Mantle convection models featuring plate tectonic behaviour: an overview of methods and progress

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Abstract

Arguably, the presence of plate-tectonic-type surface motion for periods that endure over hundreds of millions of years is the primary feature a mantle convection model must possess in order to be considered Earth-like. From the early days of mantle dynamics modelling, research has been dedicated to understanding how mantle convection produces the first order observations of plate tectonics as well as how the plates and deep mantle interact. Fledgling studies of the effect of plates on the mantle recognized the ability of imposed plate-scale surface motion to influence global temperatures and heat flux and organize convective planform. Later studies featuring model plates with dynamically determined velocities discovered that the interaction between convection and plates could result in cyclic plate motion patterns and other time-dependent behaviour that was not manifested in systems in which dynamic plates were absent. Focussing on different aspects of system realism (with respect to terrestrial mantle convection) has spawned multiple approaches for modelling convection with dynamic integrated plates. In broadest terms, the two main approaches can be categorized as rheological modelling methods and methods utilizing evolving surface boundary conditions. Over the past dozen years, studies focussing on the former approach have steadily made progress in modelling the self-generation of plate tectonics from convection dynamics. Continual advances have been encouraging, and a consensus is beginning to form regarding the necessary requirements for obtaining the primary elements of plate-like surface motion. However, despite significant progress, the generation of plates over long periods has not yet been modelled with Earth-like convective vigour. In contrast, models utilising dynamically determined boundary conditions to achieve plate-like surface motion have relatively little difficulty with emulating terrestrial convective vigour or simulations of billions of years. Instead, their weakness is more fundamental; they can only provide insight into the reciprocating dynamics of the mantle and plates once the existence of the plates is assumed and they cannot model any aspects of the dynamics responsible for the origin of the plates. This paper briefly reviews the evolution of mantle convection models featuring plates and examines the progress that has been made in our understanding of the feedback between the mantle and plate tectonics through the use of both rheological and boundary condition modelling methods. Common findings, recent advances and unbridged problems are identified and discussed.

Key words: mantle, convection, force-balance, plate tectonics, yield stress, plate velocities, global heat flux

PACS:

1 Introduction

After languishing for decades on the periphery of mainstream scientific discourse, the concept of Continental Drift was resurrected and adapted to fit our current understanding that it is a corollary to geology's modern grand unifying paradigm, Plate Tectonic theory. Plate tectonics (e.g., Wilson, 1965; McKenzie and Parker, 1967; Morgan, 1968) provides the descriptive framework that has illuminated and explained a myriad of geological and geophysical phenomena, but the theory itself does not provide a clear explanation for why plates exist or why they move. The quintessential testimony to the success of plate tectonics is that through a robust comprehensive model comprised of clear

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concepts it reconciles recognition of the similarity of the Atlantic's opposing coastlines (e.g., Bullard et al., 1965; Dietz and Holden, 1970) with countless geologic observations that verify continental drift. However, the associated notion of plate motion originating with convection in the mantle has resulted in multiple interpretations.

Nearly 80 years ago, Arthur Holmes (1931) proposed that continents might move due to convective overturning in the Earth's interior. Holmes correctly identified convection as the ultimate driving force for continental motion but his work did not define plates in the modern sense, rather it described motion at the surface (specifically continental movement) as resulting from stresses due to underlying convection (Pekeris, 1935; Hales, 1935). An unfortunate legacy of the identification of continental drift with convection, prior to the conceptualization of plate tectonics, is the often encountered treatment of the plates as being a distinct entity atop a convecting mantle. However, plates are not carried by the mantle; rather, they comprise the uppermost layer of the mantle (with the addition of a thin crustal veneer). Plate movement is not just a consequence of mantle motion but is a direct observation of the convecting system (e.g., Bercovici, 2003). Laminar convection (of the sort that occurs in an essentially momentum free fluid like the mantle) is characterised by the development of a distinct mobile thermal boundary layer at the surface. Translation of the boundary layer away from points associated with convective divergence at the surface, towards zones of convergence where cooled material plunges into the fluid interior, provides a first order description of the process that explains plate movement on Earth (e.g., Peltier et al., 1989).

Although the foregoing description suggests a rudimentary definition for a plate in terms of the thermal structure of a convection cell, it does not provide any insight into the origin of distinct plate characteristics. For example, in contrast to the Earth's surface, convecting fluids with weakly temperature-dependent viscosities generally exhibit mobile surfaces characterised by variable strain rates and nonuniform velocity (e.g., Weinstein and Christensen, 1991). Conversely, fluids with strongly temperature-dependent viscosities develop a cold lid that can decouple from a vigourously convecting underlying layer as the temperature-dependence of the viscosity is increased (Christensen, 1984; Solomatov, 1995). (When the temperature-dependence approaches the range appropriate for modelling planetary mantles a mode of convection is obtained in which surface motion is absent (Nataf and Richter, 1982; Morris and Canright, 1984; Nataf, 1991) - known in the literature as 'stagnant-lid' convection and discussed further in later sections.) In either case, the main characteristics of the Earth's surface velocity field are absent; namely, a system featuring broad regions of nearly uniform velocity separated by narrow bands of intense high strain rates (e.g., Peltier et al., 1989; Schubert et al., 2001). Intermediate grades of viscosity temperature-dependence show varying types of behaviour (to be discussed below) but come no closer to yielding plate-like behaviour without augmentation of the viscosity dependence beyond just thermal reliance. Recognition that 'plate-like' behaviour of the surface velocity and strain rate fields is not forth-readily obtained with a convecting fluid (e.g., Gurnis, 1989; Weinstein and Olson, 1992) has led to the development of a vigourous sub-discipline within the mantle convection modelling literature focused on identifying the critical ingredients required to obtain not just plate-like velocities but other characteristics that distinguish plates from 'typical' convection, such as the appearance of one-sided subduction (e.g., Zhong and Gurnis, 1995), transform faults (e.g., Bercovici, 1998; Gerya, 2010) and an associated strong toroidal component of the surface velocity field (Bercovici, 1995; Tackley, 2000b). To date, a substantial degree of progress has been made on the 'self-consistent generation of tectonic plates problem' (e.g., Moresi and Solomatov, 1998; Trompert and Hansen, 1998; Tackley, 1998, 2000; Stein et al., 2004; Van Heck and Tackley, 2008; Foley and Becker, 2009), however, obtaining long term plate-like surface motion in a model featuring Earth-like mantle convection vigour has yet to be achieved.

Models now leading the field of mantle convection studies featuring self-generated tectonic plates share a common origin with an alternative approach to modelling dynamic plates in convection calculations. These models specify time-dependent, dynamically-determined, plate-like surface motion directly, as a boundary condition (e.g., Gable et al., 1991; Monnereau and Quéré, 2001; Lowman et al., 2001; King et al., 2002; Quéré and Forte, 2006; Brandenburg and van Keken, 2007) and though limited by the fact they cannot address questions concerning the origin of plates, they allow for calculations that can explore long term plate-evolution in the Earth's convective regime. In the review to follow we shall compare advances made using a variety of modelling methods that feature plate-like surface dynamics. In particular, we consider the significance of obtaining plate-like surface motion, the history and recent advances in our understanding of the influence of plates on mantle convection and the benefits and downsides of rheological versus boundary condition methods for obtaining plate-like surface motion.

Prior to commencing on a review of the characteristics of convection featuring plates, familiarity with several fundamental parameters is required (e.g., the Prandtl, Rayleigh and Nusselt numbers). Descriptions of these fluid dynamic nondimensional quantities are given in the Appendix.

2 The influence of plate-like surface motion on convection

Plate interiors move at nearly uniform velocities (e.g., Minster and Jordan, 1978; DeMets et al., 1994) and the heat flux delivered from a thickening plate decreases with distance from the ridge axis in analogue with plate cooling models (e.g., Parsons and Sclater, 1977; Sclater et al., 1980; Stein and Stein, 1992, 1994; Carlson and Johnson, 1994) These observations alone are sufficient to suggest that the thermal evolution of the mantle must be profoundly influenced by the existence of plates. Accordingly, early studies of the effect of plate-like surface motion on the Earth's deep interior examined the convection patterns (two-dimensional) and later planforms (three-dimensional) of convecting fluids driven by a combination of thermally derived buoyancy and imposed surface velocities (e.g., Parmentier and Turcotte, 1978; Lux et al., 1979; Jellinek et al., 2004; Nettelfield and Lowman, 2007). Convection calculations with surface velocities prescribed *a priori* are widely referred to as *kinematic* models. Kinematic models provide the starting point for examining the influence of plate-like surface motion on the underlying convection. The simplest kinematic models feature imposed surface velocity conditions that are fixed in both space and time (Richter and McKenzie, 1978; Lux et al., 1979; Davies, 1988a, 1988b, 1988c).

Simple calculations featuring uniform property fluids show that when a specified surface velocity is comparable to or greater than the mean surface velocity, $\overline{v_b}$, obtained from the purely *buoyancy* driven flow (assuming a free-slip surface) convection patterns organize to reflect the profile of the surface velocity field. Both numerical and laboratory tank models have verified that the influence of the surface velocity field profile on the convection cell arrangement disappears as the surface velocity (or *velocities* if multiple plate sections are prescribed) is decreased relative to $\overline{v_b}$ (Lux et al., 1979; Davies, 1989; Han and Gurnis, 1999; Jellinek et al., 2004).

In the discussion that follows we shall refer to convergent plate boundaries and plate velocities in the context of kinematic models and therefore with the understanding that the 'plates' are modelled by specifying a constant velocity boundary condition over segments of the surface. Each distinct segment comprises a plate.

Studies with non-evolving plate boundaries indicate that a key influence of the plates is their ability to determine the site of downwelling flow (e.g., Davies, 1989; Bunge and Richards, 1996). Once plate velocities become comparable to $\overline{v_b}$, instabilities that develop in the thermal upper boundary layer start to fail to develop into full downwellings before they are entrained by surface motion to a point of convergence in the surface velocity field (Lux et al., 1979; Jellinek et al., 2004). Accordingly, downwelling flow starts to mirror sites of velocity field convergence at the surface (Bunge and Richards, 1996). Downwellings unrelated to plate convergence sites become absent or relatively weak and are swept into the regions occupied by the robust downwellings associated with convergent plate boundaries. The association of downwellings with the sites of plate velocity convergence is observed in both two-dimensional and three-dimensional and rectangular and spherical geometries (Fig. 1).

In contrast to downwellings, the upwellings that develop in convection models with specified plate velocities can greatly differ in two-dimensional versus three-dimensional calculations. Of course, in two-dimensions upwellings are restricted to a sheet-like form whereas in three-dimensional calculations featuring Earth-like convective vigour and plate-like surface motion they most often form plume-like structures, characterised by long cylindrical conduits and broad heads (Monnereau and Quéré, 2001; King et al., 2002). However, in addition to differing morphologies, the upwellings in two-dimensional models frequently appear below divergent plate boundaries (Lux et al., 1979; Jellinek et al., 2004) due to the plate scale flow imposing a roll type flow pattern on the convection. In three-dimensional calculations featuring multiple plates of different sizes and shapes, the surface motion does not produce roll-type convective patterns and upwelling locations are not typically coincident with divergent boundaries (Zhong et al., 2000; Monnereau and Quéré, 2001; Lowman et al., 2004; Quéré and Forte, 2006; Nettelfield and Lowman, 2007). Consequently, intra-plate plumes are easily formed.

The foregoing comparison of upwelling and downwelling locations and shapes remains valid in calculations featuring mantle viscosity profiles that increase by factors of 30 to 100 in the lower mantle (e.g., Bunge and Richards, 1996). A viscosity contrast between the upper and lower mantle in this range is arguably one of the most fundamental requirements for modelling the Earth's interior. A high viscosity lower mantle (e.g., Hager, 1984; Richards and Hager, 1984; Ricard et al., 1984; Forte and Peltier, 1987; Sabadini and Yuen, 1989; Spada et al., 1992; King and Masters, 1992) and the presence of plates effectively stabilize the lower and upper thermal boundary layers of a mantle convection model. However, the presence of plates in a stratified viscosity model allows multiple focussed downwellings to form whereas, when plates are absent, even in models with aspect ratios as great as 12, stratified viscosity convection typically exhibits very long wavelength flow with a single downwelling that becomes increasingly diffuse with depth (Hansen et al., 1993, Dubuffet et al., 2000; Lowman et al., 2001). Later in this review, we shall see that not only does the presence of plate-like surface motion have a particularly dramatic influence on underlying convection when the viscosity of the lower mantle is 30 or more times greater than the upper mantle viscosity but, in addition, a high viscosity lower mantle may be a vital component for producing plate tectonics at the top of the convecting mantle (Tackley, 2000b; Richards et al., 2001; Stein et al., 2004).

Over the past decade, the objectives of mantle convection studies featuring specified plate velocities and geometries (i.e., plate boundary locations) have evolved from their early focus on exploring the influence of plates on mantle convection to recent applications that concentrate on simulating the evolution of the Earth's interior (e.g., Schuberth et al., 2009a, 2009b; Zhang et al., 2010). The ability to model convection with terrestrial parameters in a full spherical geometry is now being combined with specifying the latest global plate history reconstruction models (which give plate velocities and plate boundary locations) to produce corresponding mantle density distribution history models by forward-modelling. The approach, borrowed from meteorology, is generally referred to as sequential data assimilation (Talagrand, 1997). One major limitation on the great potential of Earth history simulation to forward the interpretation of current observations from fields like seismic tomography, is our understanding of the dynamic feedback between surface motion and mantle convection. The following sections describe several findings that indicate how the relaxation of specified plate velocities allows for especially time-dependent convection resulting from the feedback between convection and plate movement.

3 Modelling dynamic plates

Although a specified plate-like surface velocity field produces results that look like convection with plate tectonics, calculations of this type have deficiencies. The surface motion is not determined by a calculation accounting for the system dynamics. Consequently, the specification of the surface velocity may not be consistent with the buoyancy derived forcing and therefore the boundary conditions can add or drain the system of energy. Not only the magnitude but also the direction of the plate velocities are unable to respond to the evolving buoyancy distribution within the convecting mantle. In addition to the inhibited dynamics of the plate velocities, the models are also limited by the prescribed geometry of the plates.

A check on matching prescribed plate velocities with the vigour of the buoyancy driven convection can be obtained by examining the laterally averaged root-mean-square of the horizontal velocity below each plate. In regions along the base of the plate where there is no vertical mass exchange, the vertical gradient of this quantity gives the shear stress on the base of the plate. Consequently, a laterally integrated velocity gradient that vanishes at the base of a plate should indicate a prescribed plate velocity that is consistent with forcing due to buoyancy. However, for time-dependent convection, a temporally constant plate velocity that yields a time-averaged stress of zero at the base of the plates differs from a temporal average of the plate velocity time series from a calculation where a mean stress of zero is required at all times at the base of the plate, rather than just as an average. Thus, snapshots of the convecting system can only be used to determine consistent plate velocities at the corresponding moment, averaging velocities in the convecting system to determine plate velocities consistent with the buoyancy forcing leads to modelling a system with constrained dynamics. The point is clarified by considering the velocity of a plate that periodically changes its direction by 180 degrees. Averaging its velocity over one period yields a velocity of zero. However, a system with a specified plate velocity of zero evolves very differently from the system with the periodically changing velocity. Systems that model plates with velocities matched to the root-mean-square of a corresponding calculation featuring a free-slip surface will similarly differ from models in which the net forcing on the plates is set to zero at all times by the choice of an evolving plate velocity (Fig. 2). The two averages will differ because of the affect of plates on mantle planform and upper mantle velocities.

The first step towards obtaining self-consistent plate tectonics in an integrated plate-mantle system focussed on achieving dynamically determined plate velocities. In such models, the surface velocity field is plate-like and potentially time-dependent plate velocity magnitudes and directions are determined by the evolving buoyancy distribution in the convecting system.

Several distinct approaches have been adopted to achieve convection solutions with dynamic plate velocities, and comparisons of the findings obtained has shown agreement between system diagnostics such as the mean temperature, Nusselt number and mean plate velocity to within a few percent (King et al., 1992; Koglin et al., 2005). Rheological methods (e.g., Olson and Corcos, 1980; Davies, 1989; Gurnis and Hager, 1988; King and Hager, 1990; Weinstein and Olson, 1992; Zhong and Gurnis, 1995; Puster et al. 1995; Zhong et al., 2000; Burkett and Billen, 2009) model plate boundary locations by including zones of narrow finite thickness low viscosity (i.e., weak zones) at the top of the convecting layer. Elsewhere, a high viscosity is specified at the surface of the model, either implicitly, through the utilization of a temperature-dependent rheology (Gurnis, 1989; Davies, 1989; Zhong et al., 2000), or explicitly, by adding a high viscosity layer of comparable thickness to the predicted thermal boundary layer thickness (King and Hager, 1990; Koglin et al., 2005). The weak zones allow focussed deformation to occur in the upper thermal boundary layer that would otherwise 'lock-up' due to its high viscosity. Consequently, confined zones of deformation occur, analogous to plate boundaries, and these separate broad regions exhibiting low strain rates and nearly uniform velocities, corresponding to plate interiors (Fig. 3). Most importantly, unlike the surface velocities of kinematic models, these models feature dynamically determined plate velocity magnitudes and directions.

Typically, models featuring plates obtained through rheological methods are referred to as *dynamic* to distinguish them from models with specified velocities. However, the kinematic approach to including plate-like surface velocity fields in mantle convection models can be adapted to also give dynamically determined plate velocities. The *force-balance* method (e.g., Gable et al., 1991; Monnereau and Quéré, 2001; Lowman et al., 2001; King et al., 2002; Quéré and Forte, 2006; Brandenburg and Van Keken, 2007, Brandenburg et al., 2008) specifies the surface velocity of the plates in a system with an arrangement of plate sizes and shapes chosen by the modeller, much like weak zones locations are specified in the rheological models described above. By treating each plate as having a bounding surface (that is, a well defined top, bottom, sides and ends) the total force acting on each plate can be determined. The forces accounted for are due to buoyancy within the plate and remaining mantle, the motion of the plate itself, and the motion of all of the other plates in the system. At any time, a unique set of plate velocities can be determined so that the total force acting on each plate is zero, given the buoyancy field at the corresponding time. A requirement in implementing this method is that the velocity profile (i.e., function shape but not amplitude) for each plate must be prescribed. However, any velocity function can, in theory, be specified. Thus, typical surface velocity fields are specified to feature nearly uniform velocity plate interiors and a rapid reduction in plate velocity near the plate boundaries. By updating the plate velocities in tandem with the stresses driven by the temperature field, the force balance method achieves dynamically determined, time-dependent, plate-like surface velocity fields.

Although rheological plate modelling methods may at first appear preferential to specifying the surface velocity of the system, closer examination reveals that the force-balance modelling approach is capable of producing very similar results to models featuring rheologically defined plates (King et al., 1992, Koglin et al., 2005). Indeed, the perceived downside of a force-balance method (that it specifies surface motion explicitly) is ultimately analogous to specifying a spatially defined model rheology in order to obtain plate mobility. Both approaches yield plate velocities that are sensitive to a group of parameters that can be used to 'tune' the model output (King and Hager, 1990; King et al., 1992). For example, both methods allow for specifying plates with different thicknesses (Koglin et al., 2005; Ghias and Jarvis, 2007; Gait and Lowman, 2007) and viscosities. Rheological methods can specify weak zones of different sizes and rheologies (King and Hager, 1990) and force-balance methods can specify different width regions over which plates make the transition from the velocity of one plate to its neighbour, or parameterize collisional forces at convergent plate boundaries. In the latter case, it is important to recognize that the gradient of the plate velocity fields at the plate boundaries affect the calculated plate velocity in an analogous way to weak zone viscosities or dimensions in rheological models. The calculated plate velocity depends on the specified width of the plate boundary regions because discontinuities in the velocity field, as would exist at an infinitesimally narrow plate boundary, result in singularities in the stress field. In order for the plates to move, the change in velocity from zero (at the plate boundary) to the plate interior speed has to be spread over a finite width region. Accordingly, the plate velocities obtained by force-balance and rheological methods are both determined by the model setup. However, findings show that appropriate choices of the plate characteristics yield good agreement between the results obtained with different modelling methods (see Fig. 4). To bring about this agreement, several simple cases can be used to calibrate model tuning (Lowman et al., 2001; Koglin et al., 2005). For example, the observation that the surface heat flux from a simple square Cartesian system with reflecting sidewalls and a single-plate can be matched to the heat flux from a similar free-slip surface model (Gurnis, 1989) provides a reference case for a plate model (Lowman et al., 2001). Once the model parameters that produce well understood behaviour in simple models have been determined (e.g., two-dimensional single plate models) these same parameters can be specified in more complex models (e.g., multi-plate models and/or three-dimensional cases).

3.1 Heat transfer and the influence of plates

The majority of mantle convection studies that feature dynamic plates have not allowed for the evolution of plate geometry. For example, studies implementing rheological methods often assign weak zone locations, and models employing a force-balance method typically specify fixed plate boundaries. These studies provide insight into the influence of plates on convection but can also exaggerate the influence of plates. Unlike the Earth, models with fixed boundaries show the effects of plates that remain unchanged during the passing of many mantle overturns. Consequently, mantle convection is given the opportunity to adjust to plate boundary locations on time scales that are non-applicable to the Earth. However, despite their limitations, these studies do provide some insight into the feedback between plates and the underlying mantle.

Like convection models with kinematic plates, dynamic plate models show that plates affect convection wavelength and that the geometry of the plates can be instilled in the mantle interior (Monnereau and Quéré, 2001). In simple two-dimensional convection models, convection cells stretch in response to increasing plate lengths (Lowman et al., 2001). In more complex three-dimensional spherical calculations featuring plates, in comparison to convection with a free-slip surface, plume and slab morphology changes and the number of upwellings and downwellings decreases (Monnereau and Quéré, 2001).

Studies show that in a convecting system, regardless of whether they have continental or oceanic properties, large plates act to retain heat (e.g., Monnereau and Quéré, 2001; Lowman et al., 2001; Grigne et al., 2005). Moreover, increasing plate size decreases mean surface heat flux (i.e., Nusselt number). A spherical system with multiple plates yields more heat than a case with a rigid surface (the single plate case) and increasing the number of plates allows surface heat flow to increase (Fig. 5a). However, the mean heat flux from a system with plates will not exceed the heat flux from a system with a free-slip surface (which can be interpretted as an infinite number of infinitesimally small plates). Consistent with their affect on surface heat flow, plates also influence mantle temperature. As plate number is decreased and mean area increased, interior temperatures increase (Fig. 5b).

Heat transfer is also affected by plate configuration not just *mean* plate size (Gable, 1989). A plate-mantle system with a mixture of large and small plates yields a lower mean heat flux than a case with plates of the same size (Fig. 6). In addition, a range of plate sizes results in more time-dependent plate velocities (Lowman et al., 2003) and heat flux (Gait and Lowman, 2007).

3.2 Time-dependence and vigorous convection

In an effectively infinite Prandtl number fluid, like the Earth's mantle, the time-dependence of convection can be enhanced by increasing the system Rayleigh number (Chandrasekhar, 1961), a measure of the vigour of the convection, or by the addition of internal heating (Jarvis, 1984). These thermal parameters also have an effect on the convective wavelength adopted by the system (as do a host of other parameters including temperature- (Tackley, 1993; Ratcliff et al., 1997; Solomatov and Moresi, 1997) and pressure-dependent (Gurnis and Davies, 1986; Cserepes, 1993; Bunge et al., 1996) viscosity and pressure-dependent thermal expansivity (Hansen et al., 1991, 1993)). By 'enhanced' time-dependence we refer specifically to characteristics such as a surface heat flux time series that exhibits great variation and a time-dependent number of convection cells or thermal plumes. Interestingly, research has shown that the transitions that occur between different time-dependent regimes in calculations featuring free-slip surfaces are also manifested as flow regime changes in both rheological and force-balance models when plates are present (Lowman et al., 2003; Koglin et al., 2005). For example, the same thermal parameters that produce narrowed convection cells in the absence of plates can produce highly time-dependent plate velocities when plates are modelled (Lowman et al., 2003). The time-dependence increases with the addition of internal heating and is characterized by rapid changes in plate direction and an accompanying change in the direction of convective overturning (Lowman et al., 2001; Koglin et al., 2005). In two-dimensions this behaviour can result in periodic or intermittent changes in plate velocity of 180 degrees. In three-dimensional convection it can result in regularly occurring but unpredictable 'plate reorganization events' (Fig. 7) marked by rapid changes in direction of two or more plates (King et al., 2002, Gait et al., 2008).

Although the relevance of reorganization events to the Earth's evolution is unclear, multiple studies and models have shown that they are a common feature of both two-dimensional (Lowman et al., 2001) and three-dimensional (King et al. 2002) models and systems featuring rectangular (Koglin et al., 2005) or curved (Ghias and Jarvis, 2007) geometries. In addition, plate reorganization events have shown that changes in the direction of plate motion driven by buoyancy forces can occur on times scales of just a few million years. The importance of this finding is that it questions the prevailing belief that the short time-scale of these reorganizations documented in the geologic record, requires that they must be caused by a tectonic or near surface process, rather than by mantle convection.

Plate reorganization events are a primary example of the importance of modelling an integrated plate-mantle system with dynamic feedback. The behaviour results from a competition between the influence of internal heating, which acts to shorten the natural convective wavelength, and the plates, which act to instill the wavelength of their geometry through the mantle. Heat built up in the shallow mantle results from the elongation of convection cells through plate motion. Heat is deposited around the system downwellings at convergent plate boundaries until a hot envelope acquires a buoyancy comparable to the downwelling itself. Eventually the pull of subduction and the push back of the buoyant envelope approximately cancel and the plate becomes vulnerable to a change in direction caused by a relatively small push or pull. Once a new downwelling forms at a young convergent plate boundary it can rapidly develop into a strong downwelling capable of overwhelming the pull of the mature downwelling locked in competition with the surrounding heat buildup (Fig. 8).

4 Plate boundary evolution

The studies described in the previous section have provided a large degree of insight into how plate tectonics might affect the mantle's thermal structure and heat loss and how, in turn, the mantle might effect plate velocities. However, they are all limited by a major constraint; they do not feature evolving plate boundaries. In this respect, mantle convection models that neglect modelling plates retain an element of realism that the models with fixed plate geometries don't; the former allow for the mobility of points of surface convergence and divergence (and therefore enable greater mobility of mantle downwellings *and* upwellings).

Plate boundaries move with velocities comparable to the velocities of the plates themselves. Consequently, a complete understanding of the influence of plates on the evolution of the Earth requires developing convection models that feature dynamic plate boundary evolution as well as dynamic plate velocities. In the same way that studies of convection models featuring specified plate velocities preceded dynamic plate models, there have been numerous studies investigating the effect of prescribed plate boundary motion on mantle convection (e.g., Davies, 1986; Ito et al., 1997; Quéré and Forte, 2006). Collectively, these studies have indicated several findings that appear to remain robust in more recent studies featuring dynamic plate evolution models. Specifically, mantle plumes do not appear to follow migrating divergent plate boundaries, even though flow patterns established with fixed plate geometries can drive hot upwellings to the regions underlying divergent plate motion in models with fixed plate boundaries (e.g., Jellinek et al., 2003; Gait et al., 2007). In addition, plate boundary motion affects heat loss and convection pattern stability (Gait and Lowman, 2007; Gait et al., 2008; Stein and Lowman, 2010).

The technical challenges associated with modelling dynamic plate velocities do not increase dramatically in moving from two-dimensional to three-dimensional modelling. However, the difficulties associated with modelling dynamic plate geometry evolution using a force-balance approach escalate considerably in three-dimensions. Two-dimensional models with plates include only divergent and convergent plate boundaries. In addition, they cannot include plate motion that is oblique to a plate boundary. Furthermore, single horizontal dimension stress fields also simplify the issue of specifying criteria for the implementation of plate rifting in two-dimensional calculations (e.g., Gurnis, 1988; Lowman and Jarvis, 1995; Butler and Jarvis, 2004; Gait and Lowman, 2007).

4.1 Two-dimensional models

In two-dimensions, a mantle convection model with dynamic self-consistent evolving plates should at least yield behaviour like subduction zone migration and divergent boundary (ridge) mobility (e.g., Puster et al., 1995). The simplest modelling approach is to specify plate boundary motion (e.g., by requiring the motion of weak zones associated with plate boundaries to move with the velocity of the neighbouring plates). In this case the dynamic evolution is dictated by the modeller based on simple requirements (e.g., symmetric sea-floor spreading). The obvious downside of such an approach is that it suppresses the possibility of obtaining non-specified behaviour (e.g., the possibility of non-symmetric seafloor spreading). However, convection models of this type (Puster et al., 1995; Gait et al., 2007) are especially robust and have the advantage of being able to run for an effectively unlimited number of mantle overturn times, yielding insight into the long term effects of mantle properties on plate dynamics, like phase changes or viscosity structure. For example, by modelling plate boundary motion and dynamic plate velocity in multiple long period mantle convection calculations featuring mobile weak zones, Puster et al. (1995) showed that a 30-fold viscosity increase in the lower mantle produced a record of plate velocity and size distribution that best matched the plate-tectonic record of the past 120 Myr.

Convection characteristics in models with and without mobile plate boundaries have been compared in calculations simulating hundreds of millions of years of evolution (Gait and Lowman, 2007) and indicate that some of the phenomena manifested in models featuring fixed boundaries may not be present once plate boundary motion is enabled. For example, Gait and Lowman (2007) modelled evolving plate velocities with a force balance method and migrated divergent plate boundaries by requiring symmetric seafloor spreading. In addition, the models featured one-sided subduction, implemented by moving convergent plate boundaries with the velocity of the younger plate at the collision zone, and a yield stress that allowed for plate rifting (e.g., Gurnis, 1988). The time-dependent plate boundary locations, plate velocities and number of plates resulted in a highly time-dependent heat flow (Fig. 9) but not the quasi-periodic time-dependence observed in models with fixed plate boundaries (e.g., Lowman et al., 2001; Koglin et al., 2005; Ghias and Jarvis, 2007).

4.2 Three-dimensional models

The three-dimensional modelling of evolving plate boundaries using a force-balance method for determining the plate velocities and a prescribed set of rules for plate boundary motion requires a re-thinking of the approaches used in two-dimensions. The fact that the plate boundary morphology does not evolve self-consistently results in a number of differences between plate tectonics on the Earth and any three-dimensional convection model featuring specified plate boundaries. Nearly all plate boundaries in force-balance models feature a mixture of convergent and strike-slip motion. On the Earth, one example of how plate boundaries evolve to accommodate the associated plate motion is the development of transform faults, offset by spreading ridge segments. However, at present, modelling the dynamic development of such a system in a force-balance model has not been attempted and remains a formidable challenge.

One possible avenue for implementing evolving boundaries in a three-dimensional force-balance method model would be to use our understanding of triple junction motion to define a set of axioms which would govern the evolution of the plate geometry in a way that emulates plate evolution. The problem for such an approach is that triple junction evolution models tell us the junction velocity when the plate boundaries are convergent, divergent or transform but not a mixture of divergent and transform boundaries. Still, the triple junction evolution approach may yet yield a dynamically evolving model that emulates plate tectonics.

Studies utilizing a simple but dynamic triple junction migration model have investigated the fundamental differences that exist in mantle convection simulations when plate boundaries are stationary compared with cases having mobile boundaries (Gait et al., 2008; Stein and Lowman, 2010). The models feature triple junctions that are moved with the area-weighted mean velocity of the three associated plates (Fig 10), somewhat in analogue with the mobile weak zones of Puster et al. (1995). Thus, although the movement is simplified compared with the movement of the Earth's triple junctions, the models do feature the ability to respond to increases in plate velocity (when the triple junctions will move more quickly) and include a time-dependent number of plates in three dimensions. Findings indicate that, as in two-dimensional calculations, the modelling of plate boundary motion increases the time-dependence of the plate-mantle system (Gait et al., 2008) and, in particular, the surface heat flux (Stein and Lowman, 2010). In addition, plate velocities exhibit reorganization events in three-dimensional calculations with evolving plate boundaries as they do in cases with fixed plate boundaries. This finding is fundamentally different from the findings of two-dimensional models where plate boundary evolution extinguished the appearance of mantle flow reversals. However, three-dimensional calculations with evolving plate boundaries agree with another finding from two-dimensional studies, namely, that mobile downwellings appear weaker in models with mobile boundaries and a stiffened lower mantle relative to the upper mantle viscosity. A possible explanation is that, in contrast to upwellings, because downwellings form at plate boundaries their locations (as well as morphology) continually change in response to plate boundary motion. As a result, downwellings become less robust as the plates evolve, sometimes failing to penetrate into the lower mantle. In a calculation with fixed plate boundary locations, entrainment by material descending in the high viscosity lower mantle aids sinking matter in the upper mantle to penetrate the lower mantle (Jarvis and Lowman, 2007). When plate boundaries evolve, cold upper mantle downwellings no longer descend directly over cold lower mantle downwellings so that the lower mantle viscosity increase becomes a greater barrier to sinking upper mantle material.

To date, the most consistent finding of the force-balance models featuring dynamic boundary evolution appears to be time-dependence. Plate velocities and surface heat flux are much more variable in models featuring evolving plates than they are in systems featuring a fixed arrangement of plate boundaries or free-slip surface conditions (Gait et al., 2008; Stein and Lowman, 2010). However, increased time-dependence complicates plate evolution and amplifies the technical challenges of producing realistic dynamic plate boundary evolution. The need to simulate a rapidly evolving surface contributes to thwarting the ultimate goal of the force-balance models (i.e., modelling dynamically evolving plate velocities with Earth-like evolving triple junctions). Consequently, self-consistent plate generation models based on temperature-, stress- and pressure-dependent rheologies may still end up being the fastest path to obtaining mantle convection models featuring continuous plate-like surface motion on time-scales of hundreds of millions of years.

5 Progress on modelling the Earth's mantle with self-consistently generated plate tectonics

The plates are rheologically and thermally distinguishable from the underlying mantle by their rigidity and temperature. Their origin is the cool upper thermal boundary layer of the convecting mantle and their stiffness is explained by the mantle's temperature-dependent rheology. However, it has been known for nearly 40 years that unless some other control on viscosity is present, a temperature-dependent rheology alone cannot produce plate-like behaviour at the surface of a convecting fluid. (Other conditions that augment rheology controlled surface motion and deformation are required to model plates generated by a temperature-dependent viscosity, such as fault modelling (Zhong and Gurnis, 1996; Zhong et al., 1998).) Assuming that convection takes place in a fluid confined between isothermal boundaries, the strength of the temperature-dependence is given by the ratio of the viscosities found at the temperatures of the top and bottom boundaries, $\Delta\eta_T$. Convecting fluids with a temperature-dependent viscosity that varies over the wide range of values appropriate for modelling the rheology of the Earth's mantle ($\Delta\eta_T > 10^4$) exhibit a mode of convection widely referred to as 'stagnant-' or 'rigid-lid convection' where a cool immobile conducting boundary layer forms above a decoupled vigourously convecting underlying fluid.

In general, three different convective regimes are observed with the employment of purely temperature-dependent viscosities (e.g., Solomatov, 1995; Moresi and Solomatov, 1995). In addition to the 'rigid-lid' mode of convection that results when a high viscosity contrast is present, for systems featuring a low viscosity contrast ($\Delta\eta_T < 10^2$) surface mobility occurs and the upper thermal boundary layer participates in the convection at all times (e.g., Ratcliff et al., 1997). For intermediate viscosity contrasts a transitional regime exists that features a mobile but 'sluggishly' moving lid (e.g., Ratcliff et al., 1997).

In a fluid with a temperature-dependent viscosity of high contrast the cool upper boundary layer becomes stiff and 'locks-up'. However, the introduction of lithospheric weak zones mobilizes the upper thermal boundary layer and allows for rudimentary plate-like surface motion (Kopitzke, 1979; Gurnis, 1989). More advanced models exhibiting plate-like behaviour in two-dimensional geometry calculations are obtained by introducing a lithospheric yield stress in models featuring large contrast temperature-dependent viscosities (Moresi and Solomatov, 1998). In these models, when stresses exceed a critical value (determined by a combination of model parameters including the temperature- and pressure-dependence of the rheology, the model's solution domain size and additional factors), the lithosphere is weakened by yielding, simulating faults and shear zones. By specifying different yield stress values stagnant-lid convection can be superseded by episodic or mobile surface convection (Fig. 11), even for systems featuring viscosity contrasts appropriate for modelling the Earth's mantle. Episodic time-dependent behaviour characterised by intermittent system overturns that punctuate upper thermal boundary layer stagnation has also been obtained with the implementation of a yield stress in three-dimensional models (e.g., Stein et al., 2004; Loddoch et al., 2006).

Implementation of a yield stress to obtain plate-like surface motion has been most successful when employing a viscoplastic rheology in which stress reaches a maximum value with strain-rate (e.g., Tackley, 2000a). However, alternative rheological mechanisms for generating weak zones in an otherwise stiff and cold lithosphere have been investigated and include the so called 'stick-slip' and pseudo-stick-slip rheologies. These rheologies feature strain-rate-weakening, where beyond a point, stress decreases with strain-rate (Bercovici, 1993, 1995; Tackley, 1998, 2000b; Ogawa, 2003). Pseudo-stick-slip rheologies have been difficult to implement in evolving mantle convection simulations (Bercovici, 2003), for example, their use can fragment plates and induce the appearance of episodic downwellings (Tackley, 2000b). However, the application of strain-rate-weakening rheology in instantaneous flow models (utilizing temperature/density field snapshots) was an early indicator that the self-generation of plate tectonics in global mantle convection models was a realizable goal. For example, utilising a strain-rate-weakening rheology enabled previously elusive results like the obtaining of self-generated strike-slip boundaries and low-strain-rate plate interiors in three-dimensional models (Fig. 12).

Early successful implementation of pseudoplastic yielding to achieve plate-like surface motion in evolving three-dimensional Cartesian geometry models was achieved by Trompert and Hansen (1998) and Tackley (2000a). These authors showed that the mobile-lid, episodic-lid and stagnant-lid regimes exist in three-dimensional systems and found that some combination of temperature- and stress-dependent rheologies can give plate-like motion over parts of the surface of a convecting layer even when the rest of the thermal boundary layer remains chiefly immobile. Moreover, their three-dimensional models include other requirements of a realistic model of plate tectonics. For example, yielding was found to be sufficient to generate narrow plate boundaries if the convecting plate-mantle system is allowed to evolve to its natural configuration.

5.1 More recent progress

Although periods exhibiting self-consistently generated plate-like surface motion was achieved in three-dimensional numerical mantle convection models a decade ago, uninterrupted mobility of the entire upper thermal boundary layer in the parameter range appropriate for modelling the Earth has yet to be modelled. However, an understanding of the important aspects of terrestrial convection that seem to contribute to the formation of plate motion is gradually accumulating. It now appears that one of the most important controls for obtaining surface mobility with plate-like characteristics is the rheology of the mantle below the plates.

The combination of a plastic yield stress with an 'asthenospheric', or low-viscosity zone (LVZ) acts (Fig. 13) to allow localized weakening of the cold thermal boundary layer that enables the formation of focussed regions of deformation (Tackley, 2000a; Richards et al., 2001). While the role of an LVZ in generating plate tectonics is not clear, it has been speculated (Chen and King, 1998; Richards et al., 2001) that the reduced tractions on the base of the lithosphere may help confine deformation to established zones of failure. A single three-dimensional spherical geometry model from the Richards et al. (2001) study indicated that LVZs may be critical in spherical shell geometries, not just two-dimensional Cartesian cases.

Systematically adding the effects of temperature-, stress- and pressure- dependence to three-dimensional Cartesian convection models, Stein et al. (2004) showed that the stratification of mantle viscosity may also be critical for producing plate tectonics. The addition of pressure-dependent rheology results in a separation of the mobile-lid regime into two regions. For low yield stresses the first of these regions is akin to the mobile-lid regime, however, for models featuring higher yield stresses (where the episodic regime prevails in the absence of any pressure-dependence) the introduction of pressure-dependence results in a transition from episodic to plate-like behaviour (Stein et al., 2004). Particularly important is that the plate-like behaviour prevails for substantial periods. The pressure-dependence required to obtain the plate regime behaviour increases with system yield stress. In addition, a critical pressure-dependence exists, above which the episodic regime vanishes between the plate regime and stagnant-lid regime (Fig. 14).

The role of pressure-dependent viscosity in the formation of plates is explained by its influence on the downwelling convection currents. Pressure increases with depth and so once pressure-dependence is present, downwellings that would have quickly descended into the deep mantle (peeling away part of the upper thermal boundary layer in the process) encounter increasing resistance that slows their descent (Stein et al., 2004). Consequently, surface velocities slow so that episodicity is damped and continuous periods of plate-like surface motion emerge.

One cause for concern that applies to many of the studies described in this section is that (except where noted) the findings were obtained in Cartesian geometries. Cartesian systems, most notably those featuring internal heating, are unrealistically hot in comparison to spherical shell convection models featuring the Earth's core to surface radii ratio (O'Farrell and Lowman, 2010). Given that calculations featuring temperature-dependent rheologies are especially sensitive to the heating mode, future calculations need to focus on the issue of realistic mantle geotherms as well as the generation of plates.

Encouragingly, recent research (van Heck and Tackley, 2008; Foley and Becker, 2009) has shown that the same surface-mobility/convection regimes observed in Cartesian systems appear to be manifest in spherical models. In addition, it has been shown that the yield stress incorporated into the rheology of the calculations can effect the wavelength of the plate geometry as well as the convection planform (Fig. 15) similarly in Cartesian ($8 \times 8 \times 1$) geometry and spherical geometry experiments.

5.2 Approaching the Earth's parameter range

The challenge ahead for self-consistent rheological plate-modelling methods is to elevate the parameters of the studies to the ranges appropriate for the Earth. The Earth's Bénard-Rayleigh number is approximately 10,000 times the critical value (e.g., Jarvis et al., 1989), the effective Rayleigh numbers of current studies are still more than an order of magnitude less than the desired values. Given the transitions in plate motion behaviour discovered in some of the high Rayleigh number studies utilizing force-balance plate formulations, there is the possibility that the rheology based modelling methods that have led to the successes of the past half dozen years may still require significant refinement to obtain plate tectonics with Earth-like convective vigour.

Changing the parameters of numerical models can cause transitions in behaviour due to the vagaries of the numerical algorithms or grids, as well as real physical behaviour. To distill the physical transitions from the numerical calculations requires benchmarking. The difficulties associated with exploring more vigourous convection and longer periods (i.e., over many mantle transit times) of evolution using self-consistent plate generation methods can potentially be eased by comparing the results with the findings from studies that use force-balance calculations or some other robust modelling method. (For example, force-balance models identified the Rayleigh number dependence of cyclic plate reversal behaviour (Lowman et al., 2001, 2003) in calculations featuring fixed plate boundaries before it was also observed in rheology based models (Koglin et al., 2005).) Despite their shortcomings, force-balance based calculations featuring high Rayleigh number convection with simple evolving plate boundaries (e.g., Gait et al., 2008; Stein and Lowman, 2010) will remain a useful tool for categorizing long term behaviour and identifying potential transitional regimes.

6 Other challenges

A dozen years after various members of the mantle convection modelling community began to focus on the issue of how to generate plates in three-dimensional convection models, reasonably robust self-consistently generated plates have been obtained in both Cartesian and spherical convection calculations. The focus in the coming decade can now shift to the refinement of these models. In addition to modelling convection with Earth-like vigour, plate-like surface motion will be required for periods in excess of a billion years of model time. Moreover, plate boundaries with terrestrial characteristics have to replace the distinctly un-Earth-like boundaries that persist in present models. Principal amongst the outstanding problems is the unresolved issue of one-sided subduction. Can it be obtained and, of equal importance, can it be sustained in a self-consistent model? Early successes modelling faults in two-dimensional models (Zhong and Gurnis, 1995) indicate that obtaining realistic plate behaviour, such as one-sided subduction with trench rollback, can be obtained by augmenting the conservation equations of fluid dynamic convection model formulations with elements taken from the physics of solids featuring brittle failure. Hybrid approaches may yield even more promising future results but are considerably more complex to model in three dimensions. Nevertheless, the inclusion of realistic tectonic forces and associated continental as well as oceanic plate boundary forces will be required in future studies. Without these elements fundamental questions cannot be answered. For example, what role does the mantle play, if any, in driving the time-dependence of plate velocities?

Of further importance is the ultimate goal of reproducing not just the primary characteristics of plate tectonics: plate-like motion; strike-slip plate boundaries (in addition to convergent and divergent boundaries); a strong toroidal flow component; plate growth and consumption and one-sided subduction, but also some of the secondary features. For example, Earth-like spreading boundaries consist of ridge sections offset by transform faults (Wilson, 1965), rather than a mixed divergent/strike-slip boundary. Moreover, stable triple junctions follow well understood trajectories determined by the characteristics of the boundaries from which they are comprised (e.g., convergent boundary). A realistic mantle convection model featuring self-consistently generated plates will need to reproduce Earth-like triple junctions and emulate their motion on a sphere. Similarly, the models will need to produce spatial heat flow and dynamic topography in agreement with terrestrial data.

Finally, a fully self-consistent mantle convection model will need to exhibit all of the Earth's surface dynamics on all time scales. Primary among these requirements will be the manifestation of continental drift and supercontinent assembly and breakup, thus requiring the modelling of distinct oceanic and continental lithosphere. Indeed, obtaining self-consistently generated oceanic and continental plates from a mantle model is an issue that must be addressed in order to answer some of the most fundamental and rewarding questions: How did the continents form? How different from mantle convection with just oceanic plates is mantle convection with distinct continental as well as oceanic plates? For example, do continents really insulate the mantle? Is cyclic supercontinent assembly inevitable and, if so, why does it happen?

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8 Appendix: The characteristics of convection in a fluid with mantle-like properties

If the density gradients that drive mantle convection could be instantaneously eliminated, motion in the mantle would come to cease in less than a millionth of a second. Thus, unlike the oceans, atmosphere or even the outer core, the mantle is an essentially momentum free fluid. In the non-dimensionalization of the flow equation for a rotating high viscosity fluid like a planetary mantle, we find that the terms representing inertia and rotation are multiplied by the inverse of the Prandtl number, $Pr = \nu/\kappa$, where ν is the kinematic viscosity and κ is the thermal diffusivity. The mantle's Prandtl number is of order 10^{23} making it effectively infinite in a scale-analysis of the hydrodynamic equations governing mantle convection. An infinite Prandtl number fluid is therefore a momentum free fluid.

Convective vigour is determined by the system Rayleigh number (e.g., Chandrasekhar, 1961). Multiple definitions of the Rayleigh number exist, depending on the variability of the system parameters and the mode of system heating. For example, heating by an isothermal bottom boundary, the presence of internal heat sources or a constant basal heat flux require different Rayleigh number definitions. However, all of these definitions measure convective vigour in terms of the ratio of the time scales for heat transfer across the system by buoyancy driven advection versus heat transfer by thermal diffusion. As Rayleigh number is increased, the dominance of heat transfer by advection across the layer grows, relative to the heat transfer by thermal diffusion.

Mantle convection studies generally adopt an isothermal surface for the top of the mantle but vary in their specification of the mantle's heat source(s). The assumption of an isothermal bottom boundary (i.e., the top of the outer core, assuming we are modelling the mantle) results in the definition of the Bénard-Rayleigh number, given by $Ra_B = g\alpha\Delta T d^3/\kappa\nu$ where g is gravitational acceleration; α is the coefficient of thermal expansion; ΔT is the superadiabatic difference in temperature between the top and bottom isothermal boundaries of the system and d is the depth of the fluid. The mantle's Bénard-Rayleigh number is estimated to be of the order of ten thousand times the critical value (i.e., the value at which the onset of convection occurs).

With the onset of advection, a thermal boundary layer develops at the top of a convecting fluid and the temperature gradient at the surface changes relative to the heat lost in a purely conducting layer. The mean heat flux obtained from a convecting fluid, normalized by the heat flux that would conduct across the system in the absence of convective motion (i.e., $k\Delta T/d$, in the case of Bénard convection, where k is thermal conductivity), is given by the Nusselt number, Nu.

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FIGURE CAPTIONS:

Figure 1: A demonstration of the ability of imposed plate-like surface motion to organize convection planform. Figures a), c) and e) and b), d) and f) show snapshots of the temperature field in the upper mantle (300 km depth) and mid mantle (1500 km depth), respectively, from three models featuring different surface conditions. Red is hot and blue cold. In all cases the core-mantle boundary is free-slip and insulating and the internal heating Rayleigh number is 4.0×10^7. The core radius and mantle thickness are scaled to the dimensions of the Earth's mantle. The lower mantle viscosity is 20 times greater than the upper mantle viscosity. The calculation depicted in a) and b) shows results for a model with a free-slip surface. Cold linear downwellings from the upper boundary layer broaden in the higher viscosity lower mantle. The calculation shown in c) and d) includes the effect of imposing the Earth's present-day plate geometry and velocities and integrating the free-slip surface solution for a further 10 transit times. The RMS value of the plate velocities has been scaled down to match the RMS of the calculation shown in a) and b) because of the reduced convective vigour of these calculations when compared to the real Earth. In e) and f) the plates have been 'strengthened' by specifying a factor of 20 linear increase in viscosity from a depth of 300km to the surface. The main effect of the imposed plate motions is to produce a 'plate scale' flow with the major upwellings and downwellings localized at present-day mid-ocean ridge and subduction zone locations. The stiffened lithosphere in the final calculation suppresses small-scale instabilities away from

plate boundaries. From Bunge and Richards (1996). Reproduced by permission of American Geophysical Union.

Figure 2: Temporally averaged mean absolute horizontal velocity plotted against height from two-dimensional aspect ratio four convection experiments featuring a pair of congruent model plates moving in opposite directions at the same speed. Plates are modelled by specifying the surface velocity as a boundary condition. The numerically labelled curves shown correspond to the nondimensionalized surface velocities (expressed in mantle transits per diffusion time) from six cases that span a range of results in which the lowest surface velocities resist the buoyancy driven flow and the highest velocities drive the flow faster than the buoyancy. Also shown are the temporally averaged velocities in cases featuring a free-slip surface and rigid surface as well as the (temporal-mean) velocity from a model where the surface velocity is determined dynamically (so that at all times the force-balance calculated surface velocity is dictated by the buoyancy driven flow). The models feature a factor of 36 increase in viscosity that begins at the 670km depth indicated. The inset shows the detail of the velocity profiles in the upper-most part of the model. Horizontal arrows above the inset are attached to markers indicating the upper (arrow pointing to the left) and lower (arrow pointing to the right) limits of the standard deviation of the mean surface velocity times series (from the dynamic plate and free-slip surface calculations). The temporally averaged, spatially integrated, shear stress is zero at the surface in the three cases corresponding to: dynamic plates, a free-slip surface and fixed plate velocities of approximately +/-1350. The mean surface velocities of

these cases all differ. One implication of the finding is that because the presence of the plates influences and organizes the mantle planform, the mean velocity from a free-slip surface calculation is greater than the mean velocity of a model featuring dynamically determined plate velocities (even though the plates do not resist the buoyancy driven flow). From Nettelfield and Lowman (2007).

Figure 3: Residual temperature field snapshots showing the formation of robust downwelling sheets (blue) and shallow passive upwellings (yellow) in a spherical convection model featuring plate-like surface motion. Also evident (in yellow) are plume-like structures rising from the core-mantle boundary (red). Plate motion is achieved by specifying weak zone plate boundaries (green lines) at the locations of the Earth's present-day plate boundaries. However, because the plate velocities are dynamically determined, the downwelling location sites that develop in the calculation do not necessarily correspond to sites of present-day subduction. The calculation features nondimensionalized viscosities of 30, 0.1, 0.0333, and 2.0 in the top 120km, weak zones, the upper mantle, and the lower mantle, respectively. The Rayleigh number defined by the viscosity in the lower mantle is 4×10^5 and approximately 60% of the surface heat flux is derived from uniformly distributed internal sources. From Zhong et al. (2000). Reproduced by permission of American Geophysical Union.

Figure 4: Plate velocity and surface topography comparisons from three two-dimensional box mantle convection models featuring identical Rayleigh numbers and geometries but different plate modelling methods. The solid curves are from a model utilising a

force-balance method, the short dashed curves correspond to a model featuring high viscosity plates separated by weak zones and the long dashed curve is from a calculation that models plates by implementing a power-law rheology. From King et al. (1992).

Figure 5: a) Demonstration of the insulating ability of large oceanic plates from a study of spherical shell convection featuring force-balanced modelled dynamic plates. As convective vigour increases (higher Ra) calculations featuring larger plates retain an increasing proportion of heat, in comparison to the surface heat flux emitted (measured by the Nusselt number) in cases with free-slip surfaces. The results shown were obtained from calculations featuring a factor of 30 increase in viscosity at a depth of 650 km and no internal heating. After Monnereau and Quéré (2001).

b) Superadiabatic temperature profiles from three spherical geometry mantle convection models. 'case 1' is from a model with a free-slip surface, 'case 2' is a model with four large plates and 'case 3' is a model featuring the Earth's present-day plate geometry; however, because the plate velocities in these calculations are determined dynamically, convergent, divergent and transform boundary locations in the model do not necessarily correspond to boundaries of the same type on the Earth (similar to Fig. 3). Large plates increase the mantle temperature. The models include a factor of 30 increase in viscosity in the lower mantle and heating by an isothermal core and internal sources. After Monnereau and Quéré (2001).

Figure 6: Mean surface heat flux from a suite of isoviscous bottom heated convection models featuring two plates in an aspect ratio two box with wrap around side boundary conditions. The plates are

modelled using a force-balance method. The horizontal axis shows the length of ‘plate 1’ relative to the box length. A second plate covers the rest of the solution domain. Minimum surface heat flow occurs if either plate has an effective length of zero, in which case the surface boundary condition converges to a rigid surface case (similar to Fig. 5a). Maximum heat flow occurs when the two plates are equal in size. In this case the mean heat flux is closest to the heat flux obtained with a free-slip surface. The result indicates that plates have only a minor effect on heat flow as long as the plate length scale is similar to the dimension of the convection cells driven by the thermal forcing. Adapted from Gable (1989).

Figure 7: Plate velocity time series from a $3 \times 3 \times 1$ rectangular geometry mantle convection model featuring four polygonal plates with dynamic velocities determined by a force-balance method. Velocity directions are illustrated by the arrows (see co-ordinate system at upper left, the z co-ordinate, implied but not shown, is anti-parallel to gravity). Feedback between the plate organized flow and convection driven by internal heating and high Bénard-Rayleigh number (5×10^7 based on the upper mantle viscosity) results in especially time-dependent plate velocities and intermittent changes in plate direction. For example, between time t_a and t_c plates 1, 2 and 4 rapidly change directions; however, only the magnitude of plate 3 is affected. From King et al. (2002).

Figure 8: Sequential temperature field snapshots from a cylindrical geometry mantle convection model featuring mobile plates with dynamically determined velocities. The nondimensional surface temperature is 0.0 and the bottom temperature is 1.0. Eight colours with

transitions at contour intervals of 0.125 indicate temperature field variation. A single rigid plate spans the surface and has a thickness of 5% of the system depth. Arrows indicate the plate velocity direction. The convective vigour and internal heating rate are essentially Earth-like but the model is isoviscous below the plates. Plate motion entrains the heat from internal sources into a hot envelope around a cold downwelling (a), the associated buoyancy of the envelope eventually exceeds the pull of the downwelling and the plate reverses direction (b). Subsequently the process repeats once a mature downwelling has formed at the right-side of the solution domain (c) and the plate reverses again (d). After Ghias and Jarvis (2007). Reproduced by permission of American Geophysical Union.

Figure 9: Observables from three, two-dimensional, aspect ratio 12 numerical mantle convection calculations. Shown from top to bottom are the time series of: the number of plates present, the mean surface velocity, the mean surface heat flux and the mean basal heat flux. The dashed and dotted curves correspond to distinct calculations that each feature four plates with dynamically determined velocities but plate boundaries that cannot evolve. (The plate widths are indicated by the model labels, e.g., TD4422 has two plates of widths 4.0 and two plates of width 2.0 spanning the aspect ratio 12 calculation). The solid curves are from a calculation featuring plates that evolve in size as well as number. This model includes dynamically determined plate velocities and plate boundary motion (with one-sided subduction and symmetric sea-floor spreading) and shows the greatest variations between maxima and minima as well as the steepest temporal gradients. (The initial condition for the model

with the evolving plate boundaries featured two plates of widths 4.0 and two plates of width 2.0.) In all three cases the calculations include internal and basal heating and a depth-dependent viscosity. The Bénard-Rayleigh number, based on the upper mantle viscosity, is 10^8. The time series cover almost 1.5 Gyr of evolution. From Gait and Lowman (2007).

Figure 10: Surface-view, plate geometry map snapshots from a three-dimensional Cartesian geometry mantle convection model featuring plate boundaries that move due to triple junction motion. Each colour represents a distinct plate. The horizontal dimensions of the calculation are $6d \times 6d$, where d is the mantle depth, and the model is periodic in the horizontal directions. Plates 1a and 1b form when plate 1 breaks (cyan), similarly 2a and 2b form from plate 2 (red). Snapshots are output at integer multiples of the mantle transit time, t. From Stein and Lowman (2010). Reproduced by permission of American Geophysical Union.

Figure 11: Surface heat flux (expressed as the Nusselt number) and RMS surface velocity from three mantle convection calculations featuring temperature-dependent viscosity and the utilization of Byerlee's law to limit the maximum stress in the lithosphere. When the yield stress is high, convection is confined below a thick, stagnant lithosphere (i.e., time series labelled 'stagnant'). At low yield stress, brittle deformation mobilizes the lithosphere, which participates in the overall circulation (labelled 'mobile'). At intermediate levels of the yield stress, there is a cycling between these two states: thick lithosphere episodically mobilizes and collapses into the interior before reforming (yielding the dashed curves labelled 'episodic'). The

dotted horizontal lines show the average values from the calculation exhibiting episodic surface mobility. From Moresi and Solomatov (1998).

Figure 12: Self-generated plates formed with strain-rate weakening and a temperature-dependent rheology in a solution domain with reflecting side-walls. Non-dimensional temperatures range from 0.0 (top) to 1.0 (bottom). a) shows temperature, 0.3 (blue), 0.7 (red); b) lithospheric viscosity (surface colour map) and velocity vectors (in white); and c) isosurfaces of horizontal divergence/convergence (+/- 50, light and dark purple) and vorticity (+/- 50, green and blue). In b) viscosity ranges over five orders of magnitude from 0.1 (magenta) to 10^4 (scarlet). With the rheology conditions implemented, a pair of plates have formed (scarlet sections characterised by high viscosity) with almost uniform velocities and no interior weaknesses. The diagonal plate boundary at the lower left (magenta) is characterized by a strong strike-slip component as the larger plate moves past the almost stationary smaller plate (left corner). From Tackley (1998).

Figure 13: Surface mobility character as a function of lithospheric yield stress and 'asthenospheric' viscosity reduction. Observations are from a two-dimensional, aspect ratio 4, convection model with Earth-like vigour heated entirely by internal sources. Where the mantle geotherm exceeds a temperature that represents the temperature of the lower extent of the lithosphere, viscosity (in the asthenospheric region extending to a depth of 500 km) is reduced by the value η^*, creating a low-viscosity zone (LVZ). Three types of surface behaviour are obtained for the different yield stress/LVZ combinations. Blue circles indicate cases featuring rigid-lid convec-

tion, red circles correspond to mobile lid convection with variable surface velocity and continuous surface deformation. Green bars indicate plate-like surface motion with broad regions of constant velocity separated by highly localized zones of intense strain. Combined symbols indicate mixed behaviour. For example, red and blue circles indicate stagnant-lid convection punctuated by episodic mobility (see Fig. 11). From Richards et al. (2001). Reproduced by permission of American Geophysical Union.

Figure 14: a) Domain diagram for a three-dimensional Cartesian geometry model study investigating convection with temperature-, stress-, and pressure-dependent viscosity. Pressure dependence increases exponentially with depth. The values given on the horizontal axis indicate the total increase in viscosity due to pressure-dependence. Four distinct convective regimes are observed as described in the text. From Stein et al. (2004).

b) An example temperature field from a statistically steady calculation featuring broad regions with plate-like surface motion (shaded grey areas). Plate-like regions are defined as having almost uniform velocity and small internal deformation. From Stein et al. (2004).

Figure 15: Viscosity at 3% of the mantle depth (a and b) in a spherical shell convection model featuring temperature-, stress-, and depth-dependent rheology. Also shown is the temperature (c and d) where it is 17% cooler than the average temperature at a given depth (i.e., residual temperature). The full sphere is shown using diametrically opposed viewing positions. Narrow blue zones correspond to naturally forming weak zones associated with focussed regions of deformation in an otherwise stiff lid. Yellow and orange regions

(plate interiors) are associated with rigidity. The number of plates and their mobility is determined by a specified yield stress, above which viscosity rapidly drops. The single case shown is for an intermediate yield stress value. At low yield stresses (17 MPa) a single 'great-circle' downwelling is formed. At intermediate yield stresses, multiple oceanic plates form and spreading centre and subduction zone locations are continually created and destroyed. At higher yield stresses ($\sim$ 240 MPa) two hemispherical plates form and there is a single elongated upwelling and downwelling. At still higher yield stresses (420 MPa) the stagnant-lid regime appears. After van Heck and Tackley (2008). Reproduced by permission of American Geophysical Union.

Figure 1
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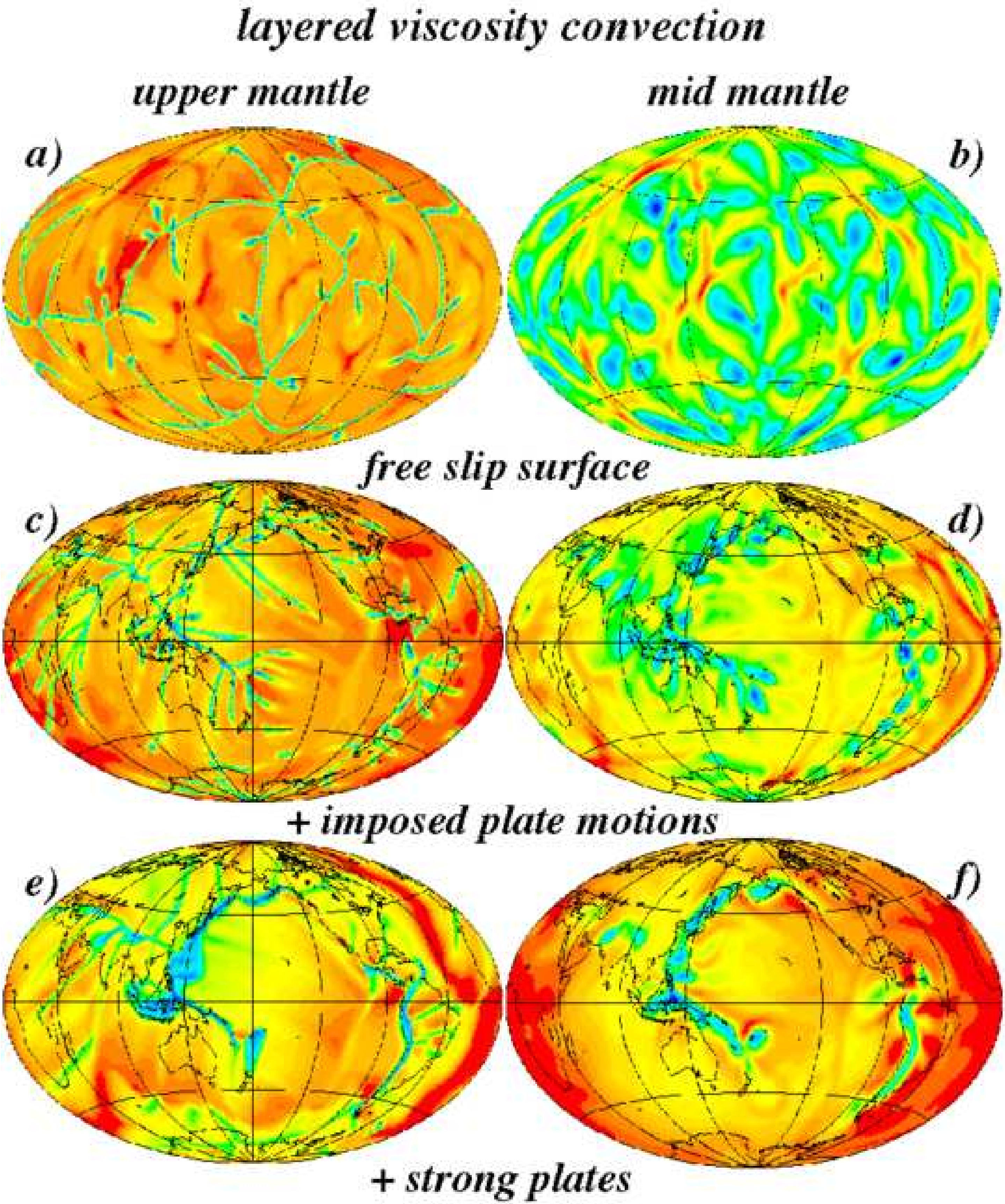
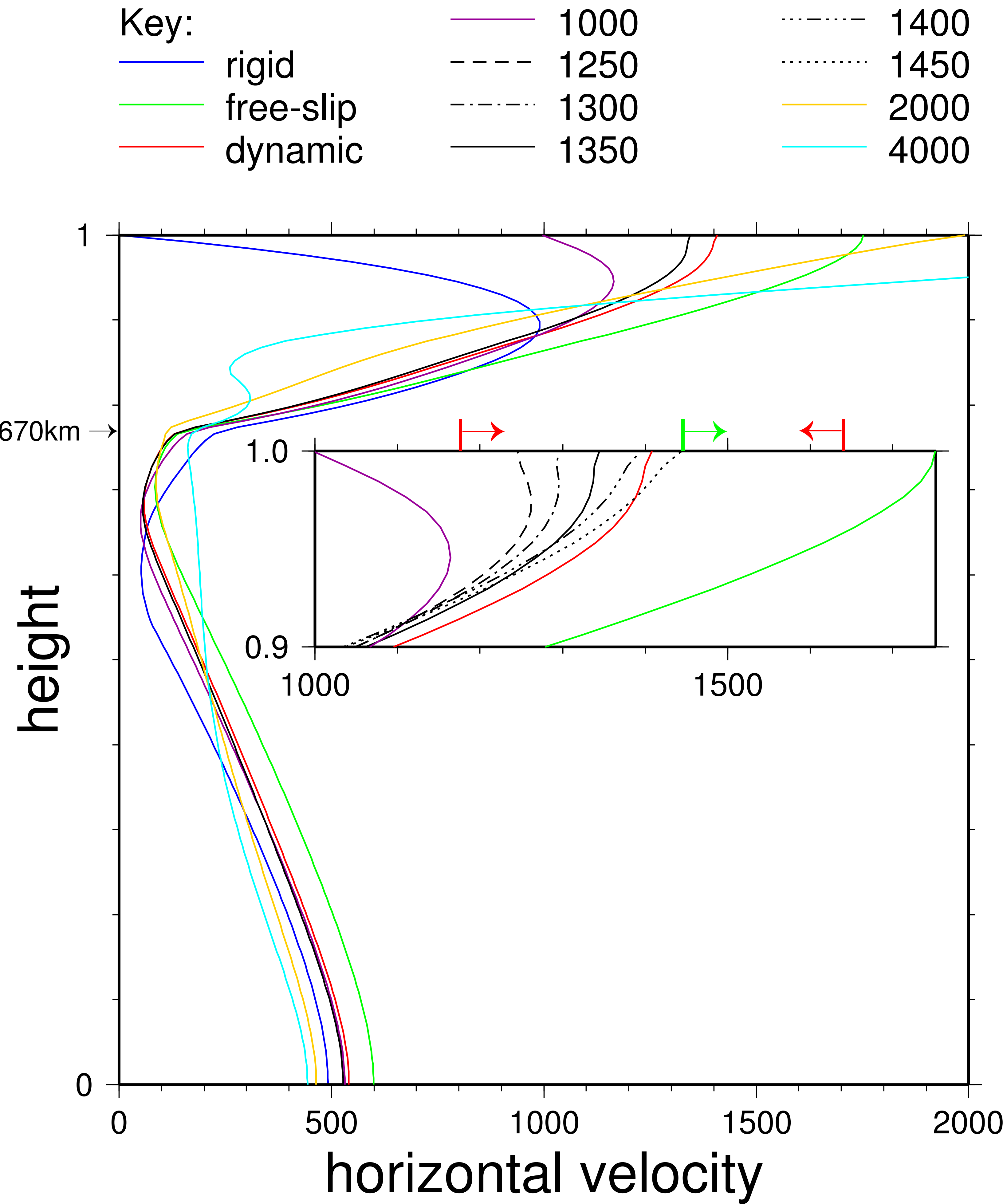


Figure 2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rigid
free-slip
dynamic
1000
1250
1300
1350
1400
1450
2000
4000
1
0
670km
1.0
0.9
1000
1500
0
500
1000
1500
2000
height
horizontal velocity

Figure 3

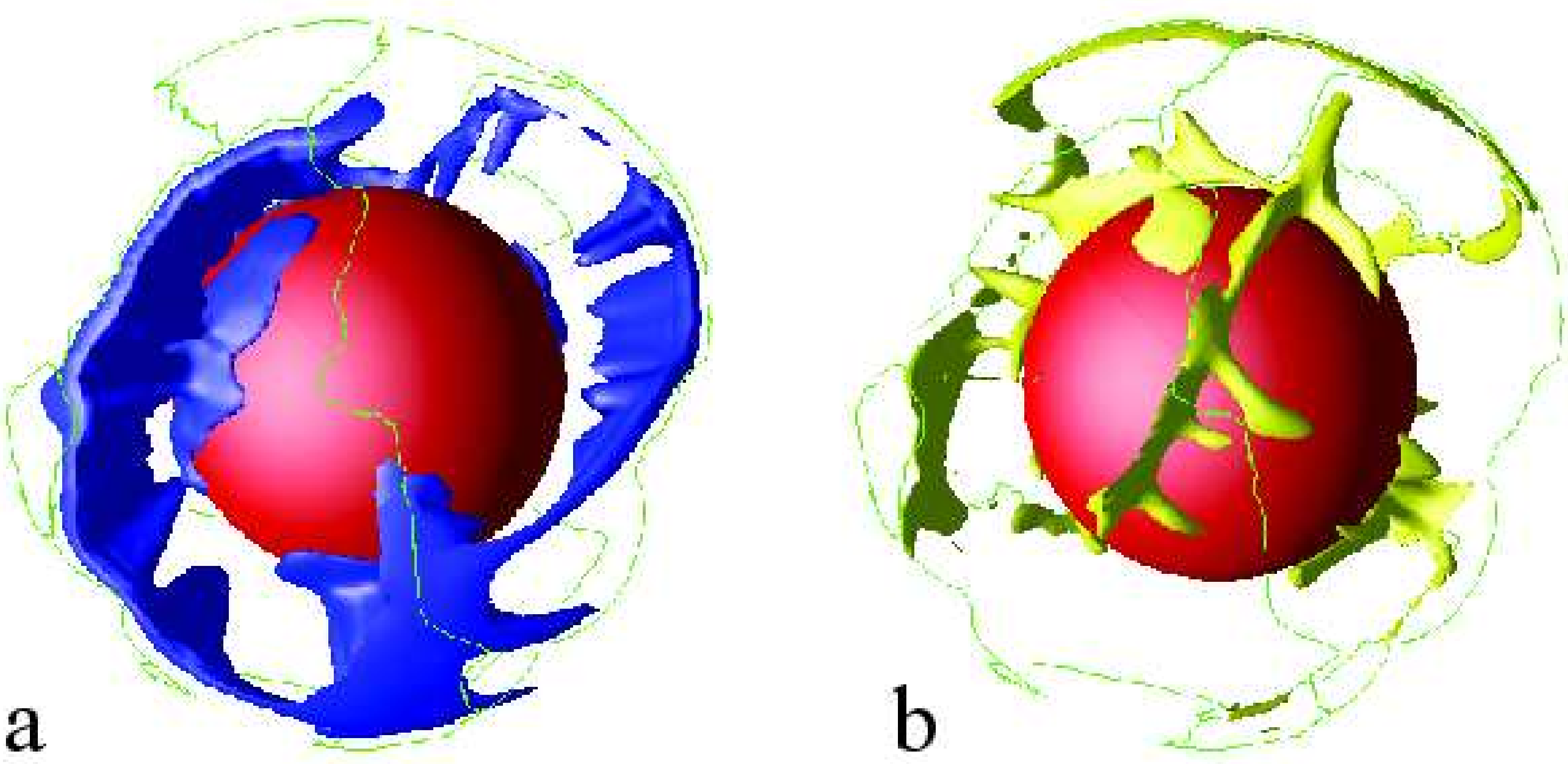

Figure 4
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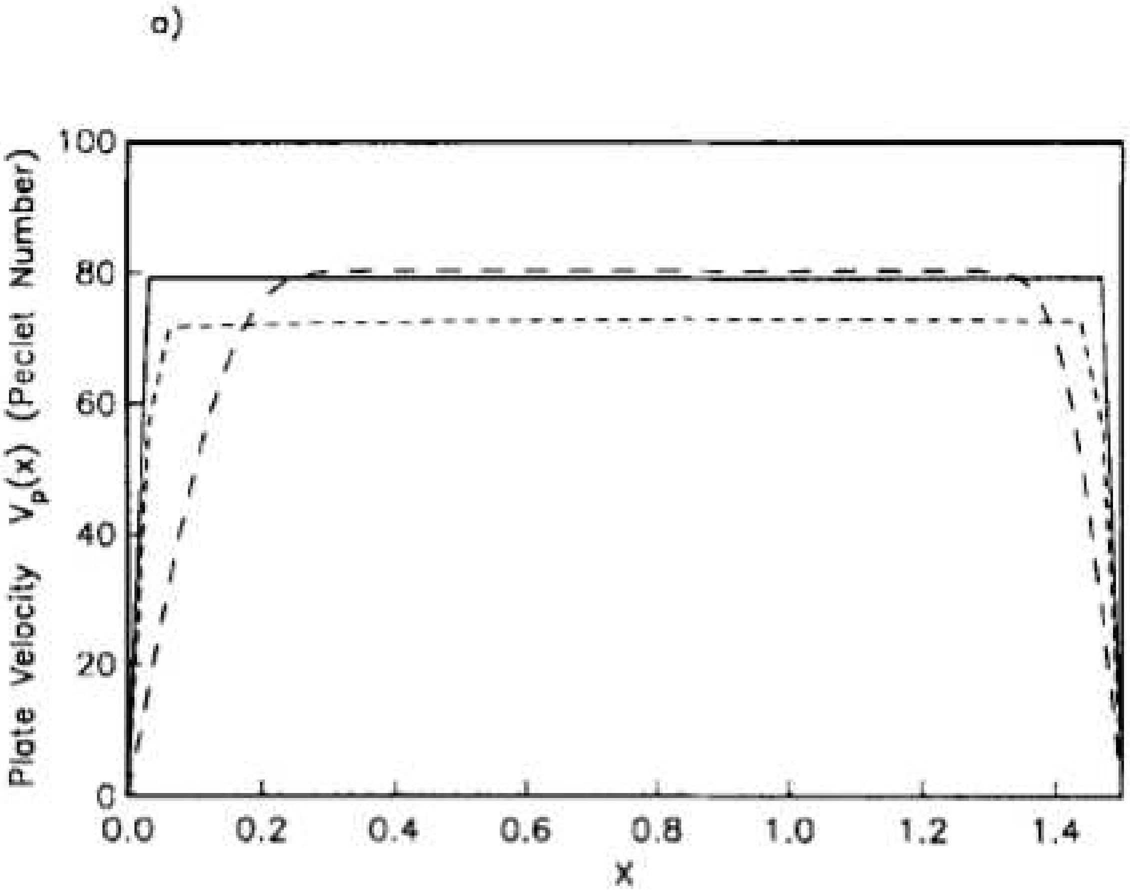


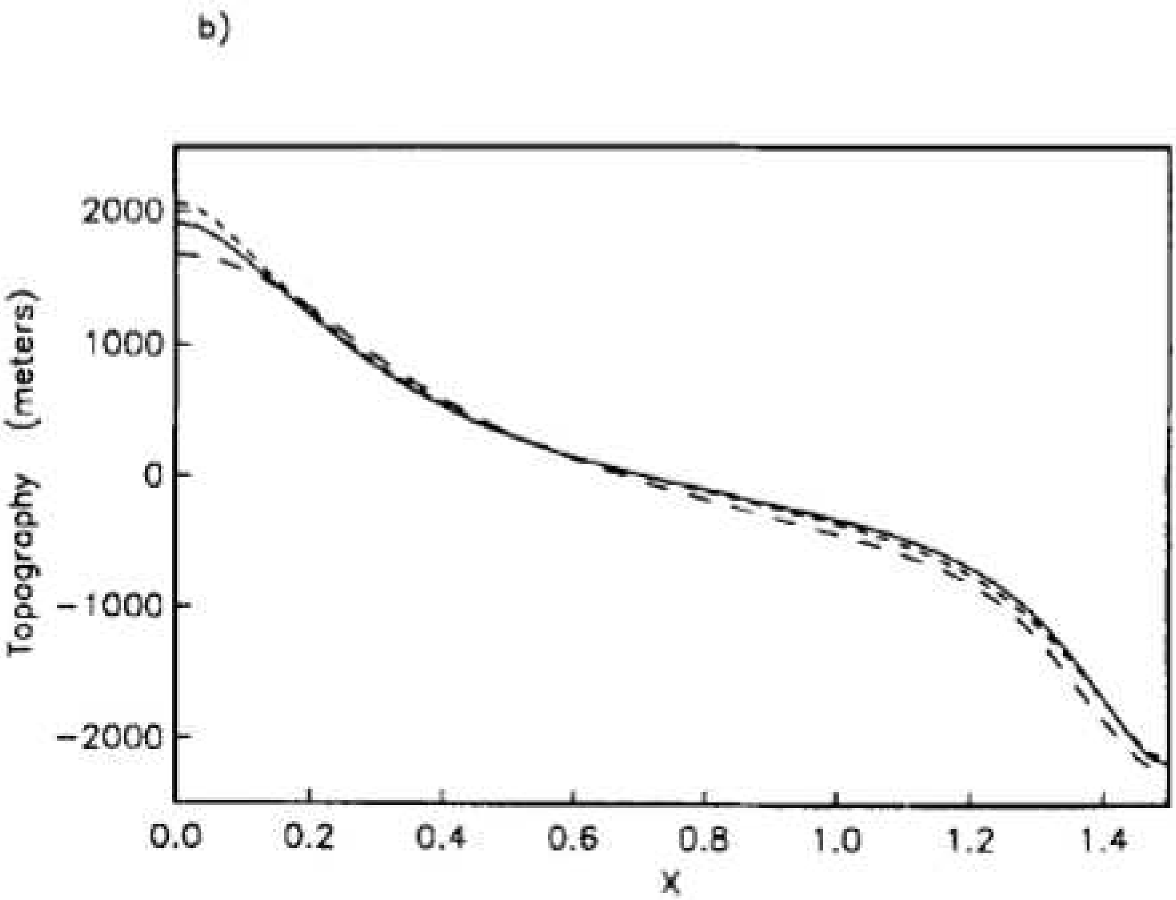
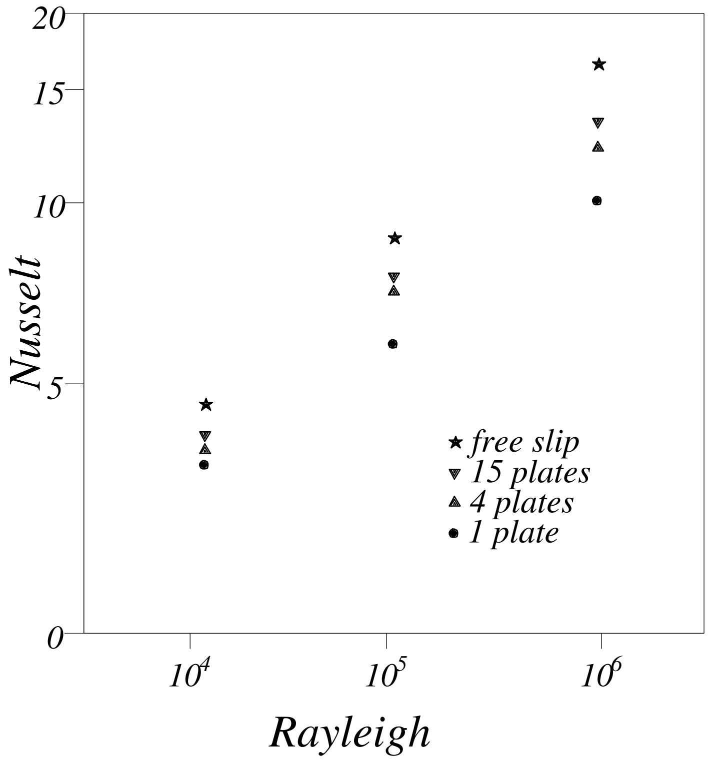


Figure 5a



20
15
10
5
0
Nusselt
10^4
10^5
10^6
Rayleigh
free slip
15 plates
4 plates
1 plate

Figure 5b

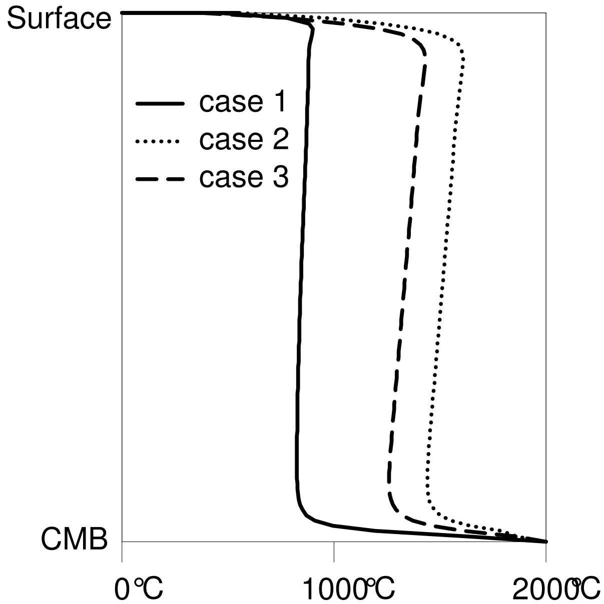
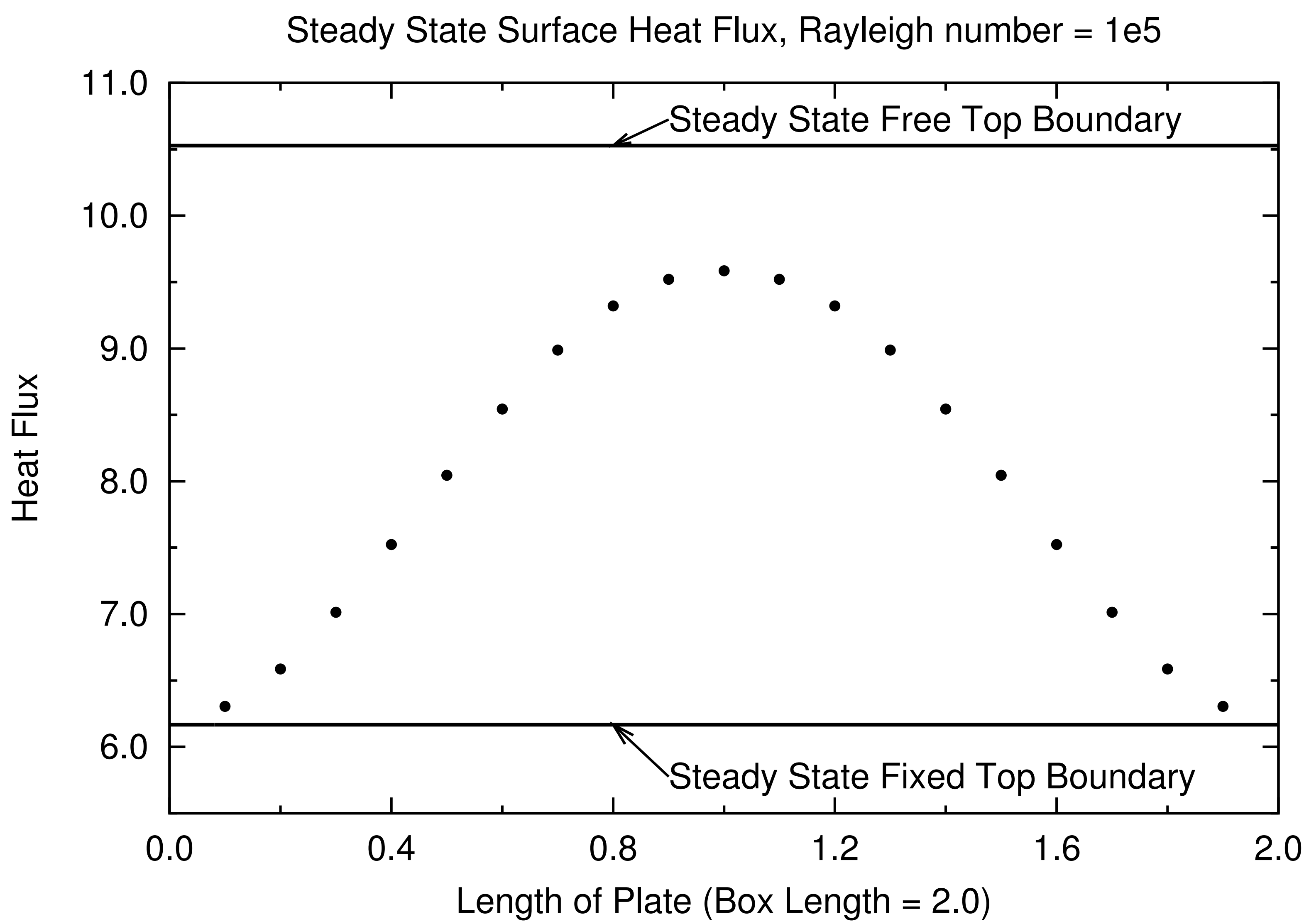
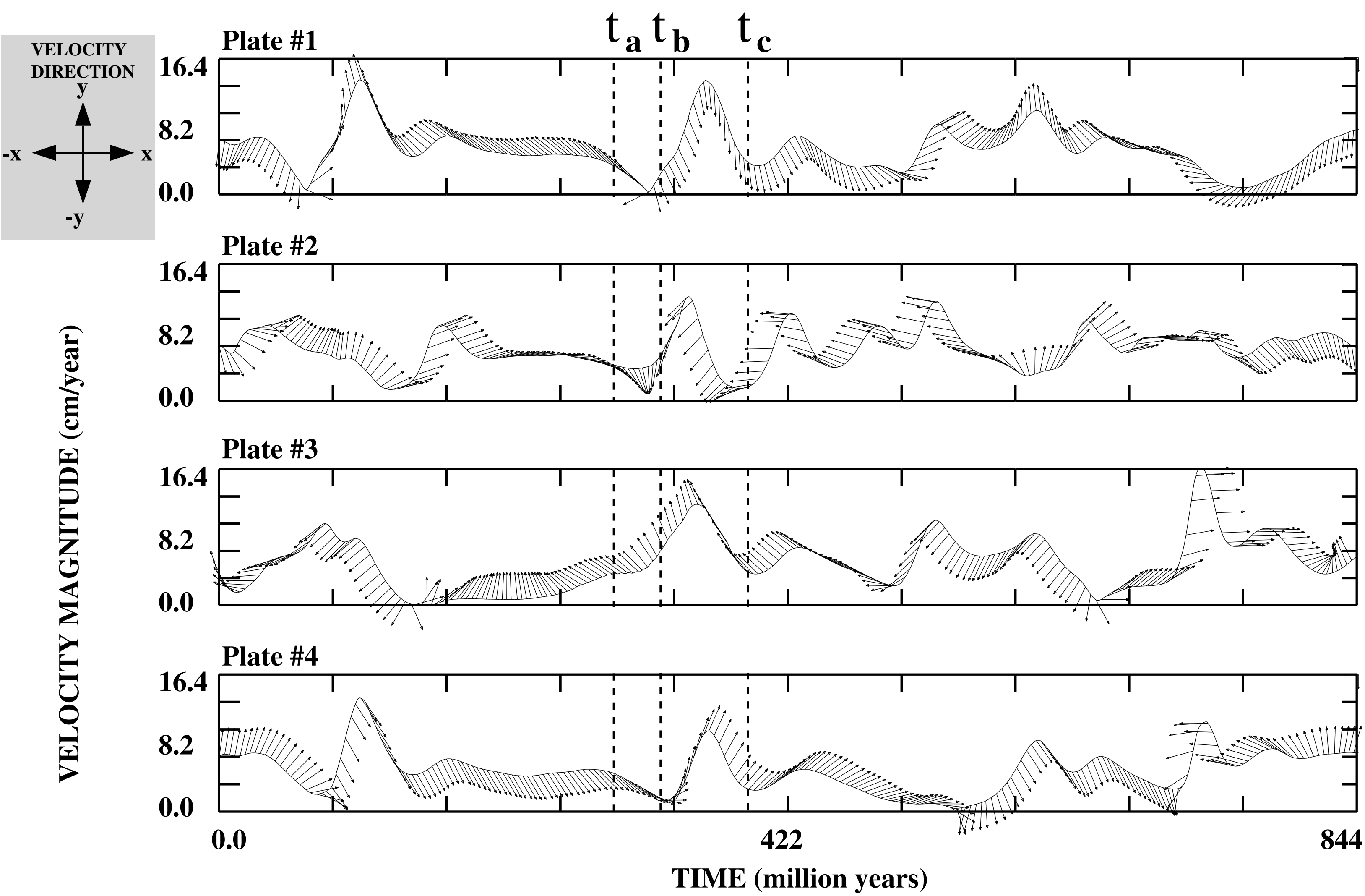


Figure 6



Steady State Surface Heat Flux, Rayleigh number = 1e5
Steady State Free Top Boundary
Steady State Fixed Top Boundary
Heat Flux
11.0
10.0
9.0
8.0
7.0
6.0
0.0
0.4
0.8
1.2
1.6
2.0
Length of Plate (Box Length = 2.0)

Figure 7



VELOCITY DIRECTION
y
-x
x
-y
t_a t_b t_c
Plate #1
Plate #2
Plate #3
Plate #4
16.4
8.2
0.0
VELOCITY MAGNITUDE (cm/year)
0.0
422
844
TIME (million years)

Figure 8

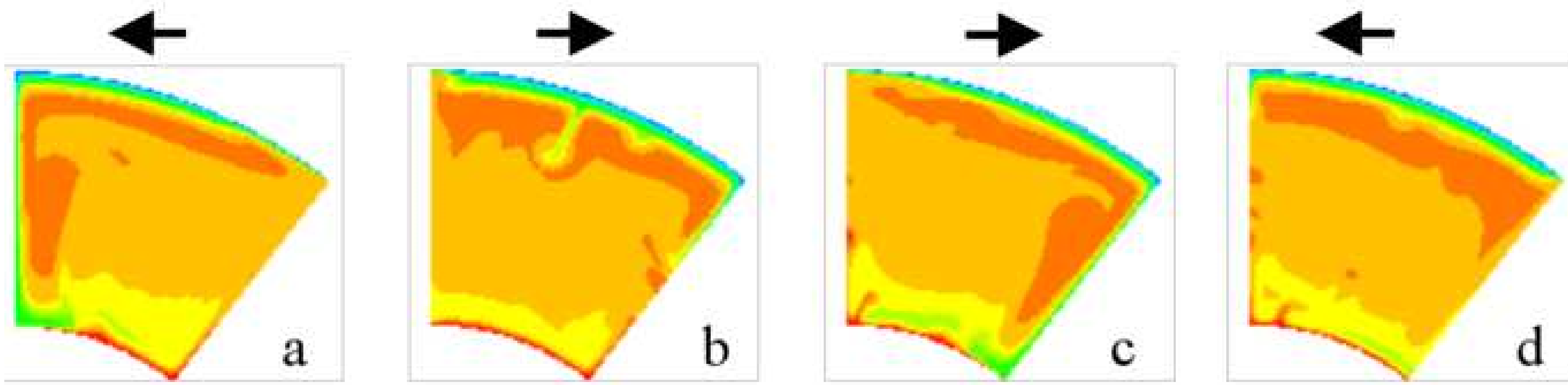
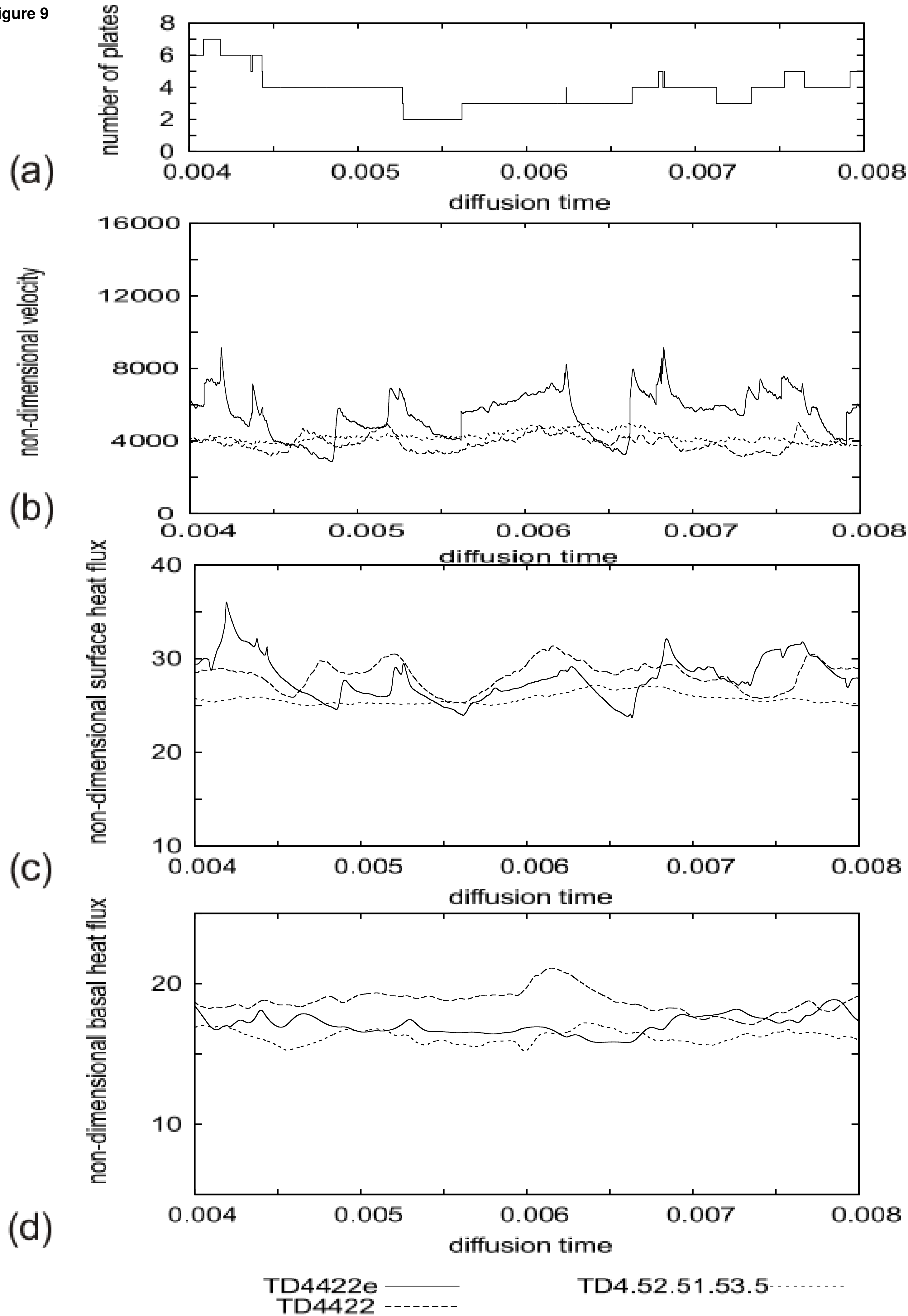
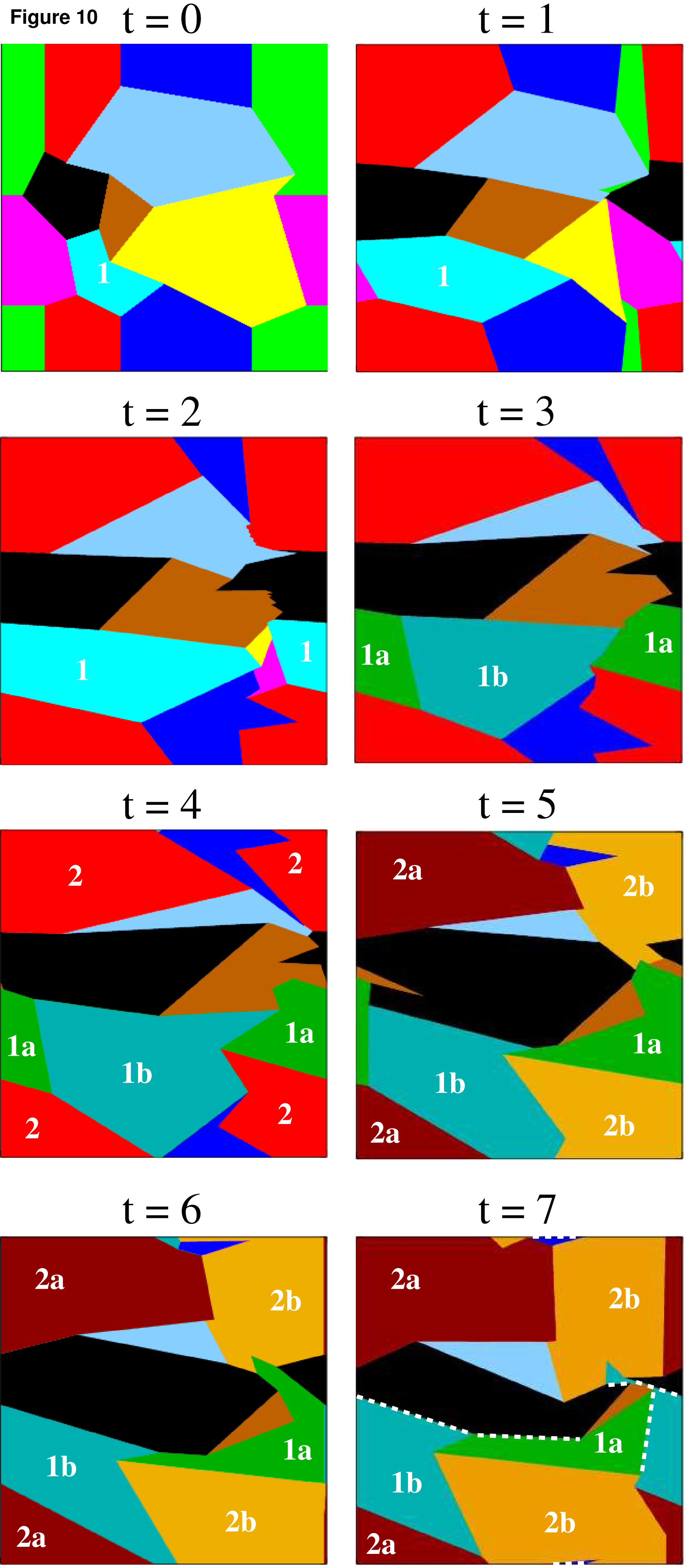


Figure 9



number of plates
diffusion time
non-dimensional velocity
diffusion time
non-dimensional surface heat flux
diffusion time
non-dimensional basal heat flux
diffusion time
(a)
(b)
(c)
(d)
TD4422e
TD4.52.51.53.5
TD4422

Figure 10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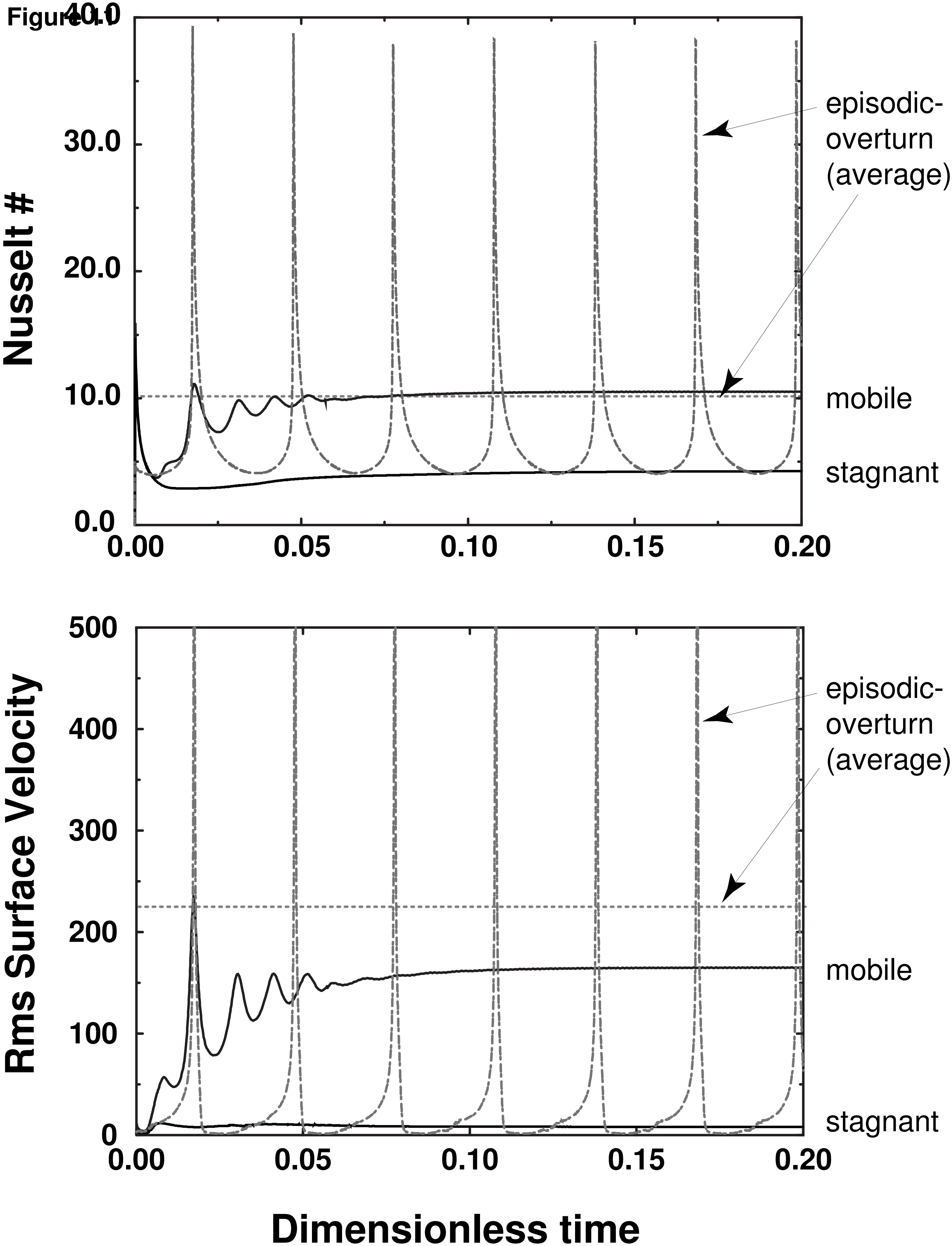
1
t = 1
1
t = 2
1
1
t = 3
1a
1b
1a
t = 4
2
2
1a
1b
1a
2
2
t = 5
2a
2b
1a
1b
2a
2b
t = 6
2a
2b
1a
1b
2a
2b
t = 7
2a
2b
1a
1b
2b
2a

Figure 11



40.0
30.0
20.0
10.0
0.0
Nusselt #
0.00
0.05
0.10
0.15
0.20
episodic-
overturn
(average)
mobile
stagnant
500
400
300
200
100
0
Rms Surface Velocity
0.00
0.05
0.10
0.15
0.20
episodic-
overturn
(average)
mobile
stagnant
Dimensionless time

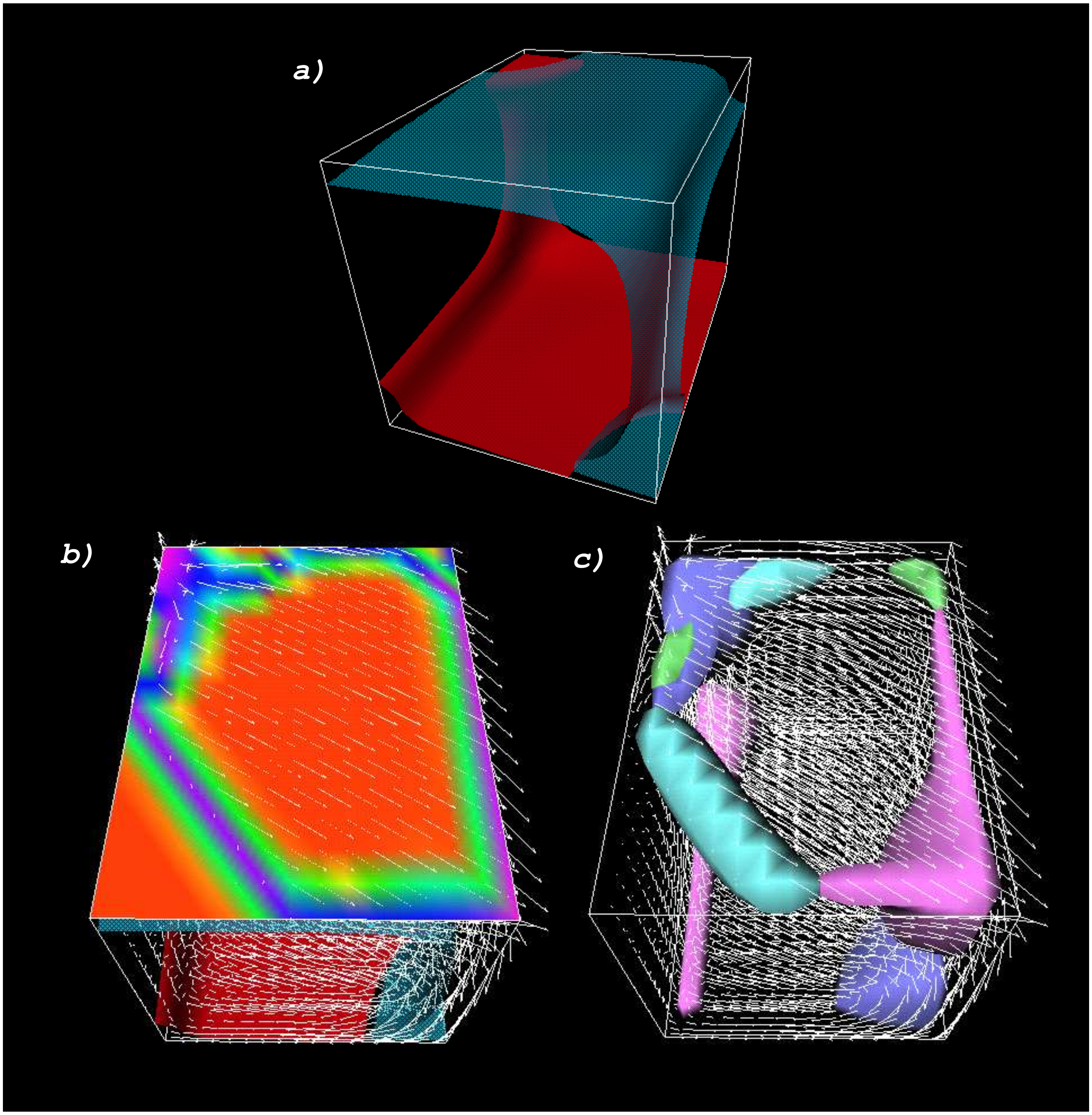


Figure 3

Figure 13

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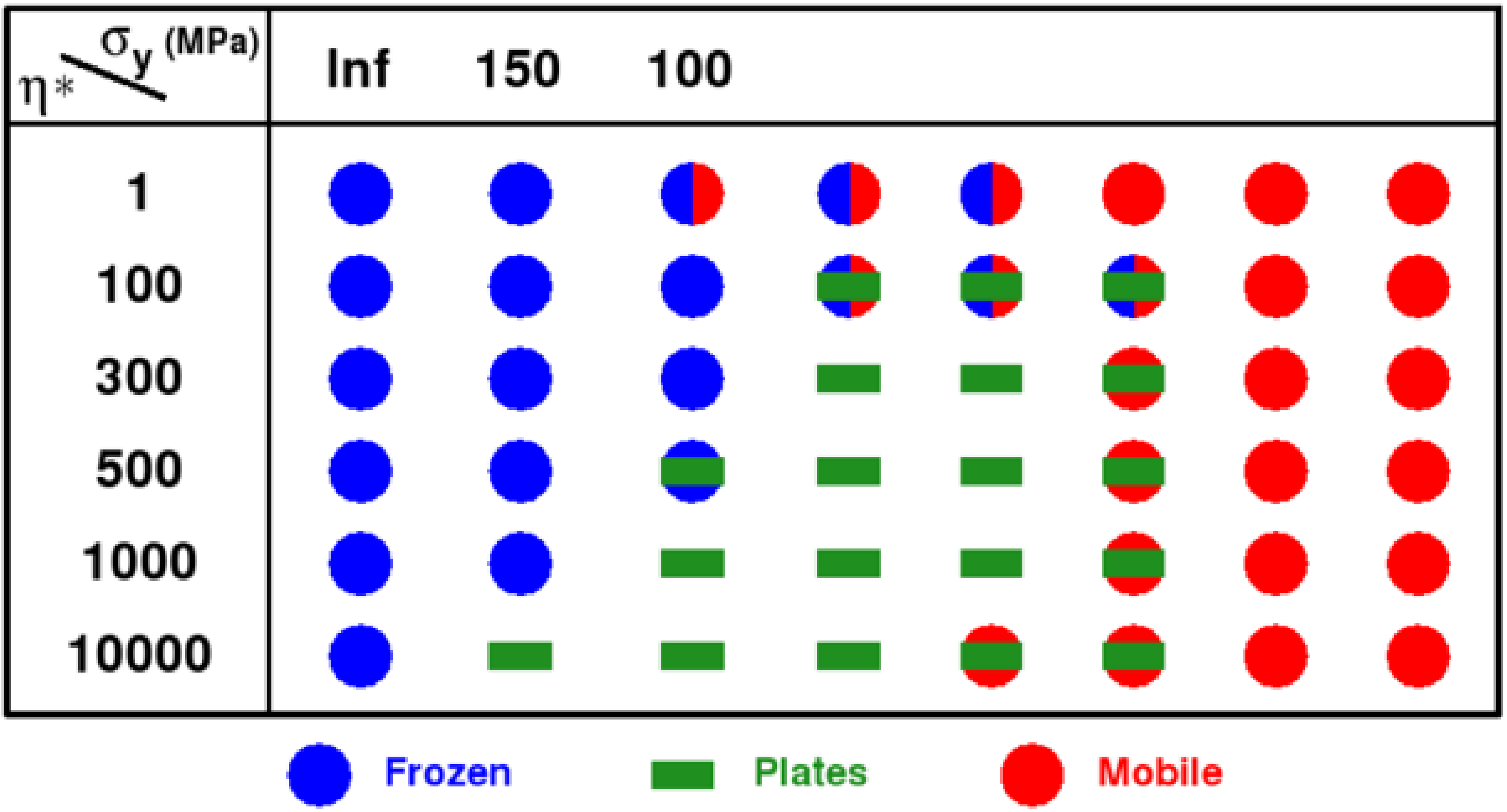

Figure 14a

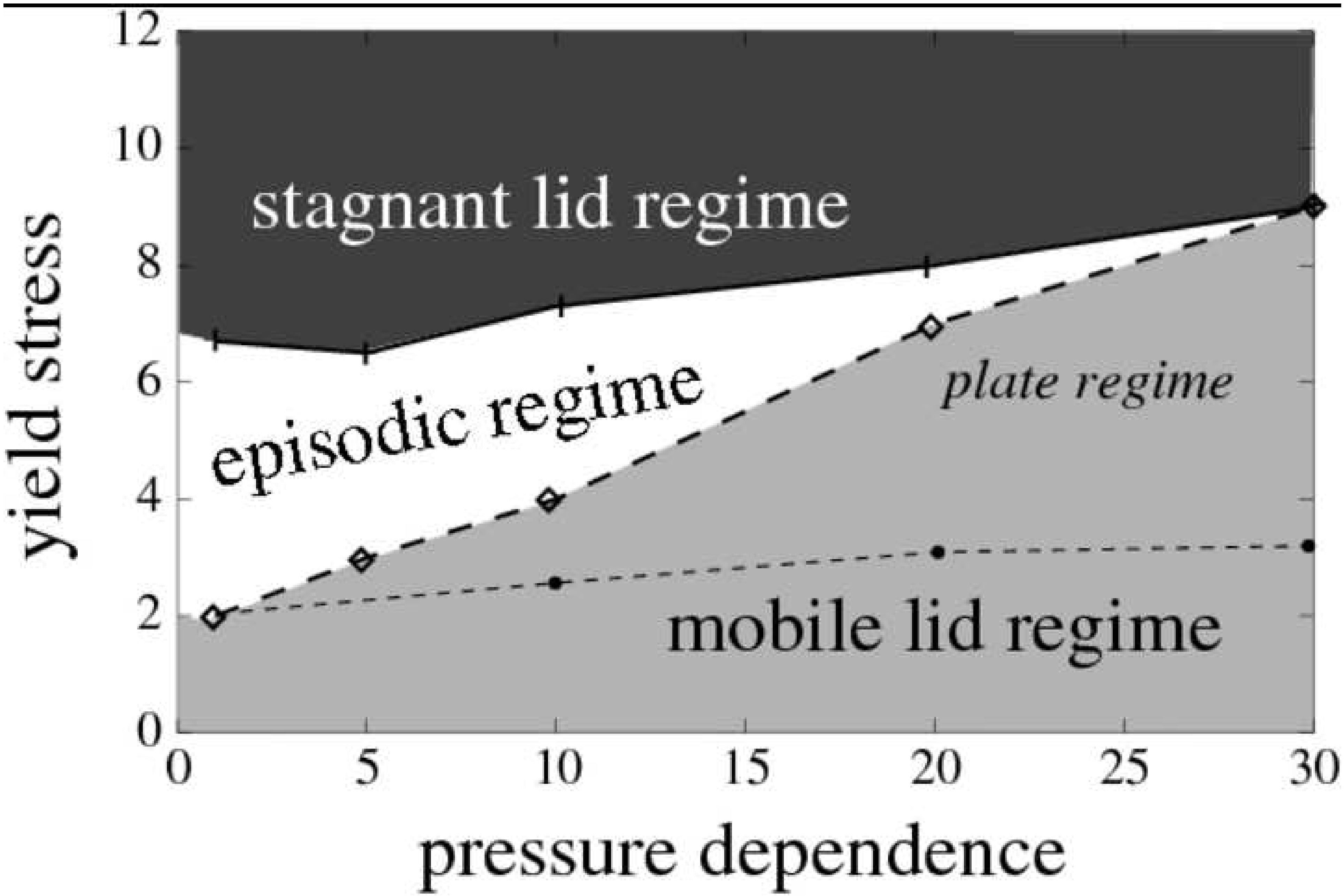
stagnant lid regime
episodic regime
plate regime
mobile lid regime
yield stress
pressure dependence
0
2
4
6
8
10
12
0
5
10
15
20
25
30

Figure 14b

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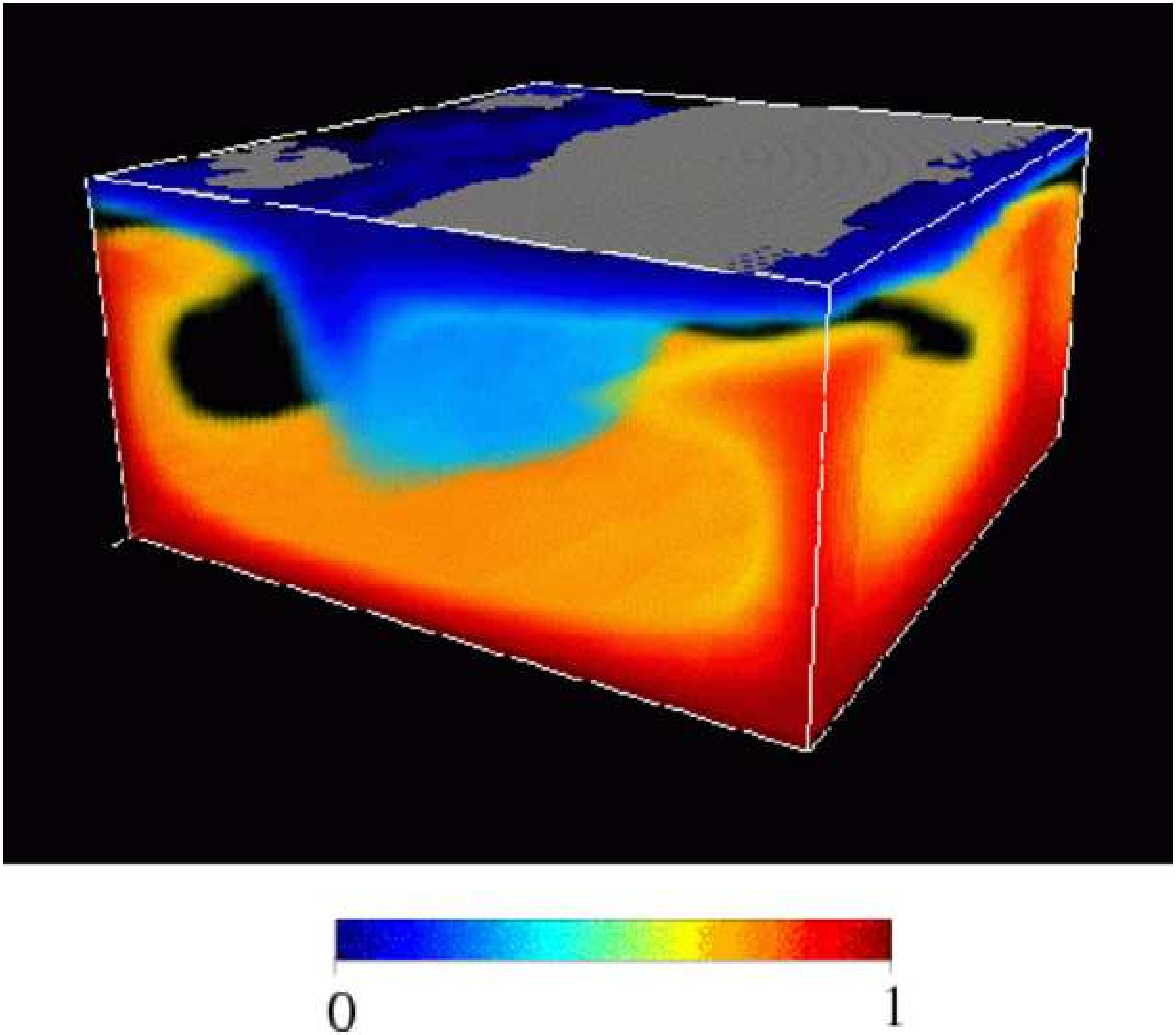

Figure 15
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