Insights on Slab-driven Mantle Flow from Advances in Three-dimensional modelling

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Abstract

The wealth of seismic observations collected over the past 20 years has raised intriguing questions about the three-dimensional (3D) nature of the mantle flow field close to subduction zones and provided a valuable constraint for how the plate geometry may influence mantle flow proximal to the slab. In geodynamics, there has been a new direction of subduction zone modelling that has explored the 3D nature of slab-driven mantle flow, motivated in part by the observations from shear wave splitting, but also by the observed variations in slab geometries worldwide. Advances in high-performance computing are now allowing for an unprecedented level of detail to be incorporated into numerical models of subduction. This paper summarizes recent advances from 3D geodynamic models that reveal the complex nature of slab-driven mantle flow, including trench parallel flow, toroidal flow around slab edges, mantle upwelling at lateral slab edges, and small scale convection within the mantle wedge. This implies slab-driven mantle deformation zones occur in the asthenosphere proximal to the slab, wherein the mantle may commonly flow in a different direction and rate than the surface plates, implying laterally variable plate-mantle coupling. The 3D slab-driven mantle flow can explain, in part, the lateral transport of geochemical signatures in subduction zones. In addition, high-resolution geographically referenced models can inform the interpretation of slab structure, where seismic data

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are lacking. The incorporation of complex plate boundaries into high-resolution, 3D numerical models opens the door to a new avenue of research in model construction, data assimilation, and modelling workflows, and gives 3D immersive visualisation a new role in scientific discovery.

Keywords: subduction, mantle dynamics, lithospheric deformation, numerical modelling, high performance computing

1. Introduction

Geologic observations (Casey and Dewey, 1984; Bird, 2003; Eberhart-Phillips et al., 2003), geochemical constraints (Hoernle et al., 2008; Gazel et al., 2009; Heyworth et al., 2011; Durance et al., 2012), and geophysical data (Russo and Silver, 1994; Gudmundsson and Sambridge, 1998; Fischer et al., 2000; Smith et al., 2001; Syracuse and Abers, 2006; Long and Silver, 2008; Müller et al., 2008) from subduction zones indicate that the two-dimensional paradigm for plate tectonic boundaries is no longer adequate to explain the observations. For example, inspection of global tectonic features reveals lateral changes in trench geometry (Bird, 2003), lateral and radial changes in slab morphology (Jarrard, 1986; Gudmundsson and Sambridge, 1998; Gutscher et al., 2000; Grand, 2002; Lallemand et al., 2005; Syracuse and Abers, 2006; Li et al., 2008; Hayes et al., 2012), and lateral gradients in seafloor age, in both modern and ancient times (Sdrolias and Müller, 2006; Müller et al., 2008) (Figure 1). Subduction zones can have (a) a high radius of curvature, such as in the Lesser Antilles and Scotia subduction zones, (b) vary significantly in slab dip along strike, such as in Japan, Alaska, and South America, (c) have tears or slab gaps, as in the Middle America-South America subduction zones, and (d) have complex interactions at triple junctions (Figure 1) (Jarrard, 1986; Gudmundsson and Sambridge, 1998; Ratchkovski and Hansen, 2002; Lallemand et al., 2005; Syracuse and Abers, 2006;

27 Nakajima and Hasegawa, 2007; Li et al., 2008; Hayes et al., 2012; Zhao et al., 2012;
28 Dougherty and Clayton, 2014; Scire et al., 2016; Syracuse et al., 2016).

29 The wealth of seismic observations collected over the past 20 years has raised
30 intriguing questions about the three-dimensional nature of the mantle flow field close
31 to subduction zones (Russo and Silver, 1994; Hall et al., 2000; Peyton et al., 2001;
32 Smith et al., 2001; Hoernle et al., 2008; Zandt and Humphreys, 2008; Long and Silver,
33 2008) and provided a valuable constraint for how the subducting plate geometry may
34 influence mantle flow proximal to the slab (Buttles and Olson, 1998; Kneller and van
35 Keken, 2007; Zandt and Humphreys, 2008; Jadamec and Billen, 2010; Song and
36 Kawakatsu, 2013). In the warm upper mantle, where A-, C-, or E-type olivine fabric
37 are expected to dominate, the deformation of olivine by dislocation creep can lead to
38 lattice preferred orientation that results in splitting of seismic shear waves, wherein
39 the azimuth of the seismic fast axis tracks the direction of shear within the mantle
40 (Nicolas and Christensen, 1987; Savage, 1999; Tommasi et al., 1999; Kaminiski and
41 Ribe, 2002; Karato et al., 2008; Long and Silver, 2009; Long and Becker, 2010).
42 Away from plate margins, in the center of the oceanic plates, the direction of plate
43 motion and the azimuth of the fast seismic axes measured from shear wave splitting
44 observations tend to be sub-parallel (Becker et al., 2003; Conrad et al., 2007; Kreemer,
45 2009; Conrad and Behn, 2010), implying simple shearing flow and alignment of the
46 mantle flow field with surface plate motion.

47 However, near subduction zones, the azimuth of the fast axes of the polarized
48 waves are commonly oriented non-parallel to the direction of local plate motion
49 (Figure 2) (Russo and Silver, 1994; Fischer et al., 1998; Peyton et al., 2001; Smith
50 et al., 2001; Pozgay et al., 2007; Long and Silver, 2008; Abt et al., 2010; Long and
51 Becker, 2010; Hanna and Long, 2012; Long and Wirth, 2013; Long, 2013). This
52 implies that the warm upper mantle proximal to subduction zones locally flows in

53 a different direction than the tectonic plates (Long, 2013) and possibly at a greater
54 speed (Conder and Wiens, 2007; Hoernle et al., 2008; Jadamec and Billen, 2010;
55 Stadler et al., 2010).

56 It is important to point out, that, because the splitting signal is cumulative along
57 the ray-path, teleseismic observations recorded near a subduction zone can include
58 effects from the subslab region, from within the slab, from the cold region of the
59 inner wedge corner, and from the overlying plate, as well as from within the dynamic
60 mantle wedge (Jung and Karato, 2001; Kneller et al., 2007; Faccenda et al., 2008;
61 Hammond et al., 2010; Song and Kawakatsu, 2012). Studies that use local S waves,
62 sourced from within the slab, are thus very valuable in resolving anisotropy of mantle
63 wedge origin, as they do not sample subslab material, and many of these studies do
64 show seismic fast axes oriented in a different direction than surface plate motions,
65 implying the decoupling of motion (Figure 2) (Yang et al., 1995; Fouch and Fischer,
66 1996; Smith et al., 2001; Audoin et al., 2004; Pozgay et al., 2007; Abt et al., 2010;
67 MacDougall et al., 2012; Long and Wirth, 2013). The local studies will, however,
68 still sample mantle and lithospheric material affected by water, melt, temperature,
69 and discontinuities that can also contribute to the splitting pattern (Savage, 1999;
70 Jung and Karato, 2001; Holtzman et al., 2003; Crampin and Peacock, 2005; Kneller
71 et al., 2007; Karato et al., 2008; Long, 2013). Thus, the ability to quantitatively link
72 the mantle flow field with a specific observed splitting pattern for a given location is
73 still actively being studied.

74 In geodynamics, there has been a new direction of subduction zone modelling
75 that has explored the three-dimensional nature of mantle flow in subduction zones,
76 motivated in part by the observations from shear wave splitting, but also by the ob-
77 served variations in slab geometries worldwide. The 3D models show that although
78 the 2D models can provide important physical realism, 2D models cannot capture

79 the truly 3D features of the mantle flow field that can occur at subduction zones, due
 80 to toroidal flow around slab edges (Stegman et al., 2006; Jadamec and Billen, 2010),
 81 lateral changes in slab geometry and buoyancy (Kneller and van Keken, 2007; Capi-
 82 tanio and Faccenda, 2012), small scale convection within the mantle wedge (Honda
 83 and Saito, 2003; Wirth and Korenaga, 2012), lateral gravitational instabilities above
 84 the mantle wedge (Behn et al., 2007), and lateral changes in thickness of the upper
 85 plate (Miller and Becker, 2012; Rodríguez-González et al., 2014). Three-dimensional
 86 slab-driven flow can be fundamental to understanding geophysical and geochemical
 87 processes in subduction systems (Kneller and van Keken, 2007; Hoernle et al., 2008;
 88 Durance et al., 2012). However, incorporating the detailed slab complexities has
 89 previously been out of reach of most computational infrastructure and software.

90 Therefore, one might ask, how feasible is it to model the geometric complexities of
 91 plate boundaries and the tectonic processes at slab edges, tears, and triple-junctions?
 92 Advances in high-performance computing (HPC) are now allowing for an unprece-
 93 dented level of detail to be incorporated into numerical models of plate boundary
 94 dynamics (Jadamec and Billen, 2010; Stadler et al., 2010; Jadamec et al., 2013; Rudi
 95 et al., 2015; Hu et al., 2016). Advances in numerical algorithms combined with the
 96 increase in computational resources are allowing high-resolution modelling and the
 97 associated examination of three-dimensional, non-linear mantle flow and slab defor-
 98 mation (Zhong et al., 2007; Burstedde et al., 2009; Jadamec and Billen, 2010; Stadler
 99 et al., 2010; Burkett and Billen, 2010; Liu and Stegman, 2012; Rudi et al., 2015; Hu
 100 et al., 2016). In addition, the coupling of the HPC with immersive 3D data visu-
 101 alisation now allows the user to bring physical meaning to the hundreds of millions
 102 of numbers that numerically represent the model, enabling the numerical modeler
 103 to grasp the model components as would a laboratory modeler working with a tan-
 104 gible physical construct (Kreylos et al., 2006; Billen et al., 2008; Chen et al., 2008;

105 Jadamec et al., 2012).

106 In this paper, I show examples of this process and demonstrate how the increase
107 in computational power has enabled a new avenue of geodynamics research, and
108 in 3D geographically referenced regional modelling of slab-driven mantle flow in
109 particular. It is not that having greater computational resources only allows the
110 same model to be run at higher resolution. Rather, the paper posits that with the
111 greater computational resources, new questions can be asked, and hence a new kind
112 of scientific discovery can be conducted in subduction zone research.

113 **2. Increase in Computing Capabilities Enables a New Direction in Sub-** 114 **duction modelling**

115 Ten years ago, the high-resolution 3D regional numerical modelling approach of
116 geographically referenced subduction zones may not have been feasible on a day-to-
117 day modelling time-frame. Only a limited number of somewhat extreme subduction
118 models utilized the computational resources necessary to run such high-fidelity mod-
119 els, $> 17,000$ compute hours per model for example (Jadamec and Billen, 2010;
120 Stadler et al., 2010). However, the rapidly changing computational landscape sug-
121 gests that this form of modelling, although less common, will become commonplace
122 in the next ten years (Keyes et al., 2003).

123 *2.1. Advances in High Performance Computing*

124 Figure 3 shows the increase in size of the most powerful super-computer systems
125 worldwide, from 1993-2015 (Strohmaier et al., 2015). The graph shows the total
126 number of cores and Rmax for the top ranked machines over this time period, ac-
127 cording to the TOP500 (Strohmaier et al., 2015). The TOP500 list is ranked by
128 performance on the LINPACK Benchmark (Strohmaier et al., 2015), and thus takes

129 into account the peak floating point operations (Rmax in Tera-Flops) as well as
130 net number of compute cores (Figure 3). In the early- to mid-1990s, the largest
131 machines were comprised of on the order of 1000 cores. Just after the turn of the
132 century (early 2000s), the size of the largest machines increased to on the order of
133 10,000 cores, followed by an increase to $\sim 100,000$ s cores in 2005-2010. The largest
134 publically accessible computers now (in 2015) contain over 1 million cores (Figure
135 3).

136 The exponential increase in the computational hardware through time is consis-
137 tent with that predicted by Moore’s Law (Keyes et al., 2003). However, it should
138 be pointed out that although one can measure the peak performance on a machine,
139 the code performance for a particular set of software is a function of not only the
140 number of floating point operations, but also the memory bandwidth and the sophis-
141 tication and portability of the numerical algorithms (Gropp et al., 1999; Keyes et al.,
142 2003). Thus, not all codes run on the largest machines will achieve peak performance.
143 Nonetheless, the increase in computational power will have dramatic effects on not
144 only the scale of problems, but also in terms of the scientific hypotheses that can be
145 posed.

146 What it means for geodynamics, is a dramatic increase in compute power available
147 for numerical simulations (Jadamec and Billen, 2010; Stadler et al., 2010; Jadamec
148 et al., 2012; Kronbichler et al., 2012; Rudi et al., 2015; Gassmller et al., 2016).
149 A little more than 10 years ago, a machine consisting of 10,000 cores would have
150 been considered state of the art and using it would have been pushing the envelop
151 of what was possible with subduction modelling (Figure 3). Now that number is
152 1,000,000 cores (Figure 3). The prospects are exciting; however, without scalable and
153 efficient numerical algorithms the hardware will remain underutilized (Zhong et al.,
154 2015; Rudi et al., 2015; May et al., 2016). Therefore, the interdisciplinary research

155 between geodynamics and computational science remains a key step in leveraging
 156 the increased computational infrastructure (Keyes et al., 2003; Kronbichler et al.,
 157 2012; Rudi et al., 2015).

158 *2.2. Governing Equations for Viscous Flow modelling*

159 To provide a framework for the governing equations implemented in the compu-
 160 tational fluid dynamics research summarized in this paper, a first order description
 161 of the governing equations for solid state flow in the mantle are described. These
 162 equations have been implemented numerically into a number of open source codes
 163 that simulate viscous flow of the Earth (King et al., 1990; Moresi and Gurnis, 1996;
 164 Tackley, 1996; Zhong et al., 2000; Moresi et al., 2007; Zhong et al., 2007; van Keken
 165 et al., 2008; Kronbichler et al., 2012; Lin, 2014). The governing equations for the
 166 viscous flow are the conservation of mass, momentum, and energy (Batchelor, 1967;
 167 White, 2003). The conservation of mass, also referred to as the continuity equation,
 168 can be expressed in differential form for an infinitesimal control volume:

$$\frac{d\rho}{dt} + \nabla \cdot (\rho \mathbf{V}) = 0, \quad (1)$$

169 which, for an incompressible fluid, simplifies to

$$\nabla \cdot \mathbf{V} = 0, \quad (2)$$

170 where ∇ is the gradient operator, $\mathbf{V}$ is the flow velocity, ρ is density, and t is time
 171 (White, 2003). In simple terms, this equation states that the mass flowing into the
 172 control volume is equal to the mass flowing out of the control volume.

173 The conservation of momentum has the following differential form for an infinites-
 174 imal control volume:

$$\rho \mathbf{g} - \nabla p + \nabla \cdot \tau_{ij} = \rho \frac{d\mathbf{V}}{dt} \quad (3)$$

175 where

$$\frac{d\mathbf{V}}{dt} = \frac{\partial \mathbf{V}}{\partial t} + (\nabla \cdot \mathbf{V})\mathbf{V}, \quad (4)$$

176 p is the pressure, $\mathbf{g}$ is the acceleration due to gravity, τ_{ij} is the viscous stress tensor,
177 and the other components are as described above. Eq. 3 sets the gravity (body)
178 force per unit volume, pressure force per unit volume, and viscous force per unit
179 volume equal to the density times acceleration (Batchelor, 1967; White, 2003).

180 In modelling the creeping flow of the mantle, we commonly assume an incom-
181 pressible viscous fluid with a high Prandtl number and Eq. 3 becomes:

$$\rho \mathbf{g} + \nabla \cdot \sigma_{ij} = 0 \quad (5)$$

182 where the stress, σ_{ij} , is equal to

$$\sigma_{ij} = -P\delta_{ij} + \tau_{ij} \quad (6)$$

183 and P is the dynamic pressure and the term τ_{ij} contains the viscosity, η , and the
184 strain-rate tensor, $\dot{\epsilon}_{ij}$, and the subscripts, ij , refer to the components of the stress
185 and strain rate tensors (Moresi and Solomatov, 1995; White, 2003). Specific viscos-
186 ity formulations, such as a strain rate dependent viscosity based on experimentally
187 derived flow laws (Hirth and Kohlstedt, 2003; Billen and Hirth, 2007) or directionally
188 dependent mechanically weakening (Lev and Hager, 2011; Sharples et al., 2016), can
189 be used and connected to the force balance through this equation.

190 In mantle convection it is common to express the initial body force ($\rho \mathbf{g}$) in Eq.
191 5 in terms of the temperature variation $[\rho_o \alpha (T - T_o)]$, wherein α , T , T_o are the
192 coefficient of thermal expansion, temperature, reference temperature, respectively
193 (Zhong et al., 2007, 2015). In addition to the body force due to density variations as
194 a function of temperature, density variations due to different rock types and phase
195 transitions can be included. These equations can be simultaneously solved to obtain

196 the flow velocity, pressure, and viscosity, if the viscosity is also an unknown. Solving
 197 the conservation of mass and momentum for viscous flow is typically what is done
 198 when the solving the instantaneous problem.

199 For the time-dependent problem, the conservation of energy is typically also
 200 solved in order to allow for the thermal field to move as a function of the viscous
 201 flow (White, 2003; Zhong et al., 2007, 2015). For incompressible flow with constant
 202 properties and ignoring internal heating and viscous dissipation, the conservation of
 203 energy can be simplified to the following differential form

$$\frac{\partial T}{\partial t} = -\mathbf{V} \cdot \nabla T + \kappa \nabla^2 T, \quad (7)$$

204 where κ is the thermal diffusivity and t is time, and the other variables are as
 205 defined above (White, 2003). Whereas, the temperature, T , is the input in the
 206 conservation of momentum equation (Eq. 3), with the conservation of energy (Eq. 7),
 207 temperature is the unknown which is solved for. The approximations, assumptions,
 208 and simplifications of these equations in a given subduction model will depend in
 209 part on the questions asked. It is important to point out that these governing
 210 equations are used to examine the solid state flow of the mantle and slab. Additional
 211 equations are required to examine the multi-physics flow of magma migration that
 212 occurs within the solid state mantle framework of the subduction zone (McKenzie,
 213 1984; Spiegelman, 1993; Katz and Takei, 2013; Wilson et al., 2014).

214 To simulate subduction and large-scale viscous flow in the mantle, the conser-
 215 vation equations are numerically approximated on a discretized model mesh (Zhong
 216 et al., 2007, 2015). The mantle convection codes vary in the numerical approach
 217 and use finite difference, finite element, and finite volume schemes most commonly
 218 (Zhong et al., 2007, 2015). As complexity, in terms of geometry, rheology, and min-
 219 eralogy, are added into the numerical models, the role of computational and applied

220 mathematics becomes critical as the choice of solvers and internal description and
221 implementation of the problem can greatly affect convergence, model runtime, and
222 scalability (Moresi and Solomatov, 1995; May and Moresi, 2008; Brown et al., 2012;
223 Jadamec et al., 2012; Brune et al., 2015; May et al., 2016).

224 **3. Incorporating Geometric Complexity into Plate Boundary Models and** 225 **Leveraging Computational Resources**

226 Although many open source software packages are available that simulate the
227 viscous flow of the Earth (King et al., 1990; Moresi and Gurnis, 1996; Tackley, 1996;
228 Zhong et al., 2000; Moresi et al., 2007; Zhong et al., 2007; van Keken et al., 2008;
229 Kronbichler et al., 2012; Lin, 2014), there are few open source programs that generate
230 the three-dimensional model input for the viscous flow codes, such that the initial
231 model configuration takes into account the geometric complexity of the earth, on the
232 scale of 2-3 km over a large model domain (Jadamec et al., 2012). This is particularly
233 true for the 3D regional subduction models that use realistic plate geometries, which
234 often contain a complex slab geometry at an arbitrary orientation with respect to
235 the model mesh grid (Billen et al., 2003; Jadamec et al., 2012; Hu et al., 2016). This
236 aim of replicating a particular plate boundary on Earth at high-resolution (< 10
237 km) in terms of the actual modern plate geometries and 3D lithospheric structure
238 separates the high-resolution geographically referenced 3D regional models from the
239 generalized 3D numerical models of subduction.

240 For simplified three-dimensional cases, the 3D model set-up can be obtained
241 with a relatively small amount of code and many simplifying assumptions (Kneller
242 and van Keken, 2008). However, the modeler is now faced with representing both
243 (a) slab complexity, including slab edges, tears, and along strike variations in slab
244 dip and depth (Gutscher et al., 2000; Ratchkovski and Hansen, 2002; Lallemand

et al., 2005; Zhao et al., 2012; Dougherty and Clayton, 2014; Scire et al., 2016), and (b) upper plate complexity, such as lateral variations in thickness, temperature, and composition (Silberling et al., 1994; Van Staal et al., 1998; Greninger et al., 1999; Bird, 2003; Artemieva, 2006; Müller et al., 2008; Hartmann and Moosdorf, 2012; Laske et al., 2013), at a resolution of < 10 km and with multiple features arbitrarily oriented with respect to the model grid (Jadamec et al., 2012). Thus, more sophisticated approaches are required (Jadamec and Billen, 2010; Liu and Stegman, 2011; Gurnis et al., 2012; Jadamec et al., 2012; Williams et al., 2012; Jadamec et al., 2013; Bower et al., 2013; Heinecke et al., 2014; Bower et al., 2015; Seton et al., 2015; Leonard and Liu, 2016).

For the high-resolution model design, the complex 3D plate boundary configurations must be configured numerically with a computational infrastructure, either a single code or series of codes connected through linked workflows (Kreylos et al., 2006; Jadamec and Billen, 2010; Liu and Stegman, 2011; Gurnis et al., 2012; Jadamec and Billen, 2012; Williams et al., 2012; Heinecke et al., 2014; Bower et al., 2015; Fischer et al., 2015). Thus, for the high-resolution 3D regional models, the model set-up can be challenging and time-consuming. This adds a non-trivial step of model design to the numerical modelling workflow (Figure 4). For this kind of modelling approach, the data collection and pre-processing phase (Figure 5a), model design phase (Figure 5b), and model simulation phase (Figure 5c) can comprise equal parts of the overall modelling effort.

3.1. *Generating Realistic 3D Tectonic Models*

An example of a recent open source C/C++ program that has been successful in building 3D tectonic configurations of geographically referenced plate boundaries is SubductionGenerator (Jadamec and Billen, 2010; Jadamec et al., 2012). A 3D model

270 mesh and tectonic configuration are constructed based on a user specified model
 271 domain, slab surface, seafloor age grid file, and shear zone surface (Figures 5a,b).
 272 The initial 3D thermal structure for the plates and mantle within the model domain
 273 is then constructed using a series of libraries within the code that use a half-space
 274 cooling model, plate cooling model, and smoothing functions (Jadamec and Billen,
 275 2010; Jadamec et al., 2012). The code maps the initial 3D thermal structure and the
 276 3D plate interface onto the mesh nodes using a series of libraries including a k-d tree
 277 to increase efficiency (Jadamec and Billen, 2010; Jadamec et al., 2012). In this way,
 278 complicated geometries and multiple plates with variable thickness can be built onto a
 279 multi-resolution finite element mesh with a 3D thermal structure and 3D isotropic
 280 shear zones oriented at any angle with respect to the grid (Figure 5b). The program
 281 has been used to build 3D configurations for the eastern Alaska subduction-transform
 282 system (Jadamec and Billen, 2010, 2012; Jadamec et al., 2013), the Cocos-Nazca
 283 slab window (Jadamec and Fischer, 2014), and the New Hebrides subduction-back
 284 arc spreading system (Jadamec et al., 2011; McLean et al., 2011).

285 The three-dimensional thermal structure, plate boundary shear zone, and model
 286 mesh can be used as input to finite-element codes that solve for the viscous flow in
 287 subduction zones, such as CitcomCU (Moresi and Solomatov, 1995; Zhong, 2006).
 288 Specifically, the initial thermal distribution (Figure 5b) can be used as input to
 289 the conservation of the momentum equation (Eq. 5) to provide the initial buoyancy
 290 forces that drive the flow (Figure 5c). The 3D plate boundary shear zone can be input
 291 in the viscous flow calculations through the viscosity formulation and constitutive
 292 relation (Eq. 6).

293 The idea of using geophysical and geologic observations to constrain the initial
 294 configuration in geodynamic modelling is not new (Ricard and Vigny, 1989; Lithgow-
 295 Bertelloni and Richards, 1998; Becker and O’Connell, 2001; Conrad and Lithgow-

296 Bertelloni, 2002; Miller and Becker, 2012). What is new is the level of detail with
 297 which this now being done, due the increased computing resources, thus further
 298 differentiating regional 3D models from global models and requiring the new approach
 299 to model design (Jadamec et al., 2012). The synthesis of detailed global data (Gurnis
 300 et al., 2012; Williams et al., 2012) in combination with the use of adaptive mesh
 301 refinement (Burstedde et al., 2009) will be invaluable in moving the high-resolution
 302 modelling to the global scale and through time.

303 **4. From 2D to 3D numerical models of subduction**

304 Before describing the impacts of three-dimensional models of subduction driven
 305 solid state mantle flow, it is important to point out that two-dimensional numerical
 306 models have, and continue to provide, a tremendous amount of insight, and in and of
 307 themselves can still be quite challenging, depending on the scientific problem posed.
 308 The progression in 2D modelling of subduction, from 2D corner flow models to models
 309 with a free slab and mobile upper plate, is illustrated in Figure 6 with a simplified
 310 rendering of three phases in the sophistication of the model set-up. Not all model
 311 development is linear, and therefore the figure represents schematically the advances
 312 in model componentry, while recognizing that parallel tracts of model development
 313 occur. In addition, the progression in modelling magma dynamics with two-phase
 314 flow is not illustrated.

315 The early approach to subduction modelling focused on solving for the corner
 316 flow in the mantle wedge, that is mantle flow field in the region between the slab and
 317 the upper plate (Figure 6a) (McKenzie, 1969; Sleep and Toksöz, 1971; Tovish et al.,
 318 1978). This line of 2D modelling continues, solving for the flow in the mantle wedge
 319 with very high-resolution (< 0.5 km) (Van Keken et al., 2002; van Keken et al.,
 320 2008; Wada et al., 2008; Syracuse et al., 2010). The approach of solving for the

321 steady-state corner flow in the mantle wedge has also been applied to 3D modelling
322 and has provided valuable insights into high-resolution ($\sim 2\text{km}$), three-dimensional
323 steady-state mantle wedge flow (Kneller and van Keken, 2007; Wada et al., 2015).

324 Because many slabs bend with increasing depth and therefore do not have a
325 single dip or slab velocity during the subduction evolution (Jarrard, 1986; Lallemand
326 et al., 2005), 2D numerical models of subduction have increased in sophistication to
327 investigate the deformation of the slab, the mantle flow beneath the slab, and the
328 time-dependent flow within the wedge (Figure 6b) (Houseman and Gubbins, 1997;
329 Gerya et al., 2004; Hyndman et al., 2005; Billen and Hirth, 2007; Cížková et al.,
330 2007; Burkett and Billen, 2009; Ribe, 2010; Jadamec, 2016). 3D numerical models
331 have also expanded the model domain and set-up to investigate the 3D deformation
332 of the slab and/or 3D deformation of the surrounding mantle, both below the slab
333 and in the mantle wedge (Zhong and Gurnis, 1996; Piromallo et al., 2006; Stegman
334 et al., 2006; Giuseppe et al., 2008; Jadamec and Billen, 2010; Liu and Stegman, 2012;
335 Faccenda and Capitanio, 2013; Paczkowski et al., 2014).

336 With increasing model complexity, the numerical models can solve for the lateral
337 motion of a mobile upper plate in addition to the motion of the slab and mantle,
338 in 2D (Figure 6c) (Sharples et al., 2014) and 3D models (Jadamec and Billen, 2010;
339 Capitanio et al., 2011; Jadamec et al., 2013; Schellart and Moresi, 2013). In this
340 way, numerical models can investigate how the upper plate evolution dynamically
341 feeds into the slab deformation and subsequent mantle flow field (Figure 6c)
342 (Sharples et al., 2014; Rodríguez-González et al., 2014; Holt et al., 2015).

343 Accurately incorporating the upper plate, and the corresponding plate interface,
344 in numerical models of subduction remains a challenge (Jadamec et al., 2012). It
345 is still non-trivial to define a rheological formulation that allows for weakening of
346 the plate boundaries and strengthening of the plate interiors, in such a way that

347 these features emerge from the dynamics and evolve in space and time with the
348 subduction zone forces (Regenauer-Lieb et al., 2001; Kaus and Podladchikov, 2006;
349 Cramer et al., 2012; Bercovici and Ricard, 2014; Sharples et al., 2016).

350 Among the 3D numerical models of subduction, there are a subset that incor-
351 porate a complex and detailed 3D slab geometry based on seismic observations into
352 the mantle domain, e.g., Moresi and Gurnis (1996). The approach therein is to
353 use the high-resolution numerical models of a particular region as a way to con-
354 strain the subduction processes, rather than running a suite of models with general
355 slab shapes and mantle properties (Figures 4 and 5). This has been done for the
356 Mariana, Izu-Bonin, Japan, and South Kurile subduction zone (Moresi and Gurnis,
357 1996), the Tonga-Kermadec subduction zone (Billen et al., 2003; Billen and Gurnis,
358 2003), the Aleutian oceanic-oceanic subduction zone (Billen and Gurnis, 2003), and
359 the oceanic-continent subduction zone in eastern Alaska (Jadamec and Billen, 2010,
360 2012; Jadamec et al., 2013). More recent high-resolution 3D numerical models of
361 specific subduction zones include the evolution of slab rupture and upper mantle
362 flow in western North America (Liu and Stegman, 2012), and the three-dimensional
363 mantle flow through the Patagonia slab window in South America (Lin, 2014), the
364 reversal of mantle flow through the Cocos-Nazca slab window beneath the western
365 Caribbean plate (Jadamec and Fischer, 2014), and the intersection of slabs beneath
366 the Taiwan subduction system (Lin and Kuo, 2016).

367 A high-resolution geographically referenced regional 3D model, solving for the
368 subslab flow as well as upper mantle flow and slab deformation, can commonly con-
369 tain over 100 million mesh nodes, over 500 million unknowns, have a local resolution
370 of 2.5 km, and cost 15,000 - 20,000 compute hours per time-step (Jadamec and
371 Billen, 2010, 2012; Jadamec et al., 2013). A fine resolution over a large portion of
372 the plate boundary is required to resolve the geometric complexity as well as the

viscosity variations (~ 7 orders of magnitude) that dynamically emerge due to a strain-rate dependent rheology. In other words, the mesh needs to be discretized sufficiently to resolve variations in viscosity occurring between the relatively strong lithospheric plates ($\sim 10^{22} - 10^{25}$ Pa s) and the weaker plate interface ($\sim 10^{18} - 10^{22}$ Pa s) and weak mantle wedge ($\sim 10^{17} - 10^{18}$ Pa s), wherein the net change in viscosity occurs over spatial distances on the order of 100 km or less and this change needs to transition smoothly for numerical stability (Moresi and Solomatov, 1995; Tackley, 1996; Jadamec et al., 2012). With adaptive mesh refinement, this high-resolution 3D slab and upper mantle non-linear modelling approach is being extended to the global scale (Stadler et al., 2010; Alisic et al., 2012; Kronbichler et al., 2012; Rudi et al., 2015).

5. Recent Results of slab-driven Mantle Flow from 3D Subduction Models

As posited earlier, a question to ask with high resolution 3D regional modelling of subduction systems is, do we really learn anything new; are new processes discovered, or just the same process resolved at higher resolution? In this section, I give examples of general tectonic processes at plate boundaries discovered or at least better characterized as a result of the high resolution 3D regional modelling approach. The section begins with a characterization of the mantle flow field determined from generalized 3D numerical models of subduction, and then highlights three processes related to slab-driven mantle flow for which the high resolution 3D models have particularly contributed to our understanding. The section will draw more heavily on my models, as those are the models with which I have the most experience and are a good example of the kind of capability and flexibility of the approach discussed in this paper and of the new kinds of results discovered. In some cases, I will highlight

397 results from 3D laboratory models, as these are increasing in level of sophistication
398 and have been a valuable constraint on the numerical models.

399 *5.1. General Patterns of Slab-driven 3D Mantle Flow*

400 Three-dimensional numerical models of subduction indicate a departure from the
401 paradigm of corner flow in subduction zones. Although the slab descent into the
402 mantle does drive corner flow in the mantle wedge drawing mantle toward the slab
403 and then entraining mantle downward as the slab sinks further downward (Figure 7a),
404 3D geodynamic models indicate that this is just one component of the mantle flow
405 field in subduction zones (Figure 7b-f) (Piromallo et al., 2006; Royden and Husson,
406 2006; Stegman et al., 2006; Conder and Wiens, 2007; Kneller and van Keken, 2007;
407 Schellart et al., 2007; Faccenna et al., 2010; Jadamec and Billen, 2010; Capitanio
408 and Faccenda, 2012).

409 Specifically, in addition to the slab-driven corner flow which is confined to within a
410 cross section through the subduction zone (Figure 7a), lateral changes in the dip and
411 buoyancy of the slab as well as in the mantle wedge viscosity can drive arc-parallel
412 flow in the mantle wedge (Conder and Wiens, 2007; Kneller and van Keken, 2007;
413 Jadamec and Billen, 2010; Capitanio and Faccenda, 2012; Jadamec and Billen, 2012)
414 and in the subslab mantle (Jadamec and Billen, 2010; Capitanio and Faccenda, 2012;
415 Jadamec and Billen, 2012) with mantle material moving perpendicular to the cross
416 section plane (Figure 7b). 3D instantaneous models with a slab edge show that slab
417 steepening can drive toroidal flow, wherein subslab mantle is transported around the
418 side of a lateral slab edge and into the mantle wedge (Figure 7c) (Piromallo et al.,
419 2006; Jadamec and Billen, 2010, 2012). Time-dependent 3D numerical models with
420 a slab edge show that this process evolves over time with the trench retreat and slab
421 rollback driving toroidal flow around the lateral slab edges that varies in its intensity

422 over the lifetime of the subduction zone (Stegman et al., 2006; Schellart et al., 2007;
 423 Faccenna et al., 2010). The 3D models with a slab edge that is free to change dip show
 424 that as the mantle material sweeps around the lateral slab edge there is commonly an
 425 upwelling component to the flow that transports subslab mantle upwards (Figure 7c)
 426 (Stegman et al., 2006; Piromallo et al., 2006; Jadamec and Billen, 2010; Faccenna
 427 et al., 2010; Jadamec and Billen, 2012). Tearing of a viscous slab can facilitate
 428 mantle upwelling through and along the slab tears (Burkett and Billen, 2010; Liu
 429 and Stegman, 2012) (Figure 7d). Three-dimensional mantle wedge flow can also
 430 occur due to structure in the mantle wedge (Honda and Saito, 2003; Conder and
 431 Wiens, 2007), structure in the upper plate (Behn et al., 2007; Miller and Becker,
 432 2012; Rodríguez-González et al., 2014), or motions in the upper plate (Hall et al.,
 433 2000). In addition, the slab-driven toroidal flow can in turn drive overriding plate
 434 deformation through mantle tractions on the base of the overriding plate (Capitanio
 435 et al., 2011; Schellart and Moresi, 2013; Menant et al., 2016) (Figure 7f).

436 Thus, recognizing the three-dimensional nature of the mantle flow field in sub-
 437 duction zones was a fundamental milestone that significantly affected the perception
 438 and conceptualization of subduction processes and numerical modelling (Hall et al.,
 439 2000; Conder and Wiens, 2007; Kneller and van Keken, 2007; Schellart et al., 2007;
 440 Jadamec and Billen, 2010; Capitanio et al., 2011; Liu and Stegman, 2012). It is
 441 important to point out that laboratory models played a critical role in characterizing
 442 this new three-dimensional paradigm for slab-driven mantle flow (Buttles and Olson,
 443 1998; Kincaid and Griffiths, 2003; Funiciello et al., 2004; Schellart, 2004; Bellahsen
 444 et al., 2005; MacDougall et al., 2014; Strak and Schellart, 2014). The laboratory
 445 models provided an invaluable framework for the 3D numerical modelling studies
 446 and made the first attempts to explain the shear wave splitting observations in a
 447 3D model (Buttles and Olson, 1998). In addition, the 3D laboratory models were

critical in first identifying the time-dependent 3D variations in trench geometry, slab dip, and slab-driven mantle flow, and in categorizing the time-dependent phases of subduction from the initial transient phases to the established modes of trench retreat and slab rollback after the slab is laid flat on the lower mantle, for example (Funiciello et al., 2004; Schellart, 2004).

5.2. Slab-driven Mantle Upwelling at Slab Edges

An interesting component of the mantle flow field in subduction zones revealed by the 3D subduction modelling is the slab-driven mantle upwelling at slab edges (Figure 7). The fluid flow pattern of toroidal flow around a pivoting structure is expected from general fluid dynamics principles (Batchelor, 1967). In the subduction zone, at a lateral slab edge, as the slab steepens or rolls back, the subslab mantle material is pushed by the moving slab and is translated around the lateral slab edge into the mantle wedge, e.g., Stegman et al. (2006). This forms the arcuate toroidal flow pattern seen in depth slices of mantle flow velocity in 3D subduction models with slab edges (Figures 7 and 8b) (Jadamec and Billen, 2010).

Associated with this toroidal flow, is an upwelling component in the mantle, such that, as the arcuate toroidal flow field moves material laterally around the slab edge, warm mantle is also transported from underneath the slab into the lower pressure mantle wedge in an upward and arcuate pattern around the edge (Figures 7 and 8b) (Piromallo et al., 2006; Jadamec and Billen, 2010). There is a combination of the slab steepening which pushes material behind the slab causing it to move laterally around the slab edge, e.g., Stegman et al. (2006), and the low dynamic pressure that is created above the slab, e.g., Manea and Gurnis (2007); Hu et al. (2016), that draws asthenosphere upwards into the mantle wedge. The zone of mantle upwelling typically occurs lateral to the slab edge, located within 500 km outboard of the slab

473 edge (Piromallo et al., 2006; Jadamec and Billen, 2012). The upward component
474 of mantle flow associated with toroidal-type return flow was first recognized in 4D
475 (time-dependent 3D) laboratory models of subduction (Schellart, 2004), and has
476 since been demonstrated in both laboratory and numerical models (Piromallo et al.,
477 2006; Faccenna et al., 2010; Guillaume et al., 2010; Jadamec and Billen, 2010, 2012;
478 Liu and Stegman, 2012; Strak and Schellart, 2014, 2016). The slab edge driven
479 mantle upwelling may lead to melting and volcanism that is distinct from typical arc
480 volcanism associated with corner flow in the mantle wedge.

481 Regional 3D numerical models can help to constrain this process through the
482 ability to spatially link the predicted mantle upwelling with the location of observed
483 slab derived melts, such as adakites. Adakitic volcanics reflect a range in petrogenetic
484 origins (Kay, 1978; Drummond et al., 1995; Yogodzinski et al., 1995; Kay and Kay,
485 2002; Kelemen et al., 2003; Castillo, 2012), including the association with slab melting
486 in subduction zones of either young oceanic material or otherwise warmed oceanic
487 material (Defant and Drummond, 1990; Johnston and Thorkelson, 1997; Abratis and
488 Worner, 2001; Durance et al., 2012). In the Alaska subduction zone, for example,
489 the Wrangell volcanics are located east of the main Aleutian-Alaska magmatic arc
490 and are situated east of a major break in slab seismicity indicative of the slab edge
491 (Ratchkovski and Hansen, 2002; Jadamec and Billen, 2010). The Wrangell volcanics
492 display a range in morphologies from strato-volcanoes to shield-like edifices and
493 are characterized by a range in geochemical signatures including, adakitic affinities
494 suggestive of melting of the edge of the subducted Pacific plate (Skulski et al., 1991;
495 Preece and Hart, 2004; Trop et al., 2012).

496 Recent geographically referenced, 3D regional models of the Alaska subduction
497 zone, with a local resolution of 2.5 km, tested the effect of high-resolution slab
498 geometry on the mantle flow field beneath mainland Alaska, including in the vicinity

499 of the Wrangell volcanics (Jadamec and Billen, 2010; Jadamec et al., 2011; Jadamec
 500 and Billen, 2012). These 3D instantaneous numerical models predicted upwelling
 501 associated with the toroidal flow around the eastern slab which entrained warm
 502 mantle material in a three-dimensional upward flow pattern, with the location of the
 503 Wrangell volcanics above the upward stroke of the 3D flow field (Figure 8b) (Jadamec
 504 and Billen, 2010; Jadamec et al., 2011; Jadamec and Billen, 2012). In addition,
 505 geographically referenced, 3D high-resolution (local resolution of $\sim 3\text{km}$) models of
 506 the Cocos-Nazca slab window also show slab edge related upwelling associated with
 507 the arc of the toroidal flow around both the southern Cocos and the northern Nazca
 508 slab edges (Figure 9) (Jadamec and Fischer, 2014). This upward motion entrains
 509 warmed asthenospheric mantle and warmed slab edge material upwards and thus
 510 could contribute to the observed anomalous volcanism in southern Central America
 511 (Defant and Drummond, 1990; Johnston and Thorkelson, 1997; Abratis and Worner,
 512 2001).

513 This detailed approach of modelling the slab-driven mantle flow field and linking
 514 the solid state flow framework with the resultant movement of geochemical signatures
 515 through out the subduction system remains an important future direction in sub-
 516 duction modelling (Jadamec and Billen, 2010; Schellart, 2010; Durance et al., 2012;
 517 Jadamec and Billen, 2012; Lin, 2014), particularly in the time-dependent evolution
 518 of the subduction system (Liu and Stegman, 2012; Bower et al., 2015; Hu et al.,
 519 2016).

520 *5.3. Slab-driven Mantle Deformation Zones and Laterally Variable Plate-Mantle Cou-* 521 *pling*

522 Another area of research where 3D high-resolution regional models have made a
 523 contribution is in constraining the slab and upper mantle viscosity, and in placing

524 constraints on lateral variability in plate-mantle coupling (Figure 10). Although
 525 the average upper mantle viscosity is the order of 10^{20} Pa·s, as determined from
 526 post-glacial rebound and global velocity models (Hager, 1991; Mitrovica and Forte,
 527 2004), experimental studies suggest the viscosity in the mantle wedge may be several
 528 orders of magnitude lower, $\sim 10^{18}$ Pa·s (Hirth and Kohlstedt, 2003; Karato, 2003).
 529 The local reduction in mantle wedge viscosity is likely due to a number of factors
 530 including increased temperature, water content, melt-fraction, and strain-rate (Hirth
 531 and Kohlstedt, 2003; Karato, 2003; Kelley et al., 2006; Wade et al., 2008). Both
 532 2D and 3D geodynamic models have constrained and continue to constrain these
 533 properties (Katz et al., 2003; Billen and Gurnis, 2001; Kincaid and Griffiths, 2003;
 534 Billen and Hirth, 2005; Hyndman et al., 2005; Currie and Hyndman, 2006; Cagnioncle
 535 et al., 2007; Conder and Wiens, 2007; Wade et al., 2008; Jadamec and Billen, 2010;
 536 Syracuse et al., 2010; Lev and Hager, 2011; Wirth and Korenaga, 2012; Wada et al.,
 537 2015), and in turn provide a range of testable viscosities in the upper mantle region
 538 of the subduction zone.

539 For example, 3D regional numerical models of the Tonga-Kermadec subduction
 540 zone and the Aleutian subduction zone using a Newtonian viscosity found a better
 541 fit to the geoid and dynamic topography by imposing a region of low viscosity in the
 542 mantle wedge within the Newtonian viscosity models (Billen et al., 2003; Billen and
 543 Gurnis, 2003). More recently, geographically referenced, high-resolution 3D geody-
 544 namic models of the Alaska subduction zone (local resolution of ~ 3 km) compared the
 545 effect of an experimentally derived flow law for diffusion creep (Newtonian viscosity)
 546 versus the effect of both diffusion and dislocation creep (composite non-Newtonian
 547 viscosity) and found a better fit to shear wave splitting observations, plate motions,
 548 and dynamic topography in models using the composite viscosity (Figures 8 and 10)
 549 (Jadamec and Billen, 2010; Jadamec et al., 2012; Jadamec and Billen, 2012; Jadamec

et al., 2013). Importantly, the models found that the non-Newtonian viscosity led to lateral viscosity variations that naturally emerged in the upper mantle beneath Alaska, with a dynamically reduced mantle wedge viscosity ($\sim 10^{18}$ to as low as $\sim 10^{17}$ Pa·s) and reduced subslab viscosity ($\sim 10^{18}$ Pa·s) that evolved in the model due to the high strain rates surrounding the slab (Figures 10a,b) (Jadamec and Billen, 2010; Jadamec et al., 2012; Jadamec and Billen, 2012).

This dynamically induced zone of low viscosity in the upper mantle surrounding the slab has also been predicted in 2D (Billen and Hirth, 2005; Andrews and Billen, 2009; Jadamec, 2016), generalized 3D (Billen and Jadamec, 2012; Hines and Billen, 2012), and global subduction models (Stadler et al., 2010; Alisic et al., 2012) that incorporate a composite viscosity and examine how the dislocation creep deformation mechanism allows for dynamic weakening of the upper mantle in regions of high strain-rate both above and below the slab. The concept of slab-driven mantle deformation zones and laterally variable plate-mantle coupling in subduction zones has been suggested conceptually by earlier seismological studies that examined misalignment of the seismic fast axes observed from shear wave splitting with the local plate motions, e.g., Russo and Silver (1994). Subsequent studies of shear wave splitting in the mantle wedge constrained by local S waves as well as teleseismic studies suggest the non-alignment of the mantle flow field with the subducting plate velocity may occur in many subduction zones, e.g., Long and Wirth (2013) and references therein, (Figure 2).

Thus, both the 3D geodynamic models and seismic observations imply local decoupling of the mantle wedge from the surface plates near subduction zones (Figures 2 and 8a-c). It is important to note that dislocation creep is the same mechanism responsible for (a) the reduced upper mantle viscosity surrounding the slab that emerges in the geodynamic models described here and (b) the lattice preferred ori-

576 entation (LPO) in the olivine, that gives rise to the seismic fabric in the upper mantle
 577 where olivine is the dominant constituent, assuming the LPO aligns with the man-
 578 tle flow in regions of high strain rate (Savage, 1999; Karato et al., 2008; Long and
 579 Becker, 2010; Jadamec and Billen, 2010). Thus, the zones of (a) slab-driven reduced
 580 mantle viscosity and (b) slab-driven mantle flow and associated complex shear wave
 581 splitting should be correlated (Jadamec, 2016), if the splitting is due to the LPO
 582 developed in olivine in regions of high strain rate aligned with the flow (Figure 2 and
 583 Figures 8-10). Away from the subduction zone, the effects of the driving force of
 584 the slab decrease, and hence there is a lateral increase in the upper mantle viscosity.
 585 The net result is a region of plate-mantle decoupling close to the subduction zone
 586 that transitions outwards to more coupling between the plates and mantle and hence
 587 more alignment of the flow with surface plate motions farther from the slab (Figure
 588 10) (Jadamec and Billen, 2010; Jadamec, 2016).

589 High-resolution geographically referenced 3D models can be useful in constraining
 590 the length scale of this decoupling zone through the comparison of the predicted
 591 upper mantle deformation with observations of shear wave splitting from specific
 592 locations on Earth, e.g., Jadamec and Billen (2010) (Figure 8). The width of the
 593 this decoupling zone seems to be on the order of a few hundred kilometers in the
 594 lateral and radial directions away from the slab, but is still being studied in both 3D
 595 (Jadamec and Billen, 2010; Faccenda and Capitanio, 2012) and 2D models (Jadamec,
 596 2016), and is an active area of research.

597 The quantitative connection between shear wave splitting observations and the
 598 mantle flow field predicted in global, generalized 3D, and regional 3D models has
 599 the potential to place a critical constraint on processes that govern the upper mantle
 600 flow field near subduction zones. Although there has been work toward this end,
 601 predicting LPO, infinite strain axes, finite strain ellipses and synthetic splitting from

geodynamic models (Hall et al., 2000; Kaminski et al., 2004; Conrad et al., 2007; Kneller and van Keken, 2007; Jadamec and Billen, 2010; Miller and Becker, 2012; Faccenda and Capitanio, 2013), a comprehensive geodynamic-seismic framework may still be in its infancy. A multi-faceted approach will be critical, wherein generalized models make predictions for deformation fabric and splitting (Faccenda and Capitanio, 2012; Di Leo et al., 2014; Li et al., 2014), and then geodynamic models of specific areas are constructed to try to match the splitting patterns and length scales in specific subduction zones (Hall et al., 2000; Fischer et al., 2000; Conder and Wiens, 2007; Kneller and van Keken, 2007; Jadamec and Billen, 2010; Miller and Becker, 2012).

5.4. *Slab-driven Mantle Flow and Mantle Flow Rates*

High-resolution 3D models have also made new strides in placing constraints on mantle flow rates in subduction zones (Jadamec and Billen, 2010; Stadler et al., 2010). The slab strength and upper mantle viscosity place a first order control on the rate at which the slab can drive mantle flow during subduction (Alisic et al., 2012; Billen and Jadamec, 2012; Jadamec and Billen, 2012). The models produce a domain of localized fast asthenospheric flow in the mantle surrounding the slab that is likely to occur at many subduction zones (Figure 10).

For example, the high-resolution 3D regional models of the Alaska subduction zone that tested the relative effect of a Newtonian versus composite viscosity, slab strength, and slab edge geometry and found that models with a composite viscosity were a better fit to shear wave splitting observations, Pacific plate motion, and patterns of subsidence and uplift in the upper plate (Jadamec and Billen, 2010, 2012; Jadamec et al., 2013), also predicted mantle flow rates an order of magnitude greater than surface plate motions for the part of the asthenospheric mantle close

627 to the subducting slab (Figures 8 and 10). The models showed that although the
628 surface motion of the plates may be on the order of 1-10 cm/yr, the asthenospheric
629 mantle beneath the plates can flow an order of magnitude faster (up to ~ 80 cm/yr) in
630 regions of reduced viscosity surrounding the slab (Jadamec and Billen, 2010; Jadamec
631 et al., 2012; Jadamec and Billen, 2012).

632 The increased asthenospheric velocities close to the subduction zone are due in
633 part to the slab induced mantle weakening and reduced viscous support of the slab
634 (Figures 8 and 10) (Jadamec and Billen, 2010; Jadamec et al., 2012; Jadamec and
635 Billen, 2012). Reducing the viscous support of the slab, requires the slab to support
636 more of its own weight which can lead to plastic failure in the slab hinge, allowing
637 for slab steepening and slab induced rapid mantle flow (Figures 8, 10, and 11).
638 The dynamic viscosity reduction and related rapid mantle flow are also observed
639 in global models that use a strain-rate dependent viscosity, with adaptive mesh
640 refinement allowing for high-resolution at the subduction zones (Stadler et al., 2010;
641 Alisic et al., 2012), as well as in generalized 3D models (Billen and Jadamec, 2012;
642 Hines and Billen, 2012) and in 2D models that vary the slab strength and upper
643 mantle rheology (Figure 11) (Jadamec, 2016).

644 Rapid mantle flow rates have also been suggested independently by geochemical
645 arguments based on the timing of the emergence of geochemical signatures in the
646 volcanic front in Costa-Rica and Nicaragua (Hoernle et al., 2008) and in the New
647 Hebrides subduction zones (Heyworth et al., 2011). The toroidal flow around slab
648 edges has been shown to lead to increased mantle velocities in the mantle wedge
649 by laboratory models that use temperature dependent glucose syrup (Kincaid and
650 Griffiths, 2003), which numerical models indicate are even faster when using a strain-
651 rate dependent viscosity formulation (Jadamec and Billen, 2010; Stadler et al., 2010).
652 This may contribute to the preservation of primitive magmas brought to the surface

653 at slab edges and slab edge-back arc spreading intersections, such as in the eastern
654 New Hebrides slab edge region (Figure 12) (McLean, 2010; Jadamec et al., 2011;
655 Durance et al., 2012).

656 This is a significant departure from the paradigm of the subduction framework
657 predicted by 2D and 3D models that fix the surface velocity and dip of the slab and
658 solve only for flow in the mantle wedge (Kneller and van Keken, 2007; van Keken
659 et al., 2008; Syracuse et al., 2010; Wada et al., 2015). A model that does not allow
660 the slab to pivot or change velocity (Kneller and van Keken, 2007; van Keken et al.,
661 2008; Syracuse et al., 2010; Wada et al., 2015) would not predict rapid flow because
662 the boundary condition on the slab surface and the inability of the slab to pivot
663 restricts the mantle flow rates in the mantle wedge (King, 2001; Jadamec, 2016). The
664 steady state models are certainly valuable, and provide important insights. A series
665 of 2D subduction models with a fixed slab dip, but that systemically vary the slab
666 inclination from model to model (Syracuse et al., 2010), could provide insights to how
667 the mantle flow field at a given location might evolve, as the different dips meant to
668 represent different locations could theoretically represent different snapshots of one
669 subduction zone through time. However, the models would not capture the non-
670 steady state part of the flow where the transition in slab dip occurs and induces the
671 more rapid flow rates.

672 This interesting phenomenon of rapid mantle flow in subduction zones is simply
673 a function of the system balancing the resisting and driving forces and is not due
674 to model side walls being too close or some complication of the 3D model setup.
675 There are no lateral sidewalls in a global models that show this result (Stadler et al.,
676 2010; Alisic et al., 2012). Furthermore, simplified 2D models, with sidewalls that
677 are over 5-8 times farther away than the length-scale of the slab, demonstrate this
678 force balance as well (Figure 11) (Jadamec, 2016). Thus, the description of the

679 rapid mantle flow field surrounding the slab largely came out of 3D models trying
680 to explain processes in a specific geographic location on Earth (Conder and Wiens,
681 2007; Jadamec and Billen, 2010; Stadler et al., 2010; Alisic et al., 2012; Jadamec
682 and Billen, 2012; Jadamec and Fischer, 2014). A key implication of this result is the
683 importance in constraining the upper mantle rheology and slab strength, as these
684 two parameters play a fundamental role in modulating the flow rates (Andrews and
685 Billen, 2009; Alisic et al., 2012; Billen and Jadamec, 2012; Jadamec and Billen, 2012;
686 Jadamec, 2016).

687 **6. Discussion**

688 In Geodynamics, processes occur over a broad range in spatial scales (nanometer
689 to thousands of kilometers), time-scales (seconds to billions of years), and are com-
690 monly governed by non-linear material behavior (Kellogg et al., 2008). In addition,
691 unlike in modelling idealized media, the Earth is inherently heterogeneous giving rise
692 to complexity in geometry as a common-place phenomenon (Li et al., 2008; Hayes
693 et al., 2012). The 3D subduction models here have shown how complexity in the
694 geometry of the plate boundary can direct and guide mantle flow within the sub-
695 duction zone (Kneller and van Keken, 2007; Jadamec and Billen, 2012; Paczkowski
696 et al., 2014). In addition, the models illustrate how processes at the grain scale
697 such as dislocation creep can have a large effect on the morphology of mantle flow
698 at the kilometer scale (Jadamec and Billen, 2010; Stadler et al., 2010). The in-
699 crease in computational resources combined with advances in numerical algorithms
700 has allowed the modelling of high-resolution, three-dimensional, non-linear mantle
701 flow and slab deformation (Jadamec and Billen, 2010; Stadler et al., 2010; Burkett
702 and Billen, 2010; Rudi et al., 2015). This next section discusses implications of this
703 approach and looks ahead to future directions in 3D subduction modelling.

704 *6.1. Moving from the 2D Subduction modelling Paradigm*
705 *to 3D Subduction System modelling*

706 With increased computational resources and advanced numerical algorithms, sub-
707 duction modelling is moving toward subduction system modelling (Govers and Wor-
708 tel, 2005; Jadamec and Billen, 2010; Lin and Kuo, 2016). In other words, rather
709 than examining a single subducting plate, 3D models are now examining the slab
710 structure in more detail, including multiple subducting plates (Jadamec and Fischer,
711 2014; Lin, 2014; Lin and Kuo, 2016), allowing for complex slab-upper plate interac-
712 tions (Jadamec et al., 2013; Malatesta et al., 2013; Rodríguez-González et al., 2014),
713 and modelling the termination of subduction zones into other plate boundary types
714 (Govers and Wortel, 2005; Jadamec and Billen, 2010).

715 It is important to emphasize that many subduction zones on Earth are not iso-
716 lated features. Modern subduction zones are discontinuous and terminate into other
717 subduction zones or into transform or spreading centers, making lateral slab edges
718 actually a common feature (Figure 1) (Gudmundsson and Sambridge, 1998; Bird,
719 2003; Syracuse and Abers, 2006; Hayes et al., 2012). Examples include the subduc-
720 tion termination into (a) a transform boundary, as in eastern Alaska and Cascadia,
721 (b) a spreading center, as in the eastern New Hebrides and the Scotia subduction
722 zones, or (c) into another subduction zone forming a right angle or some other compli-
723 cated juncture, as in the Aleutian-Kamchatka junction or the Japan-Izu-Nankai triple
724 junction (Gudmundsson and Sambridge, 1998; Bird, 2003; Syracuse and Abers, 2006;
725 Müller et al., 2008; Schellart et al., 2011; Hayes et al., 2012). The lateral portions of
726 the subduction zones and related double and triple junctions, although a pervasive
727 feature of subduction zones, tend to be less studied, modeled less frequently, and
728 hence less understood (Casey and Dewey, 1984; Govers and Wortel, 2005; Wortel
729 et al., 2009; Jadamec and Billen, 2010; Gerya, 2011; Patriat et al., 2015; Lin and

730 Kuo, 2016).

731 Therefore, there is much to be learned from how the slab and mantle respond
732 to plate boundary intersections. This cannot be achieved by modelling flow within
733 the mantle wedge alone (Kneller and van Keken, 2007), rather the mantle wedge as
734 well as subslab flow must be solved for (Schellart et al., 2007), and the slabs must
735 be able to interact with the subslab mantle (Jadamec and Billen, 2010; Capitanio
736 and Faccenda, 2012; Faccenda and Capitanio, 2013) and with each other (Lin and
737 Kuo, 2016) or features in the upper plate (Jadamec et al., 2013; Capitanio, 2014;
738 Rodríguez-González et al., 2014). In particular there are interesting implications for
739 magmagenesis (Jadamec and Billen, 2010; Liu and Stegman, 2012; Menant et al.,
740 2016). For example, the contribution of slab derived melts would be expected if
741 warm mantle is upwelling adjacent to the slab edge, particularly if the slab edge
742 intersects with the subduction zone back arc spreading center or ridge system in the
743 overriding plate, such as in the eastern New Hebrides, southern Marianas, and Scotia
744 subduction zones (Figures 1 and 12) (Defant and Drummond, 1990; Durance et al.,
745 2012; Patriat et al., 2015).

746 Quantitatively linking the solid state creeping flow of the mantle in 3D with 3D
747 melt migration is an important and complex process and an active area of research.
748 Geographically referenced high-resolution modelling of specific subduction zones can
749 directly address specific hypotheses for a given region, which is a different approach
750 than that of generalized 3D models with a generic geometry (Jadamec and Billen,
751 2012; Jadamec et al., 2013). In order to study specific areas, the numerical models
752 need high resolution, not only to resolve the detailed slab geometry but also to re-
753 solve the multi-phase flow if magma melt and migration are to be incorporated in
754 a truly three-dimensional fashion. However, incorporating magma dynamics greatly
755 complicates the solver because of the hyperbolic systems associated with the poros-

ity field, as well as more complicated nonlinearities in the conservation equations (Spiegelman, 1993; Katz and Takei, 2013; Wilson et al., 2014). These future high-resolution 3D regional multi-phase models will be computationally challenging but also scientifically rewarding in elucidating the underlying physical mechanisms for complex subduction phenomena. This will require a comprehensive subduction system modelling approach, rather than disparate mantle wedge, magma migration, and solid state flow framework models run individually.

6.2. *Constraining Slab Structure from Regional 3D Geodynamic Models*

Because high resolution regional models are geographically referenced and can incorporate entire portions of the subduction zone (> 1000 km) and asthenosphere with local resolution of 2-3 km, the models can test the effect of incorporating different detailed slab geometries and thus help to distinguish between hypothesized slab structures where seismic data are lacking or limited. Qualitatively comparing a mantle flow field produced by a generalized geodynamic model to shear wave splitting patterns from a specific geographic location on earth can be thought provoking and valuable in generating ideas (Zandt and Humphreys, 2008). However, the geodynamic models show that the scale and intensity of the toroidal flow will depend on the density contrast, viscosity ratio, and component dimensions (Stegman et al., 2006; Piromallo et al., 2006; Jadamec and Billen, 2010; Faccenda and Capitanio, 2012). Therefore, in order to match a given shear wave splitting pattern to the mantle flow field in a particular subduction zone, for example, high-resolution 3D models that incorporate detailed slab structure and mantle properties become necessary, as does the calculation of the LPO or some measure of crystal deformation for the given flow field (Jadamec and Billen, 2010) and synthetic splitting (Miller and Becker, 2012).

For example, 3D instantaneous models of the Alaska subduction zone tested two

781 different morphologies for the eastern slab edge (Jadamec and Billen, 2010, 2012).
782 The models were geographically referenced, contained over 100 million finite ele-
783 ment nodes with a variable mesh resolution as fine as ~ 2.5 km, and run for $\sim 17,000$
784 compute hours per model (Jadamec and Billen, 2010, 2012). In this region, the seis-
785 mic data indicated an ambiguous morphology for the slab edge where the Aleutian
786 megathrust terminates into the Queen Charlotte-Fairweather transform fault system
787 (Page et al., 1989; Ratchkovski and Hansen, 2002; Eberhart-Phillips et al., 2006).
788 Testing two different slab morphologies, the geodynamic models produced very dif-
789 ferent locations for the toroidal flow, that were over 500 km apart depending on the
790 slab edge morphology (Figure 8) (Jadamec and Billen, 2010, 2012). By comparing
791 the infinite strain axes produced from the different mantle flow fields to observa-
792 tions from shear wave splitting, a preferred slab shape beneath eastern Alaska was
793 identified (Jadamec and Billen, 2010).

794 In another example, high-resolution geographically referenced 3D models Cocos-
795 Nazca slab window tested the effect of a continuous slab versus a slab gap on the rate
796 and direction of the mantle flow through the slab window, thus providing constraints
797 on the modern configuration of the Cocos-Nazca slab window (Jadamec and Fischer,
798 2014) (Figure 9). Recent 3D geodynamic models have also tested the complicated
799 slab intersection beneath Taiwan and the predicted mantle flow and mantle fabric
800 deformation (Lin and Kuo, 2016). Thus, in this way the high-resolution regional
801 models can help to constrain the slab geometry where seismic imaging is limited.
802 This is an important direction in subduction modelling and could benefit from more
803 sophisticated data assimilation techniques and inversion methods (Liu et al., 2008;
804 Liu and Stegman, 2012; Baumann and Kaus, 2015).

805 6.3. Slab-driven Three-dimensional Upper Plate Motion

806 Although this paper has mostly focused on scientific investigations of slab mor-
807 phology and mantle flow in subduction zones, a 3D model set-up that has high
808 enough resolution to resolve the detailed slab morphology is also in a position to
809 examine how the regional slab morphology controls deformation in the overriding
810 plate (Jadamec and Billen, 2010; Jadamec et al., 2013). In particular, because of the
811 high resolution, geographically referenced models of this kind can examine the flat
812 slab subduction processes, such as how the flat slab couples into deformation of the
813 overriding plate in 3D for a particular location on Earth.

814 A limiter of investigating how the overriding plate is affected by flat slab subduc-
815 tion has been the resolution required to resolve the overriding plate, plate interface,
816 slab, and subslab mantle, as well as how to represent the rheology of the plate in-
817 terface (Jadamec et al., 2012; Bercovici and Ricard, 2014; Sharples et al., 2014; Holt
818 et al., 2015). 3D numerical models have examined the dynamics of flat slab subduc-
819 tion (Mason et al., 2010; van Hunen and Allen, 2011; Manea et al., 2012; Taramón
820 et al., 2015; Hu et al., 2016) and explored the effects of plateaus and oblique subduc-
821 tion on along strike transport of upper plate material, e.g. Jadamec et al. (2013);
822 Malatesta et al. (2013); Capitanio (2014); Moresi et al. (2014).

823 Not all flat slabs have the same morphology (Jarrard, 1986; Gutscher et al., 2000;
824 Lallemand et al., 2005; Syracuse and Abers, 2006; Hayes et al., 2012). In some cases
825 the flat slab portion can occur at deeper depths such that it occurs beneath a thick
826 upper plate (~ 100 km), as in parts of northern Peru and Chile (Gutscher et al.,
827 2000). Whereas in other cases, the flat slab segment occurs at shallower depths,
828 such as in the Alaska subduction zone and in Columbia (Gutscher et al., 2000;
829 Ratchkovski and Hansen, 2002; Jadamec and Billen, 2010). For cases like this, the
830 upper plate is thin (< 50 km thick), and the thin region extends laterally for up to 500

831 km inboard from the trench (Gutscher et al., 2000; Ratchkovski and Hansen, 2002;
832 Gudmundsson and Sambridge, 1998; Jadamec and Billen, 2010; Hayes et al., 2012).
833 Computationally, this means the geodynamic model must have enough resolution to
834 resolve a thin upper plate, plate interface, and top of the slab within less than 50
835 km of the model surface. A global model with resolution of 100 km, 50 km or even
836 25 km could not resolve this plate interaction. Conversely, a mantle wedge model
837 with 10 meter resolution, would not extend below the slab and therefore could not
838 incorporate the effects of subslab mantle flow.

839 Therefore, by including resolution on the order of 2-3 km, in geographically ref-
840 erenced 3D regional models, one can begin to quantitatively explore how the flat
841 slab drives three-dimensional deformation in the upper plate from below (Jadamec
842 et al., 2013; Haynie and Jadamec, In Review 2016). This is a different approach
843 to investigating spatially variable upper plate deformation due to indenter tecton-
844 ics in a thin-viscous sheet model, where the force of the slab is approximated as a
845 boundary condition on the side of the model (England et al., 1985; England and
846 Houseman, 1986; Finzel et al., 2011). Rather, with the 3D geographically referenced
847 models that include the entire slab and its buoyancy effects, the role of the slab as it
848 drives the upper plate deformation from underneath the upper plate is investigated
849 (Jadamec et al., 2013; Eakin et al., 2014; Haynie and Jadamec, In Review 2016).
850 Recent examples of the examination of the effect of a high-resolution geographically
851 referenced slab have shown the slab geometry can have a first order control on upper
852 plate deformation, including forearc sliver motion, laterally variable motion along
853 intracontinental shear zones, and laterally variable basin and uplift formation, and
854 demonstrate the importance of quantitatively investigating the effect of a detailed
855 slab shape (Jadamec et al., 2013; Eakin et al., 2014; Haynie and Jadamec, In Review
856 2016).

857 6.4. *A New Kind of Exploration in Scientific Data visualisation*

858 3D immersive virtual reality is an active area in Computational Science that is
859 well suited for parsing the influx of digital information now available to the Earth
860 Sciences (Erlebacher et al., 2001; Kreylos et al., 2006; Chen et al., 2008; Kellogg
861 et al., 2008). 3D visualisation can have a direct impact on scientific discovery as
862 hands-on 3D visualisation facilities allow researchers to gain new insights into the
863 3D modelling results and hence Earth processes (Kreylos et al., 2006; Kellogg et al.,
864 2008; Jadamec and Billen, 2010). Moreover, scientific data visualisation is becoming
865 increasingly critical in working with the design, input, and output from 3D geody-
866 namic simulations (Figure 5).

867 With Gigabytes of data involved in the geodynamic model design, efficient mech-
868 anisms must be in place to inspect the 3D tectonic configurations (Erlebacher et al.,
869 2001; Kreylos et al., 2006; Chen et al., 2008; Jadamec and Billen, 2012). Thus open
870 source interactive 3D visualisation software becomes important to enable rapid in-
871 spection and interactive exploration of the 3D plate boundary geometry, thermal
872 structure, as well as non-linear mantle flow fields (Kreylos et al., 2006; Kellogg et al.,
873 2008). With interactive 3D data exploration, the quality and smoothness of the fea-
874 tures mapped onto the model grid and results from complex 3D flow fields can be
875 assessed and explored efficiently (Kreylos et al., 2006; Kellogg et al., 2008; Jadamec
876 and Billen, 2010).

877 For high-resolution 3D geographically referenced models, the model construction
878 phase and related 3D visualisation of this construction can lead to its own form of
879 scientific discovery, particularly when different data sets are visualised in an immer-
880 sive and interactive 3D environment. In global models, where the finest element size
881 is 100 km, 50 km, or even 25 km, density anomalies could be mapped on the model
882 grid assuming a global tomography model (van der Hilst et al., 1997; Romanow-

883 icz, 2003; Li et al., 2008), and the detailed structure is less important. However,
884 for high-resolution regional models, often multiple local seismic studies are required
885 to complement the global constraints in order to characterize the slab and mantle
886 structure at the scale of kilometers (Jadamec and Billen, 2010). Therefore, some *a*
887 *priori* synthesis of the disparate data types is required for mapping a given tectonic
888 feature (slab, upper plate, plate interface) onto the grid (Figure 5) (Jadamec and
889 Billen, 2012).

890 An interesting part of this process, is that the very act of having to map a
891 complicate slab structure onto the mesh grid forces the modeler to grapple with
892 questions such as how does a slab deform, as the complex slab shape in and of
893 itself has implications for the strength and rheology (Jadamec and Billen, 2012).
894 Therefore, in a sense, the modelers are now faced with similar conceptual realms
895 of discovery as are field geologists (Figure 13a). The models can now handle the
896 complexity, and it is up to the modelers to figure out how to adequately represent
897 these structures and discover what they mean and imply for subduction dynamics
898 (Figure 13b).

899 *6.5. Limitations of the modelling and Technology*

900 Caution must be used, however, such that the new technology is embraced with
901 cognizance of limited data, numerical, and theoretical constraints. Mapping the fea-
902 tures onto a high resolution mesh can be complicated and prone to errors or artifacts
903 when the features are not aligned with respect to the mesh grid (Billen et al., 2008).
904 This requires more sophisticated programming to construct the input (Jadamec and
905 Billen, 2010). In addition, many assumptions arise in transferring the information
906 to the grid. The use of multiple and disparate data sets to constrain the high res-
907 olution tectonic features also leads to challenges, as the resolution, error, and data

908 density of the various observational constraints are typically different. Furthermore,
909 a challenge with the high resolution models is a lack of data coverage, because the
910 high-resolution mesh requires greater coverage in the observational constraints. This
911 leads to need for some form of uncertainty quantification and also requires multiple
912 phases of testing.

913 With the complicated geometries, it is even more necessary to run separate sets
914 of models on simplified set-ups to optimize the solvers, particularly when using the
915 nonlinear rheologies which can lead to viscosity variations of over 7 orders of mag-
916 nitude over short spatial scales (100 km) (Moresi et al., 1996; Jadamec et al., 2012;
917 Zhong et al., 2015). Moreover, it becomes increasingly important to understand the
918 background physics and material properties upon which the models are based, such
919 as what are the appropriate rheologies for the mantle especially when rock deforma-
920 tion experiments are extrapolated to strain rates over several orders of magnitude.
921 It is not always clear how to represent a given set of geologic or geophysical obser-
922 vations in terms of physical parameters in the model, and this choice can vary for a
923 given piece of software, or in a given numerical algorithm (Jadamec et al., 2012).

924 Another problem becomes scientific reproducibility. As various data sets are
925 integrated in inconsistent ways and used in models, and as models are run with
926 different versions of the numerical modelling software, it becomes a challenge to
927 reproduce a given set of models and modelling results. As modelling codes are
928 constantly being updated, future modelling efforts will need to archive the versions
929 of codes used, as well 3D as model input.

930 **7. Conclusions**

931 Reflecting on the development of geodynamic modelling, the changes in technol-
932 ogy have facilitated a major paradigm shift in our understanding of the nature of

933 mantle flow in subduction zones. 3D models reveal the complex nature of slab-driven
934 mantle flow, including trench parallel flow, toroidal flow around slab edges, mantle
935 upwelling at lateral slab edges, small scale convection within the mantle wedge, and
936 the effects of lateral gravitational instabilities above the mantle wedge.

937 Advances in high-performance computing allow geodynamicists to bring an un-
938 precedented level of detail into numerical models, prompting a high-resolution 3D
939 regional modelling approach and a new method for mapping the high resolution
940 data onto a mesh grid. The coupling of the HPC with immersive three-dimensional
941 data visualisation now allows the user to bring physical meaning to the hundreds
942 of millions of numbers that numerically represent the model, enabling the numeri-
943 cal modeler to grasp the model components as would a laboratory modeler working
944 with a tangible physical construct, and gives the visualisation a new role in scientific
945 discovery.

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Figure 1: Global tectonic map of plate boundaries (black lines - Bird (2003)), slab surface contours in 50 km intervals (gray lines - Gudmundsson and Sambridge (1998)), and ages of the oceanic plates (color - Müller et al. (2008)). Subduction zones: Ak - Alaska, Al - Aleutian, An - Andaman, Ba - Banda, Bl - Bolivia, Br - New Britain, Bu - Burma, Ca - Cascadia, Ch - Chile, Cl - Calabria, Cm - Central America, Co - Columbia, He - Hellenic, Hi - Hikarangi, Iz - Izu, Ja - Java, Jp - Japan, Ka - Kamchatka, Ke - Kermadec, Ku - Kurile, La - Lesser Antilles, Mk - Makran, Mr - Marianas, Na - Nankai, Nb - , Nh - New Hebrides, Pe - Peru, Ph - Philippine, Ry - Ryuku, Sc - Scotia, Sh - South Shetland, Sm - Sumatra, To - Tonga, Tr - Trobriand. For additional plate boundary labels see Schellart et al. (2008).

Figure 2: Global synthesis of shear wave splitting in the mantle wedge. Arrows represent azimuth of the average seismic fast directions and are color-coded by the pattern of the azimuths of the seismic fast axes with respect to the trench. The patterns show a range from trench perpendicular to trench parallel fast directions. Numbers give range in delay times. Figure from Long and Wirth (2013).

Figure 3: Total cores and Rmax value for supercomputers worldwide ranked Number 1, from 1993-2015, according to TOP500 list ranked by performance on LINPACK Benchmark (Strohmaier et al. (2015), <http://www.top500.org>). Both the number of compute cores and Rmax (TFlops/s) are plotted, wherein the Rmax is a measure of the peak number of floating point operations (Flops) per second, in Tera-Flops.

Figure 4: Schematic of steps in geodynamic modelling workflow. For 3D regional models incorporating extensive observationally based model design, the model design step can take as long as the model run step. In addition, the model performance and evaluation steps are typically longer due to the more complicated geometries and rheologies in regionally based models.

Figure 5: Workflow for high resolution regional 3D numeric models. (a) Data collection and pre-processing. (b) Example of three-dimensional variable-resolution model mesh and three-dimensional thermal structure in the overriding plate, subducting plate, and mantle generated by Subduction-Generator. (c) The thermal structure is used to calculate the density anomalies that drive the viscous flow in CitcomCU. Flow field shown for subset of model domain. 3D images modified from Kellogg et al. (2008); Jadamec and Billen (2010, 2012); Jadamec et al. (2013).

Figure 6: Schematic 2D subduction model set-up. (a) 2D corner flow model that solves for flow in mantle wedge, with a plate velocities applied as a boundary condition. (b) 2D free subduction model that solves for subslab flow, mantle wedge flow, and slab deformation. (c) 2D subduction model with overriding plate that solves for upper plate motion as well as subslab flow, mantle wedge flow, and slab deformation.

Figure 7: slab-driven mantle flow cartoons. (a) 2D corner flow in the mantle wedge with the downwelling parallel to the slab surface. (b) Along strike flow in the mantle wedge due to along strike slab dip variability and pressure gradients. (c) Downwelling in the mantle wedge due to steepening and/or slab rollback of a free slab such that the flow is at an angle to the slab surface, toroidal flow around the slab edge, and associated upwelling at lateral slab edge. (d) Subduction zone with slab window/tear and two opposing cells of toroidal flow and slab edge upwelling. (e) Slab edge intersection with back-arc spreading center, showing both toroidal flow and slab edge upwelling as well as upwelling associated with back arc spreading. (f) Opposing subduction zones with slab steepening and back arc spreading.

Figure 8: Predicted flow velocities in high-resolution 3D instantaneous model of the North American-Pacific plate system in Alaska, with subset of model domain shown. Predicted Pacific plate motion (black arrows)(a) is non-parallel to counterclockwise toroidal flow in mantle at 100 km depth (b). (c) Comparison of calculated infinite strain axes (colored bars) from flow field in Jadamec and Billen (2010) (b) to the fast axes of seismic anisotropy from Christensen and Abers (2010) (blue bars). Modified from Jadamec and Billen (2010).

Figure 9: Example of three-dimensional mantle flow through Cocos-Nazca slab window from a high-resolution regional slab model of the Caribbean plate region, showing counterclockwise toroidal flow around the southern Cocos slab edge and clockwise toroidal flow around the northern Nazca slab edge. The 3D thermal structure was constructed with SubductionGenerator (Jadamec et al., 2012; Jadamec and Billen, 2012). Model domain given in the inset was discretized into $897 \times 865 \times 113$ elements in the longitude, latitude, and radial directions, respectively, and mesh resolution ranges from 2.35 km to 25 km, with the finest resolution centered on the slab gap (Jadamec and Fischer, 2014). Model was run on 384 processors for 24 hours (~ 9000 core hours) using the open source mantle convection code CitcomCU (Zhong, 2006) with free-slip boundary conditions, and used the composite viscosity formulation based on the experimentally derived flow law for olivine (Hirth and Kohlstedt, 2003) implemented in Jadamec and Billen (2010). Dimensionalization and flow law parameters are the same as those used in Jadamec and Billen (2010). The examination of the length-scales of the flow is still actively being studied.

Figure 10: Illustration of slab-driven mantle deformation zone from high resolution, 3D numerical models of the Alaska subduction-transform system. (a) slab-driven toroidal flow, slab edge related upwelling, and poloidal flow in the mantle surrounding the slab. (b) Viscosity cross section through subset of model domain shows laterally emergent low viscosity region surrounding slab in composite viscosity model. (c) Viscosity profile for Newtonian model (i.e., without the strain-rate dependence in the viscosity flow law). Inset shows profiles of the viscosity for the composite viscosity models (black, gray lines) and Newtonian viscosity models (blue lines). Modified from Jadamec and Billen (2012) and Jadamec (2016).

Figure 11: Results from 2D numerical models of subduction using a composite viscosity show decreasing mantle viscosity and increasing mantle flow velocity, as a function on decreasing maximum yield stress (slab strength). (a) Minimum mantle wedge viscosity emergent in the models. (b) Percent change in maximum velocity in mantle wedge with respect to the models with a highest yield strength. The lower the maximum yield stress, the greater the slab-driven mantle weakening. The faster flow and weaker mantle viscosities are not due to sidewall proximity. Model boundaries are over 1500 km to the left of the slab and over 2000 km away to the right and below the slab, making the distance to the model boundaries over 4 and 5 times with length of the slab, respectively. Modified from Jadamec (2016).

Figure 12: New Hebrides subduction zone intersection with Central Spreading Ridge. (a) Tectonic map. Red star marks slab edge-ridge intersection where D3-1 sample was collected in Durance et al. (2012). Plate boundaries - black lines (Bird, 2003); Slab contours - thin lines, 50 km intervals (Syracuse and Abers, 2006). Holocene and Pleistocene volcanoes - magenta triangles (Smithsonian Institute, Global Volcanism Program). (b) Geodynamic cartoon of slab edge-back arc spreading intersection. (c) Comparison the geochemistry for volcanic samples, suggestive of adakitic geochemistry in sample D3-1. Modified from Durance et al. (2012).

Figure 13: Mapping actual outcrops versus constructing virtual models from digital observations. (a) Recording observations of fold and fault structures forces the field scientist to connect geologic structures in 3D (Jadamec and Wallace, 2014). (b) Mapping complexity at the plate boundary scale onto a high-resolution 3D model grid (up to ~ 2.5 km) forces the geodynamicist to connect plate-scale structures in 3D. Rainbow - Slice through thermal structure (0 - 1400° C); Purple - Isosurface of plate interface volume; Yellow - Wadati-Benioff zone from Jadamec and Billen (2010). Modified from Kellogg et al. (2008).