, gravitational and tectonic forces deform highly porous sediments as they are either accreted onto the overriding plate or carried downward with the underthrust plate. Fluids expelled from the compacting sediments influence many aspects of subduction zone geology, including heat transport [e.g., Langseth and Hobart, 1984], diagenesis and metamorphism [e.g., Etheridge et al., 1983], and benthic biology [e.g., Kelm et al., 1986]. In addition, relative rates of sediment loading and fluid dissipation determine the magnitude and distribution of excess fluid pore pressures. These pore pressures affect the shape of the accretionary wedge [e.g., Davis et al., 1983] as well as thrust fault and fold geometries [Hubbert and Rubey, 1959; Seely, 1977].

Despite the importance of fluids at convergent margins, little is known about pore pressure distributions or rates and directions of fluid flow in these settings. The presence of mud diapirs and a few measurements of pore pressures from convergent margins suggest that near-lithostatic pressures may exist within some subduction zones [Moore and von Huene, 1980; Westbrook and Smith, 1983]. However, these data are too scarce to adequately describe the regional pore pressure distribution within any accretionary complex. Estimates of fluid flow rate are rarer, and have been directly measured only during an Alvin dive at a vent site off the Oregon coast [Suess et al., 1987].

The Barbados Ridge Complex is one of the most extensively studied convergent margins. Seismic reflection profiles of the complex reveal a well-developed décollement separating the accretionary prism from subducting sediments [Westbrook et al., 1982]. Both Ocean Drilling Project (ODP) leg 110 and Deep Sea Drilling Project (DSDP) leg 78a recovered sediment samples from the complex. These samples document decreases in porosity as basin sediments are incorporated into the accretionary complex. Observations from DSDP leg 78a also suggest that pore pressures may reach near-lithostatic values just above the décollement within 5 km of the deformation front [Moore and Biju-Duval, 1984]. In addition, some pore water samples collected during ODP leg 110 contained anomalous methane, chloride, and temperature distributions, suggesting preferential fluid flow along fault zones, especially within the décollement [Moore et al., 1987; Gieskes et al., 1989].

In this paper, we present a numerical model of fluid flow in the Barbados Ridge Complex. We used the model and observations from drilling to determine feasible ranges of the intrinsic permeability of the prism décollement, and underthrust sediments. With these permeability estimates, we were able to describe possible pore pressures and fluid flow velocities throughout the complex.

Numerous studies have used models to examine the mechanics of accretionary prisms [e.g., Chapple, 1978; Davis et al., 1983; Stockmal, 1983; Ngokwey, 1984; Zhao et al., 1987; Borja and Dreiss, 1989] but few have considered fluid flow and pore pressure distributions. Bray and Karig [1985] estimated rates of dewatering in the Nankai Trough from compaction as sediments accrete into the prism. Shi and Wang [1988] developed the first model for fluid flow in an accretionary complex and used the model to demonstrate that high pore pressures could develop near the deformation front in the Barbados complex. The fluid flow model in this study is similar to that of Shi and Wang [1988]. The major difference is the manner in which the two approaches incorporate sediment compaction. Shi and Wang [1988] used a transient formulation for flow in deformable porous media and ran simulations to steady state. We estimated rates of compaction directly from porosity measurements and solved a steady state form of the flow equation. This approach is computationally efficient and does not require that the relationship between effective stress and porosity be known or

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Paper number 89JB03006.
0148-0277/90/89JB-03006S05.00
assumed, as it was by Shi and Wang [1988], but instead relies on empirical porosity estimates.

TECTONIC SETTING

The Barbados Ridge Complex, located east of the Lesser Antilles volcanic arc, formed by subduction of Atlantic ocean crust beneath the Caribbean plate (Figure 1). The rate of plate convergence has been estimated to be between 2.0 and 2.2 cm/yr [Macdonald and Holcombe, 1978; Menner and Jordan, 1978; Tovish and Schubert, 1978; Dorel, 1981], with one estimate as high as 4 cm/yr [Sykes et al., 1982]. Stein et al. [1988] reviewed these studies and concluded that a convergence rate of 2.0 cm/yr is most plausible.

A décollement, extending westward more than 100 km, separates offscraped, deformed sediments in the accretionary prism from little deformed, acoustically coherent, underthrust sediments (Figure 2). At the latitude of ODP leg 110, approximately two-thirds of the 700-m section of incoming sediments are underthrust, while one-third is accreted into the prism [Biju-Duval et al., 1984].

At the ODP leg 110 drilling sites, the Barbados Ridge Complex is dominated by fine-grained sediments with low intrinsic permeabilities [Marlow et al., 1984; Moran and Christian, 1989]. Prism sediments are predominantly pelagic calcareous mudstones and hemipelagic clays [Wright, 1984; Moore et al., 1988]. Seismic reflection profiles and core samples suggest that the accretionary prism is made up of fault-bounded packets of deformed, tectonically mixed sediments [Moore et al., 1988].

The décollement at site 671 consists of 40 m of lower Miocene radiolarian mudstones with a well-developed scaly fabric. This zone correlates stratigraphically with sediments at site 672, located 6 km seaward of the deformation front. Low-angle shear zones and normal and reverse faults at site 672 suggest incipient seaward propagation of the décollement within this stratigraphic unit. Arcward of the deformation front, seismic reflection profiles indicate that the décollement follows the same stratigraphic horizon for 12 km then enters a deeper horizon arcward of site 673 [Moore et al., 1988].

Samples from the underthrust sediments below the décollement are primarily hemipelagic and radiolarian-rich mudstones. Seismic reflection images show prominent, subhorizontal reflectors that correspond to extensive, relatively undeformed, silt and sand-rich intervals [Moore et al., 1988].

CONSTRAINTS ON PORE Pressures AND Fluid Flow Directions

Observations during ODP leg 110 and DSDP leg 78a suggest that (1) high pore pressures are present near the deformation front just above the décollement and (2) fluid flow is primarily lateral in the décollement, with small amounts of upward advection through the prism sediments. Moore and Biju-Duval [1984] concluded that pore pressures near the top of the décollement are close to lithostatic near site 542 from the high drilling fluid pressures observed at that site during DSDP leg 78a. In addition, unstable hole conditions and fluidized sediments encountered during both ODP leg 110 and DSDP leg 78a drilling programs imply the existence of high fluid pressures in the toe of the complex.

Anomalously high methane and low chloride concentrations in the décollement at sites 671 and 672 indicate that extensive lateral flow occurs in the décollement zone [Gieskes et al., 1989] (Figure 2). Isotopic evidence suggests that the methane is of

Fig. 1. Location map of drilling sites during ODP leg 110 and DSDP leg 78a [from Moore et al., 1987].
thermogenic origin, forming at temperatures of approximately 90°C, possibly at depths of 3 km (assuming a gradient of 30°C/km) [Vrolijk et al., 1989]. Subducting sediments reach these depths about 45 km arcward of the deformation front, implying that methane-bearing fluids must have migrated similar distances. The fact that the methane and chloride signals are less restricted to the décollement at site 672 than at site 671 suggests some upward leakage, from the décollement into the overlying sediments [Gieskes et al., 1989]. The spike in methane concentration just above the décollement at site 671 also indicates small amounts of upward advection of décollement fluids, possibly along a fault (Figure 2).

**FLUID FLOW MODEL**

We modeled fluid flow in a vertical cross section along seismic line CRV 128 from 10 km seaward of the deformation front to 14 km arcward of the front (Figures 1 and 2). Our analysis focused mainly on the region between site 672, 6 km seaward of the deformation front, and site 673, 12 km arcward of the deformation front. Arcward of site 673, the general geometry of the complex can be estimated from the seismic profile, but little other information is available to constrain porosity and boundary conditions. Placing the model boundaries beyond the drilling sites minimizes the influence of boundary effects and reduces uncertainties in the model simulations in the center of the flow field.

Following the approach of Karig [1985] and Shi and Wang [1988], we considered a vertical cross section in a frame of reference that migrates with the deformation front so that the geometry of the modeled region remains constant over time. We assumed that the mass flux of sediments through the complex is time invariant. Thus both the porosity distribution and sediment dewatering rates do not change with time. We neglected other possible fluid sources such as clay dehydration and other mineral phase changes because sediments near the deformation front are not deeply buried.

**Derivation of Governing Equation**

Modeling fluid flow in deforming saturated sediments requires a special form of the equation for flow through porous media. Two effects must be addressed. First, fluid motion with respect to the moving sediments must be distinguished from the fluid velocity in a fixed coordinate system. This is important since Darcy's law defines fluid velocity with respect to the sediments. Second, the fluids expelled by the compacting sediments must appear as a source term in the flow equation. The correct fluid source term follows from consideration of mass conservation in the compacting sediments. This section contains a derivation of the governing equation for fluid flow in deforming sediments following that of Bear [1972].

In a fixed coordinate system the equation for conservation of fluid mass is

\[
\nabla \cdot (\rho_t \vec{q}) + \frac{\partial (\rho_t n)}{\partial t} = 0
\]

where
- \( \rho_t \) fluid density;
- \( \vec{q} \) specific discharge;
- \( n \) sediment porosity;
- \( t \) time.

Here \( \vec{q} \) is measured with respect to a fixed coordinate system. We may express it as:

\[
\vec{q} = \vec{q}_r + n \vec{v}_s
\]

where \( \vec{q}_r \) is the specific discharge with respect to the moving sediments, and \( \vec{v}_s \) is the sediment velocity. Inserting (2) into (1) gives

\[
\nabla \cdot (\rho_t \vec{q}_r) + \nabla \cdot (\rho_t \vec{v}_s) + \frac{\partial (\rho_t n)}{\partial t} = 0
\]
Now we assume that the fluid is incompressible, or ρ_f is a constant. Using this and expanding the product derivatives in (3) gives
\[ \nabla \cdot \nabla q_s + n(\nabla \cdot v_s) + \frac{\partial n}{\partial t} = 0 \] (4)

A similar expression can be found from the equation for conservation of mass of the solids:
\[ \nabla \cdot (\rho_s (1-n) v_s) + \frac{\partial (\rho_s (1-n))}{\partial t} = 0 \] (5)

In this case we assume that the individual sediment grains are incompressible, or ρ_s is a constant. With this assumption, expanding the product derivatives in (5) yields
\[ (1-n) \nabla \cdot v_s = \nabla n + n(\nabla \cdot v_s) + \frac{\partial n}{\partial t} = 0 \]
or
\[ \nabla \cdot v_s = \nabla n + n(\nabla \cdot v_s) + \frac{\partial n}{\partial t} \] (6)

Finally, substituting the left hand side of (6) for the last three terms in (4) gives us the desired result:
\[ \nabla \cdot v_s + \nabla \cdot v_s = 0 \] (7)

Note that this equation gives specific discharge with respect to the moving sediments. This permits the use of Darcy's law to obtain
\[ \nabla \cdot \left( -\frac{k}{\mu} \nabla p + \rho_s g \right) + \nabla \cdot v_s = 0 \] (8)

where p is the fluid pressure, g is the vertical gravity acceleration vector, k is the intrinsic permeability of the sediments, and μ is the dynamic viscosity of the fluid.

Solving equation (8) requires knowledge of the fluid pressure or specific discharge along the boundaries of the problem domain. In the two-dimensional Barbados model, the boundaries are the seafloor, the ocean basement, and the seaward and arcard edges of a cross section running perpendicular to the strike of the deformation front. The sediment permeability distribution must also be estimated. Finally, solution of (8) requires that the term \( \nabla \cdot v_s \) be known. This term represents the volume of water released by the compacting sediments and can be computed if the sediment porosity distribution throughout the complex is known.

Estimation of \( \nabla \cdot v_s \)

Because sediments deform differently within the prism, décollement, and underthrust sediments, we computed \( \nabla \cdot v_s \) separately for each of the three domains. We assumed that the rates of motion, thickness, and physical properties of sediments entering each domain at the seaward boundary remain constant over time. Sediments stay within each domain and exit only at the arcward boundary. In addition, the geometries and porosity distributions of the prism, décollement, and underthrust sediments are constant for the frame of reference fixed to the deformation front. Furthermore, we assumed that the horizontal components of the rates of sediment movement along a profile are vertically uniform within each domain, although they differ between the three regions. Karig [1985] and Screaton et al. [1989] discuss the basis for these assumptions.

Computation of rates of porosity loss entailed several steps. First, we developed a porosity distribution for each of the three domains from the geometry of the complex and borehole porosity data. We then calculated rates of sediment movement in the prism and décollement, using the conservation of mass of solids. Finally, we computed \( \nabla \cdot v_s \) at discrete nodes throughout the prism, décollement, and underthrust sediments.

Porosity distribution. To describe the porosity distribution in the prism and décollement, we fit two porosity-depth relationships, similar to Athy’s [1930] equation, to available porosity measurements. Because of the scarcity of data from sediments below the décollement, we estimated porosities in the underthrust sequence with a mass-balance approximation [Screaton et al., 1989]. Figure 3 illustrates the estimated porosity distribution. Contours in the underthrust sequence represent vertical averages of porosities for the entire sequence. Prism and décollement sediments enter the complex with higher porosities than the average porosity in the underthrust sequence. However, as the prism thickness increases arcward, estimated porosities in the décollement and at the base of the prism decrease more rapidly than the average porosity in the underthrust sequence.

Rates of sediment movement. Calculations of rates of sediment movement within the prism and décollement follow from the mass conservation of solids (see the appendix). Figure 4 shows...
how the estimated rates of sediment movement, normalized by
the mean horizontal component ($V_{x_0}$) of the incoming
sediments at the seaward boundary, vary with distance from the
deforation front. The rate of movement of the prism sediments
at the deformation front is slightly less than $V_{x_0}$ because we
assumed a small amount of thickening between sites 672 and the
deforation front. In the prism, the greatest decrease in the rates
of motion occurs in the first 4-5 km from the deformation front.
The rate of movement decreases to approximately 30-40% of
$V_{x_0}$ by this distance. In the next 10 km, the rate decreases less
rapidly to about 10% of $V_{x_0}$ at 14 km from the deformation
front.

Karig [1985] estimated rates of sediment movement within the
accretionary prism at the Nankai Trough, using the instantaneous
decrease in sediment velocities (the slope of the curves in Figure
4) as a measure of horizontal shortening at a point in the prism.
The Nankai Trough has a similar convergence rate to this portion
of the Barbados Ridge Complex, a similar angle between the
trench slope and the oceanic crust, but a greater thickness of
incoming sediments [von Huene and Lee, 1983]. As in
Barbados, the instantaneous rate of shortening in the Nankai
prism is greatest in approximately the first 5 km from the
deforation front. As illustrated in Figure 4, less thickening and
slowing of sediments has been estimated for the Nankai prism,
because the incoming thickness of sediments is greater than that
of Barbados while prism geometries are similar.

Rates of porosity loss and fluid expulsion. Using the porosity
distribution in Figure 3 and computed rates of sediment
movement, porosity loss and fluid expulsion rates can be
estimated throughout the complex with a discrete approximation
of $V-V^*_{x_0}$ (see the appendix). Figure 5 illustrates the computed
fluid sources. The highest rates of fluid expulsion occur in the
prism and décollement within the first 4 km arcward of the
deforation front. Computed rates of porosity loss in the
underthrust sediments are uniform because we assigned linear
boundaries to the top and bottom of this domain, and used
constant rates of sediment movement beneath the décollement.
Beyond about 4 km arcward of the deformation front, the highest
estimated rates of fluid expulsion occur in the underthrust
sediments.

Total rates of fluid released by compaction in the prism,
décollement, and underthrust sediments are summarized in Table
1. Table 1 also shows estimates of the total amount of pore space
within the bulk sediment that is entering at the seaward boundary
of each domain, assuming an incoming velocity, $V-V^*_{x_0}$ of 2
cm/yr. The amount of pore space entering the seaward boundary
in the underthrust sediment is approximately 1.8 times the
amount entering with sediments above the incipient décollement.

| TABLE 1. Sum of Estimated Fluid Expulsion Rates and Fluid Fluxes Across Model Boundaries |
|--------------------------------------------------|--|--|--|
| **Prism** | **Décollement** | **Underthrust** |
| **Total rate of fluid release** | $4.87 \times 10^8$ | $1.53 \times 10^8$ | $6.57 \times 10^8$ |
| m$^3$/yr | 1.53 | 0.48 | 2.07 |
| **Total pore space crossing seaward boundary, $nV_{x_0}$, m$^3$/s** | $-8.78 \times 10^8$ | $-2.09 \times 10^8$ | $-1.58 \times 10^7$ |
| m$^3$/yr | -2.77 | -0.66 | -4.99 |
| **Total fluid flux across seaward boundary, $q_x + nV_{x_0}$, m$^3$/s** | $-8.76 \times 10^8$ | $1.40 \times 10^9$ | $-1.40 \times 10^7$ |
| m$^3$/yr | -2.76 | 0.04 | -4.41 |
| **Total pore space crossing arcward boundary, $nV_{x_0}$, m$^3$/s** | $-4.17 \times 10^8$ | $-6.24 \times 10^9$ | $-9.28 \times 10^8$ |
| m$^3$/yr | -1.32 | -0.20 | -2.93 |
| **Total fluid flux across arcward boundary, $q_x + nV_{x_0}$, m$^3$/s** | $-4.17 \times 10^8$ | $6.06 \times 10^9$ | $-1.90 \times 10^8$ |
| m$^3$/yr | -1.32 | 0.19 | -0.60 |

Negative values indicate flow in the arcward direction. Values are results for the case where $k_x$ varies with depth, $k_x = 10^{-14}$ m$^2$,
and $k_x = 10^{-15}$ m$^2$. 
permeabilities are largely unknown, we performed a sensitivity analysis with the model to identify ranges of permeability that are consistent with drilling observations.

**Boundary conditions.** We assumed fluid pressures at the seafloor are hydrostatic (Figure 7). The underlying oceanic crust was treated as a no flow boundary. By specifying the lower boundary in this way, we assumed that any fractures within the oceanic basement are insignificant avenues of flow and that fluid production or withdrawal due to mineral phase changes at shallow crustal levels are negligible.

We specified a hydrostatic pressure profile along the seaward boundary, 10 km east of the deformation front. This assumes that pore fluids in the sediments rafted along the North American plate have not been overpressured before reaching this boundary or that the boundary is far enough seaward to not affect simulated pore pressures in the toe of the complex.

Neither fluid pressures nor flow rates are known at the arcward boundary of the model. We treated this boundary differently for the prism, the décollement, and the underthrust sediments. Horizontal fluid fluxes at boundary nodes were defined by a discrete form of \( q_{rx} = q_x - n v_{sx} \) where \( q_x \) is the horizontal component of specific discharge in an absolute frame of reference, \( n v_{sx} \) is the fluid-filled pore space crossing the arcward boundary with the moving sediments, and \( q_{rx} \) is the horizontal, pressure-driven flow relative to the solid matrix, as given by Darcy's law. Along the arcward boundary of the prism, we assumed that flow relative to the solid matrix is primarily vertical toward the ocean floor. Thus the prism boundary is a vertical flow line and is treated as a no flow boundary where \( q_{rx} = 0 \).

We treated the arcward edge of the décollement and underthrust sediments as a specified flux boundary and estimated \( q_x \) to be a flux resulting from compaction of underthrust and décollement sediments beyond the arcward boundary. We assumed that actual boundary fluxes fall between two end-members: a minimum of no pressure-driven flow, or \( q_{rx} = 0 \) and a maximum flux that would occur if the sediments compact to a porosity of 0.10. In estimating the maximum flux, we assumed that the thickness of both the décollement and underthrust sequence continue to decrease uniformly, and that released water does not leave either domain before it crosses the model boundary. In most of the model runs, the boundary flux was assumed to be approximately half of the maximum estimated flux, or a total of 2.7 m³/yr distributed uniformly along the arcward edge of the underthrust and décollement sediments (Table 1).

**Preliminary estimates of permeability.** Because of the scarcity of permeability data, we initially assigned uniform intrinsic permeabilities within each domain. Permeabilities of discrete features, such as individual fault zones in the prism, cannot be specified because the model assumes a constant prism geometry throughout time whereas the position of faults relative to the deformation front may not be constant. Instead, the assigned values for the prism are equivalent permeabilities \( (k_p) \) that represent both fault zone and intergranular permeabilities.

**Numerical Solution**

Using the estimated source terms, \( V \cdot v = S \), we solved equation (8) numerically to find pore pressures and fluid flow velocities. We used SUTRA, a finite element code developed by Voss [1984], and a finite element grid of 442 elements and 490 nodes (Figure 7). The smaller elements along the décollement and in the toe of the prism increase numerical accuracy in these regions and preserve the properties of zones with different hydraulic characteristics. As a check on the grid, we halved the grid spacing perpendicular to the décollement and observed that computed pore pressures did not change significantly.

To solve (8), we first specified boundary conditions at the seafloor, the ocean basement, and the seaward and arcward edges of the finite element grid. In addition, the intrinsic permeability of the sediments had to be estimated. Because sediment permeabilities are largely unknown, we performed a sensitivity analysis with the model to identify ranges of permeability that are consistent with drilling observations.

**Fig. 6.** Computed cumulative pore volume flux (for a convergence rate of 0.02 m/yr) as a function of distance from the deformation front for the prism, décollement, and underthrust domains. The prism and décollement sediments lose porosity most rapidly near the deformation front, while the underthrust sediments maintain a relatively uniform rate of fluid expulsion with distance from the deformation front.

**Fig. 7.** Finite element grid and boundary conditions for fluid flow model.
Similarly, the equivalent permeability of the décollement \( (k_d) \) represents the fracture permeability of scaly mudstones found within this zone. Equivalent permeabilities of the underthrust sediments \( (k_u) \) are average permeabilities of sands and mudstones below the décollement. A steady state formulation of fluid flow cannot simulate transient permeability changes such as the episodic flow in the complex that has been suggested by several researchers [e.g., Fisher, 1987]. Instead, the permeabilities assigned in our model represent temporal as well as spatial averages.

We estimated feasible ranges of permeability for each domain, the prism, décollement, and underthrust sediments, from measurements made on similar sediments by other researchers [e.g., Brace, 1980; Morin and Silva, 1984; Morrow et al., 1984; Neuzil, 1986; Taylor and Leonard, 1989; Bryant et al., 1975]. We purposely chose a wide range of plausible values for each domain to ensure that no reasonable permeability estimates were excluded (Table 2). Wuthrich et al. [1989] discuss in detail the basis for these estimates.

**Results**

A number of fluid flow fields can be generated using the ranges of permeability values in Table 2. However, not all of the possible permeability values produce results that agree with the drilling observations. We now examine how computed flow fields vary for different assumed permeability values and compare simulation results to constraints from drilling observations.

**Effect of permeability on fluid pressures and flow directions.**

The equivalent permeability of the prism sediments \( (k_p) \) and the contrast between the equivalent décollement and prism permeabilities \( (k_d/k_p) \) greatly affect the magnitude of simulated pore fluid pressures at site 542. Fluid pressures may be conveniently described using \( \lambda * \), the excess pore pressure ratio.

The \( \lambda * \) represents the ratio of excess fluid pressures to the sediment overburden pressure [Shi and Wang, 1988]:

\[
\lambda * = \frac{(p - P_{\text{hydro}})}{(P_{\text{ litho}} - P_{\text{ hydro}})}
\]

where

\( p \) = pore fluid pressure;
\( P_{\text{ hydro}} \) = hydrostatic fluid pressure;
\( P_{\text{ litho}} \) = lithostatic pressure.

For lithostatic pore pressures, \( \lambda * = 1 \); at hydrostatic pressures, \( \lambda * = 0 \).

Figure 8 illustrates simulated \( \lambda * \) values at the top of the décollement at site 542 for a range of prism permeabilities and \( k_d/k_p \) contrasts. Equivalent prism permeabilities of \( 10^{-18} \) to \( 10^{-19} \) m² (hydraulic conductivities of \( 10^{-9} \) to \( 10^{-10} \) cm/s) produce near-lithostatic (0.8 < \( \lambda * < 1 \)) pore pressures at site 542, while equivalent prism permeabilities of \( 10^{-20} \) m² or lower produce unrealistically high pore pressures (\( \lambda * >> 1 \)). In contrast, use of effective prism permeabilities greater than \( 10^{-17} \) m² results in almost hydrostatic pore pressures.

Increasing the the décollement permeability, or the \( k_d/k_p \) contrast, while holding the prism permeability constant lowers fluid pressures (Figure 8). Thus a relatively permeable décollement can dissipate fluid pressures within a low permeability accretionary complex. The permeability contrast between the décollement and the underthrust sequence \( (k_d/k_u) \) does not affect the pressures as much as varying the prism permeability \( (k_p) \). For example, a \( k_d/k_u \) contrast of 3 orders of magnitude has less effect on fluid pressures at site 542 than an order of magnitude change in prism permeability.

The permeability contrasts between the décollement and prism \( (k_d/k_p) \), and between the décollement and the underthrust sediments \( (k_d/k_u) \) also influence fluid flow directions within the complex. If the \( k_d/k_p \) contrast is greater than 5 orders of magnitude, water flows downward from the prism into the décollement, which is inconsistent with methane and chloride

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**TABLE 2. Estimates of Intrinsic Permeability**

<table>
<thead>
<tr>
<th>Domain</th>
<th>Initial Estimates, ( \text{m}^2 )</th>
<th>Fitted Values, ( \text{m}^2 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Prism ( k_p )</td>
<td>( 10^{-17} - 10^{-22} )</td>
<td>( 10^{-17} - 10^{-19} )</td>
</tr>
<tr>
<td>Décollement ( k_d )</td>
<td>( 10^{-11} - 10^{-22} )</td>
<td>( 10^{-14} - 10^{-16} )</td>
</tr>
<tr>
<td>Underthrust sediments ( k_u )</td>
<td>( 10^{-12} - 10^{-22} )</td>
<td>( 10^{-15} - 10^{-16} )</td>
</tr>
</tbody>
</table>
The relatively permeable décollement acts as a drain, drawing water from the prism and preventing the upward migration of methane-bearing décollement fluids. To simulate predominantly horizontal flow in the décollement, the contrast in equivalent permeabilities between the prism and décollement must be at least 3 orders of magnitude. Lower contrasts between the two zones result in a large component of upward flow through the prism at site 542, which is inconsistent with the presence of the methane anomalies observed within the toe. Thus the permeability contrast between the décollement and the prism ($k_d/k_p$) must be between $10^3$ and $10^5$ to match observations. The $k_d/k_u$ contrast does not significantly affect flow directions in the décollement as long as the underthrust sediment permeabilities are lower than the décollement permeability.

Table 2 summarizes the combinations of intrinsic permeabilities that produce flow patterns and fluid pressures consistent with drilling observations. We also examined the sensitivity of the model results to variations in the amount of influx specified on the arcward boundary and the porosity distribution of the prism and décollement [Wuthrich et al., 1989]. Overall, simulated pore pressures and $\lambda^*$ values are much less sensitive to plausible changes in porosity distribution and flux at the arcward boundary than to changes in prism and décollement permeabilities.

**Variable prism permeabilities.** The assumption of uniform prism permeabilities is useful for an initial demonstration of the importance of the contrast between prism and décollement permeabilities. However, simulations with uniform prism permeabilities which otherwise match the flow constraints, produce $\lambda^*$ values in the prism that are unrealistically high ($\lambda^* > 1$) (Figure 9). We next allowed $k_p$ to change with depth according to an empirical permeability-porosity relationship for smectite developed by Morin and Silva [1984]. The resulting intrinsic permeability values vary between about $10^{-16}$ m$^2$ near the surface of the wedge to less than $10^{-18}$ m$^2$ at the base of the wedge near the arcward prism boundary (Figure 10).

An examination of the $\lambda^*$ computed using depth-varying $k_p$ and different $k_d$ values further demonstrates the sensitivity of the computed flow field to the permeability of the décollement. In all three of the examples in Figure 11, $k_p = 10^{-16}$ m$^2$, and the total arcward boundary flux along the décollement and underthrust sequence is 2.7 m$^3$/yr. When $k_d = 10^{-14}$ m$^2$, the pore pressure ratio reaches a maximum at the top of the décollement beneath the deformation front (Figure 11a). If $k_d$ is an order of magnitude higher, fluids drain rapidly through the décollement and high pore pressure ratios do not develop (Figure 11b). If $k_d = 10^{-15}$ m$^2$, the décollement ceases to act as an effective zone of preferential flow and $\lambda^*$ values become unrealistically high near the arcward boundary in both prism and underthrust sediments.
Thus, if the permeability of the prism varies with porosity as described by Morin and Silva [1984], the intrinsic permeability of the décollement must be about $10^{-14}$ m$^2$.

Figure 12 illustrates pore pressures and fluid flow velocities for a case in which $k_p$ varies with depth, $k_d = 10^{-14}$ m$^2$ and $k_u = 10^{-15}$ m$^2$. The pore pressures increase arcward and with depth and are greatest near the arcward boundary of the underthrust sediments (Figure 12a).

Figures 12b and 12c show fluid flow velocities for two frames of reference. The velocities in Figure 12b were found directly from computed fluid pressures and Darcy’s law:

$$\mathbf{v}_t = \frac{k}{\eta \mu} (\mathbf{V}_p - \rho g)$$

where $\mathbf{v}_t$ is the average linear velocity relative to the solid matrix and does not include tectonic transport of pore fluids. Water originating in the décollement and underthrusting sediments flows laterally toward the seaward boundary. In contrast, water expelled from prism sediments flows upward to the ocean floor.

The pressure-driven fluid velocities in the prism and underthrust sediments are sufficiently slow that the movement of the solid matrix should be considered in the net transport of pore fluids relative to the deformation front. The absolute fluid velocity, including tectonic transport, is

$$\mathbf{v}_{abs} = \mathbf{v}_t + \mathbf{v}_{df}$$

where $\mathbf{v}_{df}$ is the rate of propagation of the deformation front. Assuming $\mathbf{v}_{df}$ is small, $\mathbf{v}_{abs}$ is approximately the sum of the relative fluid velocity, $\mathbf{v}_t$ and the velocity of the solids, $\mathbf{v}_s$. As illustrated in Figure 12c, water in the prism and underthrust sediments moves arcward relative to the deformation front because pressure-driven flow rates are smaller than rates of tectonic transport. However, in the décollement, pore fluids migrate seaward in spite of tectonic transport.

**SUMMARY**

**Intrinsic Permeabilities**

Near-lithostatic pore pressures can be simulated in the décollement at site 542 if the décollement permeability is 3-5 orders of magnitude greater than that of adjacent prism sediments. Lower $k_d/k_p$ contrasts result in substantial upward flow from décollement into the prism, while higher contrasts produce downward leakage from the prism into the décollement. The permeability of the underthrust sequence does not affect the fluid pressure at site 542 as much as the $k_d/k_p$ contrast. If the permeability of the prism varies with porosity as described by Morin and Silva [1984], physically realistic $\lambda^*$ values require that $k_d$ must be about $10^{-14}$ m$^2$.

**Fluid Flow Directions and Rates**

Water released by sediment compaction flows laterally seaward and upward to the ocean floor. However, because fluid flow velocities relative to the solid matrix are small, pore waters in prism and underthrust sediments actually move arcward relative to the deformation front. In our simulations, pore waters migrate seaward in spite of tectonic transport only in the décollement.

**Fluid Pressures**

Computed fluid pressures increase with depth and in the arcward direction from the deformation front. However, excess pore pressure ratios, $\lambda^*$, reach a maximum in the décollement near the toe of the prism then decrease arcward as the overlying prism thickens. Average $\lambda^*$ values are lower in the underthrust sediments than in the décollement, although pore pressures are higher (Figure 12a). Because hydraulic heads are higher in the underthrust sediments, water flows into the décollement from below. The greater $\lambda^*$ values in the décollement may facilitate faulting between the prism and underthrust sediments. Thus the décollement accommodates relatively high rates of fluid flow while defining the preferred region of decoupling between overriding and underriding sediments.
**APPENDIX A: METHOD FOR ESTIMATING RATE OF FLUID PRODUCTION**

Calculations of rates of sediment movement within the prism and décollement follow from the conservation of mass of solids. From the previously discussed assumptions, the total flux of solids across any vertical plane in the domain does not change with distance from the deformation front. Thus we can equate solid flux at any two distances from the deformation front and solve for the horizontal component of the rate of sediment movement

\[
\bar{V}_{x_{i+1}} = \Delta h_i \bar{V}_{x_i} \left( \frac{1 - \bar{n}_i}{\Delta h_i (1 - \bar{n}_i)} \right)
\]

where

\[
\bar{V}_{x_i} \quad \text{mean horizontal component of rate of sediment movement for the vertical plane at } i; \\
\Delta h_i \quad \text{height of the vertical plane at } i \text{ through the prism or décollement;} \\
\bar{n}_i \quad \text{the mean porosity for the vertical plane at } i.
\]

Using the porosity distribution in Figure 3 and the computed rate of sediment movement obtained above, we can estimate the porosity loss or fluid expulsion rates throughout the complex with a discrete approximation of \( \nabla \cdot V \cdot \bar{n} \). The geometry of a computational cell, shown in Figure 13, is defined such that solid influx across the seaward boundary equals the solid outflux across the arcward boundary. This constraint allows us to iteratively compute the required seaward and arcward height of a cell such that solid mass is conserved. The rate of porosity loss within each cell is then approximated by [Scranton et al., 1989]

\[
(n_{i+1/2} \Delta x_{i+1/2} \Delta z_i - n_{i} \Delta x_{i} \Delta z_i) - (n_{i+1/2} \Delta x_{i+1/2} \Delta z_{i+1/2})
\]

where \( b \) is a unit thickness perpendicular to the cross section; \( \Delta z_i \) is the height of the cell at \( i \); and \( \bar{n}_i \) is the mean porosity over \( \Delta z_i \):

![Diagram of computational cell for estimating mass balance of solids and loss of pore volume.](image)

Fig. 13. Computational cell for estimating mass balance of solids and loss of pore volume. Positions \( i + 1/2 \) and \( i-1/2 \) at \( j \) are located at mid points between nodes. Height of cell at \( i + 1/2 \) and \( i - 1/2 \) is defined such that fluxes of solids across seaward and arcward boundaries are equal.

**Acknowledgments.** This study was supported by National Science Foundation grant OCE-8609745. Discussions with Mark Reid, Chris Neuzil, and Clifford Voss provided useful insights in the early stages of this investigation. J. Casey Moore, John Bredehoefl, Hedeef Essaid, Joris Gieskes, and two anonymous reviewers contributed many helpful comments and suggestions. We want to especially acknowledge Barbara Bekins and David Rogers for their aid in clarifying the mathematical derivations and enhancing the fluid flow simulations.

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(Received December 27, 1988; revised September 5, 1989; accepted September 18, 1989.)